Regional Geology of the St. Lawrence County Magnetite District Northwest Adirondacks, New York

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Regional Geology of the St. Lawrence County Magnetite District Northwest Adirondacks, New York

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

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Physiography, glacial geology, petrology, and structure of an area underlain predominantly by metamorphosed sediments and igneous rocks of Precambrian age

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REGIONAL GEOLOGY OF THE ST. LAWRENCE COUNTY MAGNETITE DISTRICT, NORTHWEST ADIRONDACKS, NEW YORK

By A. F. Buddington and B. F. Leonard

ABSTRACT

This report deals with a district of about 1,300 square miles in the southeastern part of St. Lawrence County in northwestern New York, lying mostly within the piedmont section of the northwest Adirondack area but partly in the true mountain section of the Adirondacks. The report describes the geology of the district—that is, the topographic features, the surficial unconsolidated materials and their origin, and the bedrock and its nature—and discusses the evidence for the suggested time sequence of geologic events. Incidental reference is made to the geology of surrounding regions for evidence pertinent to interpretation of the St. Lawrence County area.

The piedmont part of the area includes a narrow belt of the Grenville lowlands physiographic section on the northwest, where the underlying rocks are largely metasedimentary. Here the altitudes range in general from 500 feet on the northwest to 1000 feet on the southeast, and the local relief is 100–300 feet. The land is partly cleared and devoted to dairy farms. The greater part of the piedmont area, however, is 800–2000 feet in altitude and has a maximum local relief of about 400 feet; this part is almost wholly covered with forest and has few settlements. In the true mountain section in the extreme southeast the altitudes range in general between 1500 and 2500 feet and the maximum local relief is about 900 feet, although generally not more than 500 feet. It, too, is forested and has few settlements. The bedrock of the piedmont and mountain sections of the district is largely composed of syenitic and granitic gneisses and hence is referred to as the main igneous complex. The bedrock in general has a layered structure, or a well-developed foliation, which has been a major factor in controlling the direction of elongation of the hills and ridges. Locally a joint system about at right angles to the layering and foliation has determined the location of cross valleys. Also locally zones of strong northeast joints have controlled erosion and the corresponding direction of elongation of hills and ridges. In the mountain section an additional factor influencing the evolution of topography has been block faulting of post-Ordovician age.

The climate is humid, with generally mild summers and severe winters.

During the Pleistocene the area was overridden by the continental glacier, and the surface deposits are largely ground moraine and glaciofluviatile deposits. Delta plains of extinct glacial lakes, kame moraines, and eskers are common in the valleys throughout the area. There is little evidence of recessional moraines, but valleys are largely cloaked with drift, and outcrops of rock are mostly in the upper part or tops of hills.

The bedrock of most of the area consists almost exclusively of Precambrian granite, granitic gneisses, syenite and quartz syenite gneiss, and subordinate amounts of associated metasedimentary rocks of the Grenville series. There is a minor amount of metagabbro and equivalent amphibolite. Precambrian basaltic dikes are present but rare. There are also relict patches of flatlying Potsdam sandstone of Cambrian age in the northwest part of the area, resting unconformably on the Precambrian rocks.

The oldest known rocks in the area are metasedimentary rocks of the Grenville series. These rocks underlie most of a belt along the extreme northwest border of the main igneous complex and form small belts included within the main igneous complex. They include biotitic, hornblendic, and pyroxenic quartz-plagioclase or quartz-feldspar gneisses, garnetiferous quartz-feldspar gneisses, quartz-feldspar granulite, quartzite, amphibolite, marble, and skarn. The metasedimentary rocks are thought to be the metamorphosed equivalents of sandstone, in part calcareous or argillaceous; and limestone, in part argillaceous or siliceous. The metamorphosed, virtually reconstituted equivalents of the aluminous beds have a high ratio of sodium to potassium and therefore present a problem as to the nature of the original material. They correspond in chemical composition to graywacke. A "sodic shale," if such were possible, would be an appropriate term for them. An origin as tuff has been suggested, but there is no independent evidence for this.

Derivatives of the younger granite intrusions have developed rocks of migmatitic or veined types by interpenetration with the metasedimentary rocks. A major member of the migmatitic gneisses consists of a garnetiferous biotite-quartz-plagioclase gneiss containing veins of pegmatite or aplite, and intercalated thin sheets of amphibolite, some of which are metabasite.

The metasedimentary rocks are intruded by a successive series of igneous rocks.

The oldest intrusions in the area include a lens of anorthositic gabbro west of South Russell, a number of pyroxene diorite sheets in the southern part of the Tupper Lake quadrangle, and numerous sheets and lenses of olivine diabase or gabbro. Some of the olivine diabase sheets were extensive, with lengths of at least a score of miles and similar widths, but are now exposed only as disconnected lenses.

The foregoing feldspathic and mafic intrusions were followed by a widespread sheet intrusion of magma of the composition of quartz syenite, which formed the material from which the Diana, Stark, and Tupper complexes were differentiated. The Diana is more than 40 miles long and generally 1–3½ miles thick. The Stark is more than 45 miles long. The Diana and Stark complexes are thought to be parts of the same sheet, which must have had a length of more than 100 miles and a width greater than 15 miles. All the sheets have a stratiform character. Where the Diana complex is thickest it consists of relatively undifferentiated pyroxene-quartz syenite at the base, passing upwards successively through pyroxene syenite—where

zones up to 200 feet thick contain many layers of shonkinite and feldspathic ilmenite-magnetite ultramafic pyroxene rock—pyroxene-quartz syenite, transitional pyroxene-hornblende-quartz syenite, hornblende-quartz syenite, and into hornblende granite below the roof. The pyroxenic facies are green and the hornblendic facies are red. The arrangement of the facies is due to a gravity sorting of crystals in connection with fractional crystallization of the quartz syenitic magma. The pink hornblende granite of the Stark complex, where in contact with marble, has locally developed a border facies of white microcline syenite that in part is thinly interleaved with a contact-metamorphic pyroxenic wollastonite zone.

The Tupper complex forms the Arab Mountain ridge and a belt through the locality shown on the map as Inlet; this complex flanks the border of the great anorthosite mass in the eastern Adirondacks. The Tupper complex is more than 75 miles long and 40 miles wide. It is generally much thinner than the Diana, ranging from 1500 feet to less than 1 mile. It averages markedly higher in ferrous iron than the Diana and Stark complexes and is slightly less siliceous. The basal part is a mafic pyroxene syenite gneiss. This grades upwards through normal pyroxene syenite gneiss into ferroaugite-ferrohypersthene-quartz syenite gneiss. Ferrohypersthene is a common pyroxene throughout the rocks. Because of this and the nature of their metamorphism, the rocks may be called members of a charnockite suite. In the Tupper as in the Diana and Stark complexes, the diverse rocks are interpreted as the result of fractional crystallization and gravity sorting of crystals. The mafic character of the basal portion of the Tupper may have been accentuated by incorporation or assimilation of some skarn and amphibolite.

The diabase or gabbro and the quartz syenite sheets were intruded into sedimentary rocks that were at most only gently folded. A period of strong deformation followed the emplacement and consolidation of the quartz syenite sheets, whereby all the rocks were compressed into tight folds.

At a later period the deformed quartz syenite sheets were locally intruded by diabase dikes and some satellitic sills. These dikes show marked chill zones along their walls; their depth of intrusion must have been moderate.

The next event recorded in the rocks is a second major period of deformation. This deformation was so intense that practically all the rocks underwent plastic flowage with concomitant recrystallization and reconstitution, resulting in the development of orthogneisses from the igneous rocks, and of paragneisses, schists, marble, quartzite, amphibolite, and granulite from the sedimentary rocks. The development of almandite in the olivine metagabbro sheets, metadiabase dikes, and amphibolite in the area of the main igneous complex, and in the quartz syenite gneisses of the northeastern part of the Stark complex and eastern part of the Tupper complex, occurred during this period of metamorphism. There is a regional variation in the intensity of metamorphism, in the sense that reconstitution took place under deeper, hotter conditions in the northeast, east, and southeast. The metagabbro of the Grenville lowlands on the northwest and the metadiabase dikes of the Diana complex on the southwest are metamorphosed but show no development of garnet, whereas to the northeast, east, and southeast both rocks are garnetiferous. The rocks on the northwest were reconstituted under conditions of the amphibolite facies and those of the southeast under conditions of the granulite facies. The olivine metagabbro has usually yielded a hornblende-hypersthene-almandite amphibolite, whereas the metadiabase dikes commonly have been reconstituted to an almandite-augite-plagioclase granulite. The metagabbro sheets largely lie within younger granite and were deformed in the presence of permeating solutions, whereas the metadiabase dikes were reconstituted under anhydrous conditions within the older quartz syenitic sheets. Similarly, the rocks of the Diana and the southwestern part of the Stark complex are in large part granoblastic, containing a varied amount of porphyroclastic relics but no garnet, whereas in the northeastern part of the Stark complex garnet is common. Locally, though rarely, there are small relict masses of the metagabbro and metadiabase, and of the quartz syenite facies, which are practically undeformed and unmetamorphosed, and which afford evidence as to the primary nature of the original rocks.

During the latter part of this deformation or in a period of renewed deformation, granitic rocks were emplaced on a large scale. These are called the younger granites, to distinguish them from the older granite gneisses, which are facies of the quartz syenite complexes. For the most part the younger granitic rocks have phacolithic and elongate domical relations to the older rocks but locally are crosscutting. The dominant type of younger granite is a hornblende (femaghastingsite or ferrohastingsite)-microperthite granite, grading locally into a biotite alaskite that usually contains a little accessory fluorite. The alaskite occurs in large part as an upper roof facies of the hornblende granite, or as sheets in associated metasedimentary rocks. The intrusion of the granite probably occurred at a time of mild deformation. In part it is massive and has good primary flow structure; in part the groundmass is somewhat granoblastic.

In the Grenville lowlands belt of dominantly metasedimentary rocks there are two major types of granite—alaskite in the form of anticlinal phacoliths, and a porphyritic to coarse biotite granite gneiss associated for the most part with biotite gneisses. The biotite granite gneiss and alaskite are perhaps in part a more volatile-rich facies of the younger hornblende granite of the main igneous complex. In the marble and amphibolite the granite magma was, for the most part, injected mechanically, but in the biotite-quartz-plagioclase gneiss there is an intimate penetration and permeation of the gneiss to such an extent that the resulting granite gneiss is in considerable part a product of migmatization and granitization.

The younger hornblende granite may locally be contaminated with schlieren of amphibolite or metasedimentary rock and, in contact with skarn or amphibolite, may develop local syenitic facies. The alaskite may also develop a contaminated facies in contact zones with amphibolite and members of the Grenville series and may thereby become hornblendic or garnetiferous. It may also pass into a biotite granite through contamination.

Another facies of the younger granitic rocks, found almost exclusively within the main igneous complex, is a microcline-rich granite gneiss. This type of gneiss includes a hornblendic, a biotitic, a sillimanitic, a garnetiferous, and a pyroxenic facies, and also a facies with sillimanite-quartz nodules. These microcline-rich granite gneisses are interpreted as the product of (a) magmatic incorporation, with some granitization of amphibolite to yield the hornblendic facies; (b) granitization of biotite-quartz-plagioclase gneiss and magmatic incorporation to yield the sillimanite- and quartz-rich facies; and (c) granitization or magmatic incorporation of pyroxene skarn and pyroxene gneisses to yield the pyroxene facies. The granitization is possibly in connection with the emplacement of a potassium-rich fluid, perhaps of the nature of a volatile-rich potassium-rich granite magma. Each of the facies mentioned may also be in part the product of contamination of the potassium-rich fluid and migmatitization of the country rocks. There is also a local development of some biotite-microcline-plagioclase granite gneiss and andradite-microcline granite gneiss.

Locally there are small masses and sheets of a sodic granite gneiss.

The younger granites were in part emplaced under synkinematic conditions, and deformation continued after their consolidation. The deformation of the granites was greatest in the northwest—northwest of the Stark complex—and least in the south—across the southern part of the Oswegatchie, Cranberry Lake, and Tupper Lake quadrangles. In the northwest belt, the granites are almost completely recrystallized to a fine granoblastic aggregate; the primary microperthite has unmixed to yield discrete microcline and plagioclase grains; the ratio of ferric to ferrous iron in the recrystallized hornblende is slightly greater; and sphene has formed as a new mineral. Within the same geographic area, the members of the quartz syenite complexes usually show a more thorough degree of metamorphism than the associated granite, confirming the idea of two periods of strong metamorphism.

The most uniformly massive granite found by us in the Adirondacks is the arc of green fayalite-ferrohedenbergite granite north and northwest of Wanakena. Its age is uncertain, but it may be the youngest granite intrusion.

Locally, though rarely, metadiabase or amphibolite dikes cut the younger granites and are in turn cut by granite pegmatite veins.

The general structure within the belt of metasedimentary rocks of the Grenville lowlands is that of an asymmetric wedge-shaped prism, which comprises a narrow zone on the northwest, composed of isoclinal folds whose axial planes dip southeast; a narrow adjoining central zone of folds whose axial planes dip steeply; and a wide belt on the southeast, composed of isoclinal folds whose axial planes dip moderately northwest. As viewed in vertical section, the axial planes of the folds are thus oriented in the pattern of an asymmetrical fan.

The structure within the main igneous complex is extremely complex. The sheets of orthogneiss of the Stark complex and the Inlet and Arab Mountain masses have a general anticlinal character, forming an arc of relatively rigid units against which the more mobile country rocks have been overturned and isoclinally folded. There is thus on the northwest flank of the Stark anticline a belt of isoclinal folds overturned to the southeast, and on the south side of the Inlet and Arab Mountain anticlines there is a belt of isoclinal folds overturned to the north. The rocks within the arc of the quartz syenite anticlines are moderately to tightly folded, but generally not overturned. The southwest part of the Diana complex similarly has an anticlinal structure that forms a core toward which a belt of isoclinal folds is overturned on each flank. The younger granite masses are later than at least one major period of intense deformation, for they were emplaced within strongly folded and metamorphosed rocks. However, a second period of deformation accentuated the earlier structures, especially northwest of the Stark anticline, where the younger granite phacoliths are isoclinally overturned toward the anticline. The granite there shows a granoblastic character such as is produced by plastic flow in the solid state. Except locally, the younger granite is elsewhere much less deformed. The metasedimentary rocks of the Grenville series occur exclusively in narrow belts of synclinal or synclinorial structure between broad anticlinorial belts of orthogneisses or granite. The metasedimentary rocks have flowed plastically and have been subjected to intense thickening, especially on the crests and troughs of secondary folds, whose limbs are correspondingly thinned.

Linear structure is well developed in most of the gneisses. In the zones of intensely metamorphosed rocks involved in strongly overturned isoclinal folds, the linear structure is at large angles to the strike of foliation, and generally subparallel to secondary fold axes and to the dip of foliation. In zones of less intense deformation, the linear structure is about parallel to major fold axes. In the strongly deformed but more rigid units of the quartz syenite complexes, the linear structure is always parallel to the major fold axis.

Movement has taken place at a relatively late date in the junction zone between the main igneous complex and the belt of metasedimentary rocks of the Grenville lowlands. This is manifest by mylonitization of the orthogneisses, the local development of intense small-scale fractured or crackled zones, microbrecciation, slickensiding, and the development of chlorite and carbonate.

The youngest Precambrian intrusives are represented by a few basaltic dikes. They are unmetamorphosed.

The Precambrian igneous and metamorphic rocks were eroded in pre-Potsdam time to a peneplain whose relief was not more than a few hundred feet in the northwest nor more than 250 feet in the general area of the Grenville lowlands. Lower Paleozoic sediments were laid down upon this peneplain. The oldest Paleozoic formation in the Grenville lowlands is the Potsdam sandstone of Late Cambrian age. All of the Paleozoic strata have been eroded from this area with the exception of a few relict patches in the Grenville lowlands area.

In post-Ordovician time the Adirondack area was domed and block faulted. In the St. Lawrence County area such block faulting is restricted to the southeastern part, the mountain physiographic section.

INTRODUCTION

LOCATION, CULTURE, AND ACCESSIBILITY

The St. Lawrence County, N.Y., magnetite district lies in southeastern St. Lawrence County, in the northwestern part of the Adirondack area (fig. 1).

No settlements within the district have railroad passenger facilities. The nearest towns with such service are Gouverneur, Canton, and Potsdam on the St. Lawrence division of the New York Central Railroad, and Tupper Lake on the Adirondack division of the same railroad. A branch line, carrying freight only, runs from Carthage through Harrisville, Oswegatchie, and Benson Mines to Newton Falls. An extension to the Clifton mine was abandoned at the time of closure of the mine in 1952. The Grasse River Railroad, which formerly connected Cranberry Lake with the Childwold station, was abandoned in 1945.

Two main concrete highways cross the area, one from Tupper Lake through Edwards, a distance of about 60 miles, the other from Sevey to Potsdam, a distance of about 32 miles. Several secondary macadam or gravel roads branch off from the main highways. In addition,

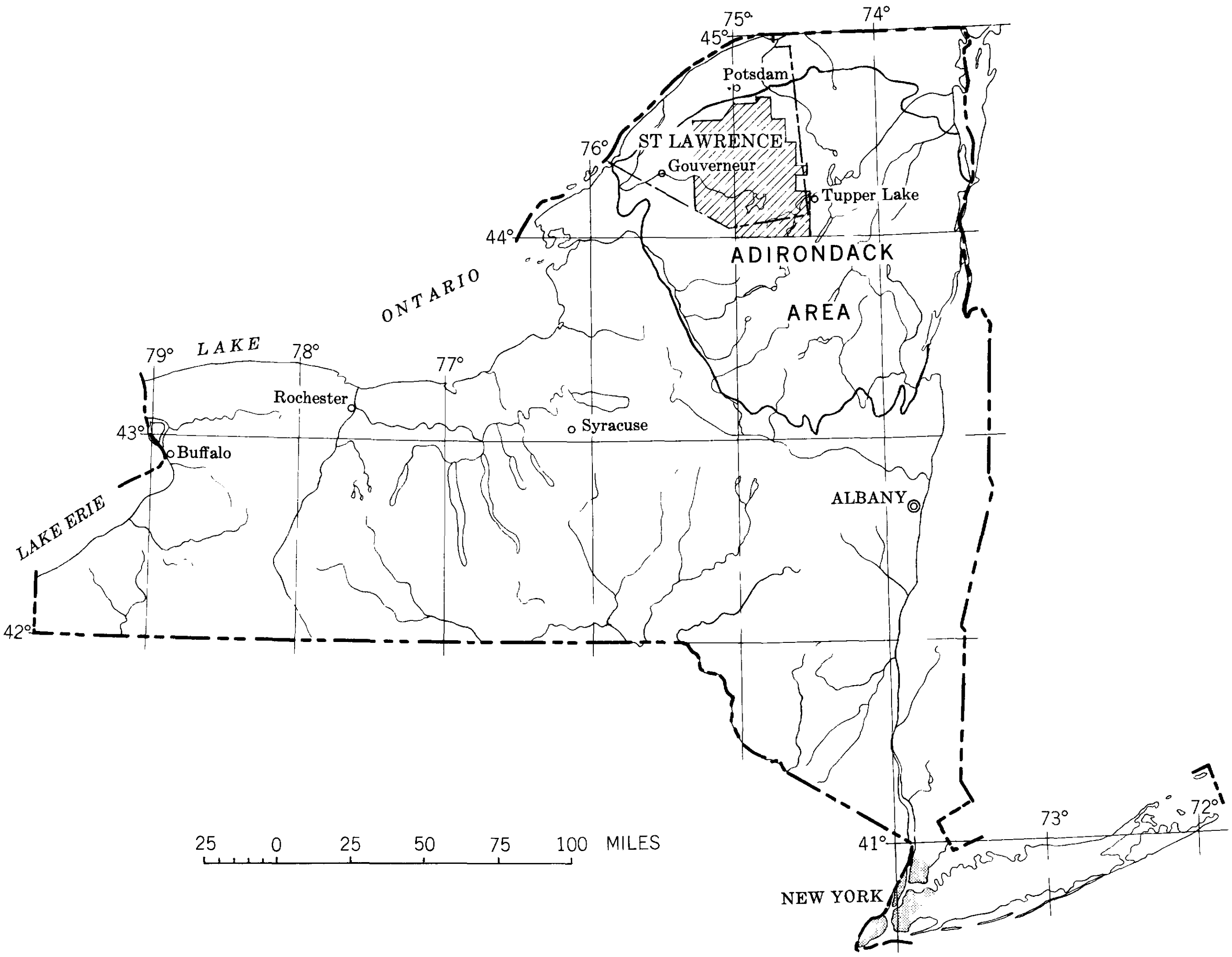

FIGURE 1.—Index map of New York, showing location of Adirondack Precambrian area, St. Lawrence County, and area of geologic map.

a number of unimproved woods roads are used for lumbering operations and access to hunting camps.

Although only a very small area within the district is more than 15 miles from a railroad, there are many local portions that are without roads and are difficult of access.

The area is sparsely settled, and in 1958 there were only 17 villages with post offices: Childwold, Colton, Conifer, Cranberry Lake, Degrasse, Edwards, Fine, Hermon, Newton Falls, Oswegatchie, Parishville, Piercefield, Russell, South Colton, Star Lake, and Wanakena. In addition, there is a post office at the hamlet of Sabattis. There is no post office in the more than 200 square miles of the Stark quadrangle.

The topography of the area is rough and hilly, though the relief, except for the Tupper Lake quadrangle, is generally not more than 300–400 feet. On the Tupper Lake quadrangle the relief is generally not more than 500 feet, though locally as much as 800–900 feet. Four principal streams—the Oswegatchie, Grass, Raquette, and St. Regis rivers, all tributary to the St. Lawrence River—drain the area.

The district is almost completely forested. Lumbering was formerly a major industry and is still carried on to some extent. The forest consists largely of maple, beech, spruce, hemlock, pine, poplar, and birch. Most of the original timber has been cut and the forest is now a second growth, locally with underbrush.

Dairy farms are located in the major valleys of the Grenville lowland belt but occupy not more than 5 percent of it. In the rest of the district there are very few farms.

The principal industries of the region are lumbering, paper manufacturing at Newton Falls, and iron mining at Benson Mines. There is substantial tourist traffic at Cranberry Lake, Star Lake, Tupper Lake, and Wanakena.

The entire area between the magnetite district and the St. Lawrence River to the northwest, except for a belt about 5–10 miles wide in the southeast portion, has a low rolling topography, relief of which is rarely greater than 200 feet. In contrast to the magnetite district, this area is largely cleared and devoted to dairy farms.

CLIMATE

Climatic data for points in or adjoining the district are given below.

Climatological data for four stations in the northwest Adirondacks

[After U.S. Weather Bureau, New York Section, 1946, Climatological Data, v.58, Albany, N.Y]

	Elevation (feet)	Length of record (years)	Temperature (°F)					Mean annual precipitation (inches)
			Mean			Minimum	Maximum	
			Annual	January	July			
Tupper Lake	1,700	36	42.1	15.7	64.9	--------	--------	37.4
Wanakena	1,510	37	42.7	15.6	65.1	−45	100	41.0
Gouverneur	450	43	45.2	17.8	68.4	--------	--------	30.8
Canton	406	58	44.8	16.6	68.6	--------	--------	35.1

The summers are usually mild, with the temperature rarely above 90°F., but the winters are severe.

The last killing frost in the spring at Wanakena usually occurs between May 15 and June 8 but in one year was as late as June 22. The first killing frost in the autumn at Wanakena usually occurs between September 14 and September 30 but in one year was as early as August 25. In 1946 there was a frost each month of the summer.

HISTORY OF SETTLEMENT

The systematic agricultural settlement of St. Lawrence County may be said to have effectively begun in 1796, following the evacuation by the British of the garrison at the present site of Ogdensburg. St. Lawrence County was established in 1802. The following dates of organization of towns will give some idea of the progress of settlement: Hopkinton (1805), Canton (1805), Potsdam (1806), DeKalb (1806), Russell (1807), Gouverneur (1810), Parishville (1814), Pierrepont (1818), Edwards (1827), Hermon (1830), Pitcairn (1836), Colton (1843), Fine (1849), Clifton (1868), and Clare (1880). The lowlands parallel to the St. Lawrence River were settled first. The towns of Colton, Fine, Clifton, and Clare are in the rough, hilly area; they were the last to be organized and are still only sparsely settled. The first settlers in the villages along the northwest border of the magnetite district were at Russell in 1805, Edwards in 1812, Pitcairn in 1824, and near Colton in 1824.

FIELDWORK AND ACKNOWLEDGMENTS

The fieldwork upon which this report is based was carried on by Buddington for a total of 21 months and by Leonard for a total of 28 months during the 8 field seasons 1943 to 1950.

The authors are indebted for very able assistance in the fieldwork to other members of the U.S. Geological Survey as follows: H. E. Hawkes, Jr., 3 weeks in 1943 and 2 weeks in 1944; Preston E. Hotz, 3 weeks in 1948; A. Williams Postel, 6 weeks in 1944; Cleaves L. Rogers, 7 weeks in 1945 and 6 weeks in 1948; A. E. J. Engel, 10 days in 1945 and 3 days in 1948; J. V. N. Dorr II, 2 weeks in 1946; George S. Koch, Jr., 7 weeks in 1949; Paul K. Sims, 4 weeks in 1949; Arthur R. Still, 12 weeks in 1949; and Robert M. Sneider, 4 weeks in 1950. Messrs. R. H. Campbell, Richard Snedeker, and R. M. Sneider, and Miss Alice Waddell Smith III, assisted in compiling material for some of the illustrations.

We are also indebted to Mr. Wen-Kuei Kuo of the Mineral Exploration Bureau of China, Dr. Felix Gonzalez Bonorino of the Geological Survey of Argentina, and Mr. Harald Major of the Geological Survey of Norway, each of whom worked with us for a period of a few weeks and with whom we had a beneficial exchange of experience.

Dr. John G. Broughton, State Geologist of the New York State Science Service, afforded us most cordial cooperation.

The authors had copies of unpublished geologic maps of the Russell quadrangle made respectively by W. J. Miller and by N. C. Dale for the New York State Museum. These maps were of the greatest assistance to us and were used freely in the fieldwork. In order to obtain more detailed structural data, however, we found it desirable to remap the Russell quadrangle and the northern part of the Oswegatchie quadrangle. The data for these two quadrangles are therefore based on the new work.

It is a pleasure to acknowledge the cordial cooperation of many persons with interests in the area. Particular thanks are due L. P. Barrett, R. M. Crump, W. M. Fiedler, John McKee, and S. A. Tyler of the Jones & Laughlin Ore Co. at Benson Mines; W. F. Shinners, A. E. Walker, K. C. Winslow, and William Ford of the Hanna Coal and Ore Co. at Degrasse; A. M. Ross of the Newton Falls Paper Mill, W. G. Srodes and the late George Collord of the Shenango Furnace Co.; G. W. and W. Clyde Sykes and Clarence McKenney of the Emporium Forestry Co.; William Doran and the late Moses La Fountain of the New York State Department of Conservation; A. Augustus Low and A. Vaillancourt of Sabattis; the late Harry Curnow of the Whitney Estate; George Collier and J. Watson Webb of Nehasane; Malcolm Broughton of Robinwood; and Ralph Friedman of Kildare Club. The Hanna Ore Co., Jones & Laughlin Ore Co., and Newton Falls Paper Mill generously gave us access to their drill cores.

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We are indebted to Princeton University for bearing the cost of a number of chemical analyses made at the Rock Analysis Laboratory, University of Minnesota.

A. E. J. Engel, having disputed with us on the outcrops, cheerfully undertook to review this report. His contributions throughout the course of our work are gratefully acknowledged.

SCOPE OF REPORT

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

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

Previous studies of the Adirondack magnetite deposits by Newland (1908, p. 29–30) and by the senior author had indicated that there was greater probability of finding new magnetite deposits in belts where metasedimentary rocks of the Grenville series and granite were associated than elsewhere. Consequently, it was deemed of first importance to find and outline as quickly as possible the general position of such belts. During the war period, the addition of details and pursuance of cognate geological problems were subordinated to the objective of securing an areal geologic map at the earliest practicable moment.

Three "Preliminary Reports" were issued (Buddington and Leonard, 1944; 1945, a and b) in which were indicated belts deemed potentially more favorable than others for the occurrence of magnetite.

The dependability and detail of the geologic mapping throughout the area is somewhat uneven, being related to the immediate importance of the area because of the magnetite mineralization, to its accessibility, and to the number of outcrops. The last is an important factor. Outcrops are relatively abundant in the northwestern border part of the district but are relatively few in the area as a whole. As has been noted, the district is almost wholly forested, and outcrops in the woods are obscured by lichens, forest litter, and moss, which must be peeled back in order to get any adequate observation on the structure and lithology. Indeed, on many outcrops, satisfactory observations are practically unobtainable. Consequently, in the time and with the resources available, fewer detailed lithologic or structural studies have been made than would normally be called for.

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

In addition to mapping the areal geology, a second objective was to make a detailed study of the known magnetite deposits in order to obtain a picture of their mode of occurrence. Such data could then be used as guides in the further development and exploitation of these deposits and in search for new ones. The economic geology of the area is discussed in Professional Paper 377.

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

The report on this district that concerns the regional geology (Professional Paper 376) has been prepared by Buddington, and the report that concerns the geology of the magnetite deposits (Professional Paper 377) by Leonard, though the authors have cooperated throughout the course of the work.

TOPOGRAPHY

The topography of the St. Lawrence County magnetite district can only be interpreted through a knowledge and an understanding of the nature of the topography in the adjoining area of the Adirondacks.

The present topography of the Adirondacks comprises so many mountains, hills, and valleys that the broader features of the relief are lost in the maze of detail shown on the quadrangle maps. If, however, we imagine the valleys of any small area refilled to the level of the higher hills, we have a surface that has many striking and systematic features. The contours shown on plate 3 are drawn by connecting the successive 200-foot contours directly across all depressions except those of major size. The position of the contours is based on the appropriate altitude of the outermost hills in the direction of the major slope. This gives a rough, generalized picture of the topography before its dissection into the present intricate detail.

The area shown in plate 3 is that part of the Adirondacks lying northwest of the main and highest topographic axis. This axis is somewhat nearer the southeast than the northwest border of the Adirondacks.

Southeast of the St. Lawrence River and parallel to it are several belts that strike northeast, each with a different type of topography (pl. 3). These are, from the St. Lawrence River eastward: the St. Lawrence lowlands, with altitudes ranging between 200 and 400 feet, in a broad belt near the river, underlain by relatively flatlying Paleozoic sedimentary beds; the northeast Adirondack foothills or Grenville lowlands, ranging in altitude from about 300 feet on the west in the Alexandria Bay quadrangle to 800–1,000 feet on the east, underlain by Precambrian metasedimentary rocks of the Grenville series and subordinate igneous intrusive rocks; the fall zone slope, ranging in altitude from 800–1,000 feet on the northwest to about 1,500 feet on the southeast in the Stark quandrangle, underlain by Precambrian rocks that are predominantly igneous gneisses; the Childwold rock terrace, ranging in altitude from 1,000–1,600 feet on the northwest to 1,800–2,000 feet on the southeast; and the Adirondack mountain belt, ranging generally in altitude from 1,800–2,000 feet on the northwest to a maximum of 5,344 feet on the southeast, but locally with basins as low as 1,300 feet. The rocks underlying all the belts except the St. Lawrence and Grenville lowlands are predominantly igneous, having only subordinate associated metasedimentary rocks of the Grenville series.

St. Lawrence Country includes parts of each of the topographic belts mentioned. The following descriptions, except where otherwise noted, refer exclusively to those portions of the topographic belts that lie within St. Lawrence County.

The St. Lawrence lowlands and the Grenville lowlands interlock and grade into each other, together constituting a belt about 30 miles wide parallel to and southeast of the St. Lawrence River.

The Grenville lowlands, Fall Zone slope, and Childwold rock terrace as defined here have all been grouped by Crowl[1] under the term "Adirondack Piedmont." The northwest part of the area covered by this report thus lies within his "Adirondack Piedmont" section, and the southeastern part within his "Adirondack Mountains" section.

ST. LAWRENCE LOWLANDS

The St. Lawrence lowlands consist of the portion of the area southeast of the St. Lawrence River that is wholly or almost wholly underlain by relatively flatlying beds of Paleozoic limestone and sandstone. The general relief is less than 100 feet. The bedded rocks tend to be eroded so as to yield flat, tabular surfaces parallel to the bedding planes, with quite steep slopes or scarps along the rivers and creeks.

GRENVILLE LOWLANDS

The belt of the Grenville lowlands lies between the St. Lawrence lowlands on the northwest and a line on the southeast running approximately through Natural Bridge, Harrisville, a point about 1½ miles north of Colton, and Allens Falls. The maximum local relief ranges in general from 100 feet on the northwest to about 300 feet on the southeast. The section is underlain by a Precambrian complex consisting of about two-thirds metasedimentary rocks of the Grenville series and one-third intrusive igneous rocks, largely granite gneiss. To the northeast and southwest the rocks of this belt are overlain by the flat-lying Paleozoic strata, of which the basal formation is the Potsdam sandstone of Late Cambrian age. The area was formerly entirely covered by the sandstone, for uneroded patches of that formation resting on the Precambrian rocks still remain throughout the belt. The present topographic surface is essentially the pre-Pots-

[1] Crowl, G. H., 1950, Erosion surfaces in the Adirondacks: Princeton Univ. Ph.D. dissertation.

dam surface, slightly eroded and dissected by post-Ordovician erosion and slightly modified by deposition of glacial and glaciofluviatile deposits. The peneplain surface was so recently stripped of its blanket of Potsdam sandstone that it has been little changed.

In contrast to the flat plains and tablelands of the St. Lawrence lowlands, the Grenville lowlands consist of a series of alternating narrow ridges or elongate hills, and flat-bottomed valleys and depressions. Hills and depressions are so numerous that the country is rough, though the relief is slight.

One of the most striking features of the Grenville lowlands is the very intimate relation among topography, structure, and kind of underlying rock. The grain of the topography generally conforms in close detail to the grain of the underlying rocks, which is predominantly alined from northeast to southwest. Local belts are underlain by medium to coarsely crystalline marble. This is a relatively soft and soluble rock and therefore generally is the terrane of valleys and low areas. Most of the lake basins of the Grenville lowlands are in marble and many, such as Sylvia Lake and Lake Bonaparte, have originated at least partly as solution basins. Where broad belts of marble are present, the topography usually comprises long, narrow, rounded lineal ridges with intervening depressions. Locally, each hill in a marble area is due to the resistance afforded by a big granite pegmatite lens. The great broad lowland from Gouverneur to Somerville lying southeast of the Oswegatchie River is in marble. The granitic gneisses and siliceous rocks and gneisses of the Grenville series usually form ridges that trend parallel to their foliation. The granite masses uniformly project as hills where they occur within the metasedimentary rocks of the Grenville series.

FALL ZONE BELT

Parallel to and southeast of the Grenville lowlands is a belt about 8–10 miles wide that is distinguished from the former by a small but distinct increase in upward slope of both the upland surface and the valleys. In St. Lawrence County this belt lies between the Grenville lowlands and the Childwold terrace, but to the north and south the terrace is absent and the fall zone merges directly into the Adirondack mountain section. This belt, called the "fall zone," is one in which falls are sufficiently concentrated and common to characterize the topography, though "stillwaters" are of course found in the belt, and falls also occur in other topographic sections. The maximum relief generally ranges from 300 to 400 feet. The decline of the major river valleys within this belt averages about 60 feet per mile, whereas across the Grenville lowlands to the northwest the decline averages about 25 feet per mile, and across the Childwold terrace to the southeast only about 12–25 feet per mile. These figures are averages along straight lines parallel to the general direction of decline of the topography as a whole.

The predominant rock underlying the belt is granite gneiss. Metasedimentary rocks of the Grenville series are subordinate.

CHILDWOLD ROCK TERRACE

The Childwold rock terrace is conspicuously developed across the Childwold and Stark quadrangles, where it is 15–20 miles wide. It narrows across the northwest half of the Cranberry Lake and the northeast part of the Oswegatchie quadrangles and gradually merges on the south into the belt of the fall zone slope of the southern part of the Oswegatchie and Number Four quadrangles. Similarly, it merges quickly into the fall zone to the northeast on the Nicholville quadrangle. The well-defined terrace occupies over 400 square miles. The maximum relief is generally about 400 feet or less. The major valley bottoms are at altitudes of 1200–1300 feet on the northwest and 1600 feet on the southeast. The upland surface is about 1500–1600 feet on the northwest and 1800–2000 feet on the southeast. The terrace nature is well shown in pl. 3 by the wide spacing of the contours between what would be the 1500-foot contour on the northwest and the 2000-foot contour on the southeast. The major valley bottoms similarly have a low gradient, as has been pointed out in the discussion of the fall zone.

An outstanding feature of the Childwold terrace is the abundance of sand plains and swamps. At least a quarter of the area of the main terrace is shown as swamp on the topographic maps, though actually a substantial part of such areas is sand plain covered with evergreen forest. Of the quadrangles in the Adirondack region, the Childwold has the largest percentage of its area in such sand plains and swamps. The part of the terrace on the Oswegatchie and Stark quadrangles is also characterized by more swamps and small lakes than is the fall zone to the northwest.

ADIRONDACK MOUNTAIN SECTION

St. Lawrence County includes only a small part of the Adirondack mountain section in its extreme southeast corner. As may be seen in plate 3, the mountain province lies to the east of the westernmost 2,000-foot contour line. The succeeding description refers only to that part of the mountain section shown in plate 3. The mountain section has the highest altitude (Mount Marcy, 5,344 feet) and greatest relief (about 3,350 feet between Mount Marcy and South Meadow, or between

Mount Marcy and Upper Ausable Lake) of any section of the Adirondacks.

The mountain section includes a number of major longitudinal topographic features that trend northeast. These are, successively from northwest to southeast, the Ellenburg Mountain range, the Santa Clara trough with multiple basins, the Lyon Mountain range, the Saranac trough with multiple basins, the broad Saranac and Beaver River valleys, and the Mount Whiteface range. There is an east-trending cross range through Mount Marcy. The ranges have been named after the highest mountain peak.

ELLENBURG MOUNTAIN RANGE

The Ellenburg Mountain range extends from the east end of Ellenburg Mountain (alt 3,654 feet) westward across the south part of the Churubusco and Chateaugay quadrangles, and then generally southwest across the Santa Clara quadrangle onto the Nicholville quadrangle as a series of disconnected ridge tops. The altitude of the crest of the ridges decreases toward the west and southwest.

SANTA CLARA TROUGH

The Santa Clara trough lies between the Ellenburg range on the northwest and the Lyon Mountain range on the southeast. It extends across the Santa Clara and Loon Lake quadrangles and includes the corners of the Chateaugay, Churubusco, and Lyon Mountain quadrangles near their junction. It includes three basins and the mountain range that runs northeast across the central part of the Loon Lake quadrangle through Debar Mountain and the Plumadore range. The bedrock of the basins is largely or wholly cloaked with drift or sand plains. The headwaters of the Chateaugay, Salmon, and St. Regis Rivers cross the Ellenburg Mountain range and drain the basins of the Santa Clara trough.

LYON MOUNTAIN RANGE

The Lyon Mountain range extends from the Mooers quadrangle southwest through Lyon Mountain (alt 3,830 ft), Loon Lake Mountains (alt 3,355 ft), St. Regis Mountain (alt 2,882 ft), Mount Matumbla (alt 2,700 ft), Arab Mountain (alt 2,535 ft), and Wolf Mountain (alt 2,420 ft) on the Cranberry Lake quadrangle. The data indicate that altitudes of the mountains of this range decline gently to the southwest of Lyon Mountain and very abruptly northeastward.

SARANAC TROUGH ("LAKE BELT")

The Saranac trough lies between the Lyon Mountain range on the northwest and the Mount Whiteface range on the southeast. This trough belt includes the broad valley of the Saranac River on the Dannemora and Lyon Mountain quadrangles, the Saranac basin, the basins on the Long Lake, Tupper Lake, and Raquette Lake quadrangles, and the valleys of the Beaver River on the Big Moose quadrangle, and of the Tupper Lakes and Raquette River on the Tupper Lake quadrangle. The terrace between 2,200 and 2,400 feet on the Big Moose and Old Forge quadrangles may also be included in this belt. The belt is separated into two major parts by the Mount Marcy cross range. The Lyon Mountain range, which forms the northwest rim of the trough, is crossed by the headwaters of the three branches of the St. Regis River, and of the Raquette, Grass, and Oswegatchie Rivers. The headwaters of the Salmon River also nearly cross the range at present and may actually have done so in preglacial time. The Raquette River also crosses the Mount Marcy cross range.

This belt has been described by Cushing (1902, p. 25–30) as the "lake belt." He writes:

> What may be called the "lake belt" is a most impressive feature in the Adirondack region. While ponds and small lakes abound throughout the district, they are much more numerous in this belt than elsewhere, and it has other peculiar features. It may be said to extend from Loon Lake in a south-southwest direction to First Lake [Old Forge quadrangle] of the Fulton chain. . . . It is sharply marked off from the district just east, the district of high Adirondack peaks, by its low relief. Rock ridges are abundant but rise very little above the general valley levels, one or two hundred feet in general as a maximum. Hills of sufficient elevation to be locally dubbed mountains and to receive names are exceptional, though such do occur. . . . Wide tracts occur of very insignificant relief, yet with abundant rock ridges.
>
> The contrast with the district on the west is less striking, though of the same kind; there is much more diversity of surface than in the lake belt, though this western district bears no comparison with the eastern in ruggedness.

MOUNT WHITEFACE RANGE

The Mount Whiteface range is the highest of the three ranges and extends almost entirely across the Adirondacks from the Dannemora to the Wilmurt and Lassellsville quadrangles. The range includes Terry Mountain (alt 2,081 ft, Dannemora quadrangle), Mount Whiteface (alt 4,872 ft, Lake Placid quadrangle), Moose Mountain (alt 3,921 ft, Saranac Lake quadrangle), Santanoni Peak (alt 4,621 ft, Santanoni quadrangle), Blue Mountain (alt 3,759 ft, Blue Mountain quadrangle), Little Moose Mountain (alt 3,630 ft, West Canada Lakes quadrangle) and West Creek Mountain (alt 2,480 ft, southwest corner of Piseco Lake quadrangle). As in the other two ranges to the northwest, the slope consisting of the crests of the mountains of the Mount Whiteface range declines quickly on the northeast and gently to the southwest.

This range forms the main drainage divide of the Adirondacks.

MOUNT MARCY CROSS RANGE

Mount Marcy is the highest peak of a range that has a generally eastward trend. It includes Dix Mountain (alt 4,842 ft, east side of Mount Marcy quadrangle), Mount Marcy (alt 5,344 ft, Mount Marcy quadrangle), Santanoni Peak (alt 4,621 ft, Santanoni quadrangle), Mount Morris (alt 3,163 ft, west side Long Lake quadrangle), Arab Mountain (alt 3,539 ft, Tupper Lake quadrangle) and Wolf Mountain (alt 2,420 ft, Cranberry Lake quadrangle). The slope of the crests of the mountains in this range declines toward the west. The range, although crossed by the Raquette River, forms the divide from which the headwaters of the Saranac and Ausable Rivers flow northeast, and those of the Hudson south.

ORIGIN OF TOPOGRAPHY

The origin of the topography may be considered under two headings—the large-scale features, such as those which characterize the physiographic sections, and small-scale features, such as individual mountains, hills, and valleys.

It is the writer's belief that the large-scale features, such as the physiographic sections, are chiefly consequent on the deformation, warping, and faulting of the pre-Paleozoic peneplain surface on the Precambrian rocks. Some modifications of the primary features, the present small-scale features, and some features of moderate size were for the most part superimposed upon the deformed peneplain surface by erosion in Tertiary time.

RELATION OF TOPOGRAPHY TO TECTONIC HISTORY

PRE-PALEOZOIC SURFACE OF LOW RELIEF

The surface of the Precambrian rocks underlying the Potsdam sandstone and other lower Paleozoic formations is thought to have been one of only low relief, and in part to have approximated a peneplain. This is indicated by the relatively flat surface of the Precambrian rocks, where they emerge from beneath the overlying sedimentary cover or in areas stripped but adjacent to still-covered areas.

In the belt between the boundaries of the continuous Paleozoic formations on the north and southwest and the 1,000-foot contour line (northeast of Natural Bridge, pl. 3) on the east, there are remnants of Potsdam sandstone, mostly in marble valleys. The present relief between the base of the sandstone beds and the top of the adjoining hills is quite generally as much as 100 feet throughout the area between the 500-foot contour line (not shown on map) and the 1,000-foot contour line. The relief may be as much as 150–220 feet along the southeast border. The hill tops must have been somewhat eroded below the original pre-Potsdam surface, so that the present amount of relief is a minimum for the topography upon which the sandstones were deposited. However, it is also probably not far from the maximum relief, so that the pre-Potsdam surface may be considered to have been a low, rolling one, essentially a peneplain. A relief of at least 200 feet is indicated by Waterman Hill (Canton quadrangle) and Burns Flat (Russell quadrangle).

Similarly, everywhere throughout the Adirondacks and around their borders, the pre-Paleozoic surface of the Precambrian rocks has but little relief where it emerges from beneath the Paleozoic formations. For the most part, the beds immediately overlying the Precambrian rocks are sandstones of the Potsdam, which include interbedded conglomerates; but in the southwest, from Natural Bridge south, Ordovician limestones directly overlie the peneplain surface.

At the present time there are no remnants of Paleozoic formations left between the 1,000-foot contour line northeast of Natural Bridge (pl. 3) and the eastern flank of the Mount Whiteface range. Relics of infaulted blocks of Potsdam sandstone, however, are found on the Mount Marcy and Thirteenth Lake quadrangles, and of lower Paleozoic limestone on the Piseco Lake quadrangle. There is reason to believe, therefore, that lower Paleozoic strata were originally deposited in the intervening area and have since been eroded.

TACONIC DOMING AND FAULTING

The pre-Paleozoic surface of the Adirondacks is thought to have been gently warped during early Paleozoic time, so that parts of it were at successive periods alternately below and above sea level, and at other periods wholly below sea level. According to Kay (1942, p. 1627), the pre-Paleozoic surface and the overlying Paleozoic rocks were deformed in Late Ordovician time, yielding a dome, and in Early Silurian time they were block-faulted along the eastern and southeastern flank. This deformation may well have given rise to the major structural features of the pre-Paleozoic peneplain, which are now reflected in the different major topographic characteristics of the several physiographic sections. This original deformation, however, has certainly been somewhat modified by tilting, warping, and faulting of more recent date.

To the northwest of a line that would about correspond to a 450- or a 500-foot contour line (pl. 3), the altitudes of the Precambrian valleys beneath the Potsdam sandstone decline very gently (less than 10–15 ft per mile) to the northwest. To the southeast, however, the inferred pre-Potsdam valley surface rises much more steeply (about 25–40 ft per mile across the Antwerp, Lake Bonaparte, Hammond, Gouverneur, and Russell quadrangles, and about 75 ft per mile across the Canton and Potsdam quadrangles).

To the southeast of the 1,000-foot contour line across the Lake Bonaparte and Russell quadrangles, the slope of the hilltop and of the valley topography is steeper than to the northwest and constitutes the fall zone slope on the northwest border of the Childwold terrace. On the Canton and Potsdam quadrangles, the fall zone slope is steeper than on the quadrangles to the southwest, and its foot starts from a lower altitude (about 600 ft). It is significant that the steeper topographic slope in these northeastern quadrangles is also correlated with a steeper slope for the pre-Potsdam surface, affording proof that at least part of the present topographic surface is the result of warping and subsequent stripping of a pre-Paleozoic peneplain surface.

FRONTENAC AXIS AND GRENVILLE LOWLANDS

The belt of Precambrian rocks from which the Potsdam sandstone has been stripped to yield the Grenville lowlands narrows toward the northwest and forms a neck continuous with the Precambrian of the Canadian shield northwest of the Thousand Islands. The belt coincides with a broad, gentle swell that is known as the Frontenac axis. The axis is about parallel to the southwest boundary of St. Lawrence County, and it may be related in structure to the Mount Marcy cross range, which lies about along its eastward projection.

FALL ZONE MONOCLINE

Around the entire northwestern, western, and southwestern outer flank of the Adirondacks there is a marked increase in slope above the 600- and 800-foot contours in the north, and the 1,000-foot contour in the west and southwest.

The base of this slope in St. Lawrence County coincides with the line that delimits the Grenville lowlands, underlain predominantly by the generally less resistant metasedimentary rocks, from the fall zone slope, underlain by predominantly resistant rocks such as granite. This alone might suggest that the steeper slope was consequent upon the greater resistance of the rocks which underlie it. However, this is inconsistent with the fact that south of Natural Bridge the gentle slope of the pre-Potsdam peneplain extends south across uniformly resistant rocks, almost exclusively granite and quartz syenite gneisses. Again, to the northeast of St. Lawrence County, the steep slope is underlain by the relatively weak Paleozoic sedimentary rocks. It is therefore concluded that the steep slope is not due to a difference in the resistance of its underlying rock from that of the adjoining belt. In the light of its relation to adjoining structural deformation, the slope seems best interpreted as the stripped surface of the pre-Paleozoic peneplain, which had been deformed by a monoclinal warp.

CHILDWOLD ROCK TERRACE

The Childwold rock terrace is underlain by a complex of rocks similar to those which underlie the fall zone slope. Its peculiar character, therefore, is not readily interpreted as resulting from more easily erodible rocks.

The terrace lies just to the southwest of the Santa Clara trough and along its structural trend and adjoins the Saranac basin, which has the structure of a graben. In the light of the evidence for structural deformation in the adjoining area to the northeast and southeast, it seems logical to interpret this terrace likewise as at least partly structural in origin—the product of warping. It may be noted that the terrace on the Big Moose and Old Forge quadrangles likewise lies to the northwest of a proven faulted zone, and to the southwest of a major structural trough belt.

The south border of the terrace has an eastward trend. This conforms to the strike of the foliation and structure of the bedrock, which in general for the rest of the terrace strikes northeast or north. The eastward trend would then seem to be a reflection of the bedrock structure in locally controlling the later warping. It will also be shown later that there is some evidence that the area of the Childwold terrace corresponds roughly to an area whose substructure may be partly controlled by a plate or platform of relatively rigid rock.

BLOCK FAULTING AND WARPING

Block faulting of the lower Paleozoic formations and the underlying Precambrian rocks has been demonstrated to be present along the north border of the Precambrian east of the Chateaugay quadrangle, to have yielded a graben structure now occupied by the Saranac basin and the broad reentrant to the northeast on the Dannemora quadrangle, and to be prevalent throughout the Adirondacks southeast of the Mount Whiteface range. Remnants of Potsdam sandstone, preserved because of downfaulting, have been found in the Piseco Lake, Thirteenth Lake, Mount Marcy, and probably the Lake Placid quadrangles. The amount of displacement ranges from a few feet to as much as 4,000 feet. The faults are thought to be steep, of the normal type. Southeast of the Mount Whiteface range most of the faults show downward displacement on the southeast side. Tilting of the blocks has accompanied the faulting. Faults with a northeast, north-northeast, or north strike are dominant and exceptionally well developed, though faults with other strikes are common.

Practically no normal faulting has been observed or inferred in the area northwest and west of the Adirondack mountain section. It effectively dies out in this direction. The Saranac intramontane basin is certainly controlled by faulting, for its borders cut indiscriminately across many different types of rock and across the bedrock structure. The structural nature of the other basins, however, is not equally well understood. It seems probable that post-Taconic faulting or warping deformed the pre-Paleozoic surface, yielding basinlike structures; but possibly one or more are primarily the result of erosion of less resistant rocks without local deformation.

In a later section, a detailed description will be given of the manner in which the lowlands of Tupper Lake, of the Dead Creek Flow belt on the Cranberry Lake quadrangle, and of the Tamarack Creek belt on the Oswegatchie quadrangle are correlated with a strong development of northeast joints, which are interpreted as the structural but less intense equivalent of the northeast-trending fault system of the Adirondack mountain section.

MESOZOIC PENEPLAIN

There are no middle or upper Paleozoic or Mesozoic sedimentary formations in the Adirondacks. We may infer that for at least much of this period the area was above sea level and subject to erosion. Many geologists believe that during this time the Adirondack mass was worn down to a peneplain surface, except perhaps for certain monadnock areas. It may be inferred that during this time the Paleozoic sedimentary strata were stripped from the pre-Paleozoic surface on the Precambrian rocks, except on the borders or where they were below the level of erosion as a result of the Taconic downwarping or downfaulting. For example, at the time of the Mesozoic peneplain the Paleozoic strata may have covered the Grenville lowlands, the fall zone slope, and the Childwold terrace, and may have occupied the basins of the Adirondack mountain section. However, the Precambrian rocks of the mountain ranges would have been exposed and to a large extent eroded below their former pre-Paleozoic surface.

TERTIARY DOMING, RENEWED FAULTING AND EROSION

By analogy with the history of adjoining regions, it is thought that the Adirondacks were uplifted in late Mesozoic or early Tertiary time, yielding an asymmetric dome whose highest axis was along the locus of the Mount Whiteface range and whose crest was in the Mount Marcy region.

It has been noted previously that the headwaters of the Salmon, St. Regis, Raquette, Grass, and Oswegatchie Rivers cross the Lyon Mountain range and have their source in the present Saranac trough. This would be an appropriate arrangement if we assume that the course of these streams was consequent on the northwest slope of the Mount Whiteface range, and that the Saranac trough was still filled with Paleozoic sedimentary rocks. Subsequent erosion of the less resistant sedimentary rocks has permitted the Saranac River to develop the Saranac basin and capture the former southeastern headwaters of several of these rivers. The Raquette headwaters still drain the west side of the Mount Whiteface range. If the Saranac basin were interpreted as due in large part directly to faulting of the Mesozoic erosion surface, it would be very difficult to explain how the major rivers could have developed their valleys across the Ellenburg and Lyon Mountain ranges. The Saranac basin is better interpreted primarily as an exhumed graben resulting from erosion of a previously infaulted block of Paleozoic sedimentary rocks.

However, there is good evidence that some renewed faulting in Tertiary and Recent time continued to the present. The earthquake of 1944, the focus of which was between Cornwall, Ontario, and Massena, N.Y., presumably resulted from slipping on a fault plane in the bedrock.

Under the hypothesis favored here, the present bedrock surface of the Grenville lowlands, fall zone slope, Childwold rock terrace, floors of basins having an origin like that of the Saranac, and perhaps some of the highest peaks of local areas, is the warped and faulted pre-Paleozoic peneplain surface, which has been stripped of Paleozoic sedimentary rocks and more or less rebeveled and etched by erosion in Tertiary time. The upland surface of the mountain section in general would represent relics of the Mesozoic erosion surface (called Cretaceous peneplain by many), though local peaks above the general level might be but little below the pre-Paleozoic peneplain surface.

The Tupper Lake, Raquette Lake, Long Lake, and Lake Placid basins are all localized on areas in which the bedrock consists largely of metasedimentary rocks of the Grenville series. These basins might therefore, in the absence of other evidence, be interpreted as due to the erosion of relatively less resistant synclines of metasedimentary rocks of the Grenville series, surrounded by the more resistant igneous rocks. However, the basins lie in a region where Taconic folding or faulting may reasonably be expected, and further data will be necessary to understand their origin properly. The basins in the Santa Clara trough likewise lie in a region where Taconic deformation might be expected, but the bedrock geology either is obscured by overburden or has not been surveyed.

On the Number Four and McKeever quadrangles the 200-foot contours for the regional slope of the fall zone strike north, whereas the hills delineated by the 20-foot contours of the topographic sheets uniformly strike northeast. The former is interpreted as representing essentially the flank surface of the domed pre-Paleozoic peneplain, whereas the present hills with northeast strike result from subsequent erosion and etching of the rocks conformably with the direction of their foliation and layering.

TERTIARY(?) OR CRETACEOUS(?) EROSIONAL TERRACES

The preceding discussion has been based on the hypothesis that the large-scale regional variations in the character of the topography are for the most part controlled by the form of the surface of a structurally deformed pre-Paleozoic peneplain on Precambrian rocks. However, another hypothesis (Crowl*) has attributed the Big Moose terrace (south part of Big Moose quadrangle and north part of Old Forge quadrangle) and the Childwold rock terrace to the intersection of a second erosion surface or partial peneplain, of Cretaceous or Tertiary age, with the pre-Paleozoic peneplain. This is based upon the fact that the surface of the Tug Hill plateau correlates in altitude with the Big Moose terrace, and that both decline systematically toward the north. The Tug Hill plateau is underlain by Paleozoic sedimentary rocks whose northeast border is the border of the plateau. One line of reasoning would thus make the Childwold and Big Moose terraces simply the stripped surface of a differentially warped pre-Paleozoic peneplain, whereas another line of evidence would ascribe them to a later period of erosion and partial peneplanation superimposed upon a general domical deformation of the pre-Paleozoic peneplain. Which of these alternatives is correct must await further studies. It may be that, whereas the large scale features of the topography were initiated and localized by structural deformation, subsequent erosion is responsible to a substantial degree for the present elevations of certain of the major features.

RELATION OF TOPOGRAPHY TO BEDROCK

Within each of the topographic sections previously described are topographic belts that differ somewhat from each other.

*Crowl, G. H., 1950, Erosion surfaces in the Adirondacks: Princeton Univ. Ph.D. dissertation.

KIND OF ROCK

Generally, the igneous gneisses (granite gneiss and quartz syenite gneiss) yield belts of hills that are prevailingly higher than those within belts underlain by metasedimentary rocks. This is particularly well shown on the Stark and Cranberry Lake quadrangles, where the hills of the Clare-Clifton-Colton and Bog River belts of metasedimentary rocks are usually lower in height than those of the adjoining belts of gneiss. Isolated bodies of granite gneiss or quartzite within the belts of metasedimentary rocks also yield isolated hills or groups of hills generally higher than those on the metasedimentary rocks, as in the case of the Buck Mountain–Marble Mountain group, on the Cranberry Lake quadrangle. A narrow valley eroded in metasedimentary rocks entirely surrounds the Spruce Mountain–Tunkethandle Hill granite mass on the Stark quadrangle.

The most easily eroded rocks are marble and calcareous rocks. Outcrops of marble within this area are rare, yet every drill hole put down on skarn has encountered local marble lenses in places where no outcrop of marble has been observed. Marble beds, where they occur as thin lenses within other metasedimentary rocks, usually yield no outcrop because of their much greater solubility. However, where thick marble beds underlie broad areas, as in the Grenville lowlands belt, they yield numerous outcrops.

LAYERING AND FOLIATION

The elongation of individual hills is generally about parallel to the trend of the layering and foliation of the bedrock, though there are also several large belts in which the elongation is controlled by a fracture system. With the exceptions noted, and some local discrepancies, the structural trend lines of the foliation and layering can be inferred from the direction of elongation of the hills. For example, the synclinal structure of the Darning Needle syncline was accurately inferred from the topographic map and air photos before the authors visited the area. The steep scarp slopes on the west side of Indian Mountain and gentler slope on the east side indicated a gentle or moderate eastward dip for the foliation. In reverse, the steep scarp slope of the east side of the hill west of Darning Needle Pond and its gentler westward slope indicated a westward-dipping foliation.

FRACTURES

Although the topography over most of the district is in large part conformable with the layering and foliation of the underlying bedrock, in local areas the topographic control is by joints and fractures.

The belt on the Tupper Lake quadrangle that is underlain by the syenitic Tupper complex is an example. Here the foliation usually has an eastward trend, whereas the elongation of the individual hills is athwart this structural trend and is controlled by a strongly developed joint system. West of the railroad and north of the road from Horseshoe to Lake Marian, there is a strongly developed joint system striking north-northeast to north, with a corresponding orientation in elongation of the hills. Again, northeast of the road from Horseshoe Lake to Tupper Lake there is a belt, parallel to the northeast part of Tupper Lake, in which both the hills and a strongly developed system of joints have a northeast strike. Throughout this belt there are strongly developed cliff faces parallel to the elongation of the hills and the corresponding joints. Arab Mountain is the only major ridge generally coincident with the direction of the foliation. North and south of the anticlinal belt of syenitic rocks, however, the topography is again controlled by the direction of foliation and layering in the underlying granitic and metasedimentary gneisses.

The elongation of hills underlain by the quartz syenite and granite complex on the Stark anticline (Nicholville and Stark quadrangles) has a substantially similar relation to strongly developed fractures or joints across the foliation. Southwest of Lake Ozonia (Nicholville quadrangle) there are very conspicuous topographic elements that trend northwest across the foliation; on the east, foliation has an east strike, and on the southwest, a northeast strike. On the Stark quadrangle the north-striking element in the topography between Albert Marsh Hill and Chapp Hill, on the west, and Sellecks Lower Camp and Little Cold Brook, on the east, is paralleled by a strongly developed set of north to north-northeast joints. In part, the hills underlain by the Stark complex on the Stark quadrangle have an elongation parallel to the foliation, and to the south across the Russell quadrangle this is generally true.

The northeast ninth of the Cranberry Lake quadrangle and the extreme southeast corner of the Stark quadrangle is another area in which the elongation of the hills is athwart the general direction of the foliation. The direction of several arms of Cranberry Lake—such as Dead Creek Flow, Wanakena Flow, Brandy Brook Flow, and the north arm—is, in part at least, across the strike of the foliation. The general strike of the topography is north to northeast, again parallel to a system of strong joints.

Similarly, farther to the west, in a belt about 3 miles wide lying parallel to and southeast of a line running from Long Lake to Star Lake (Oswegatchie quadrangle), there are topographic elements that have a northeast strike conformable with a strongly developed series of northeast joints. The joints are oblique to the direction of much of the foliation, which is from east to west.

On the Stark quadrangle, also, ridges such as Buckhorn and that northeast of Rainbow Falls have an elongation at an angle to the foliation of the underlying bedrock. This likewise is attributed to control by strong northeast joints.

On the Russell quadrangle the South Branch of the Grass River east of Degrasse and the main drainage line to the northwest as far as Whippoorwill Corners, is parallel with, and probably consequent upon, a set of intensively developed northwest joints.

In conclusion, we may say that joints are present everywhere throughout the bedrock of the area, afford lines of weakness, and have doubtless been a minor factor in controlling the direction of development of all the topography. Locally they have played a dominant part.

Several long, relatively straight, narrow valleys in the southeastern part of St. Lawrence County are, we think, related to erosion along fault zones. These include the valley of Little Tupper Lake and Sperry Brook; Tupper Lake and Cold Brook; the narrow rectilinear valley from Town Line Pond to Little Pine Pond; the rectilinear depression along the north side of Wolf and Silver Lake mountains; and the valley along Dead Creek, Dead Creek Flow, and the north arm of Cranberry Lake.

GEOLOGY OF GLACIAL DEPOSITS

The St. Lawrence County magnetite district lies within the region of the northern United States that was buried, during temporary periods in Pleistocene time, beneath the ice of continental glaciers.

DRIFT

The major part of the district is obscured by a veneer of drift left after the retreat of the ice. All the major features of the topography are thought to be the product of stream erosion, largely preglacial in age, but many minor features of the topography are the result of superimposition of deposits from the ice and its melt waters. These include ground moraine, recessional moraines, eskers, kames, and deltas.

The drift has commonly been eroded from the tops of the higher hills, but drift still fills some valleys deeply and, except where the hills are steep sided and cliff faced, usually obscures the bedrock slopes. In the course of drilling for ore, as much as 50 feet of drift is often encountered in the valleys; occasionally, thicknesses up to 150 feet are found. In the valley of the

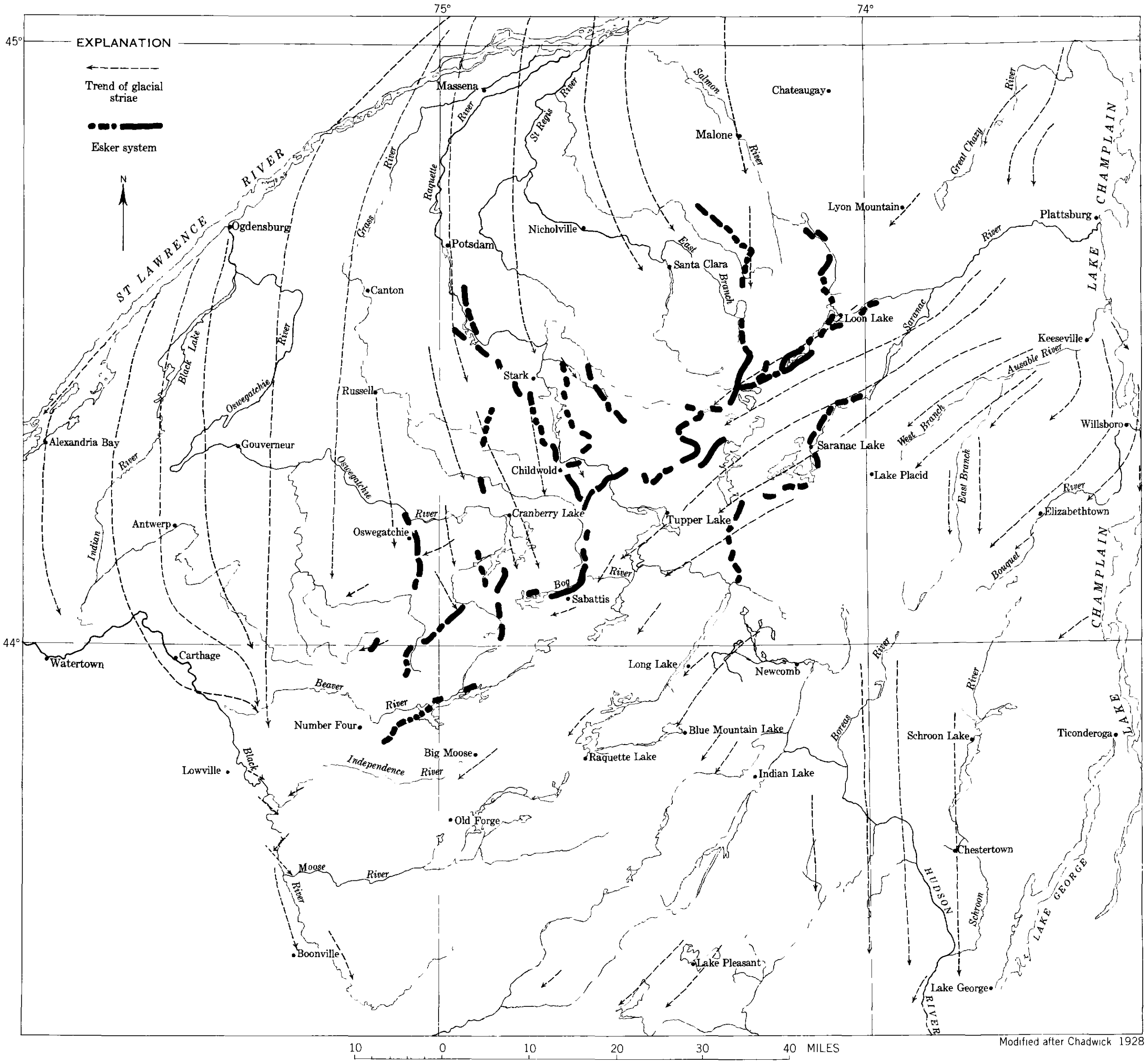

FIGURE 2.—Generalized sketch map showing the trend of glacial striae in the Adirondack area and the esker systems of the northern and northwestern Adirondack area.

Oswegatchie River southeast of Jarvis Bridge (1.3 miles east of Newton Falls), 138 feet of drift was found in one drill hole, and in the valley bottom north of Twin Mountain (Cranberry Lake quadrangle), 141 and 149 feet respectively were found in two holes.

The drift blankets the bedrock far more completely, and presumably more thickly, in the area of the Childwold rock terrace than in the fall zone slope.

GLACIAL STRIAE AND DIRECTION OF ICE MOVEMENT

A generalized picture of the direction of the movement of the ice in the Adirondacks as indicated by recorded orientations of glacial striae on bedrock is shown in figure 2. Practically all the recorded data are from outcrops in the valleys.

The evidence of the glacial phenomena in general is interpreted as showing that at a period of maximum intensity during the Pleistocene, the whole Adirondack area was buried under an ice cap, which extended as a continuous sheet back to its center in Labrador. The direction of flow of the basal ice was much influenced by the gross features of the topography. Thus the ice flowed generally southwest along the valley of the St. Lawrence, changing to south or a little east of south across the north and northwest slope of the Adirondacks. Along the intramontane lowland of the Saranac trough, which extends southwestward across the central part of the Adirondacks (pl. 3), the ice similarly moved in a southwest direction parallel to this local major trend of relatively low topography.

Within much of the St. Lawrence County magnetite district the ice moved almost wholly in a south to south-southeast direction. Strikes of striae between S. 10° E. and S. 25° E. are prevalent throughout the Stark, the northern part of the Cranberry Lake, and the northwestern part of the Tupper Lake quadrangles, and between S. 5° and 15° E. on the Russell and the northern half of the Oswegatchie quadrangles.

Most of the Tupper Lake quadrangle, and the southern half of the Cranberry Lake and Oswegatchie quadrangles, however, lie in the trough where the ice flowed southwest to west-southwest. About a mile southeast of Star Lake, where the trail to Alice Brook crosses Little River, two directions of movement are indicated on the same outcrop. The glacial striae strike S. 5° E., whereas the axis of a series of chatter marks strikes S. 40° W.

Cobbles and pebbles of anorthosite are common throughout the drift of the area. They are thought by Martens (1925) to have come in part from the Morin anorthosite area north of Montreal. In part, however, the anorthosite cobbles and boulders of the Tupper Lake, Cranberry Lake, and Oswegatchie quadrangles must have come from the main Adirondack anorthosite mass to the east-northeast.

In general, the nature of the ground moraine is such as to indicate that much of it, especially the coarser part, had not been transported very far from its source—not more than a few miles, often less than a mile, and even as short as a few hundred feet.

KAMES

Kames are hillocks or short ridges of stratified gravel and sand formed by the agency of glaciers and glaciofluviatile waters. Several or many kames may occur together in areas or belts and may be associated with generally closed depressions called kettle holes. Kames are in part formed as one facies of morainal accumulations or recessional moraines at the irregular margin of the ice sheet; in part they constitute parts of esker systems and occur at the termination or junction of eskers or in the interrupted intervals between the ends of successive eskers; again, they occur as isolated hillocks or as kame terraces locally along one or both sides of valleys. Most of the kames in this region are parts of esker systems or are in belts that have a pattern and relation to ice movement similar to esker systems.

Kame groups or kame belts that clearly form parts of esker systems will be described with the latter. There are, however, narrow belts of kames, of various lengths, which are oriented roughly parallel to the direction of motion of the ice yet are not associated with well-developed eskers, but which may, nevertheless, have originated under conditions somewhat similar to those controlling eskers. One such belt is particulary well developed in the southeast ninth of the Stark quadrangle. It extends for 5 miles from Balsam Pond northwest to include the area around Sampson Pond, Wolf Ridge, and the belt between Lem and Wolf Ponds as far as Dismal Swamp. To the north is another belt of kames, lying between Beaver and Clear Ponds and extending for 1½ miles to the north.

Another line of kames parallels Tracy Pond outlet just north of Tracy Pond.

On the Oswegatchie quadrangle a belt of kames and local esker ridges extends from the east side of the valley of the Oswegatchie at Scotts Bridge south across the river and along the west side of the valley, along the east side of Little River and east of Oswegatchie to Star Lake.

ESKERS

A typical esker is a narrow, elongate winding ridge or linear series of interrupted ridges, composed of stratified sand and gravel. Eskers commonly have a steep-sided form, which leads them to be called "embankments," "hogbacks," or "horsebacks," and are usually oriented roughly parallel to the direction of motion of the ice, in connection with which they were formed. Locally, the ridges give way to a linear complex belt of kames and depressions, or to deltas. These, together with the more typical eskers, constitute an "esker system."

In general, eskers have been interpreted as having formed in several different ways—in subglacial tunnels, in crevasses without roofs, and during retreat of the ice front as a result of successive yearly additions of deposits formed at or near the mouths of subglacial streams in bodies of standing water. Esker systems may line up to form a pattern similar to stream drainage patterns, as is the case in the northern Adirondacks (fig. 2). Unlike deposits of normal rivers, however, the eskers may increase in altitude downstream and pass over divides, as do those of the esker system that crosses the area under discussion. Many eskers branch and reunite. In some places where two esker systems join, there is a zone of kame topography.

Some of the eskers in this area have been previously referred to by Chadwick (1928). The eskers are restricted to the area of the Childwold terrace and Adirondack mountain sections.

A line of esker ridges crosses the Adirondacks from northeast to southwest along the line of the Saranac trough, following parts of the valleys of the North Saranac, Bog (fig. 3), and Beaver Rivers, and has been previously described (Buddington, 1953). It has a minimum length of about 85 miles. All the tributary eskers so far delineated come from the north, and the manner in which they join indicates that the drainage forming the main esker came from the northeast. Thus this esker, which begins at an altitude of about 1,300 feet along the valley of the North Branch of the Saranac River, crosses very low divides between the tributaries of the St. Regis River, crosses the St. Regis River at an altitude of about 1,620 feet and the Raquette River at an altitude of about 1,520 feet near Gale, and rises to an altitude of 1,800 feet where it crosses the divide between the Bog River (of the Raquette River watershed) and the headwaters of the Oswegatchie River. It then crosses the divide between the Oswegatchie and the Beaver River at its maximum altitude of 1,860 feet, whence it declines to about 1,600 feet in the southwest.

A part of this esker is well developed across St. Lawrence County as indicated by the following details. Just south of Childwold Park on the Childwold quadrangle, a fine esker extends for about 3 miles to the southwest and runs between Massawepie Lake and an unnamed lake to the east, between Boottree Pond and Horseshoe Pond and between Town Line Pond and Deer Pond. The two lines of lake basins thus form two interrupted troughs on either side of and parallel to the esker. Such esker troughs or "marginal creases" are quite characteristic of eskers in general. The road from Childwold Park follows this esker as far as the fork to Grass River Club.

The Adirondack esker system to the northeast of Catamount Pond (Childwold quadrangle) is indicated by a belt of elongate kames and kettle holes about 5 miles long, well developed southwest of the Raquette River but fading out to the northeast.

South of Town Line Pond (Tupper Lake quadrangle) along the general line of the Usher Farm road, there are local ridges of sand and gravel of poorly developed eskerlike form, which indicate the line of the esker system as far south as Little Pine Pond. From here south-southwestward and westward for 8 miles to Spruce Grouse Pond (Cranberry Lake quadrangle) the esker is again splendidly developed, almost continuously. This is shown only in part by the contouring of the topographic maps (cf. fig. 3). From the east side of Little Pine Pond, the esker runs along the west side of Hitchins Pond, thence parallel to and along the south edge of Bog River to the narrows west of Second Pond, through the islands, and then south of and parallel to the trail to Spruce Grouse Pond. To the west, on the Cranberry Lake quadrangle, there is a gap of about 4 miles where but little indication of the esker has been seen. Northwest of Big Deer Pond there are extensive sand deposits, and just east of Nicks Pond there is a conspicuous esker running from north to south. This is the junction area of the esker system along the east side of Sixmile Creek with the main Adirondack esker, which here turns south for several miles and passes off the south border of the Cranberry Lake quadrangle. The esker ridge east of Nicks Pond is well developed for a length of more than 1 mile. Another section of the Adirondack esker is well developed along the west side of the Oswegatchie River near the Herkimer and Hamilton County line, beginning about 0.8 mile north of Partlow Milldam. North of the milldam it forms an almost unbroken ridge, 40–60 feet above the river, though on the south it passes into loosely joined kames and kettles.

RECESSIONAL MORAINES

Several recessional moraines, as mapped by Taylor (1924, p. 646), should cross this region, but in this hilly country they are not well defined, and their exact position and continuity is open to question.

Local morainal deposits, possibly indicating a line of recessional moraine, are found at the following localities near the border of the Stark and Cranberry Lake quadrangles: 1 square mile beginning about 0.5 mile southwest of Newbridge (Stark quadrangle); a morainal loop around the south end of Moosehead Pond and a ridge of sand and gravel along the east side of Moosehead Pond Outlet; the ice contact slope on the north side of the Dillon Pond delta; a belt of kame moraine between Irish Brook and Dead Creek; the ridge 1–1½ miles north of Windfall Brook; and the morainal area to the east along the Raquette River. Most of these deposits are of sand and gravel, and of a character indicative of deposition in standing waters, presumably a temporary lake. Much of the moraine is at an altitude around 1,500 feet, or somewhat less.

A very well defined boulder ridge crosses Hardwood Island (Stark quadrangle) about 0.3 mile south of the east-trending portion of Dismal Swamp. The moraine strikes about east and is composed almost wholly of boulders. Boulders 2 feet or more in diameter are abundant.

On the Russell quadrangle, just east of Upper Degrasse School, there is a line of morainal hillocks across the valley of Plumb Brook.

The uniformity of level of the valley fill and many lakes on the Childwold quadrangle presents an interest-

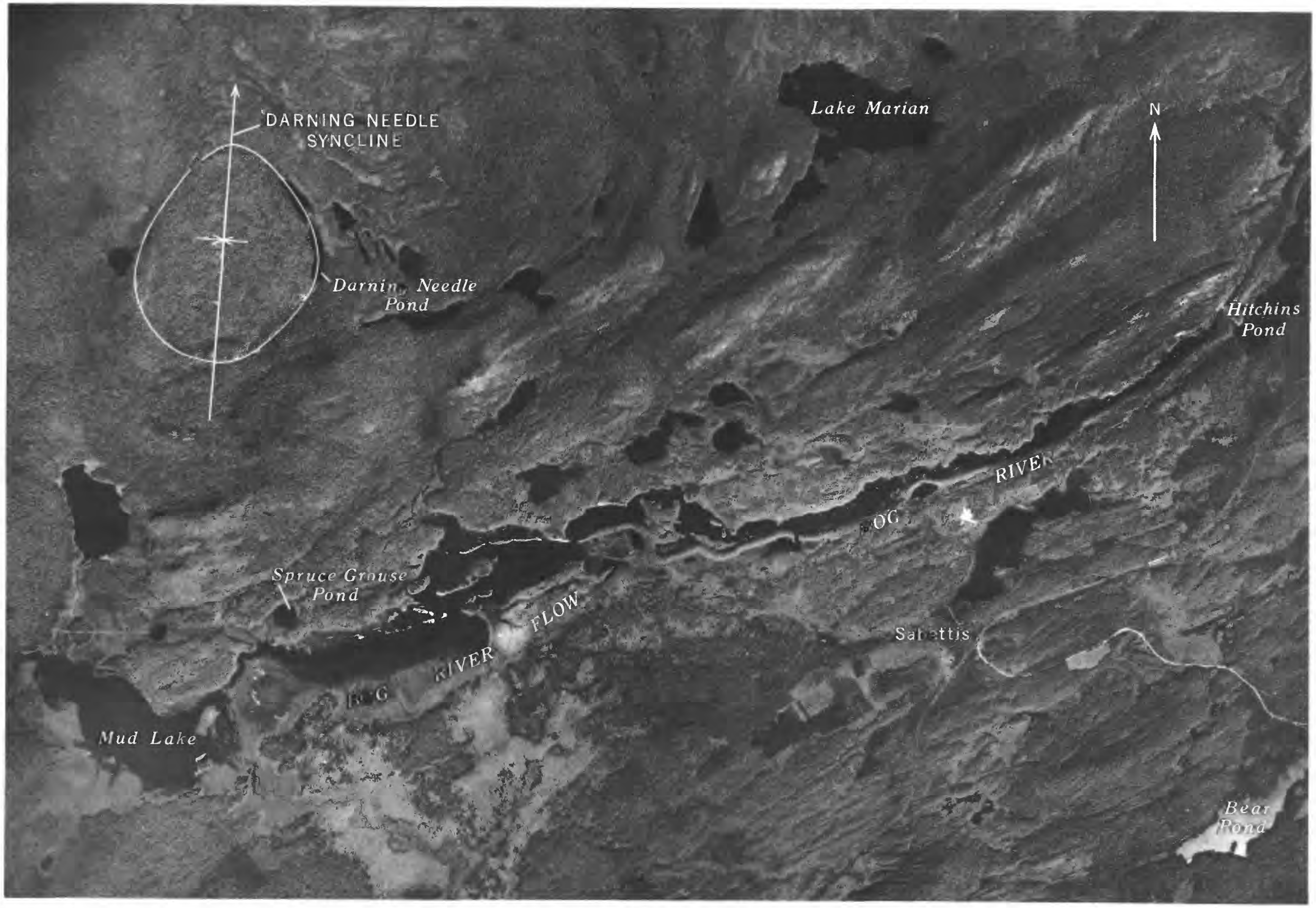

FIGURE 3.—Esker along Bog River from Hitchins Pond (upper right) to Bog River Flow and Spruce Grouse Pond. Topography in upper right, west of Lake Marian (just to right of top center), is in general conformable with foliation and structure of Darning Needle syncline. Ridges between Lake Marian and Bog River are parallel to foliation of granite. The topography between Bear Pond (lower right corner) and Sabattis (junction of road and railroad) is on sillimanitic quartz-microcline gneisses.

ing problem. The valley fill has the form of sand plains and swamps that slope upward from about 1,500 feet along the Raquette River to 1,530 feet at Kildare, and about 1,540 feet along the east side of the quadrangle. Most of the lakes and ponds have a similar variation in elevation. The most likely explanation for such uniformity of level of the deposits is that their surface was controlled by a former lake level. In order to explain a lake level above 1,500 feet, it is necessary to postulate a dam of ice across the St. Regis, Raquette, and Grass Rivers such that the outflow of the impounded waters was by way of the Oswegatchie River. A position of the ice outlined as follows would perhaps satisfy the necessary conditions. Ice must occupy the valleys of Windfall and Fallon Brooks (Childwold quadrangle) to block an outlet of the lake through this pass at an altitude of about 1,460 feet. The Dillon Pond delta (Cranberry Lake quadrangle) stands at 1,500 feet, and an ice front is indicated by the steep north slope. An ice front facing a lake is also indicated by the pitted sand plains north of Fallon Brook along the west side of the Raquette River. A tongue of ice must also have extended up the valley of the Raquette River and was perhaps continuous as a narrow connection with the tongue lying west of Long Rapids.

DELTA PLAINS

During the melting away of the glacier, its margin locally dammed back northward-flowing waters, yielding temporary lakes into which flowed streams from the land and from the ice to build deltas. Since retreat of the ice and drainage of the lakes, these deltas are now left standing above the adjoining country on one or more sides. One such delta deposit is present at and to the west of the village of Cranberry Lake and on the north side of the Oswegatchie River. Dillon Pond and Silver Pond are in the sand plain and their basins are assumed to represent the site of former buried ice blocks. Another similar sand plain and associated lake (Muskrat Pond) at a similar altitude to that at Cranberry Lake lies west of Heath Pond.

Star Lake (Oswegatchie quadrangle) is similarly in a sand plain that stands above the valley of Little River, and the basin of the lake is interpreted as occupied by an ice remnant while the surrounding deposits were made. Sand-plain levels around Star Lake, at altitudes of 1,420, 1,440, and 1,500 feet, mark ancient water levels which have been described by Dale (1935, p. 8).

"The Plains" (Cranberry Lake quadrangle), between the lower part of Glasby Creek and the Oswegatchie River, at altitudes between 1,600 and 1,620 feet, is an area of sand plain built into a temporary ancient lake by the Oswegatchie River. Just above the point where the Five Ponds trail crosses the Oswegatchie River, varved clays underlying sand are exposed in a bank along the river. These are the bottomset beds of the delta at The Plains. Another small delta built into this same lake lies between the head of Alice Brook and the outlet stream of Otter Pond. This delta rises toward the northwest and must have been built by streams coming from ice that temporarily was in this position. The outlet of this temporary lake may have been westward into the Middle Branch of the Oswegatchie River, over the divide west of Otter Pond.

Varved clays were also seen in a small gravel pit 800–900 feet east of the north end of Robin Lake (Bog Lake on Cranberry Lake quadrangle), where the wagon road swings from northeast to east; they have been described by MacClintock (1954, p. 12–13) as occurring in highway cuts 2 miles south of Russell, and also 1 mile south of Whippoorwill Corners.

A belt of deltaic deposits is well developed on the Russell quadrangle from Upper Degrasse School north through Degrasse to Lower District School. A delta (alt 1,080–1,100 ft) extends for 1 mile northwest of Upper Degrasse School. It has a steep slope on the north and a gentler slope on the south. This delta was probably built when the ice front stood along its north side and waters from the ice built out the delta into a temporary lake that was drained southwestward. South of Randall Brook there is a delta surface, 1,020–1,060 feet in altitude. The village of Degrasse and the Degrasse cemetery are on a flat sand and gravel plain, at altitude 860–900 feet, which is interpreted as a delta of the South Branch of the Grass River, built into a temporary lake. At a lower level there is the broad flat plain of the valley bottom from north of Degrasse to Lower District School, at an altitude of 800–820 feet. A high-level sand plain, at altitude 1,200–1,250 feet, occurs in the vicinity of the Orebed Ponds, near the Clifton mine. There is no drainage adequate to form such a plain, and it is thought that the plain is a delta built out by drainage from ice that lay along its western side. The basins of Orebed Ponds had their origin through melting of ice blocks buried in the delta.

The high-level (1,300–1,360 ft) sand plain at Stark has a steep north slope of the character that would have resulted from a previous ice contact. A line of kames and eskeroid ridges leads south towards it. The plain is thought to be a delta formed by drainage from ice at the north into a glacier-dammed lake. The lowest level of escape from the waters of such a lake would be a little above 1,320 feet, just south of Stark. The Raquette River would itself have been at least tem-

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porarily dammed by such an ice obstruction, and the sand plains near Hollywood (Childwold quadrangle) may represent the delta deposits of the river in such a temporary lake.

The level of the sand plains along Cold Brook (northeast Stark quadrangle) declines from a maximum of 1,220–1,240 feet north of Cold Brook School and at Irish Settlement to 1,208 feet at The Plains. The surface thus declines upstream in the opposite direction from the present valley. This indicates that again we have a delta deposit built into a temporary lake by drainage from a glacier to the north, whose front stood just southeast of South Colton. The abundance of pebbles of Potsdam sandstone in these deltaic deposits also indicates drainage from the north. There is a pass at the head of Little Cold Brook at an altitude of 1,200 feet that would have permitted waters at this level to drain to the Grass River.

Another conspicuous Pleistocene delta forms the area between Horseshoe Lake and Tupper Lake. The north side of this sand plain southeast and southwest from Black Pond is steep and must have been formed at contact with the ice front. The material of the flat between altitudes 1,840 and 1,880 feet is sand, in which boulders are distributed here and there. No ledges were found north of the road from Horseshoe to Tupper Lake. Between Black Pond and the north side of the headland opposite Rock Island Bay there are landslips on the steep north side of the delta.

There is also a small delta at the mouth of the Bog River at the head of Tupper Lake, west of South Bay at an altitude of about 1,680 feet, corresponding to a former higher level of Tupper Lake.

BOULDER AREAS

In several local areas large boulders are very abundant and appear to be almost wholly great blocks of rock which have not been moved far. One hillock after another has the appearance of a ledge broken up into blocks that have been rotated and somewhat displaced. The boulders are the same kind of rock that is found in place in the area. The resulting topography is rough and somewhat chaotic on a small scale. An area of such rock occurs between the two trails south of The Plains (Stark quadrangle) and on Buckhorn Ridge.

GLACIAL LAKE IROQUOIS

At one stage during the wasting away of the Pleistocene ice, the waters of the St. Lawrence were dammed by the ice front, so that they formed a lake whose waters spread as far south as the Potsdam and Russell quadrangles. This lake, which has been called glacial Lake Iroquois, was the much-expanded ancestor of the present Lake Ontario. The deltas formed in Lake Iroquois on the Russell quandrangle have been described by Fairchild (1919, p. 59) as follows:

> On this area the Iroquois waters reached far up the Grass River valley, among the Adirondack hills. The plains at Burns Flat and northward, on the east side of the quadrangle, probably represent the delta of the Grass.
>
> Wave erosion of Iroquois is well shown on the hills about West Pierrepont, and on the hill at Stone School, a mile southwest. Cliffs and bars appear on the northwest and the south faces of Kimball hill, north of Russell. An excellent development of beaches is found on the Hatch (Hamilton) hill, over a mile southwest of Russell . . . The summit bar of Iroquois is here 850 feet, and bars range down to 815 feet. The vertical range of 35 feet appears to represent the amount of land uplift at this locality during the life of Iroquois.
>
> The Oswegatchie river crosses the southwest corner of this quadrangle, and its delta is about South Edwards and Pond Settlement. The theoretic altitude here is 780 feet for the closing level, and the features range from 826 down to 780 feet, a vertical range of 46 feet.

The extensive sand plain whose highest altitude is about 820 feet and which extends from Smith Pond to north of Harmon School (southwest Russell quadrangle) evidently represents the delta of the Oswegatchie in Lake Iroquois.

The deltas of Lake Iroquois on the Potsdam quadrangle have been described by Fairchild (1919, p. 59) as follows:

> A good development of bars, with clear relations to the drift surfaces, gives fairly exact altitudes [for the old lake level]. One occurrence is a mile northeast of Parishville, where three elegant bars have altitude, 928, 905 and 885 feet. At Colton and the Clafin School the strong shore features have not been measured.

Recent work by P. MacClintock (oral communication) has led him to conclude that, in general, ice-contact phenomena show that many of the "deltas" are really kame terraces deposited in local lakes in the presence of stagnant ice masses and that the correlation of altitudes has little regional significance.

LAKES AND DERANGED DRAINAGE

The deposition of drift in quite uneven and irregular thickness in the valleys has resulted in the formation of many lakes and some conspicuous changes in drainage.

Many lakes have formed in the sand plains; their basins are the result of the melting of huge ice blocks, relics of the glacier, which were temporarily isolated and buried beneath the deltaic sands. Conspicuous examples of lakes of this mode of origin are Star Lake, Silver Pond, Dillon Pond, and Orebed Ponds. Other lakes are in depressions associated with belts of kame and esker deposits. Such are Sampson Pond (Stark quadrangle); Horseshoe Pond, Massawepie Lake, Town Line Pond, and Deer Pond (Tupper Lake quadrangle); Cowhorn, Clear, Grassy, Slender, Tamarack, and Big Deer Ponds (Cranberry Lake quadrangle); and Twin Lakes (Oswegatchie quadrangle). Many lakes are due to local obstruction of the valleys by local thicker deposits of drift.

The level of Cranberry Lake has been raised by an artificial dam, and the lake is much larger than it was originally. It is a relatively shallow lake and the deepest part is reported to be not over 40 feet.

The great bend in the course of the Oswegatchie River across the Cranberry Lake quadrangle—from several miles southwest of Wanakena through Cranberry Lake to Newton Falls—is quite inconsistent with the general slope of the topography and is probably not the preglacial course. The headwaters of the Oswegatchie probably originally flowed into the Middle Branch, perhaps by way of the lowland south of Otter Pond and Francis Hill (Oswegatchie quadrangle). The removal of only a small thickness of drift north of Wanakena would permit Cranberry Lake to drain by way of Little River, and it is possible that the old valley of Little River may have run where the kame belt now is—along Twin Lakes and Twin Lakes Stream (Oswegatchie quadrangle). The headwaters of the Middle Branch of the Oswegatchie River above Scanlons Camp doubtless also formerly flowed northward, through what is now the drift-clogged valley of Tamarack Creek, Star Lake, and Twin Lake Stream, to the main Oswegatchie.

The present course of the Grass River on the Russell quadrangle is equally inconsistent with the general topography. The South Branch of the Grass River would logically have flowed by way of the present valley of Plumb Brook and was probably deflected northwards by deposits of sand at Degrasse.

The Raquette River near Moosehead Rapids (Childwold quadrangle) is in a rock gorge for a length of 1 mile. A rock bench about 20 feet above the bottom of the gorge is locally glaciated, and the gorge is thus of postglacial origin.

POSTGLACIAL UPWARP

New York State north of New York City has been uplifted and differentially upwarped during and succeeding the disappearance of the Pleistocene ice, as shown by uplift of former marine shoreline features along the Lake Champlain valleys, correlation with deltas of marine origin in the northern Adirondacks, and by other similar kinds of data. The nature of this upwarp has been described by Fairchild (1919), and a revision is in preparation by Paul MacClintock for the New York State Science Service.

MacClintock (oral communication) states that marine shoreline phenomena of beaches and deltas are found up to an altitude of 525 feet at Covey Hill in the northeastern Adirondacks and at about 500–520 feet at Hermon (Russell quadrangle). Marine waters are also inferred to have been formerly at the level of the north edge of the delta at Hannawa Falls (Potsdam quadrangle). The marine isobases trend east-northeast and rise from about 500–520 feet on the north border of the Russell quadrangle to about 700 feet just north of Ottawa. They are inferred to decline in altitude to the south-southeast.

GEOLOGY OF BEDROCK

The bedrock of the St. Lawrence County magnetite district consists almost exclusively of Precambrian igneous rock and orthogneisses, and associated metasedimentary rocks of the Grenville series; a few Precambrian basaltic dikes are present. The Precambrian rocks are overlapped on the north and northwest by relatively flat-lying Potsdam sandstone of Late Cambrian age. However, only a few outlying residual patches of the sandstone have been found in place in the area of this report. Unconsolidated Quaternary gravels, sands, clay, and moraine overlie all the preceding rocks unconformably.

An outline of the geologic succession is presented in table 1.

IGNEOUS AND METAMORPHIC ROCKS

There is a very marked contrast between the nature of the predominant rocks of the Grenville lowlands (lying northwest of a line approximately through Carthage, Natural Bridge, Harrisville, Russell, and Allens Falls) and the Adirondack highlands (lying southeast of this line). The Grenville lowlands are underlain essentially by a metasedimentary complex and the Adirondack highlands by an igneous and orthogneiss complex.

The rocks underlying the Grenville lowlands consist of about two-thirds metasedimentary rocks and one-third igneous rocks, predominantly granite gneiss. In contrast, that part of the Adirondack highlands in St. Lawrence County and adjacent borders shown in plate 1 is underlain by a complex consisting of only about 13 percent metasedimentary rocks and 87 percent igneous rocks or equivalent orthogneiss. The approximate percentages of the total area occupied by the various kinds of igneous rocks in the main igneous complex of the area shown in figure 4, exclusive of the anorthosite mass, follows:

Percentage of area occupied by different rocks in main igneous complex of the northwestern Adirondacks

	Percent
Hornblende granite, biotite alaskite, and equivalent gneisses	47
Microcline granite and microcline granite gneiss (with some associated alaskite)	11
Phacoidal hornblende-quartz syenite gneiss and phacoidal hornblende granite gneiss (quartz syenite series)	12
Pyroxene syenite gneiss and pyroxene-quartz syenite gneiss (quartz syenite series, charnockitic series)	14
Metagabbro and amphibolite	3
Metasedimentary rocks and migmatites	13

The igneous rocks of the Adirondack province are similar to those of several other areas of old Precambrian terrane throughout the world, but in their nature and field relations they have many characteristics which contrast strongly with igneous complexes of younger orogens.

Andesine and labradorite anorthosite, in masses of batholithic size and of apparently intrusive character such as form the eastern core of the Adirondacks, is not positively known in formations younger than the Precambrian. Charnockitic rocks in extensive development, such as the Tupper complex, are restricted exclusively to the older Precambrian, though rocks similar to one or the other members of the charnockitic group do occur locally in younger formations. The main members of the granite family in the Adirondack province are almost exclusively normal granite containing normative oligoclase or albite, and a percentage of normative orthoclase about equal to or slightly greater than the normative plagioclase. Except as very local facies due to contamination in local contact zones with amphibolite, there are no associated granodiorites or quartz diorites of the sort found in the Cretaceous Sierra Nevada and Coast Range batholiths of western North America, the Paleozoic batholiths of the Appalachian orogen in eastern North America, and most other batholithic complexes of late Precambrian and post-Cambrian age. Also, no lamprophyric or peridotitic intrusives have been found in the Adirondacks. Only the gabbro sheets are similar in nature and field occurrence to those of younger age at many places throughout the world.

TABLE 1.—*Rock formations in the St. Lawrence County magnetite district, New York*

System	Series	Rocks	
Quaternary	Recent	Swamp accumulations and local river and lake deposits.	
	Pleistocene	Gravel, sand, clay, and marine deposits.	
		Unconformity	
Cambrian	Upper Cambrian	Potsdam sandstone; sandstone and conglomerate.	
		Unconformity	
Precambrian	Upper Precambrian	Basalt, as dikes.	
		Intrusive contact	
	Upper (?) Precambrian		
	Period of deformation and metamorphism, locally intense		
		Amphibolite, locally as sheets and dikes.	
	Granite and granite gneiss series (syntectonic emplacement).	Alaskite and alaskite gneiss, commonly with accessory fluorite, locally contaminated with garnet, biotite, or plagioclase; subordinate biotite granite gneiss.	Microcline granite and microcline granite gneiss, in part contaminated with quartz-sillimanite aggregates, biotite, plagioclase, almandite, hornblende, or pyroxene derived from metasedimentary rocks and amphibolite; subordinate albite-oligoclase granite masses. Age relations to other granite masses and to quartz syenite gneiss series is unknown.
		Hornblende granite and hornblende granite gneiss.	Porphyritic biotite granite and augen gneiss (Hermon granite gneiss), locally hornblendic; occurs only in "Grenville belt" of metasedimentary rocks.
		Intrusive contact	
	Age uncertain	Fayalite-ferrohedenbergite granite.	
		Intrusive contact	
	Period of orogenic deformation and metamorphism		
		Metadiabase, as dikes.	
		Intrusive contact	
	Period of orogenic deformation		
Contemporaneous	Quartz syenite gneiss series (Diana and Stark complexes).	Hornblende granite gneiss. Hornblende-quartz syenite gneiss. Pyroxene syenite gneiss with lenses of shonkinite and feldspathic ultramafic gneiss, both carrying much ilmenite and magnetite.	
	Charnockitic gneiss series (Tupper complex).	Ferrohypersthene-ferroaugite-hornblende-quartz syenite gneiss. Ferrohypersthene-ferroaugite syenite gneiss.	
		Intrusive contact	
		Anorthositic gabbro gneiss, gabbro gneiss, diorite gneiss, and amphibolite.	
		Intrusive contact	
	Grenville series	Metasedimentary rocks and migmatites; biotitic, hornblendic, and pyroxenic plagioclase-quartz or feldspar-quartz gneisses, garnetiferous feldspar-quartz gneisses, feldspar-quartz granulite, quartzite, amphibolite, marble, and skarn.	

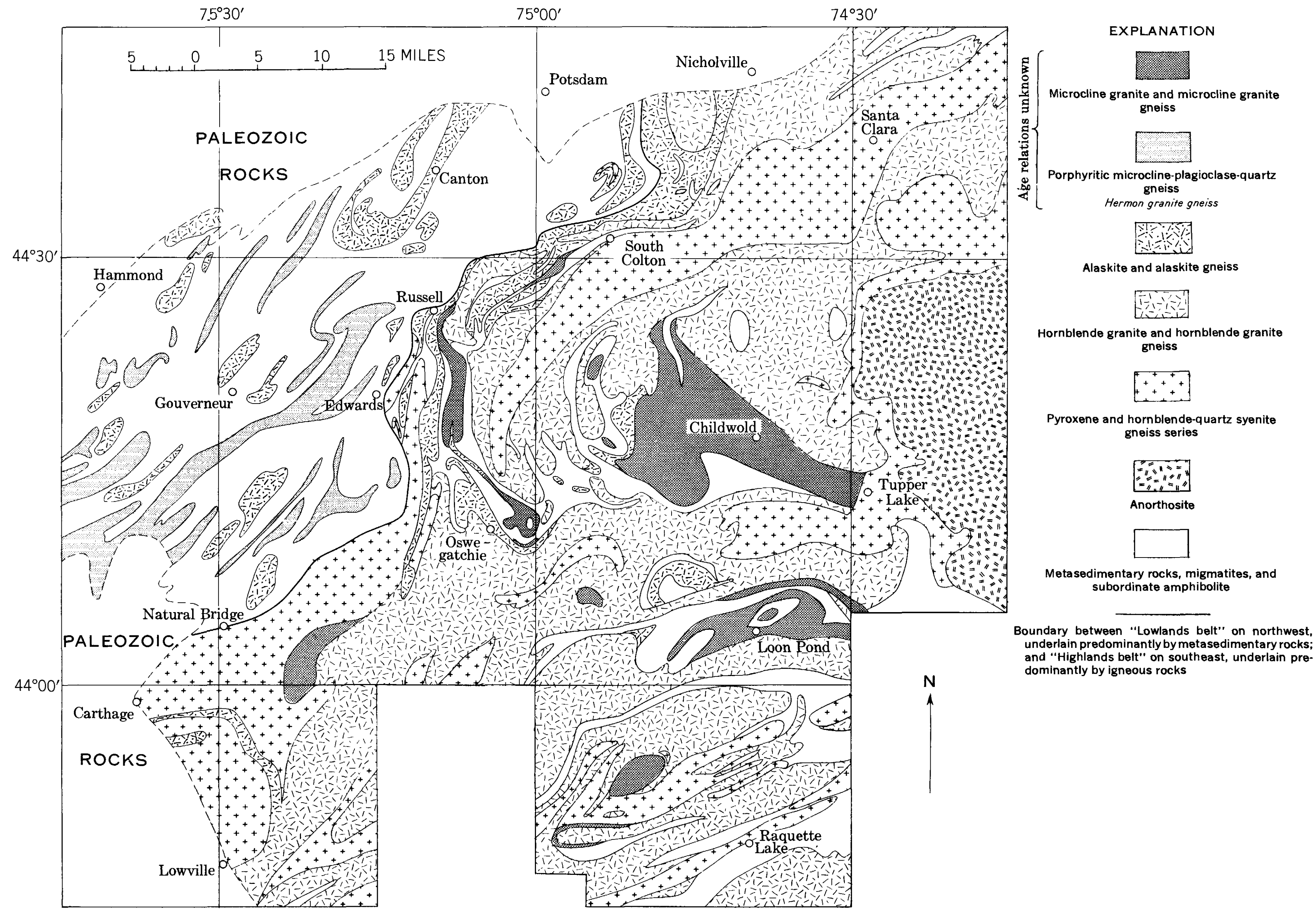

FIGURE 4.—Map showing distribution of major rock types in the northwest Adirondacks.

The major belts of metasedimentary rocks and their associated microcline granite or granite gneiss, alaskite or alaskite gneiss, and subordinate amphibolite are thought to be in synclinoria or synclines which have been named as follows: Boyd Pond synclinorium from east of South Colton southwest toward Russell; South Russell synclinorium south of Russell; Benson Mines synclinal structure east and north of Oswegatchie; Bog River synclinorium east-west through Loon Pond (Tupper Lake quadrangle); Dead Creek and Darning Needle synclines (Cranberry Lake quadrangle), which form small structures north of the Bog River synclinorium; the Clare-Clifton-Colton belt, which starts west of Tupper Lake and extends in a great arc west and then north and northeast to points about 7½–10 miles south of South Colton; and the McCuen Pond syncline, about halfway between South Colton and Childwold. The microcline granite gneiss area around Childwold also appears to have a synclinorial structure. Major anticlinal structures include the quartz syenitic mass north of Lowville and extending to south of Natural Bridge, with one limb of an anticline extending north to within a few miles of Russell; the Stark anticline, formed by the quartz syenite gneiss series extending through Santa Clara to about 4 miles north of Oswegatchie; and the Arab Mountain anticline of quartz syenitic rocks west and southwest of Tupper Lake.

The best evidence we have as to the age of any of the Precambrian rocks is based on the age of an allanite crystal associated with a granite pegmatite from Mount Whiteface (Essex County) and of a uraninite crystal from a granite pegmatite near Richville, St. Lawrence County. These granite pegmatite veins are thought to be genetically related to the granite and granite gneiss series. The age of the uraninite crystal from the McLear granite pegmatite quarry near Richville was determined by Shaub (1940) to be 1,094 million years on the basis of the lead: uranium-thorium ratio. Holmes (1948, p. 180) later gave a corrected apparent age of 1,020 million years. The age of the allanite crystal from Whiteface Mountain was determined by Marble (1943) to be about 1,200 million years also on the basis of lead: uranium-thorium ratios. G. R. Tilton (personal letter to H. D. Holland) determined the age of a zircon from marble at contact with syenite of the Diana complex about 1 mile east of Natural Bridge to be 1,110±50 million years by the Pb^{206}/Pb^{207} isotope method.

The metasedimentary rocks of the Grenville series may be very much older than the granite pegmatites, perhaps by as much as hundreds of millions of years.

A series of unmetamorphosed Precambrian basaltic dikes, similar to those of the Adirondacks, are a characteristic feature of large areas across the St. Lawrence River in eastern Ontario and western Quebec. The Canadian dikes are referred to by Moore (1929) as of Keweenawan age, and the Adirondack basaltic diabase dikes may be tentatively correlated with them.

AGE RELATIONS OF IGNEOUS ROCKS AND PERIODS OF DEFORMATION

Anorthosite forms a great massif in the eastern Adirondack province and is known to be cut by dikes of gabbro, pyroxene syenite, quartz syenite, hornblende granite, and alaskite. The anorthositic rocks must therefore belong to an early period of intrusion. Whether there are any older igneous rocks in the Adirondacks is not yet proved. The diorite sheets of the Tupper Lake quadrangle are within granite and are cut by it, but no relations to other intrusives have been seen. It is possible that the diorite is genetically related to the anorthositic rocks.

Olivine gabbro is known to have intrusive relations to the main anorthosite massif, and the anorthositic gabbro near Russell is cut by dikes of pyroxene gabbro. Southeast of Rock Pond (Childwold quadrangle), layers of pyroxene-quartz syenite gneiss alternate with amphibolite. The relations are obscure, but the amphibolite appears to be in the form of lenses included in the quartz syenite gneiss. The metagabbro is therefore thought to be in part younger than the anorthositic rocks and older than the quartz syenite complexes.

In a succeeding discussion of the Diana complex, evidence will be given that the magma of this complex was intruded into the sedimentary rocks of the Grenville series as a more or less comformable, fairly flat lying sheet. This hypothesis also carries with it the implication that at the time of intrusion the sedimentary rocks too were only gently folded and tilted, except perhaps where adjacent to the anorthosite massifs. It is believed that the anorthositic, dioritic, gabbroic, and quartz syenitic magmas were all intruded as conformable sheets into generally flat lying sedimentary rocks. However, considerable brecciation, transgression, and disturbance of the country rock did accompany these intrusions, as is the case with most intrusive sheets.

The first period of widespread major deformation probably followed the emplacement of the quartz syenite complexes and preceded the intrusion of a set of hypersthene metadiabase dikes. The older rocks were probably folded before the emplacement of the hypersthene metadiabase, because (Buddington, 1939, p. 2) "the limbs of the folds of the Diana complex where they strike NE are cut by a series of hypersthene metadiabase dikes striking NW, 'but . . . where the limbs of the fold strike NW the metadiabase dikes are parallel to the banding and foliation. . . ." The metadiabase dikes show fine-grained chilled border facies and could not have been intruded at great depth.

Following the emplacement of the dikes, the entire group of rocks was again subjected to intense orogenic deformation under such conditions that within the main igneous complex of this area the gabbroic rocks were largely granulated and reconstituted to amphibolites and garnet amphibolites, the diabase dikes were converted to amphibolite and to granoblastic garnet-pyroxene-plagioclase granulites, and the microperthite-quartz syenitic rocks were changed to augen, flaser, and granoblastic gneisses, locally garnetiferous in the eastern portions and with the microperthite partly or wholly recrystallized to plagioclase and potassic feldspar.

Evidence from the geology of the Nicholville, Santa Clara, and Lowville quadrangles has previously been presented (Buddington, 1939, p. 149–152) to show that the quartz syenite complexes and older rocks were folded and strongly deformed before the intrusion of the hornblende granite. This conclusion is based upon the fact that the hornblende granite cuts the folds in which the quartz syenite complexes are involved. The belt of pyroxene syenite gneiss that occurs locally on each side of the phacoidal granite gneiss across the Stark quadrangle is terminated at the ends as though cut out by a transgression of the hornblende granite. On the eastern part of the Russell quadrangle there is good evidence that the west limb of an anticline in phacoidal granite gneiss has been partly cut away by the hornblende granite. This is especially evident for 2 miles south of Big Swamp, where the contact of the younger granite is at an angle to the foliation of the older phacoidal granite. It is also shown east of Degrasse by the layer of older granite included in the younger granite, and 2½–5 miles south of Degrasse by the contact of the younger granite, which again transgresses the foliation of the older structure. In the vicinity of Center Pond Mountain (Tupper Lake quadrangle) and Cranberry Lake, the continuity of the anticline of pyroxene syenite gneiss is interrupted. This may be explained in part as a result of erosion of the syenite gneiss on a "culmination" of the crest of the anticline, but must also be in part a consequence of disruption of the syenite gneiss by the younger granite, for the continuity is completely broken on the limbs as well as on the crest. Attention is also called to the isolated masses of pyroxene-quartz syenite gneiss that are included in the granite south of Muskrat Pond (Cranberry Lake quadrangle).

The pregranite age of one intense period of orogenic deformation is shown not only by the transgression of hornblende granite bodies across the earlier folded structures, but also by the more intense deformation texture of the older rocks as contrasted with that of the granite. This will be described in detail under the heading, "Quartz syenite series and younger granite: contrast in metamorphism."

The alaskite locally grades into the hornblende granite and is thought to represent merely a late facies of it. Dikes of alaskite have been found at a number of localities in members of the quartz syenite series, but not in the hornblende granite to which it is related.

The relations of several of the igneous rocks are excellently shown in the hill just north of Alice Brook and 0.75 mile west-southwest of the south end of Nicks Pond (Cranberry Lake quadrangle). The oldest rock here is a maple-sugar-colored, finely granular gneiss with a positive but indistinct coarse phacoidal structure and quartz leaves. In thin section the feldspars are found to form a granoblastic mosaic and the hornblende is in flattened granular aggregates, locally somewhat porphyroblastic. This layer of gneiss is cut by a dike of mafic gneiss with a granoblastic texture. The foliation of the phacoidal gneiss strikes N. 40°–50° E., whereas the dike strikes about N. 30° W. The phacoidal gneiss together with the dike occurs as a layer included within a pink gneissoid granite that shows much less crushing and deformation than the gneiss. The mafic dike is cut by an aplitic granite dike. The granite is predominantly a fluorite-bearing alaskite, but it is locally hornblendic from contamination with amphibolite. The phacoidal gneiss is interpreted as a layer of pyroxene-quartz syenite gneiss torn off by the alaskite and so metamorphosed that the pyroxene is all changed to hornblende. The mafic dike is thought to belong to the group of north-northwest- to northwest-trending hypersthene metadiabase dikes which occur widely in the Diana complex. The diabase of the dike is here strongly metamorphosed to a granoblastic aggregate of garnet, hornblende, hypersthene, augite, and plagioclase.

The relative age of the fayalite-ferrohedenbergite granite presents a most puzzling problem. Rock of this type has so far been found at only two localities in the Adirondacks, and in both cases it is in direct association with the quartz syenite gneiss series, to which it is similar in mineralogy. Yet it is so much less deformed and recrystallized than the associated gneiss that it does not seem possible that it can be of the same age. Its massive character seems to necessitate that it was intruded after the period of deformation of the quartz syenite gneiss series. The hornblende granite was inferred at one locality, though not on conclusive evidence, to cut it. Actually, it is possible that the fayalite-ferrohedenbergite granite is younger than the hornblende granite.

No evidence has been found within this area to give a clue to the age relations of the hornblende granite and alaskite to the Hermon granite gneiss (porphyritic biotite granite gneiss). The problem has been previously discussed (Buddington, 1939, p. 158–161) and the conclusion reached that "The correct solution of the age and relationships of the granites yet remains to be demonstrated." Pending definitive data the Hermon granite gneiss may be thought of as developed from a volatile-rich equivalent of the hornblende granite magma of the main igneous complex, both by intrusion and by replacement, in which case the alaskite is a somewhat younger facies.

The age relations of the microcline granite gneisses have not been satisfactorily determined. They are known to be younger than the metasedimentary rocks and amphibolite, and generally older than or contemporaneous with the last major deformation that affected the hornblende granite and alaskite. No contacts between the microcline granite gneisses and the hornblende granite and alaskite or the quartz syenitic rocks have been found.

On the Lake Bonaparte quadrangle (Smyth and Buddington, 1926, p. 42; Buddington, 1939, p. 139) there is a fine-grained hornblende- or biotite-bearing microcline-rich granoblastic granitic gneiss (called granosyenite gneiss in those reports), which is intruded by granoblastic metadiabase dikes of the same character as those that intrude the rocks of the Diana complex. Too little is known about this granitic gneiss to evaluate its significance in the history of the geology of the Adirondacks.

The relative degree of metamorphism of the rocks in the St. Lawrence County district has been studied for any evidence it may contribute to the problem of age relations. All the microcline granite gneiss has a mosaic texture. Part of the microcline granite gneiss with this texture is known to have undergone intense deformation and recrystallization, and since all has similar texture, all might be ascribed to a similar origin. It is equally probable, however, that granitization of metasedimentary rocks could give rise to a similar texture without deformation. A qualitative study of the orientation of the crystal structure of the iron-titanium oxide minerals suggests that substantial crystallization or recrystallization must have occurred after the last deformation, for the basal plane of much of the ilmenohematite is at any angle up to right angles to the foliation. Since the origin of a large part of these gneisses is ascribed to granitization, the texture alone cannot be taken as a criterion as to the degree of deformation. Again, the microcline granite gneiss in all large areas locally shows plications and slips, and shears across the foliation, which are more intensely developed than in the surrounding or associated hornblende granite and alaskite. This would definitely suggest that the microcline granite gneiss was older than the hornblende granite, if it were certain that both had similar strengths. The microcline granite gneisses, however, have many schlieren and layers of metasediment, and therefore were probably weaker than the more uniform hornblende granite. The microcline granite gneiss may have yielded to a greater degree than the hornblende granite with the same intensity of deforming forces, though this seems difficult to believe.

Consideration of the origin of the microcline granite gneiss has indicated that on theoretical grounds it is possible for it to have originated as a subordinate potassium-rich and volatile-rich facies of the same magma as that which yielded the hornblende granite and alaskite. The almost exclusive restriction of the microcline granite gneiss to the main igneous complex in association with major volumes of hornblende granite and alaskite is consistent with this idea. On this hypothesis the microcline granite gneiss might preferably be thought to be a younger facies. However, since it is so largely a metasomatic rock, it could be developed contemporaneously with or well ahead of the emplacement of the hornblende granite and alaskite magmas at their present levels.

Definite determination of the age relations of the microcline granite gneisses to the hornblende granite and alaskite must await better data.

One mass of alaskitic gneiss in the Grenville lowlands has been found by R. V. Dietrich (oral communication) to be cut by sheets and dikes of amphibolite, which are in turn cut by granite pegmatite veins. How much of the amphibolite within the main igneous complex is of this age is unknown but is thought to be very small.

A repetition of orogenic deformation accompanied and followed the emplacement of the granitic rocks. The effect of this deformation was most intense west of the belt of the Stark complex of quartz syenite and phacoidal hornblende granite gneiss. This complex forms a belt across the eastern part of the Russell quadrangle and north part of the Stark quandrangle. The deformation of the younger granitic rocks (hornblende granite, alaskite, and perhaps microcline granite) was in general moderate south and east of the Stark complex, and moderate to slight in the granites south of the quartz syenite sheet through Inlet and south of the belt of syenite through Arab Mountain.

Northwest of the main igneous complex, in a wide belt of metasedimentary rocks of the Grenville lowlands, the porphyritic biotite granite gneiss (Hermon granite gneiss) sheets and the alaskite phacoliths—which are thought to have been emplaced in already folded rocks—have in turn been intensely deformed and compressed in strongly overturned isoclinal structures with accompanying recrystallization. Within the main igneous complex in the belt west of the Stark complex, all the younger granites have similarly been deformed into isoclinal structures with accompanying recrystallization.

It is certain that the rocks were strongly deformed before the emplacement of the granite series, and that the granites in turn were locally intensely deformed both contemporaneously with and subsequently to their emplacement. It is possible, however, that these periods of deformation should be considered as two of a series of epochs of a single great orogeny rather than as two separate orogenies.

The youngest igneous rocks in the area are the basaltic dikes, of which there are very few.

GRENVILLE SERIES

The Grenville series includes both metamorphosed sedimentary rocks that have been recrystallized and reconstituted to yield new minerals with but little change in gross chemical composition, and metamorphosed sedimentary rocks that have had their chemical composition substantially changed in addition to undergoing recrystallization and reconstitution. The latter are of mixed origin and, with the exception of skarns, commonly have a distinct veined or thin-layered structure, though certain permeation gneisses may be more or less homogeneous.

The rocks of the Grenville series are extremely varied in character. They include pyroxene-quartz-plagioclase gneisses, usually carrying several percent of potassic feldspar and a little sphene; pyroxene-microcline or pyroxene-orthoclase gneiss with a little scapolite; biotite-quartz-plagioclase gneiss, containing a varied but subordinate amount of potassic feldspar and up to several percent of iron oxides; quartz–potassic feldspar gneisses and granulites with varied amounts of biotite or pyroxene; hornblende-quartz-feldspar gneisses and amphibolite; garnetiferous biotite-quartz-feldspar migmatitic gneiss; quartzite with various accessory minerals; marble, usually containing disseminated silicates; and local skarn.

Weathering of the rocks of the Grenville series usually produces a distinctly laminated appearance and a ribbed surface, arising from the differential resistance of alternating folia and layers. The pyroxene gneisses, especially the calcareous facies, often yield a rust-colored, loosely granular aggregate at the outcrop. The quartzitic rocks of the Robinwood area (Cranberry Lake quadrangle) are generally massive, though the weathered surface is often leached and pitted, so that quartz is the only visible mineral.

METASEDIMENTARY ROCKS

PYROXENE- AND HORNBLENDE-QUARTZ-FELDSPAR GNEISS

Pyroxene-quartz-plagioclase gneiss and pyroxene–quartz–potassic feldspar gneiss together constitute a major facies of the rocks of the Grenville series in this area. Gneisses of this type are found abundantly in every one of the belts of metasedimentary rocks within the main igneous complex. They are also associated with the belt of quartz-feldspar granulite on the northwest flank of the main igneous complex. Hornblende-quartz-plagioclase gneisses also occur, but in subordinate volume.

The pyroxene gneisses are greenish or grayish, light to dark, commonly medium grained though locally fine grained, and on weathering they usually yield a rusty, limonite-stained, granular sand or a rusty, slightly porous outcrop. The outcrops commonly show a foliation or rough-ribbed character as a result of the variation in percentage of pyroxene and other minerals. On the fresh surface the rock has a speckled or foliated appearance, owing to the medium to dark green pyroxene grains or aggregates. Locally the pyroxene gneiss is calcareous. Common accessory minerals are iron oxides, sphene, scapolite, and apatite, and occasionally a little sulfide. The mineral composition of representative examples is given in table 2.

Some of the pyroxene gneisses have a veined character. These will be described below under the heading "Mixed or veined gneisses."

At many places, layers of hornblende-quartz-plagioclase gneiss a few inches to several feet thick are interbedded with the other gneisses. These hornblendic gneisses have the appearance of amphibolites and might be called para-amphibolites. They differ from the more common and characteristic amphibolites in carrying some quartz, quite independently of any evidence for granitization. They grade into the pyroxene-quartz-plagioclase gneisses on the one hand and into biotite-quartz-plagioclase gneisses on the other, and they are intermediate in character between the two. The mineral composition of several representative examples is given in table 2. They are a subordinate facies of the metasedimentary rocks.

An amphibolite 1.2 miles north-northeast of Partlow Lake (Cranberry Lake quadrangle) is of particular interest because it contains 27 percent quartz and the

TABLE 2.—*Modes of pyroxene gneisses and hornblende gneisses: facies of metasedimentary rocks of the Grenville series*

[Volume percent]

Sample No.	Quartz	Scapolite	Plagioclase	Microcline, orthoclase, or both	Pyroxene	Hornblende	Biotite	Garnet	Sphene	Iron and iron-titanium oxides	Zircon	Apatite	Calcite	Sericite
Pyroxene-quartz-plagioclase gneiss														
1	15.0	0.4	53.0	9.1	7.3	7.7	4.7			2.4	0.1	0.3		
2	16.0		66.7	9.3	2.5	3.0			1.7			.6		
3	31.4	16.5	31.5		17.0				1.8	1.2		.6		
4	38.1	1.6	32.6	5.5	18.9				.7	1.0		.2	1.4	
5	42.7		39.3	2.0	8.4	3.8	.4		1.5	.1	.1	.1	1.1	
6	45.9	10.0	8.2	6.1	27.6					1.8		.4		
Pyroxene-quartz-potassic feldspar gneiss														
7	39.0	4.2	2.1	37.0	16.0				1.2			0.5		
8	24.4			26.6	23.5		4.9		.1	0.4		.1		20.0
9	47.5			38.3	6.7	5.1	.6		1.5			.3		
Hornblende-quartz-scapolite gneiss														
10	12.8	51.8				33.2	0.8			1.0		0.4		
Hornblende-quartz-plagioclase gneiss														
11	19.0		59.2		1.3	15.2			1.1	3.8		0.4		
12	25.6		33.2	13.1	7.0	8.7	12.2					.2		
13	26.5		18.2			33.3	18.5			2.5		.9		
14	30.5		21.5			27.8	17.1			2.4		.7		
Pyroxene-feldspar gneiss														
15	1.4	1.1	24.2	32.4	38.4	1.3	0.2		0.8			0.2		
16	3.2		77.9	2.0	16.0				.7			.2		
17	10.3			74.5	9.3	.3	1.2		1.7			.3	1.9	

1. 1 mile west of head of Dead Creek, Cranberry Lake quadrangle.
2. 1 mile north of Chaumont Pond, Cranberry Lake quadrangle.
3. 0.8 mile east of Gooseberry Mountain, Stark quadrangle.
4. Average of 4 similar gneisses, Gooseberry Mountain area, Stark quadrangle.
5. Average of 4, Brandy Brook, Cranberry Lake quadrangle.
6. 0.4 mile west of Dillon Pond, Cranberry Lake quadrangle.
7. Gooseberry Mountain, Stark quadrangle.
8. Deerlick Rapids, diamond-drill hole 2, Stark quadrangle.
9. 0.5 mile south-southeast of mouth of Dead Creek, Cranberry Lake quadrangle.
10. 0.9 mile northwest of Tunkethandle Hill, Stark quadrangle.
11. 0.6 mile northeast of Boyd Pond, Russell quadrangle.
12. 0.4 mile east of Dodge Pond, Russell quadrangle.
13. North of Partlow Pond, Russell quadrangle.
14. 0.3 mile west of Triangle Pond, Tupper Lake quadrangle.
15. 0.7 mile northwest of Benson Mines, Oswegatchie quadrangle.
16. 1 mile southeast of Irish Hill School, Russell quadrangle.
17. Brandy Brook magnetite prospect, Cranberry Lake quadrangle.

plagioclase (18 percent) is a bytownite. The normal plagioclase is an andesine or oligoclase.

One thin layer of hornblende-scapolite gneiss was found as an included layer in alaskite.

PYROXENE-FELDSPAR GNEISS

Pyroxene-feldspar gneiss forms a major member of the metasedimentary rocks of the Grenville series within the main igneous complex. It has been particularly noted in the Benson Mines syncline and in the drill core of the Parish, Spruce Mountain, Deerlick Rapids, and Brandy Brook magnetite zones. The ratio of pyroxene to feldspar is highly varied. The rock grades on the one hand into pyroxene skarn and on the other hand into a quartz-feldspar gneiss containing pyroxene as an accessory mineral. The feldspar is usually microcline, but local facies may be predominantly plagioclase. Quartz is usually present as an accessory mineral, and as much as 1 percent of sphene is almost always present. Apatite is also a common accessory mineral.

Chemical analyses of two representative samples of the pyroxene-potassic feldspar gneiss are given in table 3.

QUARTZ-RICH METASEDIMENTARY ROCKS

Rocks of the metasedimentary series that contain 50 percent or more of quartz are here classified as quartz rich. All the quartz-rich rocks in this district carry other minerals than quartz in substantial amounts, and there is practically no true quartzite. Feldspar, usually microcline, is present in practically all the rocks, but there is a wide variety in the nature and amount of the other minerals, as is shown by the mineral composition of representative samples in table 4.

The predominant member of the belt of metasedimentary rocks between Bog Lake and Tomar Mountain (Cranberry Lake quadrangle) is a white quartzitic rock. The quartzitic rock is in large part a calcareous variety whose exposed portions weather locally to a coarse white quartz sand. Where the rock has retained its coherence, it is deeply pitted and has only the most obscure evidence of layering. The rock is gray to

TABLE 3.—*Chemical analyses and norms of feldspathic quartz-rich gneisses and pyroxene-feldspar gneisses of the Grenville series*

	1	2	3	4
Chemical analyses (weight percent)				
SiO_2	81.45	76.24	54.12	55.18
Al_2O_3	7.83	12.88	11.95	14.40
Fe_2O_3	1.58	.35	2.51	1.45
FeO	.78	5.04	4.59	4.07
MgO	.35	.47	6.22	2.79
CaO	.71	.46	10.61	10.88
Na_2O	1.46	.14	.36	2.18
K_2O	4.64	2.68	7.08	6.35
H_2O+	.14	.45	.73	.18
H_2O-	.05	.08	.02	.03
CO_2	.39			.68
TiO_2	.42	.58	.95	.62
P_2O_5	.08	.13	.16	.59
MnO	.04	.10	.30	.13
Total	99.92	99.60	99.60	99.53
Norms				
Quartz	54.39	60.03		
Orthoclase	27.52	16.12	42.26	37.81
Albite	12.31	1.05	3.14	16.77
Anorthite	.56	1.39	10.01	10.56
Corundum	.15	9.18		
Diopside			33.30	25.29
Wollastonite				1.86
Hypersthene	.90	9.45	3.46	
Magnetite	1.51	.46	3.71	2.09
Hematite	.56			
Ilmenite	.76	1.07	1.82	1.22
Apatite	.20	.33	.37	1.41
Calcite	.90			1.55
Olivine			1.00	
Nepheline				.85

1. Quartz-feldspar-granulite, 1.6 miles southeast of East Pitcairn, Oswegatchie quadrangle (Buddington, 1939, p. 14). Analyst, A. Willman. A little chlorite present.
2. Almandite-sillimanite-biotite-quartz-microcline gneiss forms groundmass for sillimanite-quartz nodules, road cut 0.6 mile west-northwest of Bear Pond on road to Sabattis, Tupper Lake quadrangle. Analyst, Eileen H. Kane.
3. Micaceous pyroxene-feldspar gneiss, diamond-drill hole 2, Deerlick Rapids prospect, southern part of Stark quadrangle. Analyst, Lee C. Peck. The rock consists predominantly of clinopyroxene and feldspar, and several percent phlogopitic mica. The feldspar includes both orthoclase and plagioclase, the orthoclase in larger volume. The plagioclase is largely altered to sericite. Accessory minerals include quartz, magnetite, apatite, and a trace of sphene.
4. Pyroxene-feldspar gneiss, diamond-drill hole 6, Spruce Mountain prospect, northwest corner Cranberry Lake quadrangle. Analyst, Lee C. Peck.

white, glassy, pitted, and fine- to medium- or coarse-grained (generally rather coarse). A little pyroxene and calcite are locally recognizable. The impurities are conspicuously more evident on microscopic examination than in the field. Two relatively "pure-appearing" specimens have 62 and 73 percent quartz respectively, together with substantial calcite, potassium feldspar, pyroxene, and scapolite. Epidote forms thin rims between calcite and quartz in some of the rock.

Fine-grained quartz-feldspar granulite forms much of a belt of rock flanking the main igneous complex on the Russell and Oswegatchie quadrangles, as for example from east of Moores Corners (Russell quadrangle) through Hamilton's Corners, Russell, and for several miles to the southwest of Russell, and again southwest of Red School (Oswegatchie quadrangle). It also forms part of the belt of metasedimentary rocks between School No. 15 and Sucker Lake (Oswegatchie quadrangle) and occurs as included leaves in the northwest part of the Diana complex. Fine-grained quartz-feldspar granulite also occurs as included layers in the phacoidal granite gneiss on the east flank of the Stark anticline, as at Clifton, and forms interbeds within the metasedimentary rocks of the belt across the north part of the Cranberry Lake quadrangle.

Most of the quartz-feldspar granulite is fine grained (0.1–1 mm) with an indistinct gneissic structure. The color is pale gray to greenish or pink. Much of the rock, as it appears in the field and in drill core, is so similar to granoblastic alaskite gneiss that each may easily be mistaken for the other. These rocks have often been called feldspathic quartzites. In general appearance the rock resembles one called leptite in the

TABLE 4.—*Modes of quartz-rich facies of metasedimentary rocks of the Grenville series, including quartz-potassic feldspar granulites*

[Volume percent]

	Quartz	Plagio-clase	Micro-cline, ortho-clase, or both	Pyrox-ene	Horn-blende	Biotite	Garnet	Sphene	Iron oxides	Zircon	Apatite	Scapo-lite	Calcite	Sericite	Silli-manite	Tremo-lite
1	57.7	0.5	39.6			1.9			0.2		0.1					
2	52.2	16.8	23.2			6.5			1.1		.2					
3	69.0		26.0			4.0			.3					0.7		
4	73.0	4.5	15.9			.2	4.3		.5						1.6	
5	50.8		20.8			2.5	15.7		1.5						8.7	
6	53.9		37.8		4.5	1.5			.3		.3	1.4	0.2			
7	54.8	23.7	7.2		11.3				1.7				1.3	.3		
8	56.5		32.9	5.2	.2	1.9		0.3	.4				2.5			
9	53.9	9.9	32.5	2.4				1.1	.2							
10	67.9		3.8	7.8	1.0	.1		.2				.5	15.6	3.1		
11[1]	86.6			10.2												2.9
12	48.0	1.2	47.2		1.3			.5	1.7		.1					

[1] Sample 11 contains 0.3 percent graphite.

1. Biotite–potassic feldspar–quartz granulite, Brandy Brook magnetite prospect.
2. Biotite–quartz–feldspar granulite, Brandy Brook prospect.
3. Average of 2 samples, biotite–quartz–feldspar granulite.
4. Garnetiferous feldspathic quartzite, 0.5 mile south of Star Lake, Oswegatchie quadrangle.
5. Almandite–sillimanite–biotite–quartz–feldspar gneiss, 0.6 mile west of Bear Pond on road to Sabattis, Tupper Lake quadrangle.
6. Hornblende–quartz–potassic feldspar gneiss, Brandy Brook prospect.
7. Hornblende–quartz–plagioclase gneiss, 0.5 mile northeast of Cranberry Lake village.
8. Pyroxene–quartz–potassic feldspar gneiss, Brandy Brook prospect.
9. Pyroxene–quartz–potassic feldspar gneiss, 0.5 mile north of Sucker Lake.
10. Average of 2 calcareous pyroxene quartzites.
11. Graphitic pyroxene quartzite.
12. Quartz–potassic feldspar granulite, 0.5 mile east of head of Dead Creek, Cranberry Lake quadrangle.

Scandinavian literature. The texture is one of polygonal or gently curving grains of equal to unequal size.

South of Belleville School (Russell quadrangle) the quartz-feldspar granulites are extensively intruded parallel to the foliation by granite gneiss, biotite syenite gneiss, and hornblende-quartz syenite gneiss.

There are in general four facies of the quartz-rich rocks (table 4) characterized respectively by clinopyroxene, hornblende, biotite, and sillimanite with almandite. One layer of grossularitic quartzite was found. The pyroxenic facies is very common and bordering marble belts is usually thinly layered with quartz-bearing pyroxene gneiss facies much richer in pyroxene and lower in quartz, and with calcareous layers. Sphene is ubiquitous as an accessory mineral in all the pyroxenic rocks. Microcline, locally slightly perthitic, is the predominant feldspar. The hornblendic facies is subordinate in volume. The biotitic facies is common, and here too the predominant feldspar is microcline, locally slightly perthitic. Locally plagioclase is a major mineral, and it may be somewhat altered to sericite. The sillimanite-almandite facies of the quartz-rich rocks is local and usually more obviously schistose than the other varieties. Apatite, iron oxides, and zircon may be present as accessory minerals in all facies, but they are rare. The gneisses locally carry a little iron sulfide and weather to a rusty brown.

The chemical composition of a representative sample of the fine-grained quartz-feldspar granulite and of an almandite-sillimanite-biotite-quartz-microcline gneiss is given in table 3, samples 1 and 2.

BIOTITE- AND GARNET-QUARTZ-PLAGIOCLASE GNEISS

Biotite- and garnet-quartz-plagioclase gneiss forms a major element of the metasedimentary rocks of the Grenville series.

The biotite-quartz-plagioclase gneiss is usually a medium- to dark-gray, well-foliated, fine-grained rock. Practically all outcrops show thin layers, lenses, or veinlets of quartz and feldspar of granitic character parallel to the foliation. To a lesser extent, but commonly also, veinlets of pegmatite, occasionally carrying schorl, cross the foliation. The gneiss is in many places intruded by sills of granite and very locally in the contact zone has been mineralized with varied amounts of schorl. A little pyrite is present in the rock in many places.

In thin section the gneiss is seen to have a crystalloblastic texture. The grain size ranges from 0.1 mm to 1 mm. The mineral composition of several specimens is given in table 5. The plagioclase is commonly oligoclase, usually oligoclase-andesine, but plagioclase as calcic as andesine or labradorite is present in some facies. The biotite is a normal brown variety. The potassic feldspar is in part microcline and in part untwinned; to some degree its variation in quantity is related to the introduced pegmatitic veins. Garnet of porphyroblastic habit is locally present in small amount. Several percent of iron oxides is uniformly present. The iron oxide minerals occur as small grains homogeneously disseminated through the rock. The iron oxide minerals are so uniform in amount and distribution that they are considered to be a product of reconstitution of the original rock and not of later introduction from outside. The plagioclase in some specimens is partly to completely altered to an aggregate in which sericite is usually a major constituent. Very fine grained zoisite is commonly present, and some of the feldspars are altered to uralite aggregates. A little apatite and occasional small zircons are present in most specimens.

The garnet-biotite-quartz-feldspar gneiss differs from the biotite-quartz-plagioclase gneiss in the characteristic presence of garnet and a greater ratio of potassium feldspar to plagioclase. Commonly also it has a somewhat coarser foliation and a distinctly migmatitic or veined appearance.

The average and general range of chemical composition of a number of samples of biotite-quartz-plagioclase gneiss, biotite-garnet-quartz-feldspar gneiss and migmatitic equivalents within the metasedimentary rocks of the main igneous complex are given in table 6. A garnetiferous gneiss of distinctly veined structure forms a major element in the metamorphic rocks of the

TABLE 5.—*Modes of biotite-quartz-plagioclase gneisses and garnet-biotite-quartz-feldspar gneisses: facies of metasedimentary rocks of the Grenville series*

[Volume percent]

	Quartz	Plagioclase	Microcline, orthoclase, or both minerals	Hornblende	Biotite	Garnet	Sphene	Iron oxides	Zircon	Apatite
Biotite-quartz-plagioclase gneiss										
Average of 9 samples	25.8	52.8	7.7	----	9.2	----	----	3.9	-----	0.6
Average of 3 garnetiferous samples	30.9	51.0	5.6	----	4.0	4.3	----	3.8	-----	.3
Exceptionally rich in biotite, Brandy Brook, Cranberry Lake quadrangle	45.8	25.9	1.8	----	24.8	(1)	----	.5	Tr.	.2
General range of variation	15–45	40–65	1–12	----	3–15	----	----	2.5–6	-----	.3–.8
Garnet-biotite-quartz-feldspar gneiss										
Average of 10 samples	26.8	30.5	31.1	----	5.6	3.8	----	1.8	-----	0.4
General range of variation	20–40	10–40	20–45	----	2–10	1–10	----	0–6	-----	.1–.5

1 Specimen also contains 1.0 percent pyrite.

TABLE 6.—*Chemical analyses and norms of biotite-quartz-plagioclase gneisses, migmatites, sillimanite-microcline granite gneiss, and average graywackes*

	Biotite-quartz-plagioclase gneiss				Graywacke			Migmatite					Granite gneiss
	1	2	3	4	5	6	7	8	9	10	11	12	13
Chemical analyses (weight percent)													
SiO_2	63.01	71.00	70.50	70.90	69.69	69.67	61.52	64.18	65.25	67.18	72.49	71.34	71.44
Al_2O_3	16.60	14.00	12.44	12.17	13.53	14.46	13.42	16.19	14.23	15.65	13.82	13.64	14.89
Fe_2O_3	3.14	.55	1.57	1.31	.74	1.60	1.72	1.87	4.76	.13	.13	.77	2.40
FeO	3.42	3.16	3.92	4.12	3.10	1.93	4.45	3.98	2.80	3.31	2.97	3.97	1.38
MgO	1.41	1.26	2.27	2.32	2.00	1.31	3.39	2.17	.81	1.25	1.01	.29	.30
CaO	3.50	2.26	2.13	1.55	1.95	1.51	3.56	2.69	2.39	1.70	1.48	1.45	.50
Na_2O	3.62	3.20	3.57	3.74	4.21	4.00	3.73	4.64	2.54	3.42	3.08	1.76	1.36
K_2O	3.02	2.70	2.56	2.87	1.71	2.35	2.17	3.48	4.76	5.24	3.22	5.26	5.99
H_2O+	.34	.61	.27	.21	2.08	1.82	2.29	.24	.44	.99	.65	.27	.52
H_2O-	.03	.09	.02	.05	.26	.32	.06	.06	.06	.06	.06	.05	.08
CO_2					.23	.03	3.04						
TiO_2	1.06	.53	.32	.32	.40	.51	.62	.34	1.06	.62	.50	.64	.65
P_2O_5	.35	.09			.10	.14	.06		.38	.10	.06	.08	.22
MnO	.07	.05	.06	.04	.01	.07		.06	.11	.06	.03	.05	.03
Total	99.57	99.50	99.63	99.60	100.01	[1] 100.11	100.03	99.90	99.59	99.71	99.50	99.57	99.76
Norms													
Quartz	21.63	33.33	29.46	28.77	29.91	31.68		12.51	26.91	19.98	35.88	35.40	39.54
Orthoclase	17.79	16.12	15.28	16.68	10.01	13.90		20.57	28.08	30.86	18.90	30.86	35.58
Albite	30.39	27.25	30.39	31.70	35.63	34.06		39.30	21.48	28.82	26.20	14.67	11.53
Anorthite	15.29	10.56	10.29	7.78	9.04	6.67		13.07	9.73	7.78	7.23	6.67	1.11
Corundum	.82	1.84		.15	1.43	2.80			1.28	1.53	2.55	2.65	5.71
Hypersthene	7.50	7.64	11.12	11.80	9.29	4.88		10.63	2.00	8.12	7.12	6.38	.80
Magnetite	2.89	1.86	2.32	1.86	1.07	2.32		2.78	6.15	.23	.23	1.16	2.55
Ilmenite	2.05	.61	.61	.61	.76	.97		.61	2.05	1.22	.91	1.22	1.22
Apatite	.84				.24	.34			.91	.24	.13	.20	.50
Diopside			.23					.23					
Calcite					.50								
Hematite									.48				.64
Ratio Na_2O/K_2O	1.2	1.18	1.4	1.3	2.4	1.7	1.7	1.3	.53	.65	.96	.33	.23

[1] Includes some other minor constituents not shown in table.

1. Biotite-quartz-plagioclase gneiss, diamond-drill hole 8, Spruce Mountain prospect, northwest corner Cranberry Lake quadrangle. Analyst, Lee C. Peck.
2. Biotite-quartz-plagioclase gneiss, southwest corner Gouverneur quadrangle. Analyst, A. F. Buddington.
3. Biotite-quartz-plagioclase gneiss, south end of traverse line for No. 4. Analyst, Ledoux and Co. (Engel and Engel, 1953, p. 1063).
4. Biotite-quartz-plagioclase gneiss, composite sample of 24 specimens across 2,000 ft at right angles to strike, along line traverse from West Branch of Oswegatchie River north to Oswegatchie River; 0.9 mile east of Emeryville, Gouverneur quadrangle. Samples taken as far from pegmatite veins as practicable. Analyst, Ledoux and Co. (Engel and Engel, 1953, p. 1063).
5. Graywacke, average of 3 analyses, from Franciscan group, California (Taliaferro, 1943, p. 109–219).
6. Schists (quartz-albite-sericite-cholrite-epidote), derived from graywacke, average of 8 analyses (Williamson, 1939, p. 30).
7. Graywacke (Timiskamian), average of 3 analyses (Todd, 1928, p. 20).
8. Porphyroblastic facies of Hermon granite gneiss, a migmatite, 0.8 mile northwest of Emeryville, Gouverneur quadrangle (Engel and Engel, 1953, p. 1077). Analyst, Ledoux and Co.
9. Migmatite of biotite-quartz-plagioclase gneiss and granitic material, road cut 1.4 miles northwest of Parishville, Potsdam quadrangle. Analyst, Eileen H. Kane.
10. Biotite-quartz feldspar gneiss with a little garnet, somewhat modified by granitization, road cut 2.5 miles west-northwest of Colton, Potsdam quadrangle. Analyst, Eileen H. Kane.
11. Migmatitic garnet-biotite-quartz-feldspar gneiss, road cut 2.5 miles west-northwest of Colton and 0.4 mile southwest of No. 10. Analyst, Eileen H. Kane.
12. Garnet-biotite-quartz-feldspar gneiss partly granitized and with garnetiferous microcline granite pegmatite layers, road cut on highway to Sabattis. 0.25 mile west of Bear Pond, Tupper Lake quadrangle. Analyst, Eileen H. Kane.
13. Sillimanite-quartz-microcline granite gneiss, diamond-drill hole 1 at 120-ft depth, on Skate Creek, Oswegatchie quadrangle. Analyst, Lee C. Peck.

Grenville lowlands, but will be described in connection with the mixed and veined gneisses.

Rarely a sillimanite-biotite-quartz-feldspar gneiss is found, as at the road crossing of Tracy Pond Outlet beyond Gooseberry Mountain (Stark quadrangle), but usually the sillimanite-bearing rocks are of such a nature that they are here classified as sillimanite-microcline granite gneiss.

Sillimanite-biotite-quartz-feldspar gneisses, with intensely developed foliation and some granite pegmatite veins, form the hills 0.5 mile south of East Road School, and east of the road on the Russell quadrangle.

MARBLE AND SKARN

The marble and skarn within the main igneous complex will be discussed in detail under the subject of magnetite deposits, with which they are in part so intimately connected. (See Leonard and Buddington, Prof. Paper 377*.) It is remarkable that although only two layers of marble were found to crop out within the several hundred square miles of the main igneous complex, marble was cut in drill holes at 8 of the 10 skarn deposits drilled for ore, and at 2 of 6 granite gneiss deposits similarly explored. No marble is visible at the surface in the vicinity of any of these deposits. It is therefore certain that many more marble layers are present in this area than have yet been discovered, and that marble beds (including the modified equivalent of marble, skarn) formed a widespread major member of the series of metasedimentary rocks as a whole.

*Leonard, B. F., and Buddington, A. F., Ore deposits of the St. Lawrence County magnetite district, northwest Adirondacks, New York (in preparation).

Marble is a major member of the metasedimentary rocks in the Grenville lowlands, where it has been previously described as "Grenville limestone" (Martin, 1916, p. 17–23), Reed (1934, p. 12–14), Dale (1934; and 1935, p. 13–16).

Only the major belts will be referred to here. Details on local areas will be found in the publications just cited.

The marble in general is a coarsely crystalline carbonate rock and carries disseminated graphite and silicates such as pyroxene, feldspar, scapolite, and phlogopite, and locally nodules of serpentine or tremolite. Graphite is common in the associated pyroxene granulites and gneisses.

A belt extends almost continuously (except for an interruption east and west of Hamiltons Corners) across the Russell quadrangle from the northeast corner, through Endersbees Corners, Stalbird, and Scotland School to the west edge of the quadrangle, south of Edwards. The sand plains along the Oswegatchie River obscure the geology south-southeast of Edwards, but it is possible that this belt of marble is more or less continuous to the south with the belt in the northwest corner of the Oswegatchie quadrangle.

The part of the belt through West Pierrepont and Moores Corners consists of marble with nodules of tremolite or serpentine, and includes local lenses of diopside granulite and of pyroxene gneiss containing disseminated sulfide. Large tremolite crystals occur one-third mile east of West Pierrepont, south of the road.

The area through Endersbees Corners consists of marble with zones of silicate rock. North-northeast of Endersbees Corners silicated marble layers consist of diopside granulite, diopside-tremolite rock, and feldspathic pyroxene granulites—each with thin interbeds of white quartzite. A pit has been sunk on the marble where good tremolite crystals are present, in the hill 0.5 mile northeast of Owens Corners and 0.2 mile east of the road to Moores Corners. In general, the marble also locally contains lenses of thin-layered white quartzite and marble, of quartz-mesh marble, of white quartzite, and of diopside granulite. The silicate and quartzite layers are crumpled and deformed by flowage of the enclosing marble.

The part of the belt through Stalbird and Scotland School is mostly obscured by drift. The outcrops indicate marble with lenses of diopside granulite, diopside-tremolite rock, pyroxene gneiss containing sulfide disseminations, and local white quartzites.

The wide marble belt south-southwest from White School on the Oswegatchie quadrangle is the largest area of relatively pure marble within the district covered by this report. Small thin silicate lenses a few inches to a foot long are present but rare, and occasional thin layers contain considerable disseminated silicates. A few short, narrow granite lenses are present. The marble from southwest of Snyder Lake has a very distinct odor of hydrogen sulfide when struck with a hammer.

The marble belt east and south-southwest of Jacksons Falls (northwest Russell quadrangle) consists of marble, much of it containing disseminated silicates and local lenses of granite. A little disseminated brown tourmaline is common in much of the marble of this belt. In the vicinity of Brownsville School the marble includes beds of pyroxene gneiss containing disseminated sulfides, and beds of garnet-biotite schist, quartzitic schist, and quartzose pyroxene gneiss. Similar gneiss comprising interlayered graphitic pyroxene-quartz-feldspar gneiss and marble, and beds of garnet and biotite gneiss, also forms a narrow belt west of Knott School. The gneiss at Brownsville is intricately contorted and folded.

Marble is exposed 0.5 mile north of Pestle Street School (northwest Russell quadrangle), in the valley of Gibbons Brook just south of the Devils Elbow, and southwestward up the valley where it is gradually replaced by pyroxene skarn. About 0.8 mile southwest of the road crossing of Gibbons Brook, the rocks of the belt consist of medium-green pyroxene skarn and silicated marble. One and two-tenths miles from the road this belt of rock consists predominantly of medium-green pyroxene skarn with a little mica. There is also some interlayered garnet gneiss and skarn here.

On the Tupper Lake quadrangle there is a roughly elliptical belt of marble and associated rocks in the Loon Pond syncline west of Round Lake. The belt consists chiefly of marble but includes local interbeds of white quartzite, thin-layered quartzite and marble, diopside granulite, and some tremolite schist and biotite gneiss. The marble contains varied amounts of disseminated silicates and quartz. Diopside is the most common silicate. Graphite is present in some of the marble and diopside granulites and is locally abundant. Part of the diopside granulite carries disseminated iron sulfides. Tremolite-diopside schist was observed 1.0 mile southwest of the Round Lake dam, where the tremolite schist is associated with siliceous schist and white quartzite.

PYRITIC GNEISSES

Pyrite (and locally pyrrhotite) is locally a common minor accessory in some of the gneisses and skarns of the main igneous complex; but the migmatitic gneisses of the Grenville lowlands may locally carry as much as several percent or more of iron sulfides. These rocks weather to such a characteristic brown crumbly outcrop that they have been called "rusty gneisses"; and such zones may have considerable length and thickness. Reed (1934, p. 17) describes them as occurring in the Potsdam quadrangle, near the eastern boundary of the garnet gneiss area southwest of Browns Bridge; near the western border of the quadrangle west of Colton, in the gorge between Colton and Browns Bridge; and in the valley of the Raquette River a little more than 1 mile below Rainbow Falls. We have also noted them 0.5 mile north of Browns School. A thin layer of pyritic gneiss in the garnet gneiss near Van Rensselaer Creek (Canton quadrangle) has been described by Martin (1916, p. 43).

The hills southwest of the bridge at Hermon (northwest Russell quadrangle) are composed of very conspicuous rusty-weathering pyritic gneisses. These gneisses are partly garnet-biotite-quartz-feldspar gneiss and partly pyroxene gneiss. Some of the gneiss shows chloritization as well as pyritization. Pyrite-chlorite schist outcrops in the bank of Elm Creek 0.4 mile north of the bridge at Hermon. The gneisses between Hermon and Stellaville generally carry many pyritic layers; marked concentrations of pyrite in some of these layers will be further described under the heading, "Stella Mines" in Professional Paper 377*. Pyritic pyroxene gneiss layers may occur as lenses within the marble—as west of the road 1.2 miles northeast of Edwards—and at numerous places within the belt of quartz-feldspar and pyroxene gneisses—as on the southeast side of the road, 0.6 mile southwest of Owens Corners (north Russell quadrangle), and a mile or so southeast of Edwards.

Much of the normal pyritic gneiss may reasonably be interpreted as metamorphosed sedimentary beds or lenses rich in iron sulfide. However, local concentrations of iron sulfides within the pyritic gneisses (see discussion under heading "Iron sulfide deposits," Leonard and Buddington, Prof. Paper 377) are best interpreted as hydrothermal replacement veins, and it is to be expected that the solutions which effected the concentrations would also introduce some lean pyrite impregnation into the country rock. An example of such occurrence is the exceptionally strong development of pyritic gneisses in the vicinity of Stella Mines. In part, therefore, the pyritic gneisses are thought to be formed by introduction of hydrothermal pyrite into normal metasedimentary rocks, especially so where there is associated chloritization.

ORIGIN OF THE METASEDIMENTARY GNEISSES

The biotite-, pyroxene-, and hornblende-quartz-plagioclase gneisses pose a special problem as to their origin because of the high ratio of plagioclase to potassium feldspar. Under conditions of weathering, potassium feldspar is normally more resistant to decomposition than plagioclase and under normal conditions of diagenesis, clay minerals absorb more potassium than sodium. For these reasons, potassium dominates over sodium in most shales and sandstones, and this would be the normal relation in their metamorphic, reconstituted equivalents. In the quartz-plagioclase gneisses, however, the reverse is true.

Biotite-quartz-plagioclase gneisses are common in the metasedimentary rocks of the Grenville series throughout the Adirondacks. Locally, as described by Alling (1918) in the eastern Adirondacks, this type of gneiss is more quartzose and carries several percent graphite and pyrite. He infers that the layered character, the variability in composition from layer to layer, the local presence of graphite, and the local variation to rocks richer in quartz than any igneous rocks—all indicate a sedimentary origin for the primary rocks.

Sodium-rich gneisses such as these are common in many terranes of metamorphic rocks and have been variously ascribed to reconstitution of pelitic rocks or graywacke of equivalent composition, to metamorphism of dacitic pyroclastics, and to modification of normal pelitic sediments by introduction of sodium from magmatic or deep-seated solutions. There is reasonable evidence of local introduction of potassium feldspars into the biotite-quartz-plagioclase gneisses in this area, but no evidence has been recognized of the introduction of sodium into an originally more potassic rock to yield the sodic gneisses. A better case could actually be made for the differential solution and leaching of potassium from a normal sediment to yield the quartz–potassium feldspar veinings of the gneiss on the one hand, and a potassium-impoverished, relatively sodium-rich residue on the other hand. No valid bases for this hypothesis have been recognized, however. The metasedimentary rocks are so thoroughly reconstituted that they have wholly lost any identifying features of dacitic volcanic tuffs—a possible type of rock suggested by Gilluly (1945) as the original—if ever they were derived from such rocks. We should also expect, on this hypothesis, to find metamorphosed dacitic lavas associated with the plagioclase gneisses thus assumed to be derived from volcanic tuffs; but no rocks identifiable as derived from felsic lava flows have been found, though here again metamorphism may have destroyed the features by which they might be identified.

Aside from volcanic tuffs, the type of common sedimentary rock that is the closest approximation in chemical composition to the biotite-quartz-plagioclase gneisses is graywacke. The term graywacke is used for

*See footnote on p. 31.

a type of dark sandstone composed of detrital quartz and feldspar set in a prominent to dominant silty or clayey matrix, and with a variable quantity of rock fragments such as chert, slate or phyllite, siltstone, quartzite, or volcanic rock. Mafic minerals such as pyroxene and hornblende may also be present. A review of the data on graywacke has been presented by Pettijohn (1943), whose data show that in graywackes the percent Na_2O exceeds percent K_2O, whereas in normal feldspathic sandstones the reverse is true. The rock of the thick graywacke beds of the Franciscan in California, in particular, is similar in composition to the biotite-quartz-plagioclase gneiss. Five samples from north of San Francisco Bay averaged in percent: quartz, 43.2; orthoclase, 13; albite, 11; oligoclase, 29; and rock fragments and composite grains, 3.8. The norm of the average of three chemical analyses of graywacke is quartz, 30.4; orthoclase, 10; oligoclase ($Ab_{71}An_{29}$), 43.4; hypersthene, 9.4; magnetite, 1.1; and corundum, 1.8.

Taliaferro (1943, p. 136–139) believes that the graywackes of the Franciscan were derived from a high, rugged, recently uplifted land mass under rigorous climatic conditions—a land characterized by high rainfall, possibly a cold climate in the highlands, and well-wooded lower slopes. Granodiorite is assumed to be a major element of the source rock. Pettijohn notes the common association of graywacke with pillow lavas, and follows others in attributing both to a phase of tectonics associated with orogeny. He further states that development of graywacke requires an environment in which erosion, transportation, and deposition are so rapid that complete chemical weathering of materials does not take place.

The pyroxene- and hornblende-quartz-plagioclase gneisses, like the biotite-quartz-plagioclase gneisses, are peculiar in that the percent plagioclase far exceeds the percent potassium feldspar. If some of them approximate in composition the original sediments, and we do not have any positive evidence to the contrary, then they may be interpreted as metamorphosed calcareous or dolomitic graywackes. Most graywackes do not carry the amount of calcium and magnesium carbonate necessary to yield rocks of the composition of those under discussion. However, many of the graywackes in the extensive beds of southeastern Alaska are somewhat calcareous and probably of a chemical composition similar to that of some of the pyroxene and hornblende gneisses of this district (Buddington and Chapin, 1929, p. 75, 78–79, 89–90).

The quartz-plagioclase gneisses then, insofar as their composition is involved, might represent metamorphosed calcareous and argillaceous graywacke beds. In quantity, however, feldspathic graywacke beds are often associated with lava flows. We have not found rocks associated with the gneisses that we could definitely interpret as metamorphosed lavas, though a part of the amphibolites could perhaps be ascribed to such an origin. Graywacke beds commonly are interbedded with shale. In part, the biotite-quartz-plagioclase gneisses are homogeneous throughout considerable thicknesses and therefore afford no evidence of the original presence of shale beds. Furthermore, though metamorphic equivalents of such normal sediments as limestone, sandstone, and impure siltstone are represented, there is no satisfactory metamorphic equivalent of normal shale, such as might well be expected to have occurred in quantity.

Although most shales have a higher percent K_2O than Na_2O, there are some, in small quantity, in which this relation is reversed. A sodium-rich shale could yield on metamorphism a biotite-quartz-plagioclase gneiss.

Sodium-rich gneisses, which occur widely in other metamorphic terranes similar to those of the Adirondacks, have been interpreted by many geologists as normal shales modified by the introduction of sodium and removal of potassium, often with a homogenization of the formation.

The present data do not enable a definite choice to be made between these hypotheses for the origin of the metasedimentary plagioclase-rich gneisses. The problem is further discussed in detail by Engel and Engel (1953) and Engel (1956).

The pyroxene-quartz-potassic feldspar gneisses occur partly as layers within the quartz-plagioclase gneisses and are equivalent in chemical composition to calcareous illitic sandstones or sandy illitic shales. They are also comparable to calcareous potassic feldspar-quartz sands, but if the pyroxene-quartz-plagioclase gneisses are interpreted as metamorphosed graywackes, arkoses or potassic feldspar–quartz sands would not normally be expected.

The fine-grained, feldspathic, generally quartz-rich granulites are in part quite uniform in thick layers, but in many places they have a thin-layered structure, owing to a change in the nature or content of accessory minerals from layer to layer. This is particularly apparent in zones near marble beds. The pyroxene-quartz-feldspar granulites and gneisses also have some facies that contain little or no quartz, and others that are exceptionally rich in quartz. This structure and these variations seem best interpreted as an inheritance from primary sedimentation. Illitic or potassic feldspar–quartz sandstones, locally including calcareous layers, have a chemical composition that upon metamorphism could yield rocks of the type under discussion, with little or no change.

The pyroxene-feldspar gneiss, which usually has a layered character, could be variously interpreted as a metamorphosed bedded sediment with but little change in bulk chemical composition, a feldspathized and silicated metasedimentary rock with substantial change in chemical composition, a migmatitic metasedimentary rock, or a combination of all three.

The rocks grade from fine-grained, thin-layered feldspathic quartz-rich granulite, partly biotitic and containing thin pyroxene-rich or pyroxene skarn laminae, to pyroxene skarn in thick zones interlayered with feldspathic skarn (fig. 5). The first type clearly has the appearance of a metasedimentary rock that has undergone little change in chemical composition, whereas the skarn-rich facies seems best interpreted as a metasomatic replacement of marble, resulting in a very substantial change in composition. The constant occurrence of sphene in the pyroxene gneisses is also indicative of a fairly homogeneous introduction of titanium throughout the rocks. Locally the rock has a veined character suggestive of pegmatitic injection.

Our interpretation is that the pyroxene-rich gneisses represent bedded illitic siliceous limestones that have been reconstituted with varied introduction of material by solutions working through them.

In summary, the primary sediments might have consisted of limestone and impure limestone (now represented by marble, silicated marble, or skarn replacements of marble); calcareous graywacke or calcareous shale (now represented by pyroxene- or hornblende-quartz-plagioclase gneisses); graywacke or, less probably, shale (now represented by biotite-quartz-plagioclase gneiss); illitic or potassic feldspar sandstone, in part calcareous (now represented by quartz-rich pyroxene-, hornblende-, or biotite-feldspar gneisses and granulites); illitic siliceous shales or illitic siltstones (now represented by biotite-quartz-potassic feldspar granulites); and strongly illitic or argillaceous limestones (now represented by pyroxene-feldspar gneisses). Certainly the original rocks have largely been modified in chemical composition by solutions permeating through them, but the extent to which this has occurred is in most cases an unsolved problem.

MIXED OR VEINED GNEISSES

Mixed or veined gneisses are a major facies of the rocks in the areas mapped as metasedimentary rocks of the Grenville series. Such gneisses have a thin-layered, foliated character, the alternating layers being relatively more mafic, or more feldspathic, or more granitic. The granitic layers may be similar in grain to the other laminae, or they may be coarser grained and

FIGURE 5.—Thin-layered metasedimentary gneiss; drill core, Clifton mine. Upper half is pyroxene-quartz-feldspar gneiss; lower half is biotite-quartz-feldspar gneiss.

pegmatitic in character; pegmatite veins very commonly may occur at intervals of a few inches in otherwise even-grained, layered gneiss.

The veined gneisses developed preeminently from the biotite-quartz-plagioclase gneisses, to a subordinate extent from the pyroxene gneisses, and practically not at all from the feldspathic quartz-rich gneisses and quartzites or from marbles or skarns.

The origin of the mixed and veined gneisses has not been solved. These rocks are thought to include true arteritic migmatites or injection gneisses, pseudomigmatites, and permeation gneisses. The arteritic migmatites are the product of injection of thin layers of magma along the foliation planes of the metasedimentary rocks. The pseudomigmatites result from permeation, modification, and replacement of alternate laminae of the metasedimentary rocks in such a way that they come to have the appearance of granitic veins similar to true arteritic migmatites. The permeation gneisses have a distinct foliation but an indistinct layering or vein structure. They are the product of solutions seeping, diffusing, and penetrating quite uniformly through the entire sediment, with accompanying introduction of new material and removal of old. The term "solutions" as used here may include highly fluid volatile-rich magma, water solutions, gases, and disperse migrating atoms or ions. In all cases the rocks here described have a mixed origin and have a layered or veined structure to a varied degree. All the different mechanisms of formation may well have cooperated, but to different degrees, in the development of different facies of these gneisses.

In the contact zones between granite and metasedimentary rocks, or granite and amphibolite, there is usually a belt of interbanded mixed rock; all gradations exist between contaminated granite containing schlieren, and a nearly normal metasedimentary rock or amphibolite.

The granitic vein matter can locally and often be proved on a small scale to be of replacement origin, for the veins expand and contract in size at the expense of the country rock without any evidence of displacement of its laminae. Elsewhere, granitic veins expand and contract with obvious evidence of displacement and flowage of the adjacent laminae, but in such cases one cannot be sure whether this arose from injection of magma or from crystal growth of the granitic veins in place. Again, many veins show no evidence that can be interpreted in terms of either replacement or displacement, and much of the more nearly uniform gneiss is of this character. We cannot distinguish whether the rock is largely the result of permeation with replacement, or of permeation with inflation and displacement.

BIOTITE- AND GARNET-QUARTZ-FELDSPAR GNEISS WITH VEINED STRUCTURE

Biotite- and garnet-quartz-feldspar gneiss with veined structure differs from the biotite- and garnet-quartz-feldspar gneisses previously described in having the more uniform presence of garnet, usually a much greater ratio of potassium feldspar to plagioclase, a somewhat coarser foliation, and usually a more distinctly veined structure.

Gneiss of this type is present only locally within the main igneous complex of the magnetite district proper, but it is developed on a large scale within the metamorphic rocks of the Grenville lowlands. It forms a great belt passing through Kimball Hill (Russell quadrangle) and southwestward to the Gouverneur quadrangle. Another major belt extends northeast from O'Malley Brook and southwest from Colton on the Potsdam quadrangle. It is also a major formation in the Pierrepont sigmoid structure in the southeast corner of the Canton quadrangle, and the eastern edge of a large mass occurs in the southwest corner of the Russell quadrangle and northwest corner of the Oswegatchie quadrangle.

There are all gradations between biotite-quartz-plagioclase gneiss, biotite-quartz-plagioclase gneiss with a few granitic veinings, and garnetiferous gneiss with pronounced veined structure.

Mixed rocks of the type under discussion have been described from the Grenville lowlands under the name of "garnet gneiss" or "garnet-biotite gneiss" by several geologists (Reed, 1934, p. 18–21; Martin, 1916, p. 24–40; Cushing and Newland, 1925, p. 28–33; Smyth and Buddington, 1926, p. 18–20). Locally, granitic and pegmatitic layers are few or absent, and garnets are then similarly few or absent; such gneiss is characteristically a biotite-quartz-plagioclase facies. Rock of this type occurs in considerable development on the Gouverneur and Lake Bonaparte quadrangles, where it has been described by Gilluly (1945), Engel and Engel (1953), and Smyth and Buddington (1926, p. 19–20).

Gilluly (1945) describes the gneiss, where least injected, as consisting of 25–50 percent (average 35 percent) oligoclase-andesine (An_{25-35}); 25–40 percent quartz; 2–50 percent microcline (average about 15 percent); 10–18 percent biotite; 2–10 percent muscovite; and locally some chlorite after biotite.

The typical garnetiferous gneiss is a medium- to light-gray or bluish-gray rock; it is thin layered, with lenses, laminae, or layers of granitic material, in part pegmatitic. The garnet occurs chiefly in or near the granitic material. Locally, granite pegmatite veinings are few or wanting, in which case the rock may have few or no garnets and be a biotite-quartz-feldspar gneiss. The layering in the garnetiferous gneisses is in many places emphasized by alternating layers containing more or fewer garnets than normal. The garnets are commonly one- to three-eighths inch in size. The pegmatitic layers locally show evidences of emplacement by replacement and locally have also bowed apart the adjoining layers as though injected. The pegmatitic material may blend with the intervening more micaceous layers. The intervening layers between the granitic layers appear to have been modified by granitic impregnation. The gneiss is essentially an arteritic migmatite.

The composition of some specimens of the biotitic and garnetiferous quartz-feldspar gneiss is given in table 5. The plagioclase is oligoclase. The potassium feldspar is microcline or orthoclase in the injected portion of the gneiss, but may be either of the potassium feldspars or microperthite in the pegmatite veins. The percent magnetite is quite variable. A little apatite is commonly present; zircon is occasionally present. Sphene and pyrite are other accessory minerals. Locally the feldspars are altered to an aggregate of sericite, albite, and other minerals.

Amphibolite in layers several inches to a score of feet thick is intercalated in the gneiss and may locally be abundant.

A series of chemical analyses is given in table 6 to show the variation in chemical composition of the primary biotite-quartz-plagioclase gneiss and its veined and partly granitized equivalents. Additional analyses are given in the detailed study by Engel and Engel (1958), which was published after this report was written.

The rocks that have veined structure (with the exception of No. 8, table 6) have a lower ratio of Na_2O/K_2O and a higher percent K_2O than the normal biotite-quartz-plagioclase gneiss. This fact is consistent with a hypothesis involving the addition of K_2O to the rock; but it is inconsistent with the idea of the development of the granitic veinings as a product of differentiation (venous pegmatites) within the rock as a whole without much chemical change. The chemical analysis (No. 13, table 6) of a sillimanite-microcline granite gneiss is given for comparison. Such rock is thought to be the end product of granitization of biotite-quartz-plagioclase gneiss under conditions obtaining within the main igneous complex.

A number of excellent exposures of migmatite and sillimanite-microcline granite gneiss occur along the road to Sabattis (Tupper Lake quadrangle) for three-fourths mile west-northwest of Bear Pond. The first outcrop west of Bear Pond is a thin-layered migmatitic biotite gneiss that is very striking because of the contrast of colors in the different layers. One-fourth mile west of the pond is another good exposure in the road cut. This is also a migmatite consisting of alternating layers of biotite-rich quartz-plagioclase gneiss, white to pale green quartz-plagioclase gneiss, and pink pegmatite. All are garnetiferous. A chemical analysis of a specimen of the migmatite of pale quartz-plagioclase gneiss and pink microcline granite pegmatite is given in table 6 (No. 12). The gneiss consists of a little less than one-third each of microcline, plagioclase, and quartz, and accessory biotite, magnetite, zircon, and garnet. The pegmatite layer consists predominantly of pink microcline (about 65 percent), considerable quartz, less than 10 percent plagioclase, and several percent red garnet. The garnet of the pegmatite veinings is thought to be due to contamination by incorporation of material from the biotite of the gneiss. The garnets are larger in the pegmatite veins than in the gneiss, and in associated green quartz-plagioclase gneiss they are more coarsely recrystallized than normal.

One-half mile west of Bear Pond are exposures of granite, garnetiferous granite, and sillimanitic granite, all interlayered.

About three-fourths mile west of the pond, just before the road descends to the level of the swamp, these interlayered rocks are well exposed in road cuts. The eastern hundred feet is a sillimanite-microcline granite gneiss with sparse garnet; on the west at the foot of the hill is a migmatitic garnet-biotite gneiss. Between these the rocks consist of feldspathic (mostly microcline) quartzites rich in garnet and sillimanite, with a varied amount of biotite and, locally, with large and conspicuous flattened nodules of sillimanite-quartz aggregates. A chemical analysis of a garnet-sillimanite-quartz-microcline gneiss is given in table 3 (No. 2). The rock is quartz rich. The sillimanite is partly altered to sericitic aggregates. There is a little accessory biotite and iron oxide.

PYROXENE GNEISSES WITH VEINED STRUCTURE

Some of the pyroxene-quartz-feldspar gneisses have a veined character. The pyroxene-quartz-plagioclase gneisses around Newton Falls are associated with the local occurrence of albite and oligoclase granite and may reasonably be interpreted as the product of modification of pyroxene skarn and metasedimentary rock by the ingress of solutions of sodium-rich granitizing character. By contrast, the pyroxene gneisses 1½ miles north of Newton Falls have a veined structure;

but the vein matter is largely potassic feldspar and quartz, and the granitizing and injecting solutions were presumably of similar character. Furthermore, sills of granite occur within the pyroxene gneisses and may locally be contaminated with pyroxene. Layering within the pyroxene-quartz-feldspar gneisses may also be of a normal bedded character without any appearance of veinlike structure.

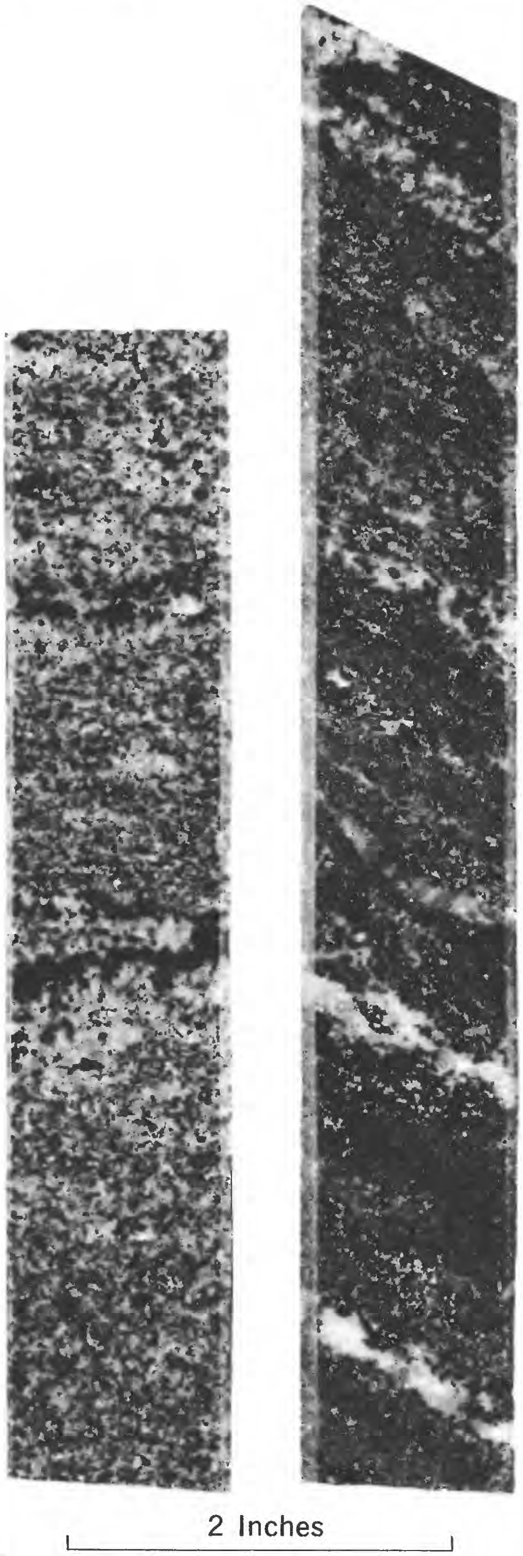

FIGURE 6.—Pyroxene-quartz-feldspar gneiss (at left); feldspathic pyroxene skarn with granitic veinings (at right). Drill core, Spruce Mountain magnetite deposit.

Photographs of two examples of pyroxene gneiss are shown in figure 6. The darker one is a feldspathic pyroxene skarn with distinct veinlets of granitic material, and the lighter one is a pyroxene-quartz-feldspar gneiss. The feldspathic skarn with vein structure grades into normal uniform pyroxene skarn and is certainly derived from it by emplacement of granitic veins (by injection or replacement, or probably both) and accompanying partial granitization of the intervening layers. The pyroxene gneiss is more uniform. Its origin is uncertain, for material of this type is found on the one hand to grade into veined pyroxene skarn as though derived from it by granitization through permeation, and on the other hand to be interlayered with other types of gneiss as though it were simply a sedimentary layer reconstituted without much chemical change.

ANORTHOSITIC GABBRO AND EQUIVALENT GNEISS

A very large mass of anorthosite, with associated subordinate amounts of gabbroic anorthosite, and several smaller bodies of similar character and variation occur in the eastern Adirondacks (fig. 4), together occupying about 1500 square miles. About 60 miles west of the main body, in the southern part of the Antwerp quadrangle, is a small isolated mass of anorthosite with gabbroic variants, about 1 square mile in area. No anorthosite has been found between these two occurrences, which means that none has been found in St. Lawrence County.

However, a lens of anorthositic gabbro and equivalent gneiss, thought to belong to the anorthositic series of intrusives, is found on the Russell quadrangle west of South Russell. This mass shows many peculiar features of scientific interest and has been described by Smyth (1896), Miller (1921), Dale (1934), and Buddington (1939, p. 53–54, 260–267). The anorthositic gabbro mass is about 4.5 miles long and up to 1 mile wide. The bulk of the rock is dark gray. Locally there are more feldspathic layers lighter in color than normal, or more mafic layers darker in color than normal. The prevailing rock has a coarse ophitic texture; the feldspars, few of which exceed one-half inch in length, are surrounded by dark green or black minerals. Local coarser portions have a mottled aspect on weathered surfaces, owing to imperfect porphyritic structure, with phenocrysts of feldspar ranging up to 3 inches in length.

Locally, the gabbro is intensely metamorphosed and reconstituted. In places, the metamorphism has resulted in new minerals and very much finer grain, though superficially an appearance of coarse texture is still preserved; elsewhere, local layers in the rock are intensely deformed and reconstituted to a medium- or fine-grained granulitic gneiss. The primary major minerals of the anorthositic gabbro are labradorite and augite, and the primary accessory minerals are hypersthene, magnetite, and ilmenite. Scapolite and hornblende, new minerals resulting from metamorphism, are the dominant minerals in the metamorphic granoblastic gneissic layers, together with a varied amount of recrystallized labradorite. Rarely there is a trace of garnet. The metamorphic facies of the rock as exposed in the bed of Bullock Creek near the western border have been studied chemically and described in detail by Buddington (1939, p. 260–267). Locally, where it borders against the younger intrusive granite, the anorthositic gabbro is impregnated by quartz and orthoclase as replacement minerals. At the north end of the mass veinings of scapolite and plagioclase, scapolite and hornblende, hornblende alone, pyroxene alone, and tourmaline pegmatite cross the foliation. Locally, at the northernmost end of the mass, there are groups of very unusual curving cracks and fault breccias which affect both the gabbro and the adjoining granite. It seems probable that the north portion of the gabbro was bent to the west during a period of deformation when the gabbro was solid and the granite was in the final stages of consolidation. One included fragment of coarse anorthosite 9 inches in diameter was seen in the metagabbro.

A small lens of anorthositic gabbro occurs about 0.7 mile northwest of South Edwards (Russell quadrangle). It carries a larger than normal quantity of ilmenite and magnetite.

The mineral composition of two facies of the anorthositic gabbro is given in table 7.

Although so small in quantity in the northwest Adirondacks, anorthositic rocks may actually underlie the St. Lawrence County magnetite district in many places, at depths within a few miles of the surface. This is indicated by the many and widespread localities at which xenocrysts of labradorite and locally angular inclusions of anorthosite occur in the igneous gneisses now exposed at the surface. Labradorite xenocrysts occur locally in metagabbro gneiss, all facies of the syenite gneisses, and the younger hornblende granite gneiss. Angular inclusions of anorthosite have been seen in a gabbro mass southwest of Jerden Falls (Lake Bonaparte quadrangle), in the Russell anorthositic gabbro, in granite east of Jerden Falls, and in pyroxene-quartz syenite gneiss south of the Grass River Club (Tupper Lake quadrangle).

Labradorite crystals have been described by Balk (1931, p. 388) from the syenite gneiss of Arab Mountain (Tupper Lake quadrangle) and have been seen by us in syenite gneiss 1 mile east and also 1.2 miles southeast of South Edwards (Russell quadrangle), in pyroxene-quartz syenite gneiss 0.5 mile northeast of Gain Twist Falls (Potsdam quadrangle), locally in hornblende-quartz syenite gneiss southwest of Kalurah (Oswegatchie quadrangle), and in the younger hornblende granite gneiss 1.3 miles northeast of Degrasse.

TABLE 7.—*Modes of anorthositic gabbros*

[Volume percent]

	1	2
Labradorite	69.1	65.4
Augite	13.5	20.3
Hypersthene	.5	
Hornblende	9.5	
Magnetite and ilmenite	1.8	10.1
Biotite	.3	2.1
Apatite	.07	2.1
Pyrite	.05	
Scapolite	4.4	
Tourmaline	.1	
Quartz	.7	

1. Anorthositic gabbro, west border of Russell sheet, 1.2 miles east of Belleville School, Russell quadrangle.
2. Anorthositic gabbro, rich in ilmenite-magnetite, 0.7 mile northwest of South Edwards, Russell quadrangle.

DIORITE GNEISS

Diorite gneiss has been found only in the southern part of the Tupper Lake quadrangle, where it occurs as lenses or long narrow sheets.

A small sheet is well exposed for 2 or 3 miles along the road to Sabattis, beginning about 1 mile west of the bridge at the foot of Little Tupper Lake. The rock of this sheet is a medium-grained, greenish, well-foliated rock. It is varied in mineral composition and texture. The least metamorphosed facies is an augite diorite gneiss. A chemical analysis and the approximate mineral composition of a sample of the least altered facies are reported in table 8. In thin section the feldspars and augite are seen in small part to form a granular mosaic in thin folia between larger grains. The plagioclase grains have exsolved orthoclase in antiperthitic intergrowth, and the augite has hyperthene lamellae as exsolved intergrowths. There is a very little interstitial orthoclase.

In contrast to this least altered facies, much of the diorite sheet along the Sabattis road has a mosaic structure and contains recrystallized porphyroblastic plagioclase. These plagioclase porphyroblasts are somewhat larger than the groundmass and contain differently oriented grains of plagioclase as well as grains of other minerals. Parts of the sheet contain garnet, which occurs as coronas around the ilmenite grains.

TABLE 8.—*Chemical analysis, norm, and mode of diorite gneiss from Little Tupper Lake*

[Locality 0.9 mile west of bridge at foot of Little Tupper Lake: Analyst, H. Baadsgaard]

Chemical analysis (weight percent)		Norm		Approximate mode (volume percent)	
SiO_2	53.93	Quartz	1.14	Andesine (antiperthitic).	74.8
Al_2O_3	19.13	Orthoclase	6.12	Augite (hypersthene lamellae).	17.8
Fe_2O_3	1.30	Albite	40.18	Hornblende	.8
FeO	5.39	Anorthite	27.70	Ilmenite	5.8
MgO	2.06	Diopside {Wo, 7.24; En, 3.44; Fs, 3.70}	14.38	Apatite	.8
CaO	9.48				
Na_2O	4.74				
K_2O	1.00	Hypersthene	3.61		
H_2O+	.22	Magnetite	1.86		
H_2O-	.09	Ilmenite	3.95		
CO_2	.04	Apatite	.87		
TiO_2	2.11				
P_2O_5	.37				
MnO	.11				
Total	99.97				

The mineral composition of two representative samples is given in table 9 (Nos. 1 and 2). The plagioclase is andesine.

Another sheet, or perhaps the faulted extension of the sheet described above, is exposed about 1 mile northwest of Moonshine Pond.

Several other sheets and lenses of diorite gneiss occur within the granite gneiss to the south. The lens southeast of Charley Pond shows relics of a distinct primary subophitic texture, which serves to prove the igneous nature of the rock from which the gneiss was derived. Much of the gneiss of these belts and lenses has a distinct layering consequent on metamorphic differentiation, such that there are intermediate plagioclase layers one-eighth to one-half inch wide, alternating with thin, dark, much more mafic laminae. The mineral composition of several specimens is given in table 9. The garnet in No. 4 occurs both as coronas and as euhedral replacements of the andesine.

TABLE 9.—*Modes of diorite gneisses*

[Volume percent]

	1	2	3	4	5	6
Andesine	69.4	67.5	72.5	71.8	73.7	74.8
Hypersthene	--------	--------	9.5	3.7	6.1	--------
Augite	1.9	7.1	6.1	8.3	3.3	17.8
Hornblende	13.0	11.5	3 0	5.4	9.5	.8
Magnetite and ilmenite	2.5	2.9	2.8	2.6	1.1	5.8
Apatite	.9	.7	.6	.9	.4	.8
Orthoclase or microcline	2.7	5.8	4.7	4.4	2.6	--------
Quartz	9.6	3.1	.8	1.6	3.3	--------
Garnet	--------	1.4	--------	1.3	--------	--------

1. Hornblende diorite gneiss, 0.9 mile west of bridge at foot of Little Tupper Lake. Andesine is An_{33}. Different sample from the one analyzed.
2. Hornblende-augite diorite gneiss, 1.6 miles west of bridge at foot of Little Tupper Lake.
3. Hornblende-augite-hypersthene diorite gneiss, 0.3 mile southeast of east end of Charley Pond. Andesine is An_{42}.
4. Hornblende-augite-hypersthene diorite gneiss, 0.5 mile north of west end of Little Slim Pond. Andesine is An_{38}, locally An_{42}.
5. Hornblende-hypersthene diorite gneiss, 0.5 mile southwest of southwest end of Charley Pond.
6. Augite diorite gneiss, least metamorphosed facies, 0.9 mile west of bridge at Little Tupper Lake. Plagioclase is microantiperthitic with orthoclase.

It is difficult to determine the extent to which the potassic feldspar and quartz have been introduced by emanations from the enclosing granite. Part of the microcline in No. 3 occurs in such textural relations with the other minerals that it appears to be primary. But another part of the microcline occurs with quartz in veinlike fashion, as though introduced. The quartz, where greater than 1 percent, also appears to corrode the other minerals as though introduced. The hornblende is at least in part secondary after pyroxene, and all of it may have had this origin. If so, the original rock was a pyroxene diorite.

The accessory iron oxide mineral of four samples of the least altered diorite gneiss has been found to be exclusively ilmenite, whereas clearly granitized facies of the diorite gneiss carry magnetite, and locally more than a normal amount of apatite.

The diorite is reported by Cleaves Rogers (oral communication) to grade locally into an anorthositic facies on the Raquette Lake quadrangle. It is possible that the diorite is related to the anorthositic series of rocks; but this is not necessarily so, as the diorite could have been independent and itself developed an anorthositic facies.

The diorite gneiss is cut by granite dikes, but its relations to the quartz syenite series are unknown.

METAGABBRO AND AMPHIBOLITE

Mafic rocks, comprising metagabbro, metagabbro gneiss, and amphibolite, form sheets widely distributed within the granite and granite gneisses, though they are subordinate in volume relative to them. Locally they occur within the metasedimentary rocks.

The term "metagabbro" is used here to designate mafic rocks which still contain relics of an ophitic or doleritic texture characteristic of gabbro, but in which part of the original minerals have been reconstituted to hornblende. Some amphibolite that grades into metagabbro, or that is thought definitely to be derived from gabbro, is here called gabbro-amphibolite. The term "amphibolite" is used for metamorphic gneisses and granulose rocks of dark color and mafic composition, in which plagioclase and hornblende are characteristic and commonly predominant minerals. Locally, however, pyroxene, biotite, or garnet may also be major minerals. Apatite, magnetite, and (or) ilmenite are always present as accessory minerals. Quartz and potassic feldspar may occur in subordinate amounts in some amphibolites, either as a product of introduction by solutions of magmatic origin or as primary minerals in amphibolites of metasedimentary origin.

Relics of gabbroic textures are rare or indistinct in the mafic gneisses of the main igneous complex of this area.

METAGABBRO GNEISS AND GABBRO-AMPHIBOLITE

There is a great variety in the mineral and chemical composition of the metagabbro gneiss. Outside the district, relics of olivine are found in many of the least metamorphosed metagabbro bodies of the main igneous complex that still have relics of ophitic structure. Hypersthene is present in all and is also common in the facies that are largely reconstituted. The metagabbro within the metasedimentary rocks of the Grenville lowlands, however, rarely contains olivine, and may or may not carry hypersthene. Much of the metagabbro and gabbro-amphibolite within the metasedimentary rocks appears to have been derived chiefly from an augite or augite-hypersthene gabbro, whereas that in the main igneous complex was formed from an olivine-hypersthene-augite gabbro. Transitional facies, however, occur in both areas. No olivine has been found in any of the metagabbro or amphibolite in the area of this report; all the rocks have been too much metamorphosed.

The largest bodies of metagabbro and gabbro-amphibolite within the metasediments form the folded sheet in the southeast part of the Canton quadrangle, the sheet on the anticline a couple of miles northeast of Stellaville, and the sheets south of Stellaville. Also, many sheets are interlayered with the rocks of the garnet- and biotite-quartz-plagioclase gneisses.

The metagabbro gneiss mass a couple of miles northeast of Stellaville locally shows conspicuous layers of a facies much more mafic than normal. These layers range from 1 inch to 150 feet in thickness, and the foliation is conformable with them. Locally there are small-scale, sharply layered zones with alternating layers more mafic and more feldspathic than normal. These are interpreted as the product of gravity sorting, and the arrangement of the layers in the south-dipping part of the sheet on the Russell quadrangle indicates that the sheet is in normal position and not overturned. The more mafic facies carry 55–60 percent mafic minerals, and the more feldspathic facies about 75 percent plagioclase. The mafic minerals include augite partly altered to hornblende, garnet in reaction zones between the other mafic minerals and plagioclase, and iron oxides. The primary rock appears to have been an augite gabbro.

The gabbro-amphibolite of the hill south of the old pyrite mine at Stellaville is characteristic of one type of these rocks, where they have been partly altered by granitization. The rock here is crossed by a network of granite pegmatite veins, and the intervening rock has been quite completely reconstituted; some silica and alkalies have been introduced and lime, magnesia, and ferrous iron have been removed by leaching.

AMPHIBOLITE

The amphibolites have various modes of occurrence and origin. Amphibolites are common in the metasedimentary rocks and granite gneisses of the Grenville lowlands, where they have been studied by many geologists. A brief survey of their results is pertinent here. Amphibolite occurs as layers several inches to 10 feet thick in the metasedimentary garnet-biotite gneisses. Indeed, amphibolite is nearly always present in these gneisses and locally may form as much as 40 percent of their total thickness. The amphibolite layers are tabular; they can only rarely be proved to crosscut the foliation but are generally conformable with it. Amphibolites have been variously interpreted as metamorphosed diabase or gabbro sills, as metamorphosed basalt or mafic volcanic rocks, and as reconstituted calcareous shales or argillaceous limestones. Amphibolite masses also occur as lenses or layers in marble; some of these masses seem to be best interpreted as reconstituted argillaceous dolomitic beds. There are many phacolithic granite masses in the marble belts of the Grenville lowlands, each of which is locally or entirely separated from the marble by a thin complex of rocks in which amphibolite is dominant. The granite also contains many included layers of amphibolite. Such amphibolite has been interpreted as replacements of marble, effected by emanations from the granite. Masses of amphibolite, ranging in size from thick bodies many miles long to small, isolated lenses a few score of feet in length, are also present, predominantly in the marble but also cutting the metasedimentary gneisses. The rocks grade from normal amphibolite to a rock that, because of its igneous texture, can definitely be determined to be a metagabbro or gabbro.

Quartzose amphibolitic layers conformably interlayered with metasedimentary beds in the St. Lawrence County magnetite district have previously been described with other metasedimentary rocks. Amphibolite layers, ranging in thickness from a few inches to a few thousand feet, also occur widely distributed as included conformable layers within the granite and granite gneisses. Locally the cores of the thicker or best preserved smaller masses may show relics of an indistinct diabasic or ophitic texture, whereas the outer parts are granoblastic, well-foliated amphibolites. Most of the amphibolite within the granitic rocks, insofar as our present knowledge goes, has not been definitely proved to be of any specific origin, but its composition and local gradation into gabbro have led us tentatively to describe most of the well-defined belts as metamorphic facies of gabbro. Nevertheless, some of the amphibolite, particularly those layers in the granite gneisses that occur near areas of metasedimentary rocks, may be related in origin to the latter rocks.

The largest belt of amphibolite mapped in the area is the arc that extends from the west side of Church Pond (Stark quadrangle) northward and eastward through Catamount Mountain to the west side of Kildare Pond (Childwold quadrangle). The amphibolite in much of this belt is not uniform but is characterized by more or less arteritic injection of granitic pegmatite and locally includes sheets of granite. This belt doubtless extends farther south, but exposures are so poor that it has not been mapped.

The hornblende granite of the Childwold quadrangle contains many lenses of amphibolite, but scarcity of outcrops precludes accurate mapping and thorough understanding of the relations of these bodies.

Usually, the amphibolite, even where abundant, occurs with such a haphazard distribution that not more than a few outcrops can reasonably be mapped together as indicating a continuous belt. Indeed, it has been found impracticable in the present work to map more than a very few of the amphibolite bodies, and those the larger ones.

Much of the hornblende granite and granite gneiss in the southern quadrangles includes only an occasional thin layer of amphibolite. The sillimanite-microcline granite gneiss only rarely contains a layer of amphibolite. The hornblende-microcline granite gneiss of the Childwold area, however, has abundant included layers of amphibolite.

The amphibolite layers associated with hornblende-microcline granite gneiss in part grade into metagabbro; but they also occur as layers in pyroxene-microcline granite gneiss and could have been derived from metasedimentary layers of pyroxene gneiss or pyroxene skarn.

The evidence from the Adirondacks in general is that the gabbro now represented by amphibolite was in part intruded as extensive, thick sheets into the rocks that are now metasedimentary; in part, the gabbro was emplaced as numerous isolated lenses of moderate to small size, more or less conformable with the structure of the country rock. Both the larger and the smaller masses were later dismembered and pulled apart by profound plastic flowage of the country rock at a time of orogenic deformation, and still later they were further dismembered and displaced by granitic intrusions. In St. Lawrence County the quartz syenite series of intrusives only rarely contains inclusions of amphibolite, but in adjoining areas amphibolite inclusions are locally common in members of the quartz syenite series.

Granitic materials have been introduced into much of the amphibolite, either as arteritic emplacements along the foliation planes, yielding a migmatite, or as partial replacements of the amphibolite. The potassic feldspar and subordinate quartz of the amphibolites, whose mineral composition is given in table 10, has been introduced as a replacement. The quantity of apatite in the amphibolite from east of Streeter Lake is far greater than normal and is similarly an aspect of granitization. The apatite is in good euhedral crystals. The content of both magnetite and apatite is commonly greater in the modified gabbro than in either the original granite or the original gabbro.

TABLE 10.—*Modes of amphibolites (metamorphosed gabbros?)*

[Volume percent]

Locality	Quartz	Plagioclase	Potassic feldspar	Augite	Hypersthene	Hornblende	Biotite	Magnetite and ilmenite	Apatite
Amphibolite									
0.5 mile east-southeast of Brouses Corners, Russell quadrangle	----	38.8	------	-----	----	52.9	5.1	2.3	0.4
1 mile east of Deerskin Pond, Stark quadrangle	----	52.6	------	-----	----	39.5	2.0	4.8	1.1
1 mile north of Lem Pond, Stark quadrangle	----	50.3	0.2	0.9	-----	35.2	11.7	.7	1.0
0.8 mile northeast of Tooley Pond, Stark quadrangle	----	47.3	------	9.7	-----	41.0	-----	1.6	.4
0.5 mile northwest of Dillon Pond, Cranberry Lake quadrangle	---	49.0	-----	4.7	19.0	19.0	----	8.1	.2
1.6 miles north of Russell	---	41.6	------	20.7	18.7	16.8	-----	2.2	------
0.9 mile southwest of Camp No. 15, Childwold quadrangle	----	46.5	------	-----	-----	48.8	-----	4.0	.7
1 mile west of Catamount Mountain, Stark quadrangle	----	48.1	------	1.8	2.2	44.8	-----	2.8	.3
Amphibolite modified by emanations from granite									
0.7 mile east of Streeter Lake, Oswegatchie quadrangle	2.6	50.8	2.8	1.3	10.2	20.3	----	5.8	6.2
0.7 mile southeast of outlet of Chair Rock Creek, Cranberry Lake quadrangle	.2	56.9	5.1	15.2	10.1	8.8	-----	3.2	.4
Southwest of Moosehead Rapids, Childwold quadrangle	2.6	42.4	[1] 17.2	6.4	10.5	16.0	---	3.7	1.2
0.5 mile east of Moosehead Pond, Cranberry Lake quadrangle	1.7	48.0	23.6	-----	----	19.1	6.6	---	1.0
0.75 mile southeast of Boyd Pond, Russell quadrangle	1.2	37.2	5.2	-----	---	38.7	12.7	4.6	.4
0.3 mile east of Buck Pond, Childwold quadrangle	3.0	45.7	-----	6.2	6.2	31.7	----	5.7	1.5
Representative migmatitic amphibolite									
0.7 mile east of Deer Pond, Tupper Lake quadrangle	21.6	38.4	26.9	2.7	3.2	6.2	-----	0.7	0.3
Foot Grass River Flow, Childwold quadrangle	17.6	13.7	53.8	-----	-----	8.6	-----	4.9	.6
Moosehead Rapids, Childwold quadrangle	9.0	45.0	[1] 33.9	1.5	3.0	2.5	-----	3.7	1.4
Hollywood School, Childwold quadrangle	12.1	49.3	16.1	-----	-----	17.6	-----	4.4	.5

[1] Mostly antiperthite.

There are three distinct mineralogical facies among the amphibolites insofar as the major mafic minerals are concerned: (a) hornblende, (b) augite-hornblende-hypersthene and (c) garnet-hornblende-pyroxene. An augite-hornblende variant is also known.

The hornblende facies uniformly occurs as thin layers within the granite or as border facies of the other types. It is thought to be the facies most thoroughly reconstituted by the emplacement of the granite. Locally there are relics of augite in the hornblende, indicating that the hornblende is partly the product of alteration of augite. Biotite is usually an accessory mineral.

The augite-hornblende-hypersthene facies occurs alone or in association with the garnet facies. The garnet amphibolite facies has nowhere been seen in contact with granite or granite gneiss, nor in arteritic amphibolite migmatite.

The garnet facies was observed in the belt of amphibolite which extends southeastward from a point about 1 mile northeast of Tooley Pond (Stark quadrangle), in abundant glacial boulders on the ridge east of Burntbridge Pond (Tupper Lake quadrangle), and near the boundary line between St. Lawrence and Franklin Counties one-half mile north of the Raquette River. The garnet at all of these localities occurs as a few large (½–1 inch) disseminated porphyroblasts, forming less than 10 percent of the rock by weight.

The amphibolite south and southwest of Bog Lake (locally known as Robin Lake, Cranberry Lake quadrangle) is, in part at least, interpreted as a metamorphosed facies of gabbro. It is a medium-grained, well-foliated, knotty-weathering rock, gray on exposed surfaces and slightly greenish on fresh surfaces. Small red garnets are common, and the rock has a few thin biotite-rich layers. The microstructure is crystalloblastic. Andesine (An_{35-40}) and dark green hornblende are essential constituents. The hornblende has olive, Z=dark brownish green; $Z \wedge c$=18°, $2V_X$ ca. 70°. the following properties: X=greenish yellow, Y= Quartz, microperthitic microcline, and biotite are present, probably introduced by emanations from the granite. Garnet may enclose rounded granules of each of the other constituents. Iron oxides form large, irregular grains intergrown with, or rimmed by, garnet; or enclosing apatite. Another variety of the amphibolite has a slightly more calcic plagioclase (An_{40-50}), and hornblende with a smaller extinction angle.

Part of the amphibolite within the granite in the southern part of the Tupper Lake quadrangle is probably derived from diorite gneiss, which also occurs as separate sheets and lenses.

The chemical and mineralogical composition of a pyroxenic amphibolite sheet in the garnet- and biotite-quartz-plagioclase gneisses is given in table 11. This rock has a crystalloblastic structure, and the hornblende is secondary after the pyroxene. The plagioclase is a labradorite ($Ab_{44}An_{56}$). Samples of amphibolite within the garnet gneiss formation on Kimball Hill and 1.3 miles southeast of Marshville are augite-hornblende-

TABLE 11.—*Chemical analyses, norms, and modes of amphibolites*

	1	2	3
Chemical analyses (weight percent)			
SiO_2	47.93	51.66	44.78
Al_2O_3	13.88	16.85	17.64
Fe_2O_3	3.37	2.48	2.80
FeO	10.99	6.48	9.63
MgO	7.42	5.94	7.11
CaO	10.67	8.70	10.13
Na_2O	2.29	2.91	2.87
K_2O	.59	1.64	.64
H_2O+	.44	1.12	1.28
H_2O-	.07	.05	.13
CO_2	.02	.59	n.d.
TiO_2	1.77	.97	2.38
P_2O_5	.22	.25	.42
MnO	.26	.13	.35
BaO	----	.05	----
Cl	----	.04	.03
Total	99.92	99.86	100.19
Norms			
Quartz	----	1.35	----
Orthoclase	3.34	10.01	3.89
Albite	19.39	24.63	21.26
Anorthite	25.85	27.94	33.50
Olivine	6.01	----	17.56
Hypersthene	14.96	19.45	----
Diopside	21.11	7.91	11.45
Magnetite	4.87	3.71	4.18
Ilmenite	3.34	1.82	4.56
Apatite	.50	.61	1.01
Calcite	----	1.35	----
Nepheline	----	----	1.56
Modes (volume percent)			
Quartz	----	6.8	----
Orthoclase	----	----	----
Plagioclase	34.5	36.5	30.0
Hypersthene	14.0	----	1.0
Augite	19.3	----	----
Hornblende	23.5	38.7	60.5
Biotite	----	11.2	.8
Magnetite and ilmenite	2.1	----	1.7
Sphene	----	.7	----
Apatite	.5	.8	1.0
Scapolite	----	4.7	----
Dactylic intergrowth	6.1	----	----
Almandite	----	----	5.0

1. Augite-hornblende-hypersthene-labradorite amphibolite as 10-ft layer intercalated in migmatitic garnet gneiss; edge of hill 1½ miles northeast of Colton, Potsdam quadrangle (Buddington, 1939, p. 12). Analyst, S. S. Goldich. Dactylic intergrowth is of clinopyroxene and plagioclase. Plagioclase is labradorite ($Ab_{44}An_{56}$).
2. Biotite-hornblende-andesine amphibolite, a slightly granitized, completely recrystallized metagabbro; hill top, just south of old pyrite mine at Stellaville, Russell quadrangle (Buddington, 1939, p. 186). Analyst, A. Willman. Plagioclase is $Ab_{70}An_{30}$.
3. Almandite-hornblende-andesine amphibolite, 1 mile south of junction of Cook Pond outlet and Grass River, Stark quadrangle. Analyst, Eileen K. Oslund. Plagioclase is $Ab_{53}An_{47}$.

hypersthene-plagioclase gneisses. These sheets are thought to be gabbro-amphibolite.

There are two iron oxide minerals in the gabbroic rocks and amphibolite—magnetite and ilmenite. Their quantity and their ratio to each other have a wide primary variation in the gabbro, and this is reflected in the equivalent gneisses. Where the gabbroic rocks have been metamorphosed to amphibolites, a secondary variation is usually superimposed upon the primary variation in content and ratio of the ferromagnetic minerals. This secondary effect is consequent upon the iron of the primary magnetite being taken up by the newly developed hornblende (and by almandite, where this is present), with the result that ilmenite remains as the sole iron oxide mineral. The normal garnetiferous and nongarnetiferous amphibolites thus commonly carry no magnetite but only ilmenite. Where, however, the amphibolites have been veined by granitic material and slightly granitized, they commonly carry an abnormally high percentage of magnetite. The magnetite of all the mafic rocks usually carries one or more lamellae of ilmenite and in part a lattice of exsolution ilmenite intergrowth. The ilmenite is in homogeneous grains.

QUARTZ SYENITE GNEISS SERIES: DIANA, STARK, AND TUPPER COMPLEXES

Three complexes of igneous rocks in the northwestern Adirondacks, each including syenite, quartz syenite, and granite gneiss facies and all thought to be genetically related, have been previously described (Buddington, 1939, p. 73–131) as the Diana, Santa Clara, and Tupper complexes. Each of these three complexes is situated in part within the St. Lawrence County magnetite district. The Santa Clara complex, as originally described, included all the associated quartz syenite rocks of what appeared to be a continuous belt across the Santa Clara quadrangle. Later work indicates the probability that the Santa Clara complex as originally mapped really includes members of two different complexes—the rocks of one complex being similar to those of the Tupper complex and charnockitic in character, whereas the rocks of the other complex normally, except for the mafic facies, carry little or no hypersthene. The complex that carries little or no hypersthene extends as a belt southwest from Santa Clara across the Stark and Russell quadrangles; it is here called the Stark complex, following the designation previously given it (Buddington, 1948, p. 25–28). Each of the three complexes (Diana, Stark, and Tupper) has an exposed length of 35–45 miles and a general width of a few miles.

The rocks of the Diana and Stark complexes range from green augite syenite and augite-quartz syenite gneisses to reddish hornblende-quartz syenite and hornblende granite gneiss. All members have a coarse, phacoidal structure, except for thin medium-grained sills isolated in the metasedimentary rocks. All the rocks of the Tupper complex are green and vary from ferrohypersthene-augite syenite gneiss to hypersthene (eulite)-ferroaugite-ferrohastingsite-quartz syenite gneiss. The content of hypersthene and the nature of the metamorphism relate the members of the Tupper complex to the charnockitic series of rocks. Much of the rock of the quartz syenite gneiss series could also be appropriately called pyroxene granite gneiss and hornblende granite gneiss. However, the term quartz syenite gneiss has been used for a long time in the Adirondacks to designate the quartzose, pyroxenic, and hornblendic rocks of this series, which carry between 10 and 20 percent quartz. The usage helps to differentiate this series from the younger hornblende granite and granite gneiss.

Each member of the quartz syenite gneiss series of rocks has peculiar characteristics that commonly but not always permit its discrimination in the field from the various facies of the younger granite gneisses.

The pyroxenic gneisses are almost uniformly green in color. Beneath the immediate surface veneer they have a peculiar maple-sugar color, and the inherently green color of the fresh rock is generally found only at a depth of several inches, though locally and exceptionally near the surface. Since the members of the quartz syenite gneiss series are so largely deformed and recrystallized to a granoblastic aggregate, they commonly have a granular character when weathered. The feldspar of much of the Tupper complex is characterized on the weathered surface by a peculiar dull whiteness. The distinctly hornblendic facies are commonly pink. The members of the Diana and Stark complexes have a coarse phacoidal structure with a varied quantity of relict porphyroclasts and in many places a small amount of porphyroblastic hornblende or pyroxene. The phacoids consist exclusively of a granoblastic feldspar aggregate, locally with varied percentages of feldspar porphyroclasts. Such quartz or mafic minerals as may be present are wrapped around the phacoids of feldspar. In the pyroxene syenite gneiss and the pyroxene- and hornblende-quartz syenite gneisses, the core of many of the large feldspars or phacoids is plagioclase and the rim is microperthite. Such feldspars or aggregates may yield a small cup-shaped structure on the weathered surface, consequent on the greater ease of weathering of the interior core of plagioclase as compared with the more resistant raised rim of perthite. In the red hornblende-quartz syenite gneisses, there are commonly green to yellow phacoids, owing to granulated large plagioclase crystals interspersed among the prevailing reddish phacoidal aggregates. In the phacoidal hornblende granite gneisses, however, a rapakivi structure is very common. In this type, the core of the phacoids is a granoblastic aggregate of perthite or microcline and plagioclase, and the rim is exclusively plagioclase.

Locally we have been unable to determine in the field precisely where certain members of the quartz syenite gneiss series stop and the younger granite begins, for in part, in the contact zone between younger granite and quartz syenite gneiss the latter takes on a reddish tint; its pyroxene, where originally present, is changed to hornblende; and where both have been strongly deformed, the textures of both rocks become similar. This development of a completely hornblendic facies is interpreted as a product of contact metamorphism of the quartz syenite gneiss by younger granite. Similarly the phacoidal hornblende granite gneiss may lose its phacoidal structure near the contact with the younger hornblende granite gneiss, and since both have locally been deformed and recrystallized, the contact between them is then indeterminate in the field. On the other hand, in most of the area, the younger granite is distinctly less deformed and granulated than the neighboring quartz syenite gneisses. The latter may then be distinguished in the field by their more granulose character and by relics of phacoidal structure.

Labradorite crystals have been found here and there in each of the various facies of the Diana and Tupper complexes throughout their extent. They have been found at a few localities in the Stark complex. These are interpreted as xenocrysts resulting from the strewing-out of crystals derived from the disintegration of incorporated blocks of anorthosite or anorthositic gabbro. The anorthosite or anorthositic gabbro from which the xenocrysts and fragments were derived is thought to have formed sheets in the metasedimentary rocks of the Grenville series at horizons below the sheets of quartz syenite gneiss.

DIANA COMPLEX

The extreme diversification of facies of the quartz syenite rocks and the best exposure of their relations are shown in the Diana complex, a part of which crops out in the southwestern border of the district. The complex has previously been described in detail (Buddington, 1939, p. 76–109). The succeeding discussion is based largely on this past work, though new data were obtained in the course of the present survey.

The Diana complex forms part of the northwest border of the dominantly igneous complex of the Adirondack highlands and extends continuously for 40 miles north and northeast from Lowville to about 2½ miles west of South Russell. Its total length is not known, for on the southwest it passes beneath a blanket of overlying Paleozoic sedimentary rocks. It has an exposed area of about 280 square miles. It is 15 miles wide on the southwest (southeast of Carthage) and thins at the extreme north to a width of 1 mile or so.

The manner in which the rocks vary in mineral composition across the strike in the general area between Bacon (Lake Bonaparte quadrangle) and the vicinity of Jayville (Oswegatchie quadrangle) is shown in table 12.

The western border facies of the complex for a width of one-half mile or more, in a belt extending from South Edwards southwestward to the west edge of the area, consists of an agmatite or igneous breccia of pyroxene-quartz syenite and fragmental layers of metasedimentary rocks of the Grenville series. The quartz syenite has intruded the metasedimentary rocks, for the most part parallel to the foliation, and many of the included fragments are long and relatively thin.

TABLE 12.—*Modes of facies of the Diana complex*

[Volume percent]

	Specimens averaged	Feldspar		Quartz	Hypersthene	Augite [1]	Hornblende [2]	Biotite	Magnetite and ilmenite	Apatite	Zircon	Sphene
		Microcline	Plagioclase									
Bacon to Jayville, Jenny Creek section, Lake Bonaparte and Oswegatchie quadrangles												
Hornblende granite gneiss	4	[3] 39.9	29.5	23.3	----	----	5.8	0.7	0.4	0.2	0.2	----
Hornblende-quartz syenite gneiss	2	[3] 50.5	26	15.8	----	----	6.4	----	.7	.4	.2	----
Transitional pyroxene-hornblende-quartz syenite gneiss	2	74.1		14.7	0.3	5.1	2.9	----	2.1	.5	.2	0.1
Pyroxene syenite gneiss	2	86.8		3.4	.3	5.5	.3	----	2.5	.7	.2	.4
Do	8	83.4		6.6	1.0	5.0	.8	----	2.3	.6	.2	.2
Shonkinite gneiss	1	25.8		3.8	.6	51.6	----	----	15.9	1.5	.5	.3
Ilmenite-magnetite feldspathic ultramafic gneiss	1	7.7		.6	2.9	58.0	----	----	27.0	2.0	.9	.9
Pyroxene-quartz syenite gneiss, western border facies	2	80		13.8	----	3.5	----	----	1.9	.6	.2	----
Stammerville School road, Russell quadrangle [4]												
Chloritic granite gneiss, 0.7 mile east of Stammerville School	1	37.1	36.0	22.0	----	----	----	[5] 2.4	0.5	0.3	0.1	1.6
Chloritic granite gneiss, 0.5 mile west of Stammerville School	1	81.3		14.4	----	----		[5] 2.3	.4	.2	.1	1.3
Hornblende-quartz syenite gneiss, 1.5 miles west-southwest of Stammerville School	3	84.4		12.8	----	----	1.2	----	1.0	.1	Tr.	.5

[1] Probably ferroaugite, near augite and salite in composition.
[2] Probably ferrohastingsite in hornblende-quartz syenite and hornblende granite gneiss, femaghastingsite in transitional gneiss.
[3] In part slightly microperthitic.
[4] Mineral variation in facies of Diana complex.
[5] Chlorite and biotite.

They range in length from a few feet to many hundreds of feet. In any given outcrop they may be sparse or abundant. The inclusions may consist of feldspathic quartz-rich gneiss, pyroxene granulite, or silicate-bearing marble. The syenite belt north of South Edwards is interleaved with layers of country rock, but the breccia structure is poorly developed.

A rock called microcline granulite, resembling a syenite aplite, was originally described by Buddington (1939, p. 14–16) as forming included layers in pyroxene-quartz syenite of the border facies of the Diana complex. Reexamination and detailed study of the type locality, however, has revealed that the analyzed microcline granulite possibly came from a dike about 2 feet wide in feldspathic quartzite and at an angle to its foliation. One such narrow dike can be seen to be varied along its length from a mylonitized aplite-like syenite to a porphyroclastic syenite like the groundmass of the major intrusive. Both the mylonitized syenite and the quartzite are pink, and careful observation is required to distinguish one from the other.

PYROXENE-QUARTZ SYENITE GNEISS, BORDER FACIES

The pyroxene-quartz syenite gneiss of the border facies forms a belt about one-half mile wide southwest of Red School (Oswegatchie quadrangle) and thins out rapidly to the north of Red School. The quartz syenite gneiss of the western border facies is pinkish to purplish, becoming greenish toward the more syenitic gneiss facies in the Jenny Creek section. Throughout, the border facies has a gneissic structure. The original structure is thought to have consisted of abundant feldspar crystals 1–1½ cm long in a coarse-grained groundmass. The typical rock now has a mortar, augen, porphyroclastic, or in small part a phacoidal structure as a result of deformation. The phacoids vary from those consisting of porphyroclasts with veinings of granulated material to lenses consisting wholly of a granular aggregate. The feldspars are largely composite, consisting of a core of oligoclase with a rim of microperthite. Some are predominantly oligoclase. The latter are green to yellow and give rise on deformation to phacoids of granular material. The phacoids are distinctive in color and weather more rapidly than the rest of the rock, giving it a pitted appearance. The oligoclase cores of the composite crystals also weather more rapidly than their perthite borders, thus giving rise to small cup-shaped depressions with raised rims. The porphyroclasts range up to a few millimeters in diameter. The quartz is in leaflike or in thin interlacing veinlike form.

The average grain of the granulated, cataclastic feldspar aggregate of the mortar is 0.1 mm and that of the granules of the granoblastic quartz leaves is 0.2 mm.

The rock is composed of the following minerals, in weight percent: feldspar, 74–82; quartz, 10–15; ferroaugite, 4–7; hornblende, 0.3–1; and combined magnetite and ilmenite, 2–5. Accessory minerals include sphene, zircon, and apatite. The hornblende is secondary after the ferroaugite.

PYROXENE SYENITE GNEISS

In the Jenny Creek section the pyroxene-quartz syenite gneiss of the border facies passes transitionally but sharply into green pyroxene syenite gneiss. This forms a belt parallel to the border facies. It is one-half mile wide southwest of Jenny Creek but thins out to the northeast. It is well developed to the southwest of Jenny Creek for the full length of the complex. The rock is slightly coarser grained than the quartz syenite gneiss and is similarly deformed. Much of it ranges in structure from a mortar gneiss to an augen gneiss. Locally it is granoblastic.

SHONKINITE AND FELDSPATHIC ULTRAMAFIC GNEISS LAYERS

On the hill 1.3 miles west of Kalurah and 0.55 mile south-southwest of BM–1031, there is a zone about 200 feet wide in which the normal pyroxene syenite gneiss is interlayered with shonkinitic syenite gneiss, shonkinite gneiss, and feldspathic ultramafic gneiss, mostly the latter three. A dip-needle survey on this belt shows a positive magnetic anomaly of 15°–25° at many places and locally up to 45°, owing to the concentrated disseminations of ilmenite and magnetite. The feldspathic ilmenite-magnetite-pyroxene gneiss has a medium-grained granular structure and is usually sharply delimited at each side. Feldspars up to 1½ inches long occur in any variety of the rocks here. The mineral composition of a specimen of the ilmenite-magnetite-pyroxene gneiss is given in table 12 (No. 7). The feldspathic ultramafic gneiss layers range from a fraction of an inch to 3 feet in thickness. This mafic and ultramafic lens was traced for more than one-quarter mile. The chemical analysis of a pyroxene (No. 2) from a feldspathic ultramafic gneiss layer is given in table 15. It is interesting to note that the pyroxenes of the normal pyroxene-quartz syenite gneiss and of the concentrations in the shonkinite gneiss and the feldspathic ultramafic gneiss have a similar chemical composition, which is that of a ferroaugite very close to the varieties augite, salite, and ferrosalite of Hess's (1941, p. 518) classification.

HORNBLENDE-QUARTZ SYENITE GNEISS

The pyroxene syenite gneiss grades through a narrow transitional belt of pyroxene-quartz syenite gneiss and pyroxene-hornblende-quartz syenite gneiss into a red hornblende-quartz syenite gneiss. The hornblende-quartz syenite gneiss forms a belt a little more than a mile wide from Greenwood Creek through Kalurah to the southwest. It is a medium-grained, roughly equigranular gneiss, except for the leaves of quartz. Occasionally a porphyroclastic fragment of feldspar is present. The feldspars constitute in general 75–80 percent of the gneiss and consist largely of granoblastic microcline and oligoclase with some relict porphyroclasts of microperthite. Quartz varies from 7 to 18 percent. Pyroxene is rarely present and then only as an accessory mineral. Hornblende commonly constitutes 4–8 percent, and combined magnetite and ilmenite, 1½–3 percent. Apatite, zircon, and occasionally sphene are accessory minerals.

HORNBLENDE GRANITE GNEISS, PHACOIDAL

The hornblende-quartz syenite gneiss is separated from the next member of the Diana complex on the east by a narrow layer of intrusive younger granite. To the east of this younger granite is a belt of coarse hornblende granite gneiss that extends from just north of Yellow Creek through Jayville to the south border of the area. It is about three-fourths mile wide at the north and 1½ miles wide at the south.

The granite gneiss of this belt in part has a coarse phacoidal structure like much of the quartz syenite gneiss series. Locally, however, the granite gneiss with phacoidal structure grades into a granite gneiss with an evenly foliated, coarse, equigranular structure, which is difficult to distinguish from the younger granite gneiss. The granite gneiss with typical phacoidal structure has a coarse lenticular structure. The individual lenses or phacoids consist of a granular feldspar aggregate embedded in a groundmass of quartz, hornblende, and some feldspar. They are often 1–3 inches long and ¼–½ inch thick. In some facies these lenses have coalesced, yielding a gneiss with the appearance of a set of veinlets of feldspar aggregate. The individual veinlike lenses may be as long as 6 inches. Some of the granite shows obscure phacoids of coarse granular feldspar aggregate in a meshwork of leaf quartz. The grain of the evenly foliated gneiss is commonly coarser than that of the granular aggregates in the phacoids. Locally the granite gneiss has the structure of an augen gneiss.

In thin section the hornblende granite gneiss is found to consist predominantly of a granoblastic aggregate of microcline and oligoclase with leaves of quartz which have uniform extinction. There is considerable variety in the size of the feldspar grains, 1–4 mm being common. The larger microcline grains usually have a little very fine grained microperthitic intergrowth of plagioclase. The microcline also encloses an occasional rounded grain of quartz. There is a little myrmekite. Hornblende is present in small grains, and a little biotite is commonly in flakes parallel to the foliation. Magnetite, apatite, and zircon are accessory minerals. Many of the deformed phacoids have granular plagioclase at the core, and microcline borders. In others evidence of an original rapakivi-like structure is often seen in the field, such as a thin rim of granular white plagioclase around phacoids of granular pink microcline.

NORTHERN PART OF DIANA COMPLEX

The part of the Diana complex north of Greenwood Creek is markedly different from that to the southwest. The sheet is much thinner; its western part is intimately interleaved with metasedimentary rocks; there is a much smaller diversity of facies; and the rock is dominantly felsic, except very locally where involved with carbonate metasedimentary rocks. The rocks are also in general much more thoroughly metamorphosed, locally with alteration of the primary augite to hornblende and (or) biotite, and of the primary hornblende to chlorite.

Between Greenwood and Jenny Creeks, there is a transition zone in which the pyroxene syenite gneiss, pyroxene syenite gneiss with shonkinite, and feldspathic ultramafic gneiss layers all die out northwards, and the pyroxene-quartz syenite gneiss of the border facies and the syenite gneiss both become somewhat more felsic in this direction. The composition of rocks from this transition zone between the rocks of the Jenny Creek section on the south and the more felsic facies to the north is given in table 13.

The belt on the Russell quadrangle consists of pinkish hornblende-quartz syenite gneiss on the west, grading into a chloritic granitic gneiss facies on the east. Porphyroclasts of feldspar are common in the gneisses

TABLE 13.—*Modes of quartz syenite gneisses and syenite gneisses near the north end of the Diana complex*

[Volume percent]

	1	2	3	4
Feldspar	80	81.7	84.1	80.3
Quartz	13.8	14.9	6.0	10.2
Hypersthene	----	----	.8	----
Augite	3.5	1.9	5.1	4.0
Hornblende	----	.2	.7	1.5
Magnetite and ilmenite	1.9	1.0	2.3	2.6
Apatite	.6		.6	.7
Sphene	----	.3	.2	.5
Zircon	.2		.2	.2

1. Average of 2 quartz syenite gneisses from border facies, Jenny Creek section, Lake Bonaparte and Oswegatchie quadrangles.
2. Average of 5 quartz syenite gneisses from border facies in section across structure, 1½ miles northeast of Jenny Creek.
3. Average of 8 syenite gneisses, Jenny Creek section.
4. Average of 2 syenite gneisses, 1½ miles northeast of Jenny Creek section.

but are usually small and few. Labradorite xenocrysts such as might have been derived from anorthosite were found three-fourths mile east of Shawville. Hornblende porphyroblasts are locally well developed in the hornblende-quartz syenite gneiss. Most of the gneiss has a phacoidal structure, but locally it is so strongly deformed as to be even grained and well foliated, without evidence of any original structure. Such gneiss is finer grained than normal.

BIOTITE SYENITE GNEISS

West of the main Diana complex and separated from it by a belt of metasedimentary rocks and alaskite gneiss there is a sheet of pink syenite gneiss that extends from Fordhams Corners to southwest of Barraford School. The composition of this syenite gneiss is somewhat varied. In general it is a biotite syenite gneiss, with inclusions of metamorphosed silicated limestone, pyroxene granulite, and quartz-feldspar granulite near the southwest end.

The rock is a fine-grained (0.5–0.6 mm), evenly foliated gneiss with a granoblastic texture. The characteristic rock averages in percent: microcline, 46.9; oligoclase, 43.6; biotite, 3.8; quartz, 2.9; magnetite, 0.5; apatite, 0.6; zircon, 0.2; and sphene, 1.5. Locally there may be a little hornblende or pyroxene as products of contamination.

No rock of this type has been found within either the Diana or the Stark complex, and its relationships and significance are obscure.

OUTLYING SYENITE GNEISS

West and northwest of the main Diana and Stark complexes there are several relatively thin sheets of syenite and quartz-bearing syenite gneiss within the metasedimentary rocks, which are thought to be related to these complexes.

A sheet of syenite gneiss extends from west of Portaferry Lake (Oswegatchie quadrangle) northward to about a mile northwest of South Edwards (Russell quadrangle). The rock is strongly deformed and consists largely of a reddish syenite gneiss, commonly containing only a little quartz and having hornblende as the prevailing mafic mineral. Locally, however, there is a little augite or hypersthene. Locally, also, the gneiss carries a little biotite. In part the biotite is in separate crystal plates; in part it is an alteration product on the edges of hornblende or pyroxene. The rock is almost completely crushed and recrystallized, and there are commonly only a few small porphyroclasts of feldspar. The feldspars are recrystallized to a mosaic of microcline and plagioclase with a grain of 0.3–1 mm. The hornblende is in considerable part in the form of porphyroblasts. The original rock may well have been a pyroxene syenite.

A sheet of pyroxene syenite gneiss lies to the west of the north-south road through Pond Settlement. This rock, too, is strongly deformed. A specimen from 0.3 mile southwest of Jones Pond has about the following composition in volume percent: microperthite, 48; oligoclase, 38; augite, 5; hypersthene, 6; hornblende, 1.5; iron oxides, 1.5; and accessory apatite and zircon. About 1.5 miles north of Jones Pond the gneissic structure swings to the northeast as though around the end of a fold. It appears that the sheet to the east and the west of Pond Settlement forms opposite limbs of a fold; these are continuous with each other about 1.4 miles north of Pond Settlement.

North and east-northeast of West Pierrepont for about 2 miles there is a narrow sill of highly foliated, dark hornblende-quartz syenite augen gneiss intercalated in limestone. The augen are porphyroclasts with plagioclase cores and perthite rims. The rock carries about 5 percent of sphene. This sheet is probably a metamorphosed facies of an augite syenite sill. A very small lens of augite-biotite-quartz syenite gneiss is also found about 1 mile north-northeast of Endersbees Corners and 1.1 miles west of Hamiltons Corners.

Another sheet of syenite gneiss occurs about 1.5 miles southeast of High Flats (Potsdam quadrangle). The rock is a green gneiss with abundant small (1–3 mm) augen and porphyroclasts of feldspar and grains of augite (largely or completely altered to hornblende) in a cataclastic groundmass of feldspar with a little quartz and granulated mafic mineral. The feldspar augen consist of a plagioclase core with microperthite borders, and the porphyroclasts are largely microperthite. Apatite is common, and iron oxides and zircon are the other accessory primary minerals. There is a little secondary sphene. The syenite gneiss sheet has a U-shaped outcrop, owing to folding. It is intruded by a chloritized biotite granite, which does not show as marked a cataclastic crushing as does the syenite gneiss.

CHEMICAL COMPOSITION

A series of chemical analyses and norms of representative facies of the Diana complex is given in table 14.

ORIGIN AND SIGNIFICANCE OF VARIATION IN COMPOSITION, DIANA COMPLEX

The preceding descriptions have shown that the quartz syenite gneiss series of the Diana complex—except for a belt of quartz syenite gneiss along the northwest border and the thin northern portion of the sheet—varies systematically from a pyroxene syenite gneiss with local feldspathic ultramafic and shonkinitic

TABLE 14.—*Chemical analyses and norms of facies of the Diana complex*

	A	1	2	3	4	5	6
Chemical analyses (weight percent)							
SiO_2	63	38.25	47.83	57.95	63.62	63.47	68.40
Al_2O_3	16	4.31	8.47	16.30	16.11	15.27	14.87
Fe_2O_3	2.7	11.91	7.72	3.46	2.15	2.40	1.77
FeO	3.1	17.71	13.19	4.45	2.51	3.43	1.58
MgO	1.0	3.62	3.01	1.31	1.09	.93	.93
CaO	3.3	10.89	7.99	4.59	2.55	2.96	1.89
Na_2O	4.5	1.25	2.29	4.75	4.37	4.02	3.86
K_2O	5.0	1.10	2.52	4.37	5.06	4.97	5.13
H_2O+		.24	.24	.32	.67	.28	.32
H_2O-		.17	.12	.10	.14	.10	.04
CO_2				.30	.55	.26	.12
TiO_2	.8	7.15	4.28	1.41	.93	1.01	.63
P_2O_5	.4	1.89	1.56	.79	.34	.38	.19
MnO	.1	.60	.38	.16	.07	.07	.06
BaO			.07			.07	
ZrO_2		.46	.28				
Total	99.9	99.55	99.95	100.26	100.16	99.62	99.79
Norms							
Quartz	10.0	5.82	7.02	4.39	13.44	14.26	21.66
Corundum					.82		.23
Orthoclase	29.5	6.67	15.29	26.13	30.02	29.46	30.44
Albite	38.2	10.48	19.39	40.08	36.94	34.06	32.49
Anorthite	9.5	2.78	5.14	10.15	7.09	8.61	7.44
Diopside	3.6	33.83	21.33	4.71		1.81	
Hypersthene	3.0	3.76	8.55	4.20	4.15	4.16	2.70
Magnetite	4.0	17.17	11.14	4.99	3.14	3.53	2.58
Ilmenite	1.4	13.68	8.14	2.74	1.75	1.90	1.19
Apatite	.9	4.37	3.70	1.84	.81	.90	.44
Calcite				.70	1.30	.55	.20
Zircon		.70	.38				

A. Estimated average chemical composition of Diana complex as a whole, based on weighted average of chemical analyses (Buddington, 1939, p. 103).
1. Feldspathic ultramafic gneiss composed of feldspar, ilmenite, magnetite, and pyroxene, 1.5 miles west of Kalurah, Oswegatchie quadrangle, New York. Analyst, Charlotte Warshaw.
2. Shonkinite gneiss, 1½ miles south of Harrisville, Lake Bonaparte quadrangle (Buddington, 1939, table 21, No. 83). Analyst, T. Kameda.
3. Pyroxene syenite gneiss, average of 3 chemical analyses (Buddington, 1939, table 21, Nos. 84, 86, 87–L).
4. Pyroxene-quartz syenite gneiss, 0.2 mile north of North Croghan, Lake Bonaparte quadrangle (Buddington, 1939, table 21, No. 90). Analyst, R. B. Ellestad.
5. Hornblende-quartz syenite gneiss, average of 2 chemical analyses (Buddington, 1939, table 22, Nos. 91 and 92).
6. Hornblende granite gneiss, average of 2 chemical analyses (Buddington, 1939, table 22, Nos. 94 and 96).

layers rich in ilmenite and magnetite on the northwest, through pyroxene-hornblende-quartz syenite gneiss and hornblende-quartz syenite gneiss, to hornblende granite gneiss on the southeast. These stratiform layers now dip uniformly to the northwest. If, however, they were rotated counterclockwise in cross section until the mass was in a more or less horizontal position, such that the pyroxene-quartz syenite was at the base and the hornblende granite at the top, they could reasonably be interpreted as differentiated facies of a thick, flat-lying intrusive sheet of igneous rock. Under this hypothesis, the pyroxene-quartz syenite gneiss of the western border would represent a relatively undifferentiated part of the original magma (because of quicker chilling due to the many included fragments of country rock); the pyroxene syenite gneiss, skonkinite gneiss, and feldspathic ultramafic layers would represent mafic facies, owing to the accumulation of pyroxene, magnetite, ilmenite, apatite, and zircon (to a minor extent) in the lower portion of the sheet as heavy minerals of a relatively early period of crystallization; and the hornblende-quartz syenite gneiss and hornblende granite gneiss would represent the complementary lighter, more felsic facies accumulated in the upper part of the magma sheet with transitional pyroxene-quartz syenite gneiss and pyroxene-hornblende-quartz syenite gneiss in between.

The entire Diana complex (Buddington, 1936; 1939, p. 88–107) has been previously studied with this type of hypothesis in mind, and the tentative conclusion was reached that the complex may be interpreted as an isoclinally folded, differentiated, gravity-stratified sheet. The section along Jenny Creek crosses the sheet where it has been overturned and now dips 40°–50° northwest.

The foregoing interpretation of the structure and origin of the relationship of the different members of the quartz syenite series will be tentatively adopted and used in interpreting the significance of masses of the quartz syenite series elsewhere, where the relationships are not so well known.

The chemical analyses of clinopyroxene and hornblende from the Diana and Stark complexes are given in table 15. The hornblende of the granite in the upper part of the Diana complex has optical properties similar to that in the granite of the Stark complex, so analysis 4 may serve as representative of both. There is a systematic increase in the ratio of FeO to MgO from that of the ferroaugite in the pyroxene syenite rocks of the lower part of the sheet, through the femaghastingsite in the hornblende-quartz syenite of the central part of the mass, to the ferrohastingsite in the granites of the upper part of the complex. This is consistent with the variation to be expected in a sheet differentiated largely by fractional crystallization. The relative concentration of magnetite and ilmenite in the pyroxene syenite facies and the decrease in the amount of mafic minerals from the lower to the upper parts of the sheet accounts for the decrease in iron and titanium in successively younger facies of the mass, even though the FeO/MgO ratio of the mafic minerals increases. The feldspathic ultramafic gneiss is noteworthy for the concentration of zircon with ilmenite, a rather unusual combination, but consistent and a necessary consequence of the mode of origin here postulated.

STARK COMPLEX

The Stark complex has been traced continuously southwestward from the southwest corner of the Santa Clara quadrangle across the Santa Clara, Nicholville, Potsdam, Stark, and Russell quadrangles to the north border of the Oswegatchie quadrangle, a length of about 45 miles. The general width for much of its length is 4–5 miles, locally up to 7 miles on the Nichol-

TABLE 15.—*Chemical analyses of some clinopyroxenes and hornblendes from the Diana and Stark complexes and from Moody Lake and Au Sable Forks*

[Weight percent]

	Diana and Stark complexes				Moody Lake	Au Sable Forks
	Pyroxene		Hornblende		Pyroxene	
	1	2	3	4	5	6
SiO_2	52.28	51.53	41.25	38.73	50.33	48.28
Al_2O_3	2.71	1.50	10.40	11.49	2.32	1.45
Fe_2O_3	2.58	2.72	3.85	5.22	1.88	3.96
FeO	12.61	13.13	16.28	22.89	18.23	27.02
MgO	8.70	8.92	8.02	4.10	6.92	.32
CaO	16.69	20.17	10.26	9.45	18.39	16.18
Na_2O	.78	.67	1.58	1.68	.61	1.51
K_2O	.28	.00	1.46	1.61	.07	.14
H_2O+	.38	.36	1.69	1.67	.16	.15
H_2O-	.56	.06	.10	.04	.09	.15
TiO_2	1.32	.19	2.90	1.70	.28	.28
MnO	.72	.82	.76	.58	.83	.76
F	-------	-------	1.17	.61	-------	-------
Cl	-------	-------	.60	.53	-------	-------
Subtotal	-------	-------	100.32	100.30	-------	-------
Less O for F+Cl	-------	-------	.62	.38	-------	-------
Total	99.61	100.07	99.70	99.92	100.11	100.20

1. Ferroaugite, from pyroxene-quartz syenite gneiss, 1.25 miles southwest of Harrisville at road crossing of West Branch Oswegatchie River, Lake Bonaparte quadrangle (Buddington and Leonard, 1953). Analyst, Lee C. Peck.
2. Ferroaugite, from feldspathic ultramafic gneiss layer in pyroxene syenite gneiss, zone 1.25 miles west of Kalurah, Oswegatchie quadrangle (Hess, 1949, p. 652, analysis 15). Analyst, Lee C. Peck. Rock from which mineral was concentrated is comparable to that of No. 1, table 14.
3. Femaghastingsite, porphyroblastic hornblende, from granoblastic quartz-bearing hornblende syenite gneiss, 3.1 miles south-southeast of Harrisville bridge, Lake Bonaparte quadrangle (Buddington and Leonard, 1953). Analyst, Eileen K. Oslund. Rock from which mineral was concentrated is comparable to that of No. 5, table 14.
4. Ferrohastingsite, from granoblastic phacoidal hornblende granite gneiss, 2 miles southeast of Degrasse, Russell quadrangle (Buddington and Leonard, 1953). Analyst, Lee C. Peck. The analysis of rock from which the mineral was concentrated is No. 2, table 18. Stark complex.
5. Ferroaugite, from mafic syenite gneiss, quarry northeast of Moody Lake, Saranac quadrangle (Hess, 1949, p. 653, analysis 17). Analyst, Lee C. Peck. The analysis of rock from which mineral was concentrated is No. 1, table 20.
6. Ferrohedenbergite, from fayalite-ferrohedenbergite granite, from quarry seven-eighths mile northeast of Au Sable Forks, N.Y. (Hess, 1949, p. 654, analysis 19). Analyst, Lee C. Peck. Analysis of rock from which mineral was concentrated is No. 7, table 20.

ville quadrangle, and considerably narrower across the Russell quadrangle, where it contracts on the southward-plunging nose of an anticline.

The Stark complex, southwest of the Clare-Colton township line, consists wholly of a pink phacoidal hornblende granite gneiss, except for a local narrow belt of green to pink pyroxene-hornblende granite gneiss along each border of the complex. To the east of the township line the width of the Stark mass is greater and the complex consists of a central core composed of green pyroxene-quartz syenite gneiss, interlayered with pink hornblende-quartz syenite gneiss, and locally with phacoidal hornblende granite gneiss. The outer part of this portion of the complex, however, also consists wholly of phacoidal hornblende granite gneiss. The core of the complex on the Stark quadrangle is not so thoroughly granulated as the border facies but consists of mortar and augen gneisses. To the east, however, the whole is granoblastic.

The mineral composition of representative facies of the Stark complex is given in table 16. Most of the potassic feldspars are slightly perthitic, with a range in appearance from orthoclase through an indistinctly twinned variety to distinct microcline. Locally there is some nonperthitic microcline in small recrystallized grains. The potassic feldspar of the granitic facies on the Russell quadrangle has in general less perthitic intergrowth than the feldspar of the Stark area. In the green pyroxene-quartz syenite gneisses, perthite-rimmed oligoclase forms conspicuous large crystals. In the hornblende granite gneiss, on the other hand, phacoids of granoblastic microcline are rimmed with granulated oligoclase.

Northeast of the main highway through Cold Brook School the quartz syenite gneisses almost uniformly contain a little garnet, predominantly as coronas around the mafic minerals but in part as small porphyroblasts. Not even a trace of garnet, however, has been seen in any of the equivalent rocks which are more than 1 mile southwest of the main road. The garnet, where it occurs, is contained in both the pyroxene- and the hornblende-quartz syenites. The garnetif-

TABLE 16.—*Modes of facies of the Stark complex*

[Volume percent]

	Potassic feldspar	Plagioclase	Quartz	Hypersthene	Augite	Hornblende	Biotite and chlorite	Magnetite and ilmenite	Apatite	Zircon	Garnet
Hornblende-quartz syenite gneiss, local roof facies											
1	49	23.4	16.4	0.5	1.2	9	-----	0.3	0.3	-----	-----
2	34.7	31.1	18.8	1.9	.6	11.8	-----	.4	.5	0.2	-----
3	37.1	42.3	14.5	-----	1.9	3.3	-----	.3	.3	.3	-----
Hornblende granite gneiss											
4	45	29.5	20	-----	-----	4.1	0.3	0.7	0.2	0.1	Tr.
5	39.1	20.5	38.9	-----	-----	-----	1.2	.3	-----	-----	-----
6	50.7	22.0	24.2	-----	0.5	1.8	.3	.4	.1	.1	Tr.
7	38.1	34.5	20	-----	-----	6.2	.3	.6	.3	.1	-----
8	45.5	33.2	16.5	-----	-----	3.4	.6	.5	.2	.1	-----
9	32	38.7	19.0	-----	-----	8.5	1.0	.4	.3	.1	-----
Pyroxene-quartz syenite gneiss, interlayered with hornblende granite gneiss											
10	44.1	32.8	17.6	0.1	2.4	1.1	0.7	0.7	0.2	0.1	0.2

1. Hornblende-quartz syenite gneiss, 2.2 miles northwest of Stillwater Club, Stark quadrangle.
2. Hornblende-quartz syenite gneiss, west of Allen Pond, Russell quadrangle.
3. Hornblende-quartz syenite gneiss, 1.5 miles east-northeast of junction of Gulf Creek and North Branch of Grass River, Stark quadrangle.
4. Average of eight samples of hornblende granite gneiss.
5. Felsic biotite granite gneiss, local facies, diamond-drill hole 45–5, Clifton mine.
6. Average of three samples of felsic hornblende granite gneiss, Stark quadrangle.
7. Average of three samples of more mafic facies of hornblende granite gneiss.
8. Average of two samples hornblende granite gneiss.
9. Hornblende granite gneiss, 1.9 miles south-southeast of Degrasse. The chemical analysis and norm of this rock are No. 2, table 18.
10. Average of seven samples of pyroxene-quartz syenite gneiss, Stark quadrangle.

erous facies is present as far to the northeast as the belt of Stark complex has been traced.

The granoblastic groundmass of the mortar and augen gneisses is quite varied; its average grain size is generally 0.25–0.6 mm, and in the more thoroughly granoblastic gneisses around 0.5 mm.

The intermediate index of refraction (nY) of pyroxenes from three samples of pyroxene-quartz syenite gneiss of the Stark complex ranges from 1.708 to 1.712. The nY of the ferroaugite of the Diana complex is 1.706, hence the pyroxene of the quartz syenite gneisses of the Stark complex must be a ferroaugite very slightly richer in iron.

The optical properties of the hornblende of several specimens of the hornblende granite gneiss were determined. They prove the hornblende in all cases to be ferrohastingsite. An analysis of a representative ferrohastingsite is No. 4, table 15.

About 0.5 mile northeast of Gain Twist Falls (Potsdam quadrangle), xenocrysts of labradorite as large as 2 inches were found in the phacoidal pyroxene-quartz syenite gneiss.

The chemical composition of a representative sample of the slightly more mafic facies of the hornblende granite gneiss and of an altered chloritic facies is given in table 18, Nos. 1 and 2.

Locally on the border of the Stark complex there is a generally more mafic and less quartzose facies of the gneiss. This facies is also commonly green, in contrast to the underlying pink hornblende granite gneiss, and usually carries a little pyroxene. Such rock is found in a belt from one-half mile west of Canton Farm (Stark quadrangle) south for 2.5 miles to the highway; on the hill 2 miles northeast of Stillwater Club, on the west side of the complex in a belt from one-half mile west of Chapp Hill southeast for about 3.5 miles including Albert Marsh Hill, and 0.6 mile north of Gain Twist Falls (Potsdam quadrangle).

ORIGIN OF DIVERSIFICATION

The diversified facies of the main body of the Stark complex, by analogy with the Diana complex, are interpreted as the product of gravity differentiation more or less in place in an originally flat-lying magma sheet. The hornblende-quartz syenite gneiss and hornblende granite gneisses are interpreted as an upper felsic differentiate, and the interlayered pyroxene- and hornblende-quartz syenite gneiss facies of the core as a transitional zone between the more felsic facies above and a more mafic facies below, which is postulated to occur at depth but is not yet exposed.

The local more mafic, less quartzose, border facies may be relics of a relatively undifferentiated roof or may have had some other origin. Exposures are not adequate to determine the relation of these rocks to others.

LOCAL MICROCLINE SYENITE CONTACT FACIES OF HORNBLENDE GRANITE GNEISS WITH MARBLE

A multitude of excellent contacts of the phacoidal granite gneiss with layers of metasedimentary rocks have been observed in the diamond-drill core from the Clifton mine. The granite gneiss shows little or no change where in contact with quartz-feldspar granulite layers, but where in contact with skarn replacements of marble the granite gneiss in many places shows development of a characteristic and unique type of skarn on the marble side and a very marked change in the nature of the adjoining granite.

The contacts of the hornblende granite gneiss with the included granulite layers are generally very sharply defined. In part adjacent to the main contact, thin layers (⅛–½ in.) of porphyritic granite gneiss alternate with granulite layers for a width of a few inches. In part, isolated porphyroblasts of feldspar have grown within the granulite, but not more than a few inches from the contact.

The phacoidal granite gneiss, on the other hand, commonly shows a diminution in the apparent intensity of development of foliation towards skarn until in the contact zone the rock appears medium grained, massive, and homophanous. The transition zone is green to white, in contrast to the pink of the normal rock. In places, the change of color and development of an apparently massive character takes place within a few inches; elsewhere, the contact zone is as thick as 25 feet. Rarely, the contact facies retains its pink color, though the rock loses its foliated character and appears massive.

Both the mineralogy and the structure of the contact skarn zone are commonly different from the normal skarn zone at contact with the younger granites. The contact skarn zone of the granite of the Stark complex is characteristically a thin-layered rock consisting of alternate leaves, ⅛–½ inch thick, of wollastonite with subordinate to minor pyroxene and of white syenite with a little quartz. The feldspar of the syenite is a granoblastic microcline aggregate, in part slightly microperthitic and usually with a partial or complete selvedge of plagioclase around each microcline grain. Scapolite is present as an accessory mineral in a little of the migmatite, and rarely a corona of garnet has developed at the contact between feldspar and pyroxene or feldspar and wollastonite. Plagioclase is present as an accessory mineral in discrete grains, and locally a very little myrmekite is also present. Sphene occurs as an accessory mineral throughout the rock. This type of rock has been found only in the contact zone between marble and the phacoidal granite gneiss. The migmatitic skarn zone may range in thickness from a few feet to 70 feet.

Within the skarn zones, sheets of massive-appearing white granite gneiss as thick as 30 feet may also occur, and independent thin layers of white syenite gneiss.

In part the syenite migmatite is a rock with lenticular laminae of pyroxene in place of wollastonite. Also, there are locally all gradations between wollastonite or pyroxene with a little microcline, and microcline syenite with a little wollastonite or pyroxene. The wollastonite and pyroxene grains are inequidimensional and well oriented parallel to the foliation. The sparse quartz is in elongate lenses. The microgranular structure of the syenite is like that of the phacoidal granite gneiss, and the whole zone is thought to have undergone deformation and recrystallization subsequent to its formation.

Two representative sections to show the relations of the various rocks are given in table 17.

Chemical analyses of pink phacoidal granite gneiss, and of the white pyroxene-microcline granite gneiss and white pyroxene-microcline syenite gneiss facies developed in a contact zone with marble, are given in table 18.

It will be noted that the white microcline granite gneiss bodies associated with the contact zones have a higher ratio of K_2O to Na_2O than any of the granite. The white microcline granite gneiss also has more quartz than the pink phacoidal granite gneiss.

The common development of a pyroxene-microcline syenite gneiss as a transitional facies of the phacoidal hornblende granite gneiss adjacent to marble is possibly due to solution by the granite magma of volatile materials given off on silication of the limestone, and the activity of the volatile materials in concentrating K_2O relative to Na_2O in the contact zones of the magma sheets and transferring the SiO_2 from parts of such zones into the adjoining marble, where it is fixed by reaction. The movement of CO_2 out of the marble may also set up a dynamic environment favorable for the agencies effecting the desilication of the granitic magma. The increase of potassic feldspar relative to plagioclase may also be favored by the development of a more anorthitic plagioclase consequent on the assimilation of CaO. No syenite facies more than a few feet thick has been observed. Independent sheets of microcline granite as thick as 30 feet are found in the limestone, but independent sheets of microcline syenite are only a few feet thick at most.

The zones consisting of alternating leaves of microcline syenite and of wollastonite (or pyroxene) may be several scores of feet thick. Subsequent metamorphism has destroyed all primary structures, and such mixed rock could be interpreted either as a pseudomigmatite resulting from replacement of alternate leaves of skarn by microcline or as a true migmatite

TABLE 17.—*Sequence of rocks involving syenite contact facies, metasedimentary rocks, and migmatites, Clifton mine*

	Igneous rocks (feet)	*Metasedimentary rocks and migmatite (feet)*
A. Sequence in diamond-drill hole 69, depth 20–463 feet:		
White syenite gneiss with laminae of pyroxene skarn	1	-----
Pyroxene skarn and garnet skarn	-----	11
White pyroxene syenite gneiss	5.5	-----
Garnet skarn	-----	1
White pyroxene syenite gneiss	1.5	-----
Pyroxene skarn with local feldspathic layers, thin marble layers, and garnet zones	-----	201.5
White pyroxene granite gneiss grading into pyroxene syenite gneiss in upper few feet below skarn. Analyzed specimens CN–69–1 and CN–69–2 (table 18, Nos. 4 and 3)	22.5	-----
Feldspathic pyroxene skarn (60–70 percent pyroxene)	-----	3.5
Pink phacoidal granite gneiss becoming white and medium grained at contact with skarn above and below	9.1	-----
Pyroxene skarn with feldspathic layers	-----	4.9
Pink phacoidal granite gneiss; a thin white granite layer at contact with skarn above. Analyzed specimen CN–69–3 (table 18, No. 1) from central part of this zone	107.5	-----
Fine-grained biotitic feldspathic quartzite	-----	21.5
Pink phacoidal granite gneiss	52.5	-----
B. Sequence in diamond-drill hole 30, depth 175 to 610 feet:		
Pink phacoidal granite gneiss with sparse included layers of metasedimentary rocks	32.6	-----
Green pyroxene granite gneiss grading into white pyroxene syenite gneiss at base. Wollastonite present locally in syenite	24.9	-----
White pyroxene granite gneiss grading into white pyroxene syenite gneiss above	28.7	-----
White migmatite of thin syenite gneiss layers and pyroxenic wollastonite layers. Sphene throughout	-----	38
Transitional white granite gneiss	2.8	-----
Pink phacoidal hornblende granite gneiss	7.2	-----
Interlayered pink phacoidal hornblende granite gneiss and feldspathic quartzite	-----	106.8
Metadiabase	16	-----
Green pyroxene skarn, locally garnetiferous, locally feldspathic, with several layers of ore. A few thin granitic zones are present	-----	130
White pyroxene syenite and transition zone	2	-----
Pink phacoidal hornblende granite gneiss	46	-----

Table 18.—*Chemical analyses and norms of phacoidal hornblende granite gneiss and its local contact facies against marble, Stark complex*

	1	2	3	4
Chemical analyses (weight percent)				
SiO_2	66.03	65.80	72.30	64.72
Al_2O_3	10.70	15.47	13.42	15.62
Fe_2O_3	6.35	1.10	.37	.68
FeO	2.03	3.59	1.02	1.93
MgO	1.00	.62	.42	.57
CaO	2.26	2.64	1.55	3.07
Na_2O	3.89	3.75	2.51	2.11
K_2O	5.24	5.35	7.10	9.98
H_2O+	.75	.30	.27	.13
H_2O-	.18	.06	.04	.06
CO_2	.62	-----	.35	.30
TiO_2	.55	.71	.32	.45
P_2O_5	.12	.21	.07	.12
MnO	.07	.10	.03	.11
Total	99.79	[1] 99.83	99.77	99.85
Molecular ratio $K_2O:Na_2O$	.88	1.07	1.90	3.12
Norms				
Quartz	22.11	16.08	27.42	8.88
Orthoclase	30.58	31.42	42.26	58.94
Albite	26.20	31.70	20.96	17.82
Anorthite	-----	9.73	4.17	3.61
Acmite	6.01	-----	-----	-----
Hypersthene	-----	5.13	1.71	-----
Diopside	5.18	1.78	.81	7.61
Magnetite	5.10	1.62	.46	.93
Ilmenite	1.06	1.37	.61	.84
Hematite	.72	-----	-----	-----
Apatite	.27	.50	.17	.27
Calcite	1.40	-----	.81	.69

[1] Also contains 0.09 percent Cl and 0.11 percent F.

1. Pink phacoidal chloritic granite gneiss, diamond-drill hole 69, Clifton mine, depth 311–315 ft, specimen CN–69–3. The rock is granoblastic and originally a hornblende granite. The ferromagnesian mineral is now completely altered to chlorite and there are thin facings of chlorite and carbonate on fracture surfaces. The feldspar is about equally divided between microcline (slightly perthitic) and plagioclase. About one-fifth of the rock is quartz. There is a little hematite, zircon, and apatite, and about 1 percent sphene. Analyst, E. Chadbourne.
2. Pink phacoidal hornblende granite gneiss, 1.9 miles south-southeast of Degrasse. The gneiss is slightly more mafic than normal. The feldspars consist of a granoblastic aggregate of slightly perthitic orthoclase or microcline and oligoclase. The hornblende is a ferrohastingsite (table 15, No. 4). The mineral composition is given in table 16, No. 9. Analyst, E. Chadbourne.
3. White pyroxene-microcline granite gneiss, diamond-drill hole 69, Clifton mine, depth 255–257 ft, specimen CN–69–2. The structure is indistinctly gneissic, having an even grain except for the quartz, which is elongate. The rock has a granoblastic structure and is mottled with a few pyroxene grains. The feldspars consist of slightly perthitic microcline and plagioclase in a ratio of about 2:1. The pyroxene is bright emerald to olive green and forms several percent. About 1 percent sphene is present, and also a very little apatite, calcite, and pyrite. The calcite appears to be a primary part of the rock. Analyst, E. Chadbourne
4. White pyroxene-microcline syenite gneiss, diamond-drill hole 69, Clifton mine, depth 243.6–245.2 ft, specimen CN–69–1. A massive-appearing, medium-grained rock, but actually with granoblastic structure. The feldspar is very largely a slightly perthitic microcline, and there is only about 13 percent plagioclase. Quartz constitutes less than 10 percent of the rock, and about 5–7 percent pyroxene is present, pleochroic from deep emerald green to olive green. About 1 percent sphene and a very little calcite, apatite, and pyrite are present as accessory minerals. The calcite appears to be a primary mineral of the rock. Analyst, E. Chadbourne.

resulting from injection of a K_2O-rich syenitic magma along foliation planes of the wollastonite or pyroxene skarn.

Leonard suggests that volatiles given off by younger granite magma, which locally invaded the skarn and the contact zones between skarn and phacoidal granite gneiss, may have contributed to the development of the zones of pyroxene-microcline granite gneiss and pyroxene (or wollastonite)-microcline syenite gneiss.

RAPAKIVI TEXTURE

There are some granites in which large potassic feldspars form ovoids mantled by a peripheral zone of plagioclase. This is called rapakivi texture or fabric. This type of texture—except that the potassic feldspars are not necessarily ovoids— is common in the granite facies of the Diana and Stark complexes. In the shonkinite, pyroxene syenite, pyroxene-quartz syenite, and much of the hornblende-quartz syenite, the plagioclase forms the core of many phenocrysts and has overgrowths of microperthite. In the granite, by contrast, microperthite or microcline forms the core of many large feldspars and has overgrowths of oligoclase.

An almost massive facies of the hornblende granite of the Stark complex (fig. 10) forms a lens near the north border of the Stark quadrangle for more than a mile on each side of Cold Brook. The feldspars are 0.5 inch to more than 1 inch in diameter. They include both potassic feldspar and plagioclase, though the former is predominant. The large plagioclase crystals are commonly cloudy with sericitic alteration at the core and have a clear mantle that is locally poikilitic with quartz, thus having the character of myrmekite. The large potassic feldspar crystals have a little perthitic intergrowth of stringlets or strings of plagioclase, are Carlsbad twins, and may have a well-defined microcline structure, a shadowy extinction, or a uniform extinction like orthoclase. They are very varied in character; they may (a) be clear and uniform, (b) enclose vein-like aggregates resembling the groundmass (hornblende, magnetite, plagioclase, or quartz), (c) hold euhedral plagioclase crystals, (d) locally have a perthitic patch fabric in which the grains of plagioclase relict from replacement extinguish together, (e) have a border locally embayed and replaced by myrmekite, (f) be partially mantled by a zone of plagioclase, or (g) show a replacement relation to their own peripheral zone of plagioclase, in such a fashion that a thin skeletal zone of isolated grains having uniform extinction occurs parallel to the outer mantle. Magnetite grains of the groundmass have sphene coronas. Much of the rock appears to the eye to be undeformed, but examination with the microscope shows that the feldspars are slightly cracked or veined with granulated material, and the borders are locally granulated. Myrmekite is developed in the groundmass.

The rock 1.1 miles west-northwest of Cold Brook School (northern Stark quadrangle) is also equally undeformed and coarse, but here plagioclase forms the core of the feldspars, with overgrowths of microperthite. The mafic minerals of this rock comprise both augite and hornblende, the latter in part secondary after pyroxene.

Two hypotheses have recently been discussed to explain rapakivi texture. Poldervaart and von Backström (1949, p. 461–462) have explained rapakivi texture and associated myrmekite as follows:

> The formation of myrmekite in the present instance may be due to repeated recrystallisation of potash feldspar, caused by a series of pulsations in physical conditions, which marked the wane of the folding movements. Initially a potash feldspar rich in soda and lime crystallised in areas where the energy level was highest. During subsequent pulsations calcic plagioclase was exsolved and by recrystallisation of potash feldspar, collected in pools to form separate crystals * * *. During later pulsations of diminishing tenor more sodic plagioclose exsolved from the potash feldspar * * *. In part the sodic plagioclase was expulsed to form rims around potash feldspar, or to migrate towards other plagioclase crystals, forming borders there.

Gates (1953, p. 55–69) has proposed that

> Brecciation of many of the potash feldspar cores in rapakivi granite suggests that, late in the crystallization history of such granite, structural activity, normally accompanying igneous activity, deformed the nearly solid granite at the time of perthite formation to create fracture zones of low relative pressure. Into these irregular low-pressure zones will migrate the sodic material unmixed from the potash feldspars in adjacent zones replacing and mantling the feldspars there * * *. This explanation requires that the *unmantled* [3] feldspar areas served as a source of sodic material [for the mantled feldspar].

Terzaghi (1940) has suggested that rapakivi structure may arise from a shifting of the cotectic ratio for crystallization of the two different feldspars accompanying changes in the amount of volatile materials present.

The highly varied relations found in the Adirondack rapakivi rocks may be interpreted in terms of magmatic crystallization of plagioclase-rich potassic feldspar phenocrysts, followed by exsolution of the plagioclase and contemporaneous extended growth of certain of the potassic feldspar phenocrysts, with the plagioclase expelled to form either a mantle or new plagioclase crystals. Parts of the groundmass were included as such in the potassic feldspars, and in part the enclosed plagioclase was replaced, as well as expelled to add to the plagioclase mantle or to separate plagioclase grains. Pulsating physical and physicochemical changes may have cooperated, for the process would have gone on at late-stage magmatic temperatures and at the lower temperatures of deuteric recrystallization and modification.

TUPPER COMPLEX

The Tupper complex forms a narrow belt on the outer flank of the anorthosite mass on the Long Lake and St. Regis quadrangles (Buddington, 1939, p. 116–123) and extends as an anticlinal prong westward from east of Simon Pond (Long Lake quadrangle) nearly across the north half of the Tupper Lake quadrangle. The latter part of the complex will here be described as the Arab Mountain sheet. It is about 15 miles long and a little less than 5 miles wide.

On the Cranberry Lake and Oswegatchie quadrangles a belt of pyroxene syenite and quartz syenite gneiss, 0.5–1.5 miles wide, extends for about 13 miles southwest from Dead Creek Flow to north of Streeter Lake and 1 mile east of Aldrich Pond. This belt will be referred to as the Inlet sheet. The rocks of which it is composed are of such character that they might belong to either the Arab Mountain sheet or the Diana complex. More detailed work would be necessary to determine this.

ARAB MOUNTAIN SHEET

The Arab Mountain sheet is composed predominantly of a green pyroxene syenite gneiss, which has a medium-grained appearance, and which contains a varied quantity of small porphyroclasts of feldspar. A phacoidal structure is commonly present but locally is not as distinct as in the Diana and Stark complexes. The Arab Mountain sheet, like the Diana and Stark complexes, is thought to be part of a strongly deformed sheet that has been intruded by later granite. The syenite gneiss in general has a strongly foliated structure. In part the mafic minerals are recrystallized as porphyroblasts. Locally the foliation is wavy and plicated. Rarely relict primary crystals occur, consisting of a plagioclase core with microperthite overgrowth.

The easternmost part of the Tupper complex overlies and is intrusive into anorthosite on the Long Lake and St. Regis quadrangles. This portion of the complex is thought to be structurally the lowest part of the sheet. It is a distinctly mafic (15–40 percent of mafic minerals) syenite gneiss and grades upwards into a normal quartz-bearing pyroxene syenite gneiss. None of the mafic facies of the sheet is exposed in St. Lawrence County, where the predominant rock is a syenite gneiss carrying less than 10 percent of mafic minerals and up to 10 percent of quartz. The Tupper sheet is thought to have been much thinner than the Stark or Diana sheets. Consequently, though there is a wide range of composition between the most mafic and the most felsic facies, the diversification is not as extreme or on as large a scale as in the Stark and Diana complexes. The facies of the Tupper complex are predominantly pyroxenic throughout and uniformly carry hypersthene as one of the pyroxenes.

No development of red hornblende-quartz syenite

[3] Italics supplied by A. F. Buddington.

or hornblende granite gneiss, such as is so largely developed in the Diana and Stark complexes, has been found in the Tupper complex; and in theArab Mountain sheet, gneiss with more than 10 percent quartz has been found only as a local facies around the plunging anticlinal nose at the eastern end. The most felsic facies (about 16 percent quartz) of the Arab Mountain sheet has been found in the upper part of the synclinal structure at the southwest end of the village of Tupper Lake.

The least quartzose (about 3 percent) and most highly recrystallized facies of the syenite gneiss of Arab Mountain is found between Tupper and Pleasant Lakes. The rock to the east and west is somewhat richer in quartz (about 8–10 percent).

The green syenite gneiss along the main highway along the east side of Tupper Lake is largely a hornblende syenite gneiss. The hornblende is in places partly altered to chlorite and some sphene.

The syenite gneiss locally has included thin layers or rounded fragments of amphibolite. Included layers are well shown in the outcrop on the railroad 0.4 mile southwest of Childwold Station (Tupper Lake quadrangle) and in the hill 1 mile to the southwest. Rounded xenoliths of amphibolite from 1 inch to 1 foot in length may be found in many outcrops of the syenite gneiss along the west side of Tupper Lake.

In thin section, the normal syenite gneiss is seen to consist of a granoblastic mortar or groundmass having a grain averaging about 0.3 mm, in which there is a varied percentage of larger porphyroclastic grains commonly 1–3 mm in size, but locally up to 4 mm. The quartz is commonly in amoeboid shapes elongate in the plane of the foliation. Typical variations in composition and the average mineral composition of a number of specimens are given in table 19. The potassic feldspar is predominantly perthite, but some of that in the groundmass is clear, and some may show an indistinct microcline structure. Locally the syenite has large porphyroclasts of plagioclase, as southwest of

TABLE 19.—*Modes of facies of the Tupper complex and fayalite-ferrohedenbergite granite*

[Volume percent]

Rock type and locality	Specimens averaged	Quartz	Feldspars: Microperthite and potassic feldspars	Feldspars: Plagioclase	Ferrohypersthene	Ferroaugite	Fayalite	Hornblende	Magnetite and ilmenite	Biotite	Apatite	Zircon	Garnet
Inlet sheet and fayalite-ferrohedenbergite granite													
Hornblende syenite gneiss	4	9.5	63.7	19.7	--------	--------	--------	6.0	0.9	--------	0.2	0.2	--------
Hornblende-augite syenite gneiss	5	10.0	57.5	23.8	1.8	2.2	--------	3.3	.9	--------	.3	.2	--------
Transitional hypersthene-augite-hornblende-quartz syenite gneiss	6	20.8	50.7	21.7	1.7	1.9	--------	2.2	.6	--------	.2	.2	--------
Transitional hypersthene-augite-hornblende-quartz syenite gneiss	4	19.9	68.1	4.8	.8	3.0	--------	2.8	.4	--------	.1	.1	--------
Fayalite-ferrohedenbergite granite	7	20.8	63.9	5.0	--------	[1] 4.7	2.7	1.5	1.0	--------	.2	.2	--------
Arab Mountain sheet													
Hypersthene-quartz syenite gneiss, quarry southwest of Tupper Lake	1	16.3	46 1	24.2	[2] 5.8	1.4	--------	2.8	2.0	--------	0.5	0.3	0.5
North and northwest of Buck Mountain, Tupper Lake quadrangle	4	9.5	58.3	25.8	1.3	.9	--------	2.6	.9	0.2	.4	.1	--------
Just south of Moody, Tupper Lake quadrangle	1	8.4	49.8	33.6	1.5	2.4	--------	2.1	1.1	--------	.8	.3	--------
0.5 mile east of south end of Simon Pond, Long Lake quadrangle	1	8.1	44.9	35.0	[2] 3.3	5.2	--------	1.6	1.0	--------	.6	.3	--------
1.3 miles south of Childwold Station, Tupper Lake quadrangle	1	6.3	53.0	34.4	2.3	2.2	--------	.8	.6	--------	.4	Tr.	--------
Between Pleasant and Tupper Lakes, Tupper Lake quadrangle	5	3.3	45.7	41.7	1.9	3.6	--------	1.8	1.3	--------	.6	.1	--------
3.4 miles north of junction of road to Litchfield Park, Tupper Lake quadrangle	--------	9.7	44.2	40.4	--------	--------	--------	5.3	.2	--------	.2	Tr.	--------
North of Kildare Pond													
North of Willis Pond, Childwold quadrangle	1	17	75		3.0	3.0	--------	1.4	0.3	--------	0.2	0.2	Tr.
0.2 mile northwest of Willis Pond, Childwold quadrangle	1	6.3	77.0		.9	4.7	--------	7.0	2.9	--------	.5	.1	0.6
Outlying masses													
Sheet west of Heath Pond, Cranberry Lake quadrangle	2	20.6	45.6	24.2	3.6	1.6	--------	2.8	0.9	--------	0.3	0.3	--------
Southwest corner Childwold quadrangle	1	16.0	72.5	--------	2.1	3	--------	5.3	.8	--------	.2	Tr.	--------
Southeast corner Stark quadrangle	1	4.0	68.1	6.9	2.8	12.2	--------	--------	4.5	--------	1.5	Tr.	--------

[1] Ferrohedenbergite.
[2] Eulite.

Otter Pond (Cranberry Lake quadrangle). In many places there is a little myrmekitic intergrowth of plagioclase and quartz in the groundmass. Part of the ferroaugite is commonly altered to hornblende. The orthopyroxene is usually fresh, but locally it is replaced by a bright red alteration product. Rarely, all the pyroxene is completely altered to hornblende.

The clinopyroxene of the normal syenite gneiss is a ferroaugite (nY=1.713–1.716), that of the fayalite-bearing granite is a ferrohedenbergite (nY=1.745). The orthopyroxene of the normal syenite gneiss is a ferrohypersthene $En_{30-37}Fs_{70-63}$ (nY=1.744–1.752), whereas that in the most siliceous facies carrying orthopyroxene is a more ferrosilitic type (eulite). The pyroxenes of a pyroxene syenite gneiss 0.5 mile east of the head of Simon Pond (Long Lake quadrangle) were examined. The rock carries about 15 percent quartz. The clinopyroxene is a hedenbergite (nY=1.734) and the orthopyroxene is the variety eulite (nZ=1.763, En_{18} Fs_{82}). The composition of the pyroxene is based on the curves of Poldervaart (1950).

Locally, at several localities, labradorite xenocrysts have been found in the syenite gneiss. Balk (1931, p. 388) has described them from the northeast spur of Arab Mountain. They also occur sparingly in the syenite gneiss of Arab Mountain and along the east side of Pleasant Lake. Three-fourths mile southeast of Grass River Club there are a few angular fragments of anorthosite in quartz syenite gneiss.

INLET SHEET

The Inlet sheet is composed of an assemblage of rocks whose relations to each other are obscured by drift and are exceptionally difficult to interpret. A belt of uniform massive fayalite-ferrohedenbergite granite occurs along the northern part of the mass. The southern part of the complex, the Inlet sheet proper, is in general of syenitic composition with local quartz syenite facies.

The syenitic rocks are varied in composition, from ferroaugite-ferrohypersthene syenite gneiss through ferroaugite-ferrohypersthene-hornblende syenite gneiss and ferroaugite-hornblende syenite gneiss to hornblende syenite gneiss. The structure ranges from that of an almost massive and unrecrystallized rock to a gneissic facies in which the microperthite is largely recrystallized to grains of potassic feldspar and plagioclase. The composition of several types is given in table 19. The hornblende in much of these rocks is clearly secondary after the pyroxene, and it is possible that all is of this origin.

About 1 mile northwest of Streeter Lake a narrow sheet of phacoidal hornblende granite gneiss about 3 miles long is included within a younger alaskite gneiss. The phacoidal gneiss resembles that of the Diana complex, but is here thought to be a part of the Inlet sheet that has been isolated by the alaskite and whose pyroxene has all changed to hornblende.

The northern part of the Inlet sheet is an augite-hypersthene-quartz syenite gneiss which passes gradationally upwards into an augite-hypersthene syenite gneiss. Several metagabbro lenses that occur within the syenite gneisses appear to be dikes.

The syenite and quartz syenite gneisses in considerable part show a coarse phacoidal structure. Locally also, though rarely, relict porphyroclasts are found, consisting of plagioclase cores and microperthite borders.

The rock of the arc of quartz syenite gneiss north of Little River is thought to have been continuous originally with the Inlet sheet but is now separated from it by an intrusive wedge of the younger granite. The quartz syenite gneiss of this arc in general is granoblastic; grains average about 0.4 mm in diameter, but in some facies average only 1 mm. The potassic feldspar of the microperthite in some of the rock has the distinct grid structure of microcline.

Another small lens of quartz syenite gneiss included in younger granite occurs about 0.5 mile west of Nicks Pond. The quartz syenite here is reddened by action of the granite and is recognized by its granoblastic structure, which contrasts with that of the granite.

An outlying mass of pyroxene syenite gneiss is situated northeast of Sampson Pond (southeast corner of Stark and southwest corner of Childwold quadrangles). This body is thought to have been a separate lens originally intruded into metasedimentary rocks of the Grenville series, which were subsequently granitized. The mineral composition of two specimens of the syenite mass is given in table 19. The feldspar is almost wholly microperthite. This is consistent with what has been determined for known isolated syenite sheets in the metasedimentary rocks of the Grenville lowlands (Buddington, 1939, p. 77), as are all the other mineralogical characters. The rocks range in composition from a mafic syenite with 21 percent mafic minerals and only 4 percent quartz to a normal pyroxene-quartz syenite. The grain size averages slightly more than 1 mm. A small part of the potassic feldspar has a peculiar indistinct perthitelike extinction pattern, suggesting anorthoclase that is in process of unmixing into separate orthoclase and oligoclase components.

The western end of the Inlet sheet has been so much torn to pieces and dismembered by younger intrusive alaskite, and the structure north of the northwest border of the main mass is so complicated, that the structural nature of the body as a whole has not been determined. One interpretation would be that the pyroxene syenite and quartz syenite gneiss of the Inlet sheet is the south limb of an isoclinal anticline, of which the pyroxene-quartz syenite gneiss north of Little River is a relic of the overturned north limb, and the nose of the fold is along the west side of Dead Creek Flow, where it has been warped around so that it plunges south. The alaskite along the north border of the fayalite-ferrohedenbergite granite has an intense linear structure and it is possible that this border is a shear zone. Drift covers much of the critical areas and the structural interpretations are indeterminate.

PYROXENE SYENITE AND QUARTZ SYENITE GNEISS NORTH OF KILDARE, CHILDWOLD QUADRANGLE

Northeast of Kildare (Childwold quadrangle), in a belt around Willis Pond, pyroxene syenite gneiss overlain by pyroxene-quartz syenite gneiss crops out on the nose of an anticline. The syenite and quartz syenite gneiss overlie anorthosite gneiss, which forms the country just east of Willis Pond on the St. Regis quadrangle. The pyroxene gneisses around Willis Pond are overlain by younger granite. The pyroxene of the syenite and quartz syenite gneiss is partly altered to hornblende, presumably by solutions related to the granite. The mineral composition of specimens of the pyroxene syenite gneiss and the pyroxene-quartz syenite gneiss is given in table 19.

South and southwest of Rock Pond (Childwold quadrangle) there is an arc of medium-grained, green pyroxene-quartz syenite gneiss with local included lenses of amphibolite. This gneiss is varied from a variety rich in hornblende (more than 15 percent) to one containing only sparse mafic minerals. In addition to hornblende there is always some augite and a little hypersthene present. The percentage of quartz ranges from 15 to 25. The rock is not as much deformed as that of the mass near Willis Pond.

CHEMICAL COMPOSITION

The chemical composition of a representative series of rocks from the Tupper complex is given in table 20. Analysis 3 is of a facies similar to the predominant type of rock in the complex. No. 1 is of the most mafic facies and Nos. 4, 5, and 6 are of the most felsic facies.

TABLE 20.—*Chemical analyses, norms, and modes of facies of the Tupper complex, and of fayalite–ferrohedenbergite granites*

	1	2	3	4	5	6	7
Chemical analyses (weight percent)							
SiO_2	54.10	61.01	62 19	67.06	65.20	70.20	69.98
Al_2O_3	17.45	15.36	16.53	14.39	14.79	12.37	12.37
Fe_2O_3	4.52	2.98	1.73	1.03	1.44	.98	1.11
FeO	6.47	7.77	4.10	4.49	5.19	4.93	5.13
MgO	2.33	.78	1.01	.40	.30	.13	.08
CaO	6.17	4.05	3 20	2.30	2.49	1.64	1.25
Na_2O	3.81	3.68	4.22	3.37	3.61	3.24	3.99
K_2O	3.06	3.90	4.94	5.33	5.41	5.15	4.91
H_2O+	.48	.49	.17	.29	.33	.28	.16
H_2O-	.09		18	.09	.16	.12	.05
CO_2				.11	.03	.11	.07
TiO_2	.19		1.01	.68	.51	.67	.46
P_2O_5	.88		.36	.13	.15	.09	.04
S	.14				.02		
MnO	.35	.08	.12		.11	.11	.14
BaO	.10				.11		
F	.05						
Total	[1] 100.13	100.10	99.76	99.67	[2] 99.84	100.02	99.74
Norms							
Quartz	2.86	10.86	9.81	19.98	15.18	25.77	22.62
Orthoclase	18.13	22.80	29.11	31.14	32.25	30.30	28.91
Albite	32.17	31.44	35.63	28.32	30.39	27.51	34.06
Anorthite	21.43	13.62	11.54	8.62	8.06	4.17	1.39
Diopside	3.05	5.58	1.87	1.09	3.66	2.48	3.96
Hypersthene	12.50	10.84	6.04	6.81	6.42	6.30	5.89
Magnetite	6.57	4.41	2.55	1.39	2.20	1.39	1.62
Ilmenite			1.98	1.29	.99	1.29	.91
Apatite	2.09		.84	.34	.34	.20	.10
Calcite				.25		.25	
Modes (weight percent)							
Microperthite	56.12	58.07	51.6	46.1	42	61	63
Oligoclase		6.04	34.0	24.2	25	5.6	4.5
Quartz	2.87	9.47	6.4	16.3	14	20	21.5
Fayalite						3.7	3
Orthopyroxene	10.57	3.01	2.8	5.8	5		
Ferroaugite or ferrohedenbergite	11.54	6.83	2.6	1.4	6	7.2	6
Hornblende	9.65	6.01	.9	2.8	6		Tr.
Iron oxides	6.57	4.33	1.1	2.0	1.4	2.0	1
Apatite	1.37	1.01	.6	.5	.3	.2	.2
Zircon		.13	Tr.	.3	.3	.3	.4
Garnet	.77	4.50		.5			
Sphene	.40	.33					
Pyrite		.43					
Fluorite							.2

[1] Corrected for O=F+S.
[2] Corrected for O=S.

1. Mafic ferroaugite-ferrohypersthene syenite gneiss, a little more than 1 mile north-northwest of Raquette Falls, Long Lake quadrangle (Cushing, 1907, p. 512–514). Analyst, E. W. Morley.
2. Ferroaugite-ferrohypersthene syenite gneiss, 3.8 miles northeast of Tupper Lake Junction on New York Central and Hudson River railroads, St. Regis quadrangle (Cushing, 1907, p. 514–518). Analyst, E. W. Morley.
3. Ferroaugite-ferrohypersthene syenite gneiss, railroad cut 1.3 miles south of Childwold Station, Tupper Lake quadrangle. Analyst, Eileen K. Oslund. Orthopyroxene=$En_{35}Fs_{65}$ (nZ=1.744) for clinopyroxene, nY=1.716.
4. Eulite-quartz syenite gneiss, old quarry on Raquette Pond at southwest end of Tupper Lake village, Long Lake quadrangle (Buddington, 1939, table 32, No. 112). Analyst, R. W. Perlich.
5. Eulite-ferrohastingsite-ferrohedenbergite-quartz syenite gneiss, 0.4 mile north-northeast of McBride Pond, Long Lake quadrangle. Analyst, Eileen K. Oslund. Sp. Gr.=2.736.
6. Fayalite-ferrohedenbergite granite, road cut 1.5 miles west-northwest of Wanakena, Cranberry Lake quadrangle. Analyst, E. Chadbourne.
7. Fayalite-ferrohedenbergite granite, Au Sable Forks (Buddington, 1939, table 32, No. 113). Analyst, R. W. Perlich.

ORIGIN OF DIFFERENCES BETWEEN TUPPER, DIANA, AND STARK COMPLEXES

The rocks of the Tupper complex differ from those of the Diana and Stark complexes in several ways. The rocks of the Tupper sheet have in considerable part a primary medium-grained texture, in contrast to the very coarse texture of the Diana and Stark rocks. The Tupper rocks do not show as extreme differentiation as the Diana and Stark rocks. Both these features are consistent with the evidence that the Tupper sheet is generally much thinner than the Diana mass. The mafic syenitic facies in the lower part of the Tupper complex uniformly has abundant lenses or schlieren of amphibolite and locally of skarn. The felsic facies also locally has similar amphibolite and skarn schlieren, but not to such a degree and so consistently as the more mafic lower facies. Locally the upper felsic facies itself is more mafic where intimately involved with amphibolite. If members of the Tupper and the Diana and Stark sheets that have similar contents of quartz are compared, it is seen that each facies of the Tupper complex is much higher in mafic minerals than equivalent facies of the Diana and Stark complexes, also that orthopyroxene is a major mineral in all members of the former, and insignificant or absent in the latter. This content of orthopyroxene, plus other features, permits the members of the Tupper sheet to be called a charnockitic series.

Chemical analyses show that each member of the charnockitic series is different from the equivalent members of the Diana and Stark sheets in being markedly higher in FeO and slightly higher in TiO_2. An estimate of the average chemical and mineralogical composition of the two sheets, which is given in table 21, will serve to bring out the contrast. In addition to a higher FeO content, the charnockitic rock has a slightly higher percentage of CaO, which results in a slightly more calcic plagioclase and a little higher percentage of augite. The average composition is also very slightly lower in SiO_2. The fundamental difference, however, is in the higher percentage of FeO, which results in a much higher percentage of iron-rich pyroxene. Part of the pyroxene, in the absence of adequate CaO to yield ferroaugite, occurs as an iron-rich orthopyroxene (ferrohypersthene to eulite), and the monoclinic pyroxene is also high in ferrous iron (ferroaugite to hedenbergite). (See analyses 5 and 6, table 15.)

It is probable that the initial magma of the Tupper complex was already exceptionally rich in FeO relative to MgO when it arrived at its site of emplacement. Rock of composition similar to the average of the Tupper complex occurs in such stratiform sheets

TABLE 21.—*Comparison of average chemical and mineralogical composition of Diana and Tupper complexes*

Oxides	Chemical composition (weight percent)		Minerals	Modes (volume percent)	
	Diana [1]	Tupper [2]		Diana [1]	Tupper [3]
SiO_2	63	62.29	Quartz	10.0	9.2
Al_2O_3	16	14.59	Feldspar	76.5	70.5
Fe_2O_3	2.7	2.19	Hypersthene	.7	4.0
FeO	3.1	6.31	Augite	4.5	6.2
MgO	1.0	.92	Hornblende	3.5	3.8
CaO	3.3	3.66	Magnetite and ilmenite	3.2	3.4
Na_2O	4.5	3.56	Apatite	.7	1.0
K_2O	5.0	4.52	Zircon	.3	.3
TiO_2	.8	1.05	Garnet	-------	1.6
P_2O_5	.4	.39	Sphene	.2	-------
BaO	-------	.10	Chlorite and carbonate	.4	-------

[1] Quoted from Buddington (1939, p. 103).
[2] Average of 11 chemical analyses.
[3] Average of 64 Rosiwal analyses of random specimens.

TABLE 22.—*Comparison of chemical composition of average of Tupper complex with differentiates of stratiform sheets*

[Weight percent]

	1	2	3
SiO_2	62.29	62.33	62.18
Al_2O_3	14.59	13.31	13.24
Fe_2O_3	2.19	2.91	2.93
FeO	6.31	7.37	5.41
MgO	.92	.36	2.02
CaO	3.66	3.88	2.67
Na_2O	3.56	3.71	3.55
K_2O	4.52	3.49	3.98
TiO_2	1.05	.80	1.47
P_2O_5	.39	.16	.41
BaO	.10	n.d.	n.d.
MnO	-------	.12	.16
H_2O+	-------	1.33	1.34
H_2O-	-------	.30	.38

1. Average of 11 chemical analyses of rocks of Tupper complex.
2. Average of two chemical analyses of hornblende-quartz syenite from "upper zone" of Bushveld complex (Hall, 1932, p. 307, Nos. 3 and 4).
3. Average of 4 chemical analyses of "red rock" from diabase sills in Duluth area (Schwartz and Sandberg, 1940, table 1, Endion sill, Nos. 4 and 6; Northland sill, Nos. 11 and 12).

as the Bushveld complex (cf. Hall, 1932) and in many of the sills in the area of the Lake Superior geosyncline. Chemical analyses of representative examples are shown in table 22. The Bushveld and Lake Superior rocks are inferred to have formed from a late-stage magma derived by fractional crystallization of a gabbroic magma. It seeems probable that the magma of the Tupper complex could have been formed at depth by a similar mechanism of differentiation from an initial gabbroic magma.

The magma from which the rocks of the Diana complex were formed is inferred to have been richer in volatile materials than that from which the rocks of the Tupper complex were derived. Both magmas may be differentiates of gabbroic magma under somewhat different physiocochemical conditions. The formation and separation of hornblende instead of pyroxene, and of magnetite and quartz instead of ferrous iron silicate, together with slower cooling, is the type of thing which may lead to a magma of the character forming the Diana complex, as distinguished from that of the Tupper complex.

The most felsic facies of the Diana and Stark complexes is a ferrohastingsite granite gneiss in which the ferrohastingsite carries some volatile materials (H_2O+, 0.83–1.67; F, 0.61–0.95; and Cl, 0.53–0.77 percent). The most felsic facies of the Tupper complex is a granite gneiss with varied amounts of eulite, ferrohastingsite, and hedenbergite. This is consistent with a hypothesis that the differentiation of the Diana and Stark complexes took place in a magma richer in volatile constituents than the magma of the Tupper complex. Differentiation of a sheet of magma to yield felsic differentiates in the upper part and mafic facies below, plus some modification by assimilation of mafic country rock, is thought to be the explanation of the diversity of rocks found in the Tupper complex.

METADIABASE DIKES

The Diana complex on the Lake Bonaparte and Lowville quadrangles is cut by a substantial number of dikes or sheets of hypersthene diabase or its metamorphosed equivalent. The Diana complex on the Oswegatchie quadrangle is similarly cut by a few metadiabase dikes. The dikes are usually not more than 20 feet wide and generally strike N. 0–30° W.

A number of metadiabase dikes also occur in the gneiss of the Inlet quartz syenite sheet. The mineralogy of these dikes is different from those in the Diana complex, but they are nevertheless thought to be metamorphosed, reconstituted dikes belonging to the same group. Dikes possibly belonging to this group are present but rare in the Arab Mountain sheet of the Tupper complex and in the Stark complex.

The least reconstituted dike lies about 1 mile west of Kalurah. This dike strikes northwest across the northeast structure of the enclosing pyroxene-quartz syenite gneiss and has a dense chill zone at its contacts. It possesses a northeast foliation parallel to that of the country rock and in thin section is partly to wholly granulated. The rock consists of about equal parts of andesine and mafic minerals. The mafic minerals comprise hypersthene and augite with a little ilmenite and magnetite, and accessory biotite and apatite. It is similar to the hypersthene metadiabase dikes for which chemical analyses and description have been given previously (Buddington, 1939, p. 133–134).

A dike north of Wanakena likewise cuts across the foliation of pyroxene-quartz syenite gneiss. It has a granoblastic structure and the mafic minerals are almost completely reconstituted to hornblende and biotite. A similar sheet is found 1.65 miles northeast of the bridge at Wanakena, on the southeast side of Wanakena Inlet.

A dike 1 mile east of Rock Island Bay (Tupper Lake quadrangle) also crosscuts the foliation of quartz syenite gneiss but in addition sends off a sheet parallel to the foliation. The foliation of the dike is parallel to that of the country rock and hence across the dike. The rock has a granoblastic structure and is intensely metamorphosed and reconstituted. Disseminated euhedral garnets occur throughout the rock. A dike 0.75 mile west-southwest of Nicks Pond (Cranberry Lake quadrangle) is also reconstituted to a garnetiferous facies. It occurs in granoblastic quartz syenite gneiss facies of the Inlet sheet and strikes north-northwest across the foliation of the country rock. The dike is in turn cut by alaskite.

Hypersthene metadiabase dikes and sheets also occur at the Clifton mine, where they appear to cut across the metasedimentary layers. These dikes, too, are garnetiferous. (See Prof. Paper 377*.)

A chemical analysis of a specimen from a dike in the Inlet sheet 1.7 miles west of Wanakena, the average of analyses of two dikes from the Diana complex on the Lake Bonaparte quadrangle, and the mode of representative samples of several dikes are given in tables 23 and 24.

TABLE 23.—*Chemical analyses and norms of hypersthene metadiabase dikes*

	1	2
Chemical composition (weight percent)		
SiO_2	48.15	48.95
Al_2O_3	15.40	15.32
Fe_2O_3	2.42	3.22
FeO	12.73	10.48
MgO	5.34	5.63
CaO	7.98	8.28
Na_2O	2.81	2.81
K_2O	1.40	1.32
H_2O+	.49	.36
H_2O-	.16	.07
CO_2	—	.79
TiO_2	2.70	2.14
P_2O_5	.24	.28
S	—	.09
MnO	.22	.19
Total	100.04	99.93
Norms		
Orthoclase	8.34	7.78
Albite	23.58	23.58
Anorthite	25.30	25.30
Diopside	10.53	8.49
Hypersthene	13.51	23.36
Olivine	8.57	—
Magnetite	3.48	4.64
Ilmenite	5.17	4.10
Apatite	.57	.67
Calcite	—	1.80

1. Pyroxene-plagioclase granulite gneiss, dike 1.7 miles west of Wanakena, Cranberry Lake quadrangle. Analyst, J. J. Engel. (See table 24, No. 6, for mode.)
2. Average of 2 hypersthene metadiabase dikes, Lake Bonaparte quadrangle (Buddington, 1939; 1934).

TABLE 24.—*Modes of metadiabase dikes*

[Weight percent]

	1	2	3	4	5	6
Andesine	57.6	45.5	55.4	59.2	47.6	64.4
Hypersthene	23.4	5.2	7.2	7.8	3.5	11.0
Augite	12.0	3.5	19.0	15.0	25.4	11.1
Hornblende	—	37.0	—	7.8	3.1	—
Garnet	—	—	7.8	4.2	4.3	1.3
Biotite	2.0	5.4	7.8	4.2	8.2	9.7
Magnetite and ilmenite	3.2	3.3	2.7	1.8	5.3	2.1
Apatite	.4	.1	.1	—	.2	.4
Orthoclase	1.4	—	—	—	2.4	—

1. Hypersthene metadiabase dike, one-sixth mile northwest of Oswegatchie Corners, Lake Bonaparte quadrangle. Similar to dike west of Kalurah, Oswegatchie quadrangle.
2. Amphibolite dike, 1.4 miles north of Wanakena, Cranberry Lake quadrangle.
3. Pyroxene-garnet-plagioclase granulite dike, 1 mile east of Rock Island Bay, Tupper Lake quadrangle. Dike 0.7 mile east-southeast of Nicks Pond, Cranberry Lake quadrangle, is very similar.
4. Pyroxene-garnet-plagioclase granulite gneiss dike, 0.75 mile west-southwest of Nicks Pond, Cranberry Lake quadrangle.
5. Pyroxene-plagioclase-garnet granulite gneiss dike. Average of 3 specimens, Clifton mine, Russell quadrangle.
6. Pyroxene-plagioclase granulite gneiss dike, 1.7 miles west of Wanakena, Cranberry Lake quadrangle. (For chemical analysis, see table 23, No. 1.)

*See footnote on p. 31.

FAYALITE-FERROHEDENBERGITE GRANITE

A green fayalite-ferrohedenbergite granite forms a belt 0.5–1 mile wide along the north border of the Inlet sheet of gneiss. This granite is unique among the green pyroxenic rocks of St. Lawrence County because of its almost massive, undeformed, unrecrystallized character. A slight gneissoid structure is common in thin section but is most indistinct in the outcrop. The rock is medium grained and bright green. The feldspar grains range from 1 to 3 mm in length, averaging about 1.5 mm. The fayalite is partly altered to bright red iddingsite. The feldspar is almost wholly microperthite, there being only a little exsolution of albite to the grain borders. The rock is almost identical in chemical composition (table 20, No. 6) with that from near Au Sable Forks (table 20, No. 7). The pyroxene from the Au Sable Forks rock has been analyzed and determined to be a ferrohedenbergite (Hess, 1949, p. 654, No. 19).

Structural relations between the fayalite-ferrohedenbergite granite and adjoining rocks are wholly indeterminate because of covered zones. The structure within the granite itself is too indistinct to mean much. The granite is so massive that it must be younger that the orthogneisses of the Inlet sheet. There is one crosscutting red band in the green granite which could be inferred to be either a dike of the younger hornblende granite or a zone of alteration. The fayalite-ferrohedenbergite granite is referred to an age definitely younger than the rocks of the Diana and Tupper complexes and very doubtfully older than the younger hornblende granite and alaskite. The fayalite-ferrohedenbergite granite could be the youngest igneous rock in the region, with the exception of the basaltic dikes.

It is noteworthy that fayalite and orthopyroxene have never been found together in either the Inlet or the Au Sable Forks rocks. This is consistent with the experimental data presented by Bowen and Schairer (1932, p. 177–213) for the system $FeO–SiO_2$. They find that in this system at high temperatures ferrosilite does not exist, but decomposes on melting and is represented by fayalite and free silica (tridymite) as its equivalent mineral assemblage on crystallization. It has been further shown by Bowen, Schairer, and Posnjak (1933, p. 193–284) that in the system $CaSiO_3$–$FeSiO_3$ mixtures between 80 percent and 95 percent $FeSiO_3$ crystallize as hedenbergite solid solutions plus olivine and tridymite within a temperature range of a few hundred degrees below 980°C. The normative ratio of $CaSiO_3$ to $FeSiO_3$ of the fayalite-ferrohedenbergite rocks of the Adirondacks lies within the range where, on the basis of the experimental data, a hedenbergite-ferrosilite solid solution, olivine, and free silica should be expected; their actual mineralogical composition is consistent with this expectation. The hedenbergite-ferrosilite solid solution according to experimental data consists of 78 or more percent of $FeSiO_3$. Actually the ratio of $FeSiO_3$ to $CaSiO_3$ in the ferrohedenbergite solid solution of the Au Sable Forks rock is 60:40, a ratio at which a simple clinopyroxene solid solution alone might be expected. There is a substantial amount of other constituents than $FeSiO_3$ present in the ferrohedenbergite, however, which would affect the range of stability. The development of fayalite and quartz instead of a ferrous metasilicate is clearly correlated with the exceptionally low MgO content of the rock. In the presence of a slightly larger ratio of MgO to FeO (table 20, No. 4) the metasilicate eulite becomes stable in association with clinopyroxene, and fayalite disappears.

It may also be noted that, on the basis of experimental data, Bowen and Schairer (1932, p. 203) write—

> Quartz and fayalite occur together in rocks, a condition that has been assumed by some investigators to be due to a special action of volatile constituents which prevents their combination to form metasilicate, but the facts pointed out above show that any such assumption is unnecessary. Indeed, it would appear that the opposite is true. . . .

Unquestionably, a magma may exist that is equivalent in composition to the fayalite-ferrohedenbergite granite of the Adirondacks, for a rock very similar in chemical composition has been described by Bain (1934, p. 218) as forming a cone sheet of fayalite-quartz porphyry in the Kudaru Hills, Nigeria. It is assumed here that the Adirondack rock similarly formed from a magma.

If the interpretation is correct that the fayalite-ferrohedenbergite granite is younger than the period of deformation and recrystallization of the associated pyroxene-quartz syenite and syenite gneisses, its similarity in mineralogy and geographic association could be ascribed to a fractional remelting at depth of a part of the pyroxenic gneisses, and its intrusion at the very last waning stage of deformation and before the intrusion of the younger hornblende granite. An alternative hypothesis would be to consider it as a residual differentiate of a tholeiitic basaltic magma; but no bodies of massive rock equivalent to such magma have been found associated, and this hypothesis therefore seems a poorer one.

GRANITE AND GRANITE GNEISS SERIES

There are four major facies of reddish granitic rocks in this region, and three minor facies, equally distinct but small in volume.

One major facies consists of hornblende, plagioclase, microcline, and quartz. This facies has a phacoidal structure in large part, is usually completely or mostly recrystallized and deformed to constitute a gneiss, and is a felsic member of the quartz syenite series of the Diana and Stark complexes, in connection with which it has been described.

Another facies of the granitic rocks, the one that is the most widespread, is a hornblende granite or equivalent granite gneiss that is younger than the Diana and Stark complexes. This younger granite or granite gneiss is varied, ranging from a gneissoid hornblende-microperthite granite to a strongly deformed, wholly granoblastic gneiss in which the microperthitic feldspar is recrystallized to microcline and plagioclase, and there is locally the development of a little sphene as a new mineral. The recrystallized gneiss facies lies to the west and northwest of the Stark complex (Russell and Potsdam quadrangles) and in the northern part of the Oswegatchie quadrangle.

A third facies is an alaskitic variety. This occurs predominantly as sheets within the metasedimentary rocks, and as a local facies at or near the upper border of the younger hornblende granite and granite gneiss masses where they adjoin belts of metasedimentary rocks. Locally, but not commonly, alaskite occurs as layers or lenses within the areas of hornblende granite and granite gneiss. In part, there is a continuous gradation between the hornblende granite and alaskite. The mafic minerals of the alaskite are commonly magnetite, ilmenite, and biotite. Fluorite is a ubiquitous accessory mineral. In the South Russell syncline, biotite granite gneiss is associated with the alaskite gneiss and with metasedimentary rocks in such fashion that much of it, if not all, may be interpreted as a facies of alaskite gneiss contaminated by incorporation of material from the metasedimentary rocks. In the Tupper Lake quadrangle there is a belt of alaskite contaminated with almandite. In the gneissoid alaskite the feldspar is microperthite, but in the deformed, recrystallized, gneissic facies of the South Russell syncline it is potassic feldspar and plagioclase.

The fourth major granitic facies is a potassium-rich microcline granite gneiss in which microcline is overwhelmingly the predominant feldspar. Most of this facies is inhomogeneous and has a layered structure suggestive of metasedimentary rocks, or has an accessory mineral or mineral aggregates in the form of knots or discs which can be interpreted as more or less modified relics of metasedimentary rocks. Layers of veined metasedimentary rocks are often included in this granite gneiss. The gneiss commonly carries 1 to 6 percent of iron and (or) iron-titanium oxides, and the nature of the other accessory minerals varies in a systematic way according to the nature of the associated metasedimentary rock. The most uniform facies is commonly biotitic, but the accessory minerals or mineral aggregates may be sillimanite or sillimanite-quartz nodules, almandite, hornblende, pyroxene, or andradite. The gneiss is fine grained, usually with numerous conformable thin pegmatitic seams.

A fifth and subordinate facies is a potassium-rich variety of the hornblende-microperthite granite which has been found only in a belt through Windfall Pond between the Raquette River and Jordan Lake, on the Childwold quadrangle. It is transitional in character between the hornblende-microperthite granite and the hornblende-microcline granite gneiss.

A sixth facies occurs only very locally and as small sills. It is an albite-oligoclase granite, in which sodic plagioclase is the predominant feldspar and microcline is subordinate.

A seventh facies, the Hermon granite gneiss, is a coarsely porphyritic granite gneiss, usually biotitic, which is widespread in the Grenville lowlands but is of very local occurrence within the area of this report.

HORNBLENDE GRANITE AND HORNBLENDE GRANITE GNEISS

Hornblende granite and hornblende granite gneiss are parts of a series of granitic masses which are widespread throughout the Adirondacks. These masses, taken as a whole, constitute a batholithic complex with local included layers of older rocks. The hornblende granite and granite gneiss underlie about one-half the area of the main igneous complex in the northwest Adirondacks.

The microstructure and mineralogy of the rock west and north of the Stark anticline differs from that east and south of the anticline and elsewhere. This is attributed to greater intensity of metamorphism northwest of the anticline and is discussed in detail under the heading "Metamorphism of granite series."

The mineral composition of representative facies of the hornblende granite and hornblende granite gneiss is given in table 25.

The rock of the main mass across the southern tier of quadrangles is a medium-grained pink granite with a foliation usually sufficiently well developed that good observations of its strike and dip can be readily obtained. The least deformed facies of the granite occurs locally in this belt. On microscopic examination, such rock shows feldspars predominantly 1.5–4 mm in diameter, with an average of about 3 mm, interlocking in a manner characteristic of crystallization from magma. The quartz is predominantly in elongate amoeboid shapes. The feldspar and hornblende grains are in-

TABLE 25.—*Modes of facies of hornblende granites and hornblende granite gneisses*

[Volume percent]

Locality and rock type	Specimens averaged	Quartz	Potassic feldspars		Plagioclase	Hornblende	Biotite (and chlorite)	Magnetite and ilmenite	Apatite	Zircon	Fluorite	Sphene	Allanite
			Microperthite	Microcline and (or) orthoclase									
Gradation of hornblende granite into fluorite alaskite gneiss on Colton Hill anticline													
0.9 mile southeast of Fine	1	36.8	34.5		25.6	------	------	0.8	0.1	------	2.0	0.2	------
Colton Hill	1	39.5	36		21.3	0.8	------	1.0	------	0.1	.7	.4	------
Just south of Oswegatchie	1	24.6	45.5	5.6	18.2	2.3	------	1.3	.4	.2	.9	1.0	------
1 mile south of Colton Hill	1	26.3	35.6		29.4	5.8	0.4	1.1	.5	.2	.2	.5	------
Vroomans Ridge to Oswegatchie	5	29.7	47.7	5.0	12.7	3.9	.2	.5	.2	.1	------	------	------
Recrystallized hornblende granite gneiss													
West limb of South Russell syncline	11	22.0	------	35.7	32.4	5.7	1.1	1.4	0.5	0.2	------	1.1	------
West flank of Stark anticline	10	23.2	------	38.4	27.6	7.2	.2	1.6	.6	.1	------	1.1	------
Gneissoid hornblende granite													
Hornblende granite with allanite, 1 mile north of Newton Falls	2	28.1	46.4		18.5	4.0	------	1.0	0.2	------	------	0.1	1.7
Hornblende granite:													
East flank of Stark anticline	9	25.3	35.2	11.4	20.1	6.4	0.7	.4	.3	0.2	------	------	------
North of Jordan River, Childwold quadrangle	4	26.5	49.7		16.7	4.9	.1	1.8	.2	Tr.	------	------	------
Between Chaumont Swamp and main highway on east, Cranberry Lake quadrangle	6	23.3	46.3	5.5	16.5	6.9	.4	.6	.3	.2	------	------	------
(Less than 27 percent quartz) south of main highway, Cranberry Lake quadrangle	14	23.6	55.1	3.0	11.8	5.5	.5	.3	.1	.1	------	------	------
North of Slim and Charley Ponds, Tupper Lake quadrangle	5	25.1	55.5		11.4	6.9	.3	.3	.3	.2	------	------	------
Ferrohastingsite granite:													
Cranberry Lake quadrangle	7	29.6	49.3	3.0	11.0	6.4	------	.3	.2	.2	------	------	------
East of Windfall Pond, Childwold quadrangle	3	32.1	50.2		13.9	2.8	------	1.0	Tr.	Tr.	------	------	------

equidimensional and arranged with their longer diameters in the plane of the foliation. The feldspar is almost wholly microperthite. Small euhedral to rounded blebs of quartz are here and there included in the microperthite. A thin selvage of plagioclase often borders the contact between microperthite and quartz. A little myrmekite locally replaces the microperthite.

However, the bulk of the hornblende granitic rocks has a mortar of granules forming much of the rock. Commonly about one-fourth to one-third of the rock consists of a mortar of grains 0.2–0.4 mm in diameter, though locally the percent of mortar may be as low as 10 or as high as 50. The granite on the core of the anticlinal nose east of the Dead Creek syncline does not contain more than 20 percent of mortar. In a similar manner, the granite near the County Line across the Oswegatchie quadrangle locally shows but little groundmass mortar. The potassic feldspar in the mortar is in small part recrystallized and lacks any perthitic intergrowth, but it commonly also consists of microperthite.

The facies of the granitic rocks just described have been subjected to some deformation, subsequent to complete or almost complete crystallization. However, in view of the small extent of their recrystallization and, by contrast, of the complete recrystallization of the similar granitic rocks northwest of the Stark anticline, they are here referred to as gneissoid granite instead of as gneiss.

The hornblende granite east of the Stark anticline, on the Stark and Childwold quadrangles, is less metamorphosed than that west of the anticline and generally more deformed than that on the quadrangles to the south.

All the hornblende granite consists in major part of microperthite and quartz with subordinate plagioclase, orthoclase, and hornblende, and accessory biotite, ilmenomagnetite, ilmenite, apatite and euhedral zircon. The plagioclase is a sodic oligoclase ranging from $Ab_{92}An_8$ to $Ab_{82}An_{18}$.

There are, however, two subfacies of the hornblende granite.

The predominant type is a hornblende granite with less than 27 percent quartz, a femaghastingsite variety of hornblende, and a little accessory biotite. The femaghastingsite has the following properties: X= light yellow green, Y=dark brown or olive green, Z=very dark green; $2V_X$=about 55°–62°; $Z=c$=14°–19°; nX generally less than 1.685. The quartz content at extremes may range from 18 to 30 percent.

Another facies of the hornblende granite generally has 27 percent or more of quartz, a ferrohastingsite variety of hornblende, and virtually no biotite. The ferrohastingsite has the following properties: X=pale olive, Y=grayish olive, Z=green with a bluish tint, $2V_X$ about 45°–55°, $Z \wedge c$=10°–13°, nX generally greater than 1.688. Rarely the ferrohastingsite granite has only the amount of quartz normal to the femaghastingsite granite (see chemical analysis 1, table 30). The ferrohastingsite granite may occur as homogeneous material in substantial volume, but it has not been mapped separately from the normal femaghastingsite granite. The quartz content at extremes may range from 20 to 35 percent. Much of the plagioclase may have myrmekitic intergrowths of quartz. Masses of quartz-rich ferrohastingsite granite, in particular, were noted locally in the southern third of the Cranberry Lake quadrangle and east of Cranberry Lake.

One facies of the hornblende granite and granite gneiss carries both ilmenomagnetite and ilmenite in similar amounts or has ilmenomagnetite predominant, whereas another facies carries ilmenite as the predominant or exclusive iron-titanium oxide.

The granite in a belt through Windfall Pond between the Raquette River on the south and west and the Jordan River on the north is a K_2O-rich facies of the normal hornblende-microperthite granite in which the microperthite carries less plagioclase intergrowth. The hornblende has the optical properties of a ferrohastingsite. A chemical analysis (table 30, No. 13) of a representative specimen shows the rock to be transitional between the normal hornblende-microperthite granite and the hornblende-microcline granite gneisses. It carries ilmenomagnetite and ilmenite as accessory minerals like the normal microperthite granite, in contrast to the ilmenomagnetite, hemoilmenite, and ilmenohematite of the hornblende-microcline granite gneiss.

Locally the normal hornblende granite grades into fluorite alaskite. This has been observed on the plunging noses of several anticlines, as at the east end of Tomar Mountain (Cranberry Lake quadrangle), north of Colton Hill (Oswegatchie quadrangle), and northwest of Newton Falls. The gradation is well shown and has been studied on the Colton Hill structure. The hornblende granite between Briggs and Star Lake has slightly less hornblende and more quartz than the granite to the south. Northward on the core of the Colton Hill anticline, the granitic rocks are all deformed and recrystallized, with the usual unmixing of microperthite to yield potassic feldspar and plagioclase. Quite independent of this recrystallization, however, there is continuous gradation from hornblende granite on the south to fluorite alaskite on the north. Fluorite first appears in a hornblende granite with a normal amount of hornblende. In the transition zone, fluorite occurs throughout, quartz increases, and hornblende decreases until it is wholly lacking in the alaskite. Fluorite has not been found in any of the hornblende granite of the main granite masses.

The hornblende granite of the Spruce Mountain dome (Stark quadrangle) is an alaskitic facies of the hornblende granite. Hornblende is present but commonly does not form more than 1–3 percent of the rock. Schlieren of amphibolite are common, and the texture of the rock is varied.

The hornblende granite about a mile north of Newton Falls locally carries accessory allanite, some of which is altered to a zoned brownish material with colloform-like structure.

The hornblende granite gneiss in the Russell quadrangle, west of the Stark anticline and within the south Russell synclinorium, is primarily a fine-grained hornblende–oligoclase–potassic feldspar–quartz gneiss. The rock is similar to that of the gneissoid granite except that the feldspar, instead of being largely microperthite, consists of plagioclase and potassic feldspar; there is a higher percentage of accessory magnetite and apatite; and sphene is almost always present, whereas it is absent from the granite. The potassic feldspar is predominantly microcline, but in part it is slightly perthitic. The ratio of microcline to untwinned potassic feldspar varies from place to place. The feldspars often contain rounded blebs of quartz. The plagioclase commonly has a rim of untwinned albite around it. The hornblende is pleochroic from green to bluish green. Sphene occurs both as disseminated rounded grains and as aureoles or coronas around magnetite. Some of the apatite is in larger grains than is normal for the gneissoid granite. The grain of the feldspars averages 0.6–1.2 mm. The optical properties of the hornblende from the granite gneiss show it to be a ferrohastingsite, differing from that in the normal granite only in a higher percentage of ferric iron (7.75 percent Fe_2O_3 in the hornblende of the gneiss as against 5.5–5.8 percent in the hornblende of the granite, as determined by chemical analysis). Magnetite is the only iron oxide. Any ilmenite present in the primary granite has reacted to form sphene.

CONTAMINATED FACIES

The belt of hornblende granite passing through Doctors Pond, Sand Pond, and south of Mud Pond, in the southeast part of the Tupper Lake quadrangle, shows the local presence of such minerals as garnet, augite, or hypersthene, which are not normally present. These are interpreted as a product of contamination by material from included layers of amphibolite and diorite gneiss. The contaminated rocks are all char-

TABLE 26.—*Modes of contaminated facies of hornblende granite*

[Volume percent]

	1	2	3	4	5	6
Quartz	23.5	25.6	26.6	22.8	25.7	26.0
Microperthite	51.6	41.1	34.8	37.3	44.4	36.2
Plagioclase	17.0	18.8	23.5	30.8	16.2	28.8
Hornblende	7.1	7.0	10.0	3.1	9.8	2.6
Biotite	.3	2.9			.2	2.6
Magnetite and ilmenite	.3	1.7	1.5	3.2	2.7	2.1
Apatite	.1	.6	.3	.4	.6	.3
Zircon	.1	.1				.1
Hypersthene			3.3			
Augite				2.2		
Garnet		2.3			.4	
Sphene						1.2

1. Average of 2 specimens of normal hornblende granite, Sand Pond belt, Tupper Lake quadrangle.
2. Hornblende granite with additional biotite, magnetite, apatite, and garnet from contamination, Sand Pond belt.
3. Hornblende granite with hypersthene and additional hornblende, magnetite, and apatite from contamination, Sand Pond belt.
4. Hornblende granite with augite and additional plagioclase, magnetite, and apatite from contamination, Sand Pond belt.
5. Hornblende granite with garnet and additional apatite and magnetite from contamination. The potassic feldspar is microcline. 0.7 mile south of Catherineville School, Potsdam quadrangle.
6. Same locality as 5.

acterized by an increase in the amount of magnetite and apatite as shown in table 26.

The belt of rock south of the road from the South Bay of Tupper Lake to Horseshoe Lake consists for the most part of a thin-layered, streaked hornblende granite gneiss that has numerous layers and schlieren of amphibolite and so many pegmatitic veinings that the rock has a migmatitic aspect. There are some layers of normal hornblende granite.

Another belt of contaminated hornblende granite occurs north and south of the road between Catherineville School and School No. 7, on the Potsdam and Nicholville quadrangles. The granite of this belt has numerous layers and schlieren of amphibolite, and local ghost structures as a product of granitization of country rock. Sphene is locally common. Thin granite pegmatite seams that are parallel to the foliation are common. This hornblende granite gneiss is crushed and recrystallized, and the sphene in part is very probably of metamorphic origin.

LOCAL SYENITE FACIES

Locally, where in contact with amphibolite or pyroxene skarn, the hornblende granite passes into a syenite facies with hornblende and augite as the mafic minerals. The amount of the syenite is usually small and it only develops quite locally.

One such locality is west-southwest of Little Charley Pond (Tupper Lake quadrangle). A hornblende-augite syenite is developed here as a mixed facies of the hornblende granite in a contact zone with garnetiferous amphibolite. It may also be noted, however, that a layer of pyroxene skarn is included in the metagabbro of the hill just southwest of the pond, and hence skarn may also be involved in the development of the syenite facies here. The syenite has the following minerals in percent: microperthite 64, plagioclase 26, augite 3, hornblende 5, magnetite and ilmenite 2.5, and accessory apatite and zircon.

Further descriptions of the development of syenite facies of the hornblende granite in contact zones with pyroxene skarn will be found in a discussion of the mineral deposits (Prof. Paper 377*).

PORPHYRITIC GRANITE GNEISS (HERMON GRANITE GNEISS)

A coarsely porphyritic granite gneiss is one of the major members of the igneous rocks present among the metasedimentary rocks of the Grenville lowlands. However, only a little of this type of rock occurs within the area of this report. West of Porter Hill School (Russell quadrangle) a mass of porphyritic granite gneiss forms the northeast termination of a long belt that extends to the southwest across the Gouverneur quadrangle. Excellent outcrops of this rock are exposed about 1 mile west of Hermon. The rock is an augen gneiss containing conspicuous pink augen of Carlsbad-twinned microcline more than 0.5 inch in length, and yellowish to white plagioclase with leaves of quartz and with biotite wrapping around the augen. Biotite forms 5–7 percent of the gneiss, and quartz about 15 percent. There is considerable granular mortar of feldspar. A little secondary sphene is present. The granite mass north of White School (Oswegatchie quadrangle) may also be a facies of the porphyritic augen gneisses. This rock carries about 3.5 percent biotite, 1 percent hornblende, and 18 percent quartz.

ALASKITE AND ALASKITE GNEISS

NORMAL FACIES

Alaskite and alaskite gneiss may occur in homogeneous masses as a normal facies, or they may contain schlieren and develop a contaminated facies in association with amphibolite and metasedimentary rocks.

The alaskite and alaskite gneiss occur as a local roof facies of the hornblende granite and hornblende granite gneiss, and as sheets within belts of metasedimentary rocks and amphibolite. Their development as a roof facies locally beneath synclines of metasedimentary rocks is well shown around the southeast end of the South Russell synclinorium, near Star Lake and Newton Falls, around the syncline south of Jayville, in the Loon Pond syncline, and in the Lake Lila syncline. Alaskite or its gneissic equivalent is very well developed as a roof facies on the northeast nose of plunging anticlinal structures north of School No. 15 (Oswegatchie quadrangle) and at Tomar Mountain (Cranberry Lake quadrangle). At other localities at the roof of the

*See footnote on p. 31.

hornblende granite no alaskite has been observed, but there may be a local diminution in the amount of hornblende in the hornblende granite and an increase in the percent of quartz. Drift covers much of the border zone between the hornblende granite and synclines of metasedimentary rocks, where development of alaskite might be expected, as around the Bog River syncline. The occurrence of alaskite or of alaskite gneiss as sheets within the metasedimentary rocks is well shown in all their major belts.

By contrast with the occurrence of alaskite as a local roof facies of the hornblende granite masses, it has not been found at the base of the granite sheets as on the Arab Mountain and Stark anticlines. It is, therefore, asymmetrically developed as an upper border facies where it occurs in association with hornblende granite.

The alaskite and alaskite gneiss are so low in mafic minerals that in the field it is often difficult to find the foliation. The microstructure, however, is characteristically gneissoid or gneissic.

The alaskite gneiss at a number of localities weathers with a conspicuous platy or tabular jointing parallel to the foliation. Examples may be seen 1½ miles south of Benson Mines, 1 mile east-southeast of Newton Falls, and northeast of Lower District School (Russell quadrangle). In such cases the rock is a strongly foliated gneissic facies.

The alaskite in the Lake Lila syncline contains numerous schlieren of material similar to amphibolite, presumably derived from included layers of diorite gneiss. There are also local schlieren of biotitic and garnetiferous metasedimentary rocks and microcline-rich gneiss.

The alaskite, like the hornblende granite, has both a primary and a metamorphic facies. The least deformed, though gneissoid, alaskite, occurs in the Dead Creek, Darning Needle, and Loon Pond synclines (Cranberry Lake and Tupper Lake quadrangles). Alaskite gneiss with maximum deformation and recrystallization occurs in the South Russell syncline, and in general west and northwest of the Stark anticline.

The average mineral composition of several facies of the alaskite and alaskite gneiss is given in table 27. It is characteristically high in quartz and low in dark minerals. The latter consist entirely of biotite, in part chloritized in the gneissic facies, and iron oxides. A noteworthy aspect of the uncontaminated alaskite and alaskite gneiss is the prevalence of a small amount of fluorite. Over one-half of the thin sections of normal alaskite and alaskite gneiss that were examined showed this mineral; hence it must be fairly homogeneously distributed throughout the rock. The fluorite occurs as intergranular material coordinate with the other minerals. The mode of occurrence and widespread homogeneous distribution indicate that it is a primary mineral. In general, the alaskite gneiss has a fine-grained appearance, as though the grains averaged around 1 mm. In detail there are larger grains, commonly 1–2.5 mm, in a finer groundmass. All the alaskitic rocks have been deformed and somewhat crushed. The primary feldspar was almost exclusively microperthite, as it still is in the gneissoid alaskite of the Stark, Cranberry Lake, and Tupper Lake quadrangles. The feldspar of the alaskite gneiss, however, as northwest of Oswegatchie, in the Russell syncline, and northwest of the Stark anticline is partly or wholly recrystallized to potassic feldspar and plagioclase. There is often a small development of sphene in such gneiss, but there is no ilmenite.

The occurrence of aplitic, alaskitic, or more felsic facies, locally at or near the roof of a batholith or thick sheet of granite and as subordinate intrusions in the country rock, has been described at a number of localities in the world (Daly, 1914, p. 368–370; White, 1940, p. 967–994; Strauss and Truter, 1945, p. 47–78).

In all the cases cited there is independent evidence—such as the presence of miarolitic structure or secondary cavities, local pegmatitic developments, or the occurrence of minerals such as fluorite or tourmaline—to indicate that the development of the felsic differentiate is correlated with an exceptional concentration of volatile and hyperfusible materials. The prevalence of fluorite as a primary accessory constituent in the alaskite of the Adirondack area similarly indicates such concentration. The alaskite is interpreted as a light segregate with a relatively high concentration of hyperfusible material accumulated locally beneath the roof of the main granite mass. The alaskite, being thus more fluid and accumulated at or near the roof of the main granite mass, was in a favorable location and of an appropriate character to be injected into the metasedimentary rock, where it is the most abundant intrusive in the form of thin to moderate-sized sheets.

In the metasedimentary rocks of the broad Grenville lowlands belt, there are 14 phacoliths of granite (Buddington, 1929, p. 51–52) from 2 to more than 15 miles in length. All of these are alaskitic in character and all but one are biotite alaskite. All have numerous tourmaline pegmatite veins associated with them, and all have produced a profound and complex series of metasomatic replacements and modifications in their adjoining country rock, indicating that they were highly charged with volatiles.

CONTAMINATED FACIES

The alaskite and alaskite gneiss locally contain included layers and schlieren of metasedimentary rocks and are contaminated by minerals derived from the included material. The mineral composition of several facies of such contaminated rock is given in table 27. In all cases it will be noted that the contaminated facies carry less quartz and fluorite, but in addition to the major minerals developed by contamination they have an increase in apatite and magnetite. The alaskite, with hornblende or pyroxene as a product of contamination, in particular shows an increase in the content of apatite as compared with the normal facies. Allanite occurs only in the contaminated facies of the alaskite.

The minerals of the contaminated facies are directly related to those of the country rock, biotite and garnet being derived from biotite gneiss, and hornblende from amphibolite or pyroxenic layers.

Biotite granite gneiss

In the South Russell synclinorium alaskite gneiss is associated with biotite granite gneiss and grades into it by increase of biotite. Some of the biotite granite gneiss is in part uniform and homogeneous, but most of it is so associated with schlieren of metasedimentary rocks as to indicate that it is a contaminated facies of the normal alaskite gneiss. Locally also, though rarely, the alaskite may be contaminated with sillimanite and garnet in addition to biotite. The alaskite gneiss and biotite granite gneiss normally carry only biotite as the ferromagnesian mineral. Locally, however, where the granite is intrusive into skarn, pyroxene gneiss, or amphibolite, it may be contaminated by incorporation of the ferromagnesian minerals of the country rock, either in their original form or modified to other minerals.

Much of the alaskite sheet passing through Beaver Meadow (Russell quadrangle) is composed of biotite granite gneiss. Locally, near the east border, south of Green Valley School, it contains included layers and schlieren of metasedimentary rocks. The granite gneiss in this belt is hornblendic and locally pyroxenic, presumably as a result of contamination by the metasedimentary rocks. Local skarn lenses and biotite-plagioclase gneiss layers are present.

A parting layer of granite gneiss separates the two belts of sillimanite-microcline granite gneiss south of Russell. This parting layer is largely biotite granite gneiss, which contains up to 5 percent biotite, depending upon the degree of contamination from included layers of metasedimentary biotite-quartz-plagioclase gneiss.

The granite gneisses in the belt northeast of Boyd Pond range in composition from biotite granite gneiss to alaskite gneiss.

The belt of granite gneiss that passes through Palmer Hill and extends for several miles southeast of Austins Corners is a biotite granite gneiss, mostly fine grained, with layers or schlieren of metasedimentary rocks here and there throughout its extent. There are also three well-defined layers of sillimanite-microcline granite gneiss and local layers of muscovite-microcline granite gneiss. Much of the biotite granite gneiss has pegmatite seams at intervals of a few inches or a few feet. Quartz seams and nodular quartz lenses are locally common. On Palmer Hill there are numerous quartz-schorl seams, and northeast of Clear Lake schorl is common.

The rock forming the hills west of the Trembley Mountain magnetite prospect is prevailingly an alaskite gneiss, locally contaminated with biotite or garnet and locally associated with sillimanite-microcline granite gneiss. The typical rock is mostly recrystallized with porphyroclasts of microperthite (up to 1 mm) in a granoblastic groundmass (0.3 mm) of microcline, plagioclase, and quartz. Biotite is varied in amount.

Another belt of alaskite gneiss and associated biotite granite gneiss (contaminated alaskite) extends for more than 25 miles southwest from a point just west of Hopkinton Pond (Nicholville quadrangle) across the Potsdam quadrangle to Boyd Pond (Russell quadrangle). The biotite granite gneiss has in considerable part a confused, messy schlieren or ghost structure of disintegrated and modified metasedimentary rocks. Much of the original country rock, as indicated by local distinguishable relics, must have been a biotite-quartz-feldspar gneiss. The part of the belt that is on the Nicholville quadrangle carries a little hornblende in addition to biotite. The contaminated rock is commonly thin seamed, with coarser pegmatitic layers. Lenses of metagabbro in the country rock have been inherited as such except for the development of a boudinage structure.

The average mineral composition of a number of specimens of the biotite granite gneiss is given in table 27. A comparison with the alaskite gneiss shows that the biotite granite gneiss has a slightly higher ratio of plagioclase to microcline; less quartz; more biotite, magnetite, and apatite; and only rarely any fluorite. All of these features are in accord with the kinds of phenomena that are commonly known to occur in zones of contamination between granite and aluminous gneiss such as biotite-quartz-plagioclase gneiss.

Hornblende granite gneiss

The alaskite gneiss, in addition to developing a contaminated biotitic facies where it is intrusive into bioti-

TABLE 27.—*Modes of alaskite and alaskite gneiss, and of their contaminated facies*

[Volume percent]

Rock type and locality	Number of specimens averaged	Quartz	Microperthite (minor microcline or orthoclase)	Microcline, orthoclase, or both minerals	Plagioclase	Hornblende	Biotite (and chlorite)	Magnetite and ilmenite	Apatite	Zircon	Fluorite	Sphene	Garnet
Alaskite and alaskite gneiss—normal:													
Alaskite:													
East of Stark anticline	10	35.5	50.2	--------	12.0	--------	1.2	0.6	Tr.	0.1	0.4	--------	--------
West of Cook Pond	2	40.5	50.2	--------	7.8	--------	1.4	.1	--------	Tr.	--------	--------	--------
Quartz-rich facies, Brandy Brook	3	41.1	45.3	--------	11.3	--------	.9	1.0	--------	Tr.	.3	--------	--------
Normal facies, Brandy Brook	7	33.5	52.1	--------	12.9	--------	.7	.4	--------	Tr.	.3	--------	--------
Dead Creek and Darning Needle synclines	6	33.5	56.5	1.5	5.4	--------	1.6	1.0	Tr.	.15	.4	--------	--------
Southwest of Lake Lila, Big Moose quadrangle	2	39.1	45.1	--------	12.9	--------	.2	2.3	--------	--------	.4	--------	--------
Alaskite gneiss:													
Hawk Ledge, Potsdam quadrangle	2	38.7	--------	29.3	28.0	--------	2.2	.2	--------	.1	.2	--------	--------
South Russell syncline	10	32.2	--------	38.1	27.2	--------	1.0	.7	Tr.	.1	.6	0.1	--------
0.8 mile northeast of Sellecks Corners	1	36.6	--------	33.1	27.5	--------	1.5	.5	--------	--------	.5	--------	(1)
Alaskite and alaskite gneiss—normal (n) and contaminated (c), arranged geographically:													
Brandy Brook:													
Alaskite (n)	7	33.5	52.1	--------	12.9	--------	.7	.4	--------	Tr.	.3	--------	--------
Hornblende granite (c)	10	19	45	--------	21	8.0	3	3	0.5	Tr.	--------	(2)	--------
Horseshoe Lake area, Tupper Lake quadrangle:													
Alaskite (n), southeast of lake	2	29.2	59.6	--------	8.3	--------	1.0	.7	--------	--------	1.2	--------	--------
Garnet alaskite (c), between Sperry Pond and lake	3	36.3	[3] 53.6	--------	6.0	--------	2.1	.4	--------	--------	.8	--------	0.8
Garnet alaskite (c), south of lake	3	23.8	[3] 62.6	--------	9.5	--------	.3	1.7	--------	(4)	--------	--------	1.8
Hornblende granite (c), 0.4 mile south of lake	1	25.2	[3] 48.8	--------	17.0	7.0	--------	1.7	.2	.1	--------	--------	--------
South Russell syncline:													
Alaskite gneiss (n)	10	32.2	--------	38.1	27.2	--------	1.0	.7	Tr.	.1	.6	.1	--------
Biotite granite gneiss (c)	10	26.4	--------	35.2	32.6	--------	2.8	2.2	.4	.1	Tr.	.3	--------
Hornblende granite gneiss (c)	4	21.3	--------	31.5	36.1	5.3	1.5	1.9	.7	.2	--------	1.4	--------
Nicholville quadrangle:													
Hornblende-biotite granite gneiss (c)	2	25.1	--------	35.7	35.4	1.7	.8	1.0	.4	.1	--------	.8	--------

[1] Contains 0.3 percent tourmaline.
[2] Contains a little pyroxene, sphene, pyrite, and leucoxene.
[3] Contains only a little perthite.
[4] Contains 0.2 percent allanite.

tic metasedimentary gneisses, gives rise to a contaminated hornblendic facies where it is intrusive into pyroxenic or hornblendic gneisses. The average mineral composition of several examples of contaminated hornblende granite gneiss from the South Russell synclinorium is given in table 27. In comparison with the normal alaskite, the contaminated facies shows a higher ratio of plagioclase to microcline; a diminution of quartz; an increase in the percentage of biotite, magnetite and apatite; the disappearance of fluorite; an increase in sphene; and the addition of hornblende.

The relationships of hornblende granite as a contaminated facies of alaskite have been studied in considerable detail in the diamond drill cores of the Brandy Brook magnetite deposit, and a description follows.

Contaminated alaskite gneiss is found in three main zones at the Brandy Brook deposit: at the base of the pink alaskite and above the upper skarn zone; locally, between the upper and lower skarn zones; and in the quartz-feldspar gneisses below the lower skarn zone. The thickness of the individual sheets of contaminated granite ranges from 3 to 30 feet; six that were studied in detail average 15 feet. Rocks effectively in contact with the contaminated granite insofar as observed are, in general, varieties of pyroxenic skarn.

In one place only is there hornblende-bearing rock (hornblende-scapolite skarn) within 3 feet of the contaminated granite. One sheet, whose footwall was not drilled, has several thin layers and schlieren of amphibolite. Elsewhere, the mafic component in the wall rocks of these alaskite sheets is pyroxene or, less commonly, biotite. This restriction of hornblende-biotite granites to pyroxenic zones will at first seem anomalous.

The contaminated granites are pink, gray, or gray-green, depending roughly on the degree of contamination. At one extreme, they approach the pink alaskite already described; at the other, they grade into highly granitic migmatites. The texture is fine to medium. A strong foliation, which is characteristic of the contaminated alaskite gneiss, is afforded by alternating mafic and felsic layers in which hornblende and biotite have subparallel alinement. The foliation may be regular or irregular in development. Mafic layers are commonly about 2–10 mm thick. Even the most regular foliation shows many minor irregularities when closely examined—the dark layers are unevenly spaced and of varied thickness, the ratio of light to dark minerals in the predominantly dark layers is changeable, and the outline of the dark areas passes from smooth to ragged. In the more homogeneous rock, the dark layers are often fuzzy, blending with the lighter material. Some specimens have a speckled appearance.

The contaminated alaskites are inequigranular. The texture shows much greater variation within a given thin section and among different sections than does the texture of the normal alaskite. The larger grains are 1.5–3.1 mm in diameter; the smaller, 0.4–0.9 mm. Fine-grained material predominates. Its distribution in rude layers may be homogeneous or heterogeneous. The dark minerals usually have a subparallel orientation. Though they may be almost uniformly scattered through the lighter groundmass, they commonly form irregular layers in which the plagioclase is sometimes concentrated. Because of the inhomogeneous distribution of mafic minerals and the difficulty of assessing the relative importance of mechanical incorporation versus reciprocal migration (see origin, below), it is impossible to estimate the degree of deformation. At least some of the contaminated granite looks granoblastic.

Potassic feldspar forms almost one-half of the average contaminated granite. Quartz and plagioclase are always present. Hornblende and biotite are the common mafic minerals; pyroxene is rarely observed. Hornbende ranges from 1 to 13 percent; biotite, from a trace to 10 percent; and magnetite, from 1 to 6 percent. Magnetite, apatite, and zircon are ubiquitous accessories. Others of minor importance are sulfides, leucoxene (?), sphene, scapolite, and garnet. Secondary chlorite, calcite, muscovite, and kaolin are found in some slides.

Most of the potassic feldspar is a slightly microperthitic orthoclase or microcline. A little nonperthitic feldspar—invariably microcline—is present in much of the contaminated alaskite. The amount of plagioclase in microperthitic intergrowths is highly varied. In the least contaminated alaskites it is the same as in the normal alaskite. In the highly contaminated facies, however, the plagioclase forms relatively few stringlets, visible only at high magnification. In one contaminated alaskite that is virtually a migmatite, nonperthitic microcline and orthoclase are present, though other migmatitic rocks of similar composition have slightly perthitic feldspars.

The plagioclase is mostly fresh. Occasionally, a few grains exhibit a myrmekitic intergrowth with quartz. In composition, the plagioclase ranges from $Ab_{90}An_{10}$ to $Ab_{68}An_{32}$; the average for 9 specimens is $Ab_{80}An_{20}$. Some generalizations, to which one specimen is an exception, may be drawn: (1) The more mafic rocks contain oligoclase or sodic andesine; the less mafic, sodic oligoclase. (2) The anorthite content increases with the mafic content. (3) Rocks high in hornblende (as opposed to biotite) have a more calcic plagioclase than those low in hornblende.

The kind of hornblende is not constant. In most of the contaminated granites, it is a common green variety with X=yellow or olive-yellow, Y=olive or olive-green, Z=dark green, $Z \wedge c=12°-16°$. Hornblendes with higher extinction angles and slightly bluish colors for Z were observed in two sections. In some instances, brown biotite has formed at the expense of hornblende; in others, the amphibole has yielded a little secondary chlorite.

The biotite of the groundmass is also brown; rarely, it contains inclusions of zircon and embaying sphene grains. A little biotite sometimes contains wormy intergrowths of clear quartz or albite. Chlorite is the common alteration product.

Pyroxene in significant quantity is present in one specimen. It is light-green and nonpleochroic, with $2V_z$ large; $r>v$, weak; $Z \wedge c$ near 35°. Grains are elongate subhedra, some showing slight alteration to green hornblende ($Z \wedge c=14°$).

Magnetite, the most abundant accessory, occurs as small anhedral to subhedral grains. In places it is elongate. Commonly it shows slight replacement of biotite, hornblende, or microperthite. Where especially abundant, the magnetite may have an amoeboid shape and include small apatites. Rarely, it has partial rims of sphene or biotite.

Apatite and zircon form small, round but locally subhedral or euhedral grains scattered through the groundmass. Zircon inclusions in biotite have been observed.

Sulfides generally fill minute cracks in the silicates, though rarely they replace either light or dark minerals.

A comparison between contaminated alaskite and migmatite shows that the migmatites contain sphene, the microperthite is slightly higher in the migmatites, and plagioclase is generally low or absent. Magnetite, hornblende, and apatite are somewhat less abundant in the migmatites. The ratio of perthitic to nonperthitic potassic feldspar shows no systematic variation. The distinction between contaminated granite and migmatites of this type is arbitrary.

A comparison of the alaskitic gneiss with its contaminated facies shows that the quartz content of the contaminated granite is greatly reduced; microperthite has decreased slightly; plagioclase content has almost doubled, and the composition is generally more calcic; biotite has increased greatly; hornblende, absent in the alaskite granite, is present to the extent of 8 percent in the contaminated granite; magnetite has increased greatly. Apatite appears (0.5 percent). Fluorite disappears and sphene, pyroxene, sulfides, and leucoxene(?) appear in some places. The increase in magnetite may be due to the position of the contaminated granite closer to the ore zone.

Many of the features that distinguish the contaminated alaskite gneiss from the normal alaskite gneiss also differentiate it from the hornblende granite of the main batholith. In addition, it is important to note that the hornblendes present in the two granites are different. Although the pleochroism is in similar colors, the hornblende from the "core granite" has greater absorption and exhibits grayish tints that are lacking in the Brandy Brook hornblende. Moreover, the hornblende from the "core granite" has a slightly smaller optic angle; dispersion $r>v$, weak; and $Z \wedge c=$ 10°–12°.

The hornblende-biotite granite that occurs as thin sheets within the metasedimentary rocks and forms a contact zone between alaskite gneiss and metasedimentary rocks has been derived from the alaskite gneiss by a normal process of contamination. S. R. Nockolds (1935, p. 289–315) has discussed these processes, illustrating them by a study of the rocks at Carlingford, Eire. The rocks with which he dealt show striking similarities to those found at the Brandy Brook deposit. His petrographic data, supported by chemical analyses, make it possible to draw from our own data certain conclusions that might otherwise seem unwarranted.

At Carlingford, a microperthite granite of varied texture and alaskitic composition invaded somewhat modified basic hybrids. By exocontamination (mechanical incorporation) and endocontamination ("reciprocal reaction"), xenoliths of the basic hybrids were partly reconstituted; and both heterogeneous and homogeneous hornblende-bearing granites were developed. As the result of exocontamination, the "normal granite" first acquired clots of mafic minerals (largely pyroxene) and calcic plagioclase from the basic rocks. Gradually the granite magma reacted with the alien material, transforming some pyroxene to hornblende and biotite, modifying the composition of mafic minerals, and reducing the quantity of plagioclase in microperthite. As the changes proceeded, the incorporated clots were further disintegrated and the material distributed as xenocrysts. "Reciprocal reaction" was facilitated thereby. More and more pyroxene was converted to hornblende, plagioclase was made over into a more sodic variety, and the quantity of potassic feldspar was reduced somewhat, the feldspar at the same time resuming a more perthitic habit. Considered chemically, the xenoliths acquired Al_2O_3 and alkalis, while CaO and (Mg,Fe)O were given to the granite magma by reciprocal migration.

The story of contamination at Brandy Brook is much the same. The initial alaskite is virtually the same, and the "normal contaminated alaskite," the end product, is similar. Our attention must first be directed to the country rock into which the alaskite came.

The scarcity of hornblendic rocks near the sheets of contaminated alaskite was emphasized when the distribution of the contaminates was set down. The following interpretations of the field relations and petrographic data are possible:

1. The "contaminated granites" are really metasedimentary rocks, slightly modified.
2. The "contaminated granites" are a facies of the hornblende-microperthite granite of the Adirondack core.
3. The "contaminated granites" represent alaskite injected into amphibolites and amphibole skarns that have been almost completely digested, leaving traces at one place only.
4. The "contaminated granites" developed from alaskite that, by incorporation and reaction, developed hornblende and biotite from predominantly pyroxenic rocks.

Against the first two hypotheses are the megascopic appearance, microstructure, and mineral composition. Against the second hypothesis is the absence of hornblende-microperthite granite from the whole Brandy Brook belt. Only one argument can be advanced against the third: the absence of unquestioned relics of amphibolite within the contaminated granite. Perhaps the contaminated granite far in the footwall of drill hole B8 did incorporate amphibolite at depth, but this cannot be checked from the cores that exist. In favor of the fourth hypothesis are the observed field relations, and this seems the most probable.

North of the Bog River south-southeast of Horseshoe Lake (Tupper Lake quadrangle), there is a belt of alaskite in large part associated with schlieren of amphibolite and contaminated with garnet. The normal alaskite carries fluorite and a strongly perthitic feldspar. In the contaminated facies south and southeast of Horseshoe Lake the potassic feldspar is much less perthitic than in the normal facies, and the rock is of a type intermediate between the contaminated microcline granite and alaskite.

MICROCLINE GRANITE GNEISS

DISTRIBUTION AND FACIES

The microcline-bearing granitic rocks, generally called microcline granite gneiss in this report, present a problem in terminology and classification. A distinction can be made between the secondary gneissic and the primary gneissoid facies of the hornblende-microperthite granite and alaskite by means of the difference in microtexture and degree of recrystallization. The microcline-bearing granitic rocks, however, have similar texture and range of mineralogy throughout the region. That a major part of the rocks have been subjected to intense deformation and recrystallization and are therefore gneisses can be proved. The question arises, however, whether all have been similarly metamorphosed, or whether, as in the case of the hornblende-microperthite granite and alaskite, there is a regional variation in the degree of metamorphism, so that part of the microcline-bearing granitic rock has a primary structure. It will be shown later that a large part of the microcline-bearing granitic rock is probably the product of replacement and transformation of metasedimentary rocks, or of migmatitic injection of metasedimentary rocks, resulting in inheritance of a foliated structure. Such rocks, involving as they do substantial changes in composition of the country rock, may also appropriately be called gneisses. The question then remains whether any part of the foliated microcline-bearing granitic rock is due essentially to crystallization from a magma, with concurrent development of a flow or foliated structure without subsequent metamorphism, and hence properly called granite. It is thought that part of the microcline-bearing granitic rock did form as true granite, but the evidence is not clear whether all such rock was subsequently metamorphosed to gneiss. Nevertheless, the term gneiss alone will be used throughout the succeeding discussion, without implying that there is no true granite.

The microcline granite gneiss occurs in part as thin lens-shaped sheets, a few feet to a few hundred feet thick, isolated within the metasedimentary rocks—such as the ore-bearing sheets at Jarvis Bridge (50–150 ft thick), Parish (85–150 ft thick), and Dead Creek No. 1 (150–225 ft thick). In much larger part, the microcline granite gneiss occurs as sheets at least several hundred feet thick in synclinal structures, occupying moderate-sized to large areas. Examples of the latter are the South Russell and Childwold synclinoria, and the Granshue, Dead Creek, and Loon Pond synclines.

A long belt of microcline granite gneiss with associated layers of metasedimentary rocks and with alaskite and biotite granite gneisses (contaminated alaskite) extends in a great arc, more than 35 miles long, from Benson Mines approximately through South Russell, across the northwest corner of the Stark quadrangle, and onto the Potsdam quadrangle. This will be called the Russell belt. In large part, the microcline granite gneiss of this belt carries sillimanite-quartz aggregates in the form of discs or nodules in varied amounts. Locally, some muscovite-microcline granite gneiss is associated with the sillimanitic facies. Quartz veinlets with muscovite crystals usually occur with the muscovite-microcline granite gneiss. In many places part of the sillimanite is altered to muscovite, and locally biotite is altered to chlorite.

Microcline granite gneiss—most of it more or less involved with material from the associated biotite-quartz-plagioclase gneiss of the Grenville series and part of it sillimanitic or garnetiferous—occupies an elliptical area of synclinal structure on the Stark quadrangle east of Stone Dam and is here described as the Granshue mass. Farther south, at Deerlick Rapids and about a mile to the west, there are sheets of sillimanite-microcline granite gneiss in the metasedimentary rocks.

Another similar synclinal area of microcline granite gneiss occurs along Dead Creek (Cranberry Lake quadrangle). This too is variably sillimanitic and locally contains schlieren of the associated biotite-quartz-plagioclase gneiss.

Again, microcline granite gneiss, in large part sillimanitic and (or) garnetiferous in association with layers and schlieren of metasedimentary rocks, forms a synclinal belt across the Tupper Lake quadrangle through Round Lake and Loon Pond.

Microcline granite gneiss occupies a large area around the junction of the Stark, Childwold, Tupper Lake, and Cranberry Lake quadrangles. This will be called the Childwold area or belt. There are three major facies of the gneiss here, a hornblende-microcline granite gneiss, a biotite-microcline granite gneiss, and a pyroxene-microcline granite gneiss. Locally there is an andraditic facies. The sillimanite type of microcline granite gneiss is present in this belt, but rare.

A narrow belt of sillimanite- and biotite-microcline granite gneiss is exposed along the east side of the McCuen Pond syncline (Childwold quadrangle), and another in association with metasedimentary rocks in the southern part of the Tupper Lake quadrangle.

The microcline granite gneisses differ from the hornblende-microperthite granite or equivalent gneisses in having microcline as the almost exclusive feldspar, and from the alaskite in having 1–6 percent or more of iron oxides (in part or in whole consisting of rutilo-ilmenohematite, ilmenohematite or hemoilmenite) and an appreciable but varied amount of apatite.

In many places the microcline granite gneisses, especially the contaminated facies, have thin alternate layers of finer and coarser grain. Occasionally small pegmatitic veinlets both parallel to and across the foliation carry a little schorl. All the microcline granite gneisses are fine grained and have a gneissic structure, which is most conspicuous in the facies that are most contaminated. Where weathered the gneisses usually have a granulose appearance. Locally, small quartz lenses occur parallel to the foliation, especially in pegmatite-seamed zones.

The microcline granite gneiss is important from an economic standpoint, for it is the host rock for the magnetite concentrations at the Parish, Jarvis Bridge, Deerlick Rapids, Dead Creek No. 1 and No. 2, Skate Creek, and part of the Benson Mines deposits.

The mineral composition of representative facies is given in table 28.

RUSSELL BELT

The microcline granite gneiss of the Russell belt is almost wholly sillimanitic. Belts of biotite alaskite gneiss, of muscovite-microcline granite gneiss, and of biotite granite gneiss with schlieren and layers of biotite-quartz-plagioclase gneiss occur interlayered with the sillimanite-microcline granite gneiss. Northeast of Palmer Hill the sillimanite-microcline granite gneiss occurs as a few interlayers in biotite granite gneiss, the former varying markedly from layer to layer with respect to the quantity and coarseness of the sillimanite-quartz aggregates. Much of both the alaskite and the various facies of the microcline granite gneiss has the superficial appearance of an injection gneiss, owing to the close spacing of thin pegmatite seams parallel to the foliation. Locally there are seams and irregular veins of quartz in the granitic gneisses. Occasionally a few of these quartz veins carry well-developed muscovite crystals.

The granitic gneiss between Brouses Corners and Shingle Pond has layers which are sillimanitic and layers and schlieren of granitized biotite-quartz-plagioclase gneiss. The predominant rock carries a little muscovite. To the southwest, on Palmer Hill, the rock is microcline granite gneiss, containing thin pegmatite seams a few inches to a few feet apart, parallel to the foliation, and numerous quartz-schorl seams.

CHILDWOLD AREA

The Childwold area differs from all the other areas of microcline granite gneiss in that (a) no associated fluorite alaskite has been found, (b) the sillimanite facies has been found in only one small belt, (c) a major facies is a hornblende-microcline granite gneiss, and (d) layers and belts of amphibolite and of a veined rock of amphibolite and granite pegmatite are widespread. Locally, there are lenses of skarn a few hundred feet long, or thin layers of granite gneiss with small knots and lenses of skarn a few inches long. Such microcline granite gneiss is pyroxenic. The biotite facies of the microcline granite gneiss is a major

TABLE 28.—*Modes of microcline granite gneisses*

[Volume percent]

Rock type	Number of analyses averaged	Quartz	Microcline	Plagioclase	Pyroxene	Hornblende	Biotite (and chlorite)	Muscovite	Magnetite and (or) titaniferous hematite	Apatite	Zircon	Sillimanite	Garnet	Sphene	Fluorite
Loon Pond area, Tupper Lake quadrangle															
Microcline granite gneiss facies:															
Biotite	3	32.6	48.0	18.4			0.4		0.9	Tr.					
Garnet	1	31.1	54.6	11.4				0.5	.8				1.6		
Garnet-rich	2	32.8	57.9	7.2			.5		.3				6.3		
Garnet-biotite	2	26.6	58.8	8.8			2.2		2.0	0.2			1.4		
Sillimanite	1	30.9	63.2	2.4			1.0		1.4			1.0			
Childwold area															
Microcline granite gneiss:															
Biotite,[1] quartz-rich	4	31.9	56.7	7.6			0.8		2.8	0.2	Tr.				
Biotite	6	21.7	59.2	13.9			2.2		2.6	.3	Tr.				
Hornblende	9	23.5	52.6	15.1		4.1	.2		3.3	.4	0.05			0.8	
Pyroxene	2	25.6	56.8	10.7	2.0				3.9	.3				.7	
Pyroxene [2]	1	21.7	68.1	7.8	.8				1.2	Tr.				.4	
Hornblende [3]	1	19.8	70.7	2.4		2.6			2.6	.3				1.6	
Parishville area															
Microcline granite gneiss:															
Fluorite	1	26	43.8	25.0			1.1		1.7	0.3				1.1	1.0
Biotite	2	24	46.5	26.6			1.3		1.1	.2				.3	
South Russell synclinorium and Dead Creek syncline															
Microcline granite gneiss:															
Muscovite	4	39.7	42.2	8.5				7.3	1.7			0.6			
Quartz-sillimanite	12	38.6	50.5	5.5			1.6		1.9	0.05	0.05	1.7			
Sphene-biotite	4	30.1	57.4	6.1			2.1		2.4	.3	.1			1.5	
Biotite	5	26.3	53.6	13.9			2.7		3.0	.4	.1				
Granite gneiss facies:															
Biotite	8	27.0	33.6	32.9			3.1		2.7	.4	.1	.1	0.1		

[1] See analysis No. 18, table 30.
[2] See analysis No. 15, table 30.
[3] See analysis No. 16, table 30.

rock. In many places it is interlayered with the hornblende facies, and locally it contains schlieren of hornblende gneiss. Locally also it contains schlieren of more biotitic gneiss. The normal biotite-microcline granite gneiss is in large part varied in grain in alternating laminae, as though of migmatitic origin. There are no sharply defined contacts between the laminae, however, as in normal migmatite.

Much of the microcline granite gneiss of all facies—biotitic, hornblendic, pyroxenic, and sillimanitic—similarly weathers with a ribbed, veinlike, migmatitic appearance but is more or less homogeneous on fresh surfaces. Much of the fresh rocks have a "phantom" structure suggesting granitized and migmatitized metasedimentary rocks, or granitic gneisses contaminated by incorporation of material of the metasedimentary rocks and inheriting their structure. However, there are also sheets of biotite-microcline granite gneiss, which are uniform on both the weathered and the fresh surface. Locally, well-defined amphibolite migmatites grade into hornblende gneisses with a phantom migmatite appearance, but most of the hornblende-microcline granite gneiss shows no evidence of such gradation.

A sillimanitic facies of the biotite-microcline granite gneiss occupies a narrow east-west belt, about 2 miles long, a little less than one-half mile south of the north border of the Tupper Lake quadrangle, west and east of the County Line. This belt is not shown on the geologic map.

On the north part of the Tupper Lake quadrangle and in the southwest corner of the Childwold quadrangle, the gneisses of dominantly granitic appearance grade on the south and southwest into a belt of rocks that to a larger extent consist of hornblendic gneisses, have a more distinct migmatitic appearance, and are mapped as part of the metasedimentary rocks and migmatites of the Grenville series.

BIOTITE-MICROCLINE GRANITE GNEISS

Biotite-microcline granite gneiss is a major member of the microcline granite gneisses in all areas of their occurrence. The range of mineral variation is small, and the composition is fairly uniform.

In thin section, the biotite facies is seen to have a granoblastic texture. The grain diameters in general range from 0.5 to 1 mm. The microcline is in grain aggregates having a mosaic structure. The plagioclase is in grains of similar size and shape. The quartz is in grains of varied size and in grain aggregates. It commonly embays the feldspars and shows rounded, scalloped borders toward them. A little quartz occurs as small rounded grains within the feldspars. Locally the microcline is slightly microperthitic on a very small scale. The iron oxides occur as small disseminated grains, usually between the grains of the other minerals, but locally enclosed within them. The grains of iron oxides are either equidimensional, or elongate parallel to the foliation. They commonly average about 0.4–0.6 mm in diameter. The biotite occurs as scales between the grains of the other minerals, with the long diameter oriented roughly parallel to the foliation. In many places, the biotite is partly or wholly altered to chlorite, and sporadically one or more flakes are altered to muscovite. A little apatite and zircon is everywhere present as disseminated grains. Many apatite grains are associated with iron oxide grains.

Locally the biotite-microcline granite gneiss of the South Russell synclinorium carries a little sphene. The sphene occurs in rounded grains with replacement relation to the other minerals.

MUSCOVITE-MICROCLINE GRANITE GNEISS

The muscovite facies of the microcline granite gneiss is similar to the biotite facies, except that muscovite takes the place of biotite and in some places is in larger proportion, the plagioclase is partly sericitized, the rock carries more quartz, apatite occurs only in traces, and a little sillimanite is present in some specimens. The average mineral composition of four specimens is given in table 28. The muscovite formed later than the sillimanite, for locally it replaces portions of bundles of sillimanite needles. Where the muscovite-microcline granite gneiss occurs, small muscovite-bearing quartz lenses are usually present also.

BIOTITE GRANITE GNEISS FACIES

Locally, biotite granite gneiss is found associated with the normal microcline granite gneiss. Contacts between these two types of gneiss have not been observed. The biotite granite gneiss is a homogeneous rock and is similar in mineralogy to the biotite granite gneiss that is thought to be a product of alaskite contaminated by incorporation of material from the biotite-quartz-plagioclase gneiss of the country rock (p. 66). The biotite granite gneiss under discussion, however, occurs with the biotite-microcline granite gneiss independent of any observed alaskite. The average mineral composition of several samples is given in table 28. Sillimanite is found in parts of many of the outcrops from which the specimens of biotite granite gneiss were obtained, but no sillimanite was actually found in the biotite granite gneiss. Some of the biotite granite gneiss has a distinct to phantom migmatite structure, and some is homogeneous.

Locally the quartz content is less than 10 percent, and the biotite granite gneiss grades into a biotite syenite gneiss, with or without a little almandite. Rock of this type has been found locally in the Benson Mines syncline, the Dead Creek syncline, and at Trembley Mountain.

GARNET-MICROCLINE GRANITE GNEISS

Most of the microcline granite gneiss of the Loon Pond syncline, and much of that of the Dead Creek syncline, is a garnetiferous facies with associated layers and films of garnet-biotite-quartz-plagioclase gneiss, the latter in part migmatitic. The garnet-microcline granite gneiss is generally fine grained and contains uniformly disseminated small garnets. Locally it is medium grained and contains coarse garnets, many of which are in pegmatitic seams. However, pegmatitic laminae are not conspicuous in the normal rock. To a small extent, layers of migmatitic hornblende gneiss are associated with the garnet-microcline granite gneiss. The garnet-microcline granite gneiss in the Dead Creek syncline is largely associated with sillimanite-microcline granite gneiss. In the Loon Pond syncline, sillimanite-microcline granite gneiss is present but subordinate to the garnet facies.

The representative mineral composition of a garnet-poor and garnet-rich facies of the microcline granite gneiss of the Loon Pond syncline is given in table 28. It is noticeable that the garnet-rich facies shows a marked decrease of magnetite relative to the garnet-poor facies. The garnet-microcline granite gneiss appears to be associated with a larger percentage of biotite-quartz-plagioclase gneiss than does the sillimanite facies and may represent a less intense degree of granitization and metamorphism.

The garnet-microcline granite gneisses described in the preceding paragraphs are associated with and derived by granitization of biotite-quartz-plagioclase gneiss. The garnet of these rocks is almandite. There is another facies of the garnet-microcline granite gneisses, however, in which the garnet is andradite. The andradite-microcline granite gneisses are associated with pyroxene-microcline granite gneisses, and the andradite is the product of incorporation of metasomatized calcareous rocks. The andraditic facies is excellently exposed in the western part of the large road cut about one-half mile northeast of Piercefield Flow. The andradite occurs both as grains disseminated in the gneiss and as knots of andradite aggregate, one-quarter inch to several inches in diameter, sporadically distributed in a pegmatitic sheath in the gneiss. There are some lenses of pyroxene skarn in the gneiss to the northwest.

SILLIMANITE-MICROCLINE GRANITE GNEISS

Much of the microcline granite gneiss of the Russell arc, the Granshue, Dead Creek, and Loon Pond synclines, and most small sills in the metasedimentary rocks is more or less sillimanitic.

The sillimanite-microcline granite gneiss as a whole is much richer in quartz and a trifle poorer in biotite than either the alaskite or the nonsillimanitic microcline granite gneiss and carries up to several percent of sillimanite. The mineral composition of several examples of the slightly sillimanitic facies is given in table 28. Locally a slight amount of red garnet occurs in the sillimanite-microcline granite gneiss.

The sillimanite occurs almost exclusively as fibers and fibrous aggregates in association with quartz. These aggregates range from sparse, almost indistinguishably small lenticles to abundant thin discs, platy lenses, or nodules as much as several inches in length and one inch in thickness. Some small flat lenticles consist largely of sillimanite and yield thin lenslike cavities on the weathered surface of the gneiss, but the discs and platy lenses usually weather in relief, appearing as projecting white fins. The lenticular nodules also weather in relief, giving a conglomeratic appearance to the rock. In their coarser phases they are 1–6 inches long and ½–1 inch thick, and in layers where they are abundant, they form as much as 25 percent of the rock. Locally, the sillimanite is partly altered to a sericitic aggregate.

Some sillimanite-quartz aggregates, occurring in very thin discs or in films, consist largely of sillimanite with a little quartz; but the larger discs and nodules are predominantly quartz with a little sillimanite, and locally a little microcline or muscovite, or sparse zircon. Locally, adjoining the sillimanite-quartz aggregates, there may be found isolated fibers or bundles of fibers of sillimanite replacing microcline, or thin replacement veinlets of sillimanite and quartz in microcline. Iron oxides are commonly present in the aggregates and may occur as disseminated grains, in concentrations at and near the center, as intergrowths with and replacements of the sillimanite, or locally as complete replacements of the discs and nodules. Perfect pseudomorphs of iron oxide after sillimanite are found at many places. Locally, the iron oxide grains have antennaelike veinlets which cut across the quartz-sillimanite aggregates and the minerals of the granitic gneiss. Some grains are compound, consisting of iron oxide and apatite. Veinlets of iron oxide and apatite also occur locally.

The discs and nodules of sillimanite-quartz aggregate usually have three unequal dimensions and are commonly so oriented as to define a linear structure. In many places, small local slip surfaces have formed along some of the longer discs. Locally, also, the discs

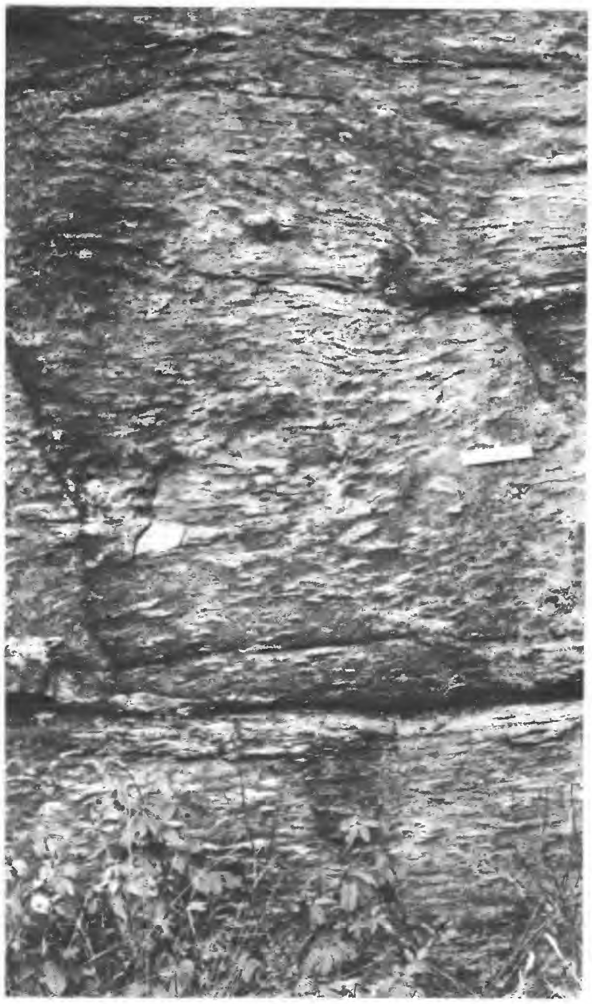

FIGURE 7.—Microcline-rich granite gneiss with sillimanite-quartz nodules and fabric of small isoclinal plications. White scale is 7 inches long. Locality 0.6 mile and a little east of north from Red School, Russell quadrangle.

are crumpled into small chevron folds, many of which are sheared out, forming a new nodular rock through dynamic deformation (fig. 7). The nodular sillimanite gneiss of the hill 1½ miles northwest of Benson Mines may have had such an origin.

Locally there are narrow zones of sillimanite-quartz schist associated with the sillimanite-microcline granite gneiss, whose origin is uncertain. One such zone lies on the east side of the road from Green Valley School to Partlow Pond and extends for more than a mile south of Stammer Brook. The rock is a sericitic sillimanite-quartz schist in which sericite has partially replaced the sillimanite fibers. Quartz also occurs in veinlike forms and locally carries well-developed muscovite crystals, accessory feldspar, a little tourmaline, or hematite grains.

Another lens of sillimanite-quartz schist lies at the head of a swamp a little north of 1.8 miles west of Upper Degrasse School. It is within a belt of sillimanite-microcline granite gneiss.

HORNBLENDE-MICROCLINE GRANITE GNEISS

The normal hornblende facies of the microcline granite gneiss is usually pink, though it is light gray where the hornblende is considerably higher than normal and the rock grades into amphibolite. The average mineral composition of the normal pink gneiss is given in table 28. The percentage of hornblende ranges from 1 to 6, and quartz from 18 to 32. The hornblende facies also differs from the biotite facies in having an almost uniformly small percentage of sphene. The sphene occurs as rounded grains, apparently with later replacement relations to the other minerals. Like the hornblende, it has resulted from contamination with mafic rocks. The hornblende commonly has a distinct pleochroism from blue to green. A very little biotite is associated with the hornblende in a few specimens.

PYROXENE-MICROCLINE GRANITE GNEISS

Pyroxene-microcline granite gneiss occurs as layers at many localities in the Childwold area. It is exposed, for example, just southeast of the observation tower on Moosehead Mountain, where it forms a layer several feet thick containing sparse thin lenses and beads of pyroxene skarn. Adjoining layers contain amphibolite inclusions a few inches to a few feet long. Granite pegmatite veins up to one-half inch thick are present every few inches. North of the bridge across the Raquette River at Piercefield, the microcline granite gneiss contains little knots of pyroxene and dark brown garnet. The rock in the old quarry just east of the bridge is a microcline granite gneiss with 2 or 3 percent augite and 1 percent sphene. The granitic gneiss here also has a heterogeneous layered texture and contains many schlieren of amphibolite. The granitic gneiss 0.7 mile north-northeast of Clark (Stark quadrangle) is a pyroxene facies, of which a chemical analysis is given in table 30 (No. 15).

MIGMATITE OF AMPHIBOLITE AND GRANITE PEGMATITE

Locally throughout the Childwold synclinorium there are layers of amphibolite migmatite within the microcline granite gneiss. Such rock consists of alternating thin layers of partly granitized hornblende gneiss and granite pegmatite. The potassic feldspar is largely microperthitic, in contrast to the microcline of the normal granite gneiss. Either augite or hypersthene, or both, may be present depending on the original composition of the amphibolite. The average composition of two types is given in table 10. Migmatite of this kind is well exposed in railroad cuts just east of

Shurtleff and in road cuts on the main highway 1 mile east of Sevey.

ALBITE-OLIGOCLASE GRANITE GNEISS

Albite-oligoclase granite gneiss forms a quite minor facies of the granitic gneisses as a whole. It occurs largely as small sheets and lenses in the metasedimentary rocks but is found widely distributed within them. It also occurs in small veinlike layers within the sillimanite-microcline granite gneiss. It has been found as part of the wall rocks of many of the magnetite deposits. The largest identified mass forms the hills one-half mile northeast of Jarvis Bridge.

The rock is usually white to light gray, varying to a pink or lighter tint than the normal alaskite gneiss. The predominant minerals are plagioclase (usually on the border line between albite and oligoclase) and quartz, with subordinate microperthite or microcline. The plagioclase may in part have antiperthitic intergrowths of potassic feldspar. Quartz ranges from 5 to 40 percent. The mafic minerals may be augite, hornblende, or biotite, or a combination of these; they range in amount from 0.5 to 6 percent. Accessory minerals are magnetite, apatite, zircon, and locally a little sphene.

Locally the granite gneiss is slightly contaminated with minerals of the country rock, and locally, as about one-half mile north of Newton Falls, it forms a thin-layered migmatite with pyroxene-quartz gneiss.

The mineral composition of a number of representative facies of the sodic granite gneiss is given in table 29.

TABLE 29.—*Modes of albite-oligoclase granite gneisses*

[Volume percent]

		Feldspars											
	Quartz	Plagioclase	Antiperthite	Microcline, orthoclase, or both minerals	Microperthite	Hornblende	Biotite (or chlorite)	Augite	Sphene	Iron and titanium oxides	Apatite	Zircon	Pyrite
1	28.3	43.5	-----	7.4	17.5	0.6	0.3	-----	-----	1.8	0.2	0.4	-----
2	5.0	69	18.1	-----	-----	5.5	(1.7)	-----	0.7	-----	-----	-----	-----
3	30.6	54.9	-----	7.6	-----	5.1	-----	-----	-----	1.7	.1	-----	-----
4	36.5	61.3	-----	.2	-----	-----	.5	-----	-----	1.3	-----	.1	-----
5	12.8	19.7	54.4	4.7	-----	2.9	-----	4.0	-----	.8	.7	-----	-----
6	12.1	64.3		-----	14.8	3.6	.6	1.0	-----	3.0	.5	-----	-----
7	11.0	53.2	7.3	12.4	-----	7.1	0	6.2	.6	.4	1.1	.2	0.5
8	12.8	48.5	17.0	-----	-----	12.3	1.7	-----	-----	7.2	.5	-----	-----
9	12.9	77.6	-----	8.0	-----	-----	.8	-----	-----	.2	-----	.1	.5

1. 0.5 mile northwest of Newton Falls, Cranberry Lake quadrangle.
2. 1.1 miles east of Scotts Bridge, Oswegatchie quadrangle.
3. Yellow Creek, 1.75 miles west of School No. 13, Oswegatchie quadrangle.
4. Diamond-drill hole 5, Brandy Brook, Cranberry Lake quadrangle.
5. 1.75 miles northeast of Newton Falls, Cranberry Lake quadrangle.
6. 0.5 mile east of Jarvis Bridge, 2 miles east of Newton Falls, Cranberry Lake quadrangle.
7. 2.5 miles northeast of Newton Falls, Cranberry Lake quadrangle.
8. 0.7 mile east-northeast of Cook Corners, Cranberry Lake quadrangle.
9. Just south of Claflin School, Potsdam quadrangle.

The microstructure of the rock ranges, dependent upon environment and geographical location, from a medium grain (1–1.5 mm) texture, in which the feldspar grains show interlocking borders, to a fine-grained granoblastic texture, in which most of the feldspar grains have a polyhedral outline. The amount of potassic feldspar as intergrowths with plagioclase ranges from zero to normal microperthite that consists of about equal amounts of potassic and sodic feldspar. Phenomena which could definitely be interpreted as indicating replacement of microperthite by plagioclase were in general not seen, though occasional grains of plagioclase with relics of potassic feldspar might be so interpreted. Rarely the grains of plagioclase are traversed by replacement veinlets of potassic feldspar. The antiperthite has in part a rough checkerboard pattern and in part a parallel stringlet intergrowth of potassic feldspar.

The feldspars of the chemically analyzed specimen (table 30, No. 19) comprise clear plagioclase, plagioclase with microintergrowths of potassic feldspar ranging from nearly zero to 50 percent, slightly microperthitic microcline, and clear microcline. The low potassic antiperthite could be interpreted as arising from a replacement of microperthite, or of antiperthite with much potassic feldspar intergrowth, by plagioclase. The mafic minerals comprise augite, hornblende, and oxides. The augite and hornblende are partly altered to chloritic products, and locally a little biotite is present.

The phenomena as a whole could be interpreted in terms of an antiperthite (or microperthite) granite more or less replaced by albite, locally with recrystallization of part of the potassic feldspar as new grains of microcline, or as a new and coarser antiperthitic intergrowth. There may also locally have been some sodic granite formed directly.

If the sillimanite-microcline granite gneiss is the result of modification of the biotite-quartz-plagioclase gneiss by potassium-rich solutions, necessarily a vast quantity of plagioclase must have been displaced. It would seem that at least a small amount of this material would be deposited in the environment. The albite-oligoclase granite gneiss could thus be reasonably interpreted as formed from a sodium-rich fluid of such an origin, sweated out of the biotite-quartz-plagioclase gneiss. Plagioclase granite veins in sillimanite-microcline granite gneiss may have the form of chevronlike plications, very much thicker on the crests and troughs than on the limbs. They have a much more massive structure than the enclosing gneiss and in part are almost pegmatitic in nature. They look like late segre-

TABLE 30.—*Chemical analyses, norms, and modes of granites and granite gneisses*

	Hornblende granite			Hornblende granite gneiss		Biotite alaskite			Biotite alaskite gneiss			Garnet alaskite gneiss	Microcline granite gneiss						Oligoclase granite
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
Chemical analyses (weight percent)																			
SiO_2	70.11	70.72	69.30	69.70	70.44	76.20	76.41	75.42	75.71	73.10	70.38	70.99	68.03	68.66	70.99	67.89	71.44	70.99	64.25
Al_2O_3	14.11	13.11	14.94	13.29	13.28	12.54	12.41	11.76	12.68	14.29	14.45	14.28	14.63	12.98	13.34	13.11	14.89	13.54	16.05
Fe_2O_3	1.14	1.85	1.02	2.90	2.07	.90	1.01	1.98	.96	1.04	1.30	1.19	2.72	2.89	2.71	3.32	2.40	4.14	1.80
FeO	2.62	1.97	1.92	2.07	2.16	.56	.50	1.19	.32	1.04	1.46	2.30	1.77	1.26	1.15	2.24	1.38	.54	1.80
MgO	.24	.50	.46	.41	.43	.08	.46	.15	.13	.53	.39	.53	.66	.76	.24	.56	.30	.23	1.31
CaO	1.66	1.36	1.46	1.55	1.57	.73	.78	.83	.67	1.18	1.37	1.12	.65	2.63	.78	1.45	.50	.31	3.30
Na_2O	3.03	3.35	3.60	3.14	2.95	3.57	3.34	3.34	3.29	3.08	3.72	2.80	3.00	2.05	2.06	1.84	1.36	2.71	5.56
K_2O	6.03	5.60	5.94	5.51	5.57	5.23	4.33	4.32	5.49	5.36	5.59	5.97	7.24	7.50	7.83	7.88	5.99	6.63	3.69
H_2O+	.23	.37	.25	.32	.42	.27	.34	.24	.24	.54	.28	.30	.36	.48	.22	.27	.52	.12	.30
H_2O-	.07	.06	.09	.03	.06	.06	.13	.10	.09	.07	.04	.08	.08	.09	.09	.03	.08	.08	.10
CO_2				.20															.63
TiO_2	.42	.41	.43	.51	.57		.03	.20	.18	.18	.42	.26	.52	.19	.43	.72	.65	.45	.64
P_2O_5	.09	.23	.12	.13	.14			.01	.02	.03	.08	.02	.12	.07	.09	.15	.22	.07	.22
MnO	.06		.05	.08	.07		.06	.03	.01	.17	.05		.05	.24	.05	.08	.03		.06
F	.09		.09		.13		.01	.29	.19	.02	.20	.09		.01			.06		
BaO				.10	.06				.02				.10	.07			.09	.06	
Subtotal	99.90		99.67		99.92		[1]99.84	99.86	100.00	[2]100.68	99.73	99.93					99.91		
Less O for F	.04		.04		.05			.12	.08		.08						.03		
Total	99.86	99.53	99.63	99.94	99.87	100.14		99.74	99.92		99.65		99.93	[3]99.96	99.98	99.54	99.88	99.87	99.71
Norms																			
Quartz	23.19	26.46	21.33	26.76	27.60	33.6	37.95	38.25	34.44	30.40	24.06	27.42	20.52	23.18	27.30	23.67	39.54	29.24	11.28
Orthoclase	35.58	33.36	35.03	32.80	33.36	31.1	25.58	25.58	32.80	31.69	33.36	35.58	43.37	44.37	46.15	46.70	35.58	38.92	21.68
Albite	28.03	28.30	30.39	26.72	25.15	30.4	28.30	28.30	27.77	26.20	31.44	23.58	25.15	17.34	17.29	15.72	11.53	23.06	47.16
Anorthite	5.70	4.17	6.39	5.56	6.12	2.5	3.89	1.95	1.81	5.80	4.73	5.42	2.50	3.97	3.39	3.89	1.11		8.06
Hypersthene	3.40	2.25	3.28	1.69	2.55		1.23	.60	.30	2.46	2.05	3.94	1.85		.60	.73	.80	.60	3.06
Magnetite	1.74	2.65	1.39	4.18	3.02	1.4	1.39	3.02	.53	1.67	1.86	1.86	3.94	4.20	2.67	4.87	2.55	1.74	2.55
Ilmenite	.76	.77	.82	.97	1.06			.38	.35	.31	.76	.46	.99		.84	1.37	1.22		1.22
Apatite	.20	.58	.27	.34	.34			.02	.04		.20	.04	.27	.15	.20	.34	.50	.17	.50
Fluorite	.19		.19		.26			.62	.39		.39						.12		
Corundum			.25				.82	.92	.66	1.50	.51	1.17	.82		.20		5.71	1.89	
Hematite									.59						.88		.64	2.96	
Diopside	1.02	.93		.11	.09									4.50		1.97			2.44
Wollastonite														1.06					
Sphene														.33				1.08	
Calcite				.45															1.40
Modes (weight percent)																			
Quartz	24	22	22.7	29	27.5	35	34.57	32.8	33.7	29.0	22.6	24	26.6		25.0	21.6		30.1	11.7
Microperthite	55.3	67.1	46.2			62	61.74	55.7	44.4			52.3	56.6						
Microcline and (or) orthoclase			6.4	38	37.4			5.3		36.5	38.2				63.3	[4]65.8		49.4	14.0
Plagioclase	11	3.2	16.2	22	25.2	1	2.01		18.9	30.5	34.8	17.5	10.1		7.8	2.4		14.7	[5]61.1
Augite															1.0				1.2
Hornblende	8.6	4.6	6.5	7	5.0	.5							.4			2.9			4.2
Biotite		.8	.8		1.5	1.5		1.9		1	2.3	1.6							.3
Iron oxides	.7	1.8	1.0	3	1.0		.34	3.1	1.6	1	.6	2.6	5.3		2.3	4.8		4.0	5.6
Apatite	.2	.4	.2	.3	.4	.1					.2				.2	.4		.1	.6
Zircon	.2	.1	Tr.	.2	.2	.1	.18	.2	.2		.2		.1					.1	
Fluorite								1.0	1.2		.2								
Sphene				.1	1.8		.03				.8				.5	2.1			
Calcite				.5															1.0
Chlorite							1.13						[6]1.0					1.2	.3
Muscovite										2									
Garnet												2						.4	

[1] Contains also 0.02 percent ZrO_2 and 0.01 S not shown in table.
[2] Contains also 0.03 percent Cl and 0.02 S not shown in table.
[3] Contains also 0.08 percent S not shown in table.
[4] Slightly perthitic.
[5] Antiperthitic, in part.
[6] Chloritized hornblende.

NOTE.—Chemical analyses 1, 3, 5, 8, 9, 11–13, and 15–19 are from Buddington (1957).

1. (B–189). Ferrohastingsite granite, 0.5 mile north of Cranberry Lake State Park, Cranberry Lake quadrangle. Analyst, Lee C. Peck. Hornblende is a ferrohastingsite ($nX=1.694$, $Z\wedge c=12°-13°$). For chemical analysis of hornblende, see Buddington and Leonard (1953, analysis 4).

2. (W–14). Femaghastingsite granite, 1.2 miles north-northeast of Bushs Corners, Lowville quadrangle (Buddington, 1939, table 37, No. 133, p. 138). Analyst, R. B. Ellestad. Hornblende is a femaghastingsite ($nX=1.674$, $Z\wedge c=14°\pm3°$). For chemical analysis of hornblende, see Buddington and Leonard (1953, analysis 2).

3. (L–285). Hornblende granite, at road crossing of creek 0.4 mile southeast of High Rock, Cranberry Lake quadrangle. Analyst, Lee C. Peck.

4. (4603). Hornblende granite gneiss, abandoned road-material quarry on south side of State Road, 0.15 mile west of intersection of Waverly, Dickinson, and Hopkinton township lines, Nicholville quadrangle (Buddington, 1939, table 37, No. 134, p. 138). Analyst, T. Kameda.

5. (956). Femaghastingsite granite gneiss 1 mile south of Palmerville, Blanchard Hill, Russell quadrangle. Analyst, Lee C. Peck. Hornblende is a femaghastingsite ($nX=1.682$, $Z\wedge c=16\frac{1}{2}°$). For chemical analysis of hornblende, see Buddington and Leonard (1953, analysis 3).

6. Biotite alaskite, Bishas Mill, Lake Bonaparte quadrangle (Smyth and Buddington, 1926, p. 43). Analyst, A. F. Buddington.

7. Alaskite, Mount Morris, Long Lake quadrangle (Cushing, 1907, p. 510–512). Analyst, E. W. Morley.

8. (B–1599). Biotite-fluorite alaskite, southeast of Dead Creek near Bench Mark 1901, Cranberry Lake quadrangle. Analyst, Lee C. Peck.

9. (B–1287). Biotite-fluorite alaskite gneiss, 1.3 miles southeast of Star Lake, north of Alice Brook, Oswegatchie quadrangle. Analyst, Lee C. Peck.

10. Biotite alaskite gneiss, taken 0.25 mile south of Alexandria Bay, Alexandria Bay quadrangle (Cushing, 1910, p. 177). Analyst, E. W. Morley.

11. (B–1136). Biotite-fluorite alaskite gneiss, road cut 2.5 miles east of Pond Settlement, Russell quadrangle. Analyst, Lee C. Peck.

12. (B–2087). Garnet alaskite, a facies contaminated by metasediments, 0.4 mile south of southwest end of Horseshoe Lake, Tupper Lake quadrangle. Analyst, Lee C. Peck.

13. (B–2383). Potassium-rich ferrohastingsite-microperthite-microcline granite gneiss, 0.5 mile southwest of Windfall Pond, Childwold quadrangle. Hornblende is a ferrohmastingsite ($nX=1.688$, $Z\wedge c=11°$, $2Vx=56°\pm4°$, $r<v$, X=yellowish green, Y=dark grayish olive, Z=dark bluish green, $Z=Y>>X$). Density of rock=2.664. Transitional in character between normal hornblende-microperthite granite and the microcline granite gneisses. Analyst, James J. Engel.

14. Augite-microcline granite gneiss, near Lake Catlin, Long Lake quadrangle (Cushing, 1907, p. 526). Analyst, E. W. Morley.

15. (B–209). Augite-microcline granite gneiss, 0.7 mile north-northeast of Clark, Stark quadrangle. Analyst, Eileen H. Kane.

16. (B–555). Femaghastingsite-microcline granite gneiss, abandoned crushed-stone quarry, north side of Dead Creek, Tupper Lake quadrangle. Analyst, Eileen K. Oslund. Hornblende is a femaghastingsite ($nX=1.666$, $Z\wedge c=21°\pm5°$). For chemical analysis of hornblende, see Buddington and Leonard (1953, analysis 1).

17. (SC–1–120). Sillimanite-microcline granite gneiss, diamond-drill hole 1, Skate Creek, 120-ft. depth, Oswegatchie quadrangle. Analyst, Lee C. Peck.

18. (B–126). Biotite-microcline granite gneiss, road cut 0.8 mile north of Shurtleff, Childwold quadrangle. Analyst, Eileen H. Kane.

19. (B–1245). Oligoclase granite, abandoned crushed-stone quarry, 0.5 mile east of Jarvis Bridge or 2 miles east of Newton Falls, Cranberry Lake quadrangle. Analyst, Lee C. Peck.

gation veins. Some are present in the rock of figure 7 but are not identifiable in the picture.

CHEMICAL COMPOSITION OF ROCKS OF GRANITE AND GRANITE GNEISS SERIES

Chemical analyses of representative samples of the granite and granite gneiss series are given in table 30, together with the norms and the modes where the latter have been determined. The modes do not correspond with the norms to the degree that might be expected in some cases. This is in part because the granite itself is somewhat varied in composition, and the number of thin sections studied was inadequate to give a close correspondence.

Some of the ferrohastingsite granites have a higher ratio of K_2O to Na_2O than the femaghastingsite granites (Nos. 2–5) and are higher in quartz. This is reasonably explained in terms of magmatic differentiation and fractional crystallization, whereby the later magmatic fraction becomes richer in K_2O and SiO_2, and the hornblende is enriched in FeO. There is also ferrohastingsite granite No. 1, however, which has K_2O and SiO_2 similar to the femaghastingsite granites and equivalent gneiss. This may be explained by the fact that the composition of the hornblende also depends on the degree of oxidation of the iron as well as on the degree of fractionation of MgO and FeO. For rocks with the same content of total iron, the iron may appear either in part as magnetite and in part as hornblende, or wholly as hornblende, depending upon the degree of oxidation. With the same content of iron and titanium we may thus have either femaghastingsite granites with magnetite and ilmenite, or ferrohastingsite granites with ilmenite as the only accessory opaque oxide.

The femaghastingsite-microperthite granite (No. 2) is very similar to the hastingsite-microcline-oligoclase granite gneisses (Nos. 4, 5). The microcline and oligoclase represent the unmixing and recrystallization of the microperthite during the deformation that yielded the gneissic structure. The sphene in the gneisses (Nos. 4, 5) must also be the product, in part at least and probably in whole, of reconstitution during deformation, for the gneiss has nearly the same percentage of TiO_2 as the normal granite.

The alaskites are predominantly much higher in SiO_2 than the hornblende granites. The ratio of normative orthoclase to plagioclase is about the same in both rocks. The plagioclase in the alaskites is somewhat more sodic than that in the hornblende granites. Nos. 8, 9, and 11 carry more fluorine than the hornblende granite. The fluorine is present in the alaskite partly as fluorite and partly in the biotite. In the hornblende granite the fluorine is normally present wholly in the hornblende. The percent of magnetite in No. 8 is exceptionally high for alaskite and is probably in part the product of later mineralization. The specimen came from a position only a short distance away from a low-grade magnetite deposit. It will also be noted that microperthite alaskite and microcline-albite alaskite gneiss may have similar chemical compositions.

The garnet alaskite has a slightly higher percentage of FeO in consequence of assimilation of material from the metasedimentary rocks.

The microcline granite gneisses are high-K_2O rocks. The average ratio of normative orthoclase to total normative feldspar is 69 in the microcline granite gneisses, as contrasted with 48.5 in the alaskites and 50 in the hornblende granites. The microcline granite gneisses are also characterized by a higher content of iron oxides, especially Fe_2O_3, than any of the other granitic rocks. The high normative corundum in No. 17 is due to the presence of considerable sillimanite in the rock. This rock also shows a higher quartz content than normal. Both the sillimanite and the quartz are thought to be the product of modification of relics of metasedimentary rocks.

BASALT AND KERATOPHYRE DIKES

Diabasic or basaltic dikes are common in the Precambrian rocks in the vicinity of the St. Lawrence River (Alexandria Bay quadrangle) and in the northern Adirondacks (northern third of the Santa Clara quadrangle, southern part of Malone quadrangle and the Lyon Mountain quadrangle). With the exception of the Alexandria Bay area they are rare in the northwestern Adirondacks and only four have been noted within the area of the St. Lawrence County magnetite district. The basaltic dikes of this area are wholly undeformed, show distinct chill zones, locally glass, and must have formed distinctly later and much nearer the surface than any of the other Precambrian rocks.

The dikes all strike N. 45°–60° E. The prevailing strike for the diabase dikes of the northern Adirondack area in general is east-northeast.

The basaltic dikes were noted on the Stark quadrangle in the gorge of the South Branch of the Grass River about 1 mile east of Newbridge, on the top of Catamount Mountain, and one-half mile west of Catamount Mountain. A narrow 2-inch dike was seen in the bed of the Raquette River one-half mile southwest of South Colton, in the Potsdam quadrangle.

The dike on Catamount Mountain is 9 inches wide. In thin section it is seen to be composed of an interlocking mat of microscopic plagioclase laths with interstitial dark groundmass, presumably pyroxene shot through with a skeletal network of hairlike ilmenite and magnetite. There are a few microphenocrysts of mafic minerals that are completly replaced by calcite.

A dark dike is exposed on the Cranberry Lake quadrangle, 0.7 mile northwest of the main road on the trail to Moosehead Pond. This dike appears to be a keratophyre, composed largely of plagioclase and a little skeletal magnetite, biotite, and chlorite. Keratophyre dikes of Precambrian age are not certainly known in the Adirondacks. Keratophyre dikes have been found in the Willsboro quadrangle (Buddington and Whitcomb, 1941, p. 80), where they have been inferred to be of post-Ordovician age.

SEDIMENTARY ROCKS

The sedimentary rocks of the St. Lawrence County magnetite district are represented by a few outlying patches of the Potdam sandstone of Late Cambrian age. These are residual remnants of Paleozoic sediments once deposited on the previously eroded sub-Cambrian peneplain.

POTSDAM SANDSTONE

Potsdam sandstone occurs locally in the northwest border portion of the area as small relict patches of flat-lying sedimentary rock unconformably overlying the Precambrian rocks. It has been described by Chadwick (1920) and by Reed (1934, p. 29–33).

The predominant member of the Potsdam rocks is normally a well-bedded white to red sandstone. However, the outlying relics, such as are found in this area, include conglomerate beds, breccias (of taluslike character), and autoclastic breccias as major elements.

All the relics of sandstone in this area overlie marble beds. The marbles during Potsdam time formed the valleys as they do now, and it is only at these lower levels that the sandstones have been preserved from subsequent removal by erosion.

An excellent example of reddish sandstone and autoclastic sandstone breccia is present on the south side of Parkhurst Brook, 0.35 mile south of High Flats, Potsdam quadrangle. The sandstone is underlain by marble. There is also a sandstone dike in the marble here. This has resulted from sand infilling a solution fissure in the marble during Potsdam time. The sandstone of the fragments of the breccia, of the matrix of the breccia, and of the normal unbrecciated sandstone beds is all similar. It is composed of well-rounded quartz grains with a surface film of hematite and an overgrowth of silica cement. A few tourmaline grains are present.

Excellent exposures of normal Potsdam sandstone with conglomerate interbedded near the base occur below the dam at Whitaker Falls (Potsdam quadrangle). This locality has been described by Reed (1934, p. 30–31). Locally, a lens of conglomerate with a maximum thickness of 4–5 feet in part rests directly on Precambrian granite gneiss and in part is separated from the gneiss by a bed of white sandstone. The pebbles in the conglomerate are almost wholly quartz, granite gneiss, and feldspathic gneiss. The feldspar of the gneisses is almost wholly altered to clay minerals. There are rare jasper pebbles in the upper part of the conglomerate. The sandstone above the conglomerate is a well-bedded white variety with reddish interlayers. The granite gneiss for a foot beneath the unconformity has a shaly character, and the feldspars are altered to clay minerals. The sandstone carries a few tourmaline and zircon grains. Authigenic potassic feldspar has developed locally within small, irregular aluminous aggregates forming part of the matrix in the sandstone and conglomerate.

The area west of Pierrepont mapped as Potsdam sandstone is completely covered with drift except for an old quarry exposure 1.5 miles northwest of Pierrepont in the bed of a brook 0.3 mile north of the schoolhouse. The boundaries of the area have been based upon the work of Chadwick (1920, p. 15 and geologic map). Chadwick reports that at a point 1.5 miles west-southwest of Pierrepont, just north of the road, a well was drilled through 28 feet of till and 150 feet of typical red Potsdam sandstone.

Sandstone and breccia of the Potsdam are also exposed in patches along the northwest side of the valley of Van Rensselaer Creek, between the road from West Pierrepont to Beach Plains Church and the road from West Pierrepont to Pierrepont.

One set of exposures is on the Allen farm, 0.4 mile northwest of West Pierrepont. The basement rock unconformably underlying the Potsdam relics and forming most of the valley bottom consists of marble with zones of bedded white coarsely crystalline quartzite, thin-layered quartzite and silicate rocks, and quartz-mesh marble. The marble generally carries some serpentine knots and nodules, or disseminated grains of serpentine. The silicate layers are largely composed of diopside, tremolite, or diopside and tremolite. The thin-layered zones are often intensely twisted and crumpled. Four outcrops of reddish breccia and sandstone of the Potsdam occur in a belt extending about 800 feet east-northeast from the barn on the Allen place, and another line of outcrops of pink breccia of the Potsdam starts about 1,300 feet northeast of the barn and extends for about 800 feet east-northeast. (Illustrated in Prof. Paper 377*.) There are also many small

*See footnote on p. 31.

patches of breccia north and northeast of the house. The breccias of the Potsdam are usually purplish in hue. The breccia consists of a purplish sandstone or pebbly conglomerate matrix, with rounded to subangular fragments of white quartz and quartzite. The quartzite fragments are similar to the quartzite in the underlying marble, and are presumably derived from it by erosion and sedimentation. Part of the quartzite breccia of the Potsdam is of the nature of a talus breccia. Some of the fragments are up to 1.5 feet in diameter. Fragments and blocks of purplish sandstone also occur in some of the breccia; they resemble the Potsdam sandstone itself and may represent autoclastic fragments. There are local lenses and beds of reddish sandstone interbedded with the breccia. The bedding in some of the lenses has a moderate to steep dip.

A fine example of a coarse autoclastic breccia of the Potsdam is exposed in outcrop and large boulders at the Dillabaugh place, 1.9 miles south of Pierrepont (Canton quadrangle), where the Potsdam unconformably overlies marble of the Van Rensselaer Creek valley. Here too there are all gradations between well-bedded sandstone and autoclastic breccias in which the fragments are similar to the matrix and to the normal bedded parts. There are no foreign fragments in the breccia. The fragments are up to one foot in diameter. The breccia has druses lined with quartz crystals and specular hematite. There are similar druses in the unbroken sandstone, but they are very small and much less abundant. The sandstone varies from that in which the boundaries of the grains may still be observed to almost completely recrystallized material.

About 275 yards east-northeast of the old house is an old pit, about 18 feet long and 10 feet deep, which was sunk on ferruginous sandstone in a search for hematite veins. The sandstone appears to dip about 50° SW. The basal part is a red, ferruginous, completely cemented and recrystallized quartzitic sandstone, with hematite veins overlain by a 2-foot bed of yellow-brown limonitic siltstone. The upper beds are normal sandstone.

A small patch of thin-bedded, locally brecciated Potsdam sandstone unconformably overlies marble in the bed of Little River, 1.15 miles northwest of Hamiltons Corners (Russell quadrangle).

About 0.6–0.7 mile south of Pierrepont there are numerous great slabs of thin-bedded calcareous sandstone which probably represent material moved but little from the place of deposition.

A small area of sandstone and conglomerate of the Potsdam, largely obscured by Pleistocene drift, lies about one-fourth to one-half mile north and northeast of Stellaville. Reddish, cross-bedded sandstone and conglomerate are exposed near the western end of the area. A drill hole put down by the operators of the Stellaville Mines near the middle of the southern side of the area is reported to have passed through about 150 feet of sandstone.

A small outcrop of sandstone and conglomerate is also present about a mile north-northwest of Stellaville.

ACCESSORY OXIDE MINERALS IN THE COUNTRY ROCK

The accessory oxide minerals in the country rock received special study because of their relation to variations in the earth's local magnetic field and to the origin of the ores. This study, results of which are summarized below, was a separate project, begun after the geologic study of the district had been completed. Mineralogical work by Buddington was supported by mineral separations and chemical analyses by J. J. Fahey and Angelina Vlisidis. The interpretations are Buddington's. The geophysical data were provided by J. R. Balsley and associates. The accessory oxide minerals include several species not known to occur in the ore deposits, and relations among these minerals are complex. How abundant the accessories are, the reader may determine by referring to the modes for the various rock types in the tables of this report. Generally, the accessory oxides constitute tenths of 1 percent to a few percent of the country rock, whereas oxide minerals in the ore deposits usually are ten or more times as abundant.

MINERALOGY AND DISTRIBUTION OF ACCESSORY OXIDES

The accessory oxides found in some of the important rock types are listed in table 31. Compound terms have been used where one mineral occurs as a microscopic intergrowth in another. The host mineral (noun) is named last; an adjective describing the intergrown guest precedes it. Thus, ilmenomagnetite is magnetite containing a microscopic intergrowth of ilmenite, ilmenohematite is titanhematite (see below) with intergrown ilmenite, hemoilmenite is ilmenite with intergrown hematite, and rutiloilmenohematite is titanhematite with intergrowths of rutile and ilmenite. The fabrics and relations of the microscopic intergrowths to their host mineral and host rock are such that they are interpreted as products of exsolution from originally homogeneous material then existing at higher temperature. Subtitanhematite is hematite with 1.5–5 percent TiO_2, and titanhematite is hematite with 5–10 percent TiO_2 in solid solution; the TiO_2 may be recomputed as though it occurred as the compound ilmenite, as rutile, or as both compounds. Ilmenomagnetite is

TABLE 31.—*Variation of accessory oxide minerals with kind of rock, Adirondack massif*

[X, Mineral is the only accessory oxide, or the predominant one; x, mineral is a minor accessory, locally absent]

Rock	Magnetite host		Ilmenite host		Hematite host				Rutile
	Magnetite	Ilmeno-magnetite	Ilmenite	Hemo-ilmenite	Titan-hematite	Ilmeno-hematite	Ilmeno-rutilo-hematite	Rutilo-hematite	
Hornblende-microperthite granite and alaskite, or equivalent gneisses:									
A. Fe_2O_3 greater than or somewhat less than FeO		X	X						x
B. FeO much in excess of Fe_2O_3; rock commonly contains a ferrohastingsite hornblende		x	X						x
K_2O-rich hornblende-microperthite and microperthitic microcline granite transitional to granitized gneiss, and some slightly granitized metasedimentary rock		X	x [1]	x [1]		x			
Hornblende-microcline-plagioclase granite gneiss, recrystallized in lower temperature range of amphibolite facies with development of sphene	X								
Sillimanitic and biotitic microcline granite gneiss, formed by granitization, with varied degrees of oxidation (A, B, C, and D) of oxides:									
A. Moderately oxidized		X				X			
B. Strongly oxidized					X [1]	X [1]			
C. Very strongly oxidized							X		
D. Most intensely oxidized								X	X
Migmatized, granitized metasedimentary pyroxene gneiss:									
A. Several oxides		X		X		X			
B. One oxide only					X	X		X	
Garnet-biotite-quartz-plagioclase-microcline gneiss (slightly granitized), and pyroxene-plagioclase granulite				X [2]					
Phacoidal granite gneiss, Stark complex:									
A. One oxide			X						
B. Two oxides		X	X						
Pyroxene syenite gneiss and pyroxene-quartz syenite gneiss		X	X						x
FeO-rich pyroxene-quartz syenite gneiss and fayalite-ferrohedenbergite granite, Tupper complex		x	X						x
Amphibolite:									
A. Pyroxenic or pyroxenic-amphibolitic metagabbro gneiss		X	X [1]	X [1]					
B. Amphibolite			X						
Anorthosite and gabbroic anorthosite gneiss		x		X		X			X
Pyroxenitic melanocratic segregations in anorthositic rocks		X [3]	X [3]	X [3]					
Pyroxene diorite and pyroxene diorite gneiss		x	X						

[1] Either mineral present.
[2] A little magnetite and ilmenohematite commonly accompanies the hemoilmenite.
[3] Either ilmenite, hemoilmenite, or ilmenomagnetite may be the predominant mineral.

almost always associated with some magnetite grains that do not show intergrown ilmenite. In part, this may be due to the angle at which the grain is cut, but in part there seems to be an inhomogeneous distribution of ilmenite among the separate grains.

In respect to accessory oxides, four major facies of hornblende-microperthite granite and its equivalent gneiss are recognized (see table 31). One facies has ilmenomagnetite and ilmenite as the predominant oxides, and another facies has ilmenite as the only or nearly the only oxide. The ratio FeO : Fe_2O_3 is higher in the facies containing ilmenite only, and the ratio Fe_2O_3 : FeO is higher where both ilmenomagnetite and ilmenite occur, as shown in table 32. In the rocks having relatively high FeO, the hornblende is able to take into its structure most of the Fe_2O_3, whereas magnetite forms in the rocks having relatively high Fe_2O_3.

In several large belts of granite, ilmenite is the predominant or only oxide; these belts give rise to mild negative aeromagnetic anomalies. One of the belts extends from the east-central part of the southeast rectangle of the Stark quadrangle northeastward for about 18 miles through Stark Falls to the north part of the northwest rectangle of the Childwold quadrangle, and then southward along the east border of the north-central rectangle of the Childwold quadrangle. The entire belt is about 25 miles long. This belt of ilmenitic hornblende granite is continuous with a branch belt that extends across the Stark quadrangle southwestward from the west side of the Brunner Hill magnetite deposit, through Sellecks Lower Camp and Parmeter Pond to Rainbow Falls. The belts are about

TABLE 32.—Fe_2O_3, FeO, *and* TiO_2 *content of major facies of hornblende-microperthite granite and equivalent gneiss, St. Lawrence County magnetite district*

Rock type	Weight percent in rock		
	Fe_2O_3	FeO	TiO_2
1 Hornblende-microperthite granite with ilmenomagnetite and ilmenite	1.60	1.40	0.37
2 Hornblende-microperthite granite with ilmenite predominant (commonly a ferrohastingsite facies)	1.32	1.93	.32
3 Ferrohastingsite-microperthitic microcline-oligoclase granite gneiss with ilmenite only	1.59	4.17	.53
4 Hornblende-microperthitic microcline-oligoclase granite gneiss with ilmenomagnetite predominant over ilmenite	2.43	3.29	1.24
5 Hornblende-microcline-plagioclase granite gneiss with sphene and magnetite only	3.19	1.96	.60

¾–1¾ miles wide. Ilmenitic hornblende-microperthite granite forms most of the granite area in the southern part of the Cranberry Lake quadrangle, south of the Bog River syncline of metasedimentary rocks. Another mass of ilmenitic hornblende-microperthite granite forms a north-south belt just west of Briggs, central rectangle, Oswegatchie quadrangle.

A third facies, richer than normal in K_2O, is in part a microcline-rich hornblende-microperthite-microcline granite gneiss associated with included lenses and schlieren of amphibolite. The oxides of this type of rock are ilmenomagnetite and ilmenite, or ilmenomagnetite and a little associated hemoilmenite and ilmenohematite. The oxides of this facies are thus transitional to a more oxidized suite of accessory minerals, found in microcline-rich gneisses representing granitized metasedimentary rocks.

The fourth facies of the granitic rocks is the hornblende-microcline-plagioclase granite gneiss. This rock is the product of metamorphism and recrystallization of a hornblende-microperthite granite. No ilmenite or rutile is present; magnetite is the only oxide. Most of the titanium has reacted to yield sphene. The original ilmenomagnetite has been recrystallized to a magnetite relatively poor in titanium. This type of magnetite- and sphene-bearing hornblende granite gneiss forms the belt through Stone School, northeast rectangle, Russell quadrangle.

The phacoidal hornblende granite gneisses of the Stark complex have two facies, so far as the accessory oxides are concerned. In one facies, ilmenomagnetite and ilmenite are the predominant oxides. In the other facies, ilmenite is the predominant oxide. In the ilmenomagnetite, the intergrown ilmenite is in two sizes—coarse lamellae and a fine lattice.

The normal pyroxene syenite and pyroxene-quartz syenite gneisses contain ilmenomagnetite, ilmenite, and minor rutile. However, local facies of the Tupper complex are exceptionally rich in FeO relative to Fe_2O_3 and MgO. These facies include eulite-quartz syenite gneiss, with or without ferrohastingsite; and fayalite-ferrohedenbergite granite. The oxides of these rocks are ilmenite, ilmenomagnetite, and minor rutile. Ilmenite is the predominant or, locally, the only oxide.

The pyroxene gabbros and pyroxene-hornblende metagabbros (with or without garnet) contain ilmenomagnetite and ilmenite, or ilmenomagnetite and hemoilmenite. Where the rocks have been wholly reconstituted into hornblende-plagioclase amphibolite, magnetite and any Fe_2O_3 are largely or wholly taken up by the hornblende, and the only oxide preserved is ilmenite. However, ilmenomagnetite and ilmenite may occur in migmatitic amphibolite, in association with granitic veinings.

The oxides of the anorthosite and gabbroic anorthosite gneisses are characteristically more oxidized than those of any other igneous rocks or equivalent gneisses in the district. The oxides are hemoilmenite, ilmenohematite, and rutile, with a subordinate to minor amount of ilmenomagnetite. Some melanocratic pyroxenitic segregation layers within the anorthositic rocks contain ilmenite or hemoilmenite as the predominant or only oxide, whereas other layers contain ilmenomagnetite as a major to subordinate mineral. The pyroxene diorite and pyroxene diorite gneiss of the Tupper Lake quadrangle in part contains ilmenite as the only oxide, and in part both ilmenite and ilmenomagnetite.

The biotitic and sillimanitic microcline granite gneisses are the product of granitization of biotite-quartz-plagioclase gneiss. There are all gradations in intensity of oxidation of the iron, from a biotite-quartz-plagioclase gneiss in which the iron is wholly in the biotite, to an intensely granitized facies in which the only oxides are rutilohematite and rutile. (See table 33.)

TABLE 33.—*Variation of minerals with intensity of oxidation during granitization of biotite-quartz-plagioclase gneiss, northwest Adirondacks*

Rock type	Weight percent			Minerals
	FeO	Fe_2O_3	TiO_2	
Average composition of biotite-quartz-plagioclase gneiss.[1]	4.12	1.31	0.32	Biotite.
A. Sillimanite-microcline granite gneiss, moderately oxidized.[2]	.92	3.99	.54	Ilmenomagnetite, ilmenohematite.
B. Sillimanite-microcline granite gneiss, strongly oxidized.[2]	.45	4.64	.76	Titanhematite with slight ilmenite intergrowth.
C. Sillimanite-microcline granite gneiss, most intensely oxidized.[2]	.07	4.86	.68	Rutilohematite, rutile.

[1] Analysis from Engel and Engel, 1953, p. 1063.
[2] Analyst, Angelina Vlisidis.

NOTE: The analyses of A, B, and C are of the oxide minerals only. A little biotite or chlorite, where present in these rocks, would increase the percent iron and titanium slightly for the rock as a whole. A and B are averages of 12 and 16 separate samples.

The assemblage rutilohematite, titanhematite, and rutile has been found exclusively in the sillimanite-microcline granite gneiss of a narrow belt between L Pond and the trail to Cranberry Pond, and in biotite-microcline granite gneiss 0.7 mile southeast of Lem Pond, both in the Stark quadrangle.

The more or less migmatized and granitized metasedimentary pyroxene gneisses may also have moderately oxidized assemblages such as ilmenomagnetite, hemoilmenite, and ilmenohematite; or they may have a single strongly oxidized mineral such as ilmenohematite, titanhematite, or rutilohematite.

The ilmenohematite varies from hematite grains having only the slightest trace of ultrafine microscopic ilmenite intergrowths to composite grains of broadly interlayered ilmenohematite and hemoilmenite. The size of the ilmenite in the intergrowths in hematite varies as the abundance of the ilmenite. Where the amount of ilmenite is slight, the ilmenite is in ultrafine stringlets; where the amount is moderate, the ilmenite occurs in two generations—as strings and stringlets; where ilmenite is abundant, the intergrowths are in three generations and sizes—as broad lamellae or plumes, strings, and stringlets.

Rutile is present very sparingly in most of the granitic and syenitic rocks as a primary mineral associated with ilmenomagnetite and ilmenite. In the anorthositic and gabbroic anorthositic rocks, rutile is a primary mineral of a more oxidized primary assemblage that includes hemoilmenite and ilmenohematite. In the microcline granite gneisses of granitization origin, rutile appears only in the more intensely oxidized assemblages. As the percentage of ilmenite decreases, the percentage of hematite and rutile increases, as though the latter two formed in place of ilmenite.

Magnetite very low in titanium and without intergrown ilmenite, occurring as the only oxide mineral, is restricted almost entirely to the skarn deposits, to a small part of the vein concentrations in migmatitic and granitized gneiss, and to the sphene-bearing hornblende-microcline-plagioclase gneiss. Very locally it occurs as a disseminated product introduced into the country rock by late-stage solutions.

Subtitanhematite containing 1.5–5 percent TiO_2 in solid solution occurs exclusively in certain ore vein replacements in sillimanite-microcline granite gneisses in the Benson Mines area, and as replacement veins of disseminated type in similar rock elsewhere, as, locally, at the Parish deposit.

Hemoilmenite has been found as the sole oxide in only one specimen, an amphibolite. In general, hemoilmenite is restricted to only slightly granitized metasedimentary rock such as biotite-garnet-quartz-plagioclase-microcline granite gneiss, to metasedimentary pyroxene-plagioclase granulite, to gabbro and metagabbro gneiss, to anorthosite and gabbroic anorthosite, to amphibolite in contact with granitic material, and to K_2O-rich hornblende-microcline granite and granite gneiss.

Locally, the magnetite minerals are partly altered along their parting planes to hematite (variety martite), but this occurs in substantial amounts only locally. The ilmenite is also partly altered in places to a very fine grained secondary aggregate which in part consists of rutile (or anatase), hematite, and relict ilmenite. Both types of alteration are, for the most part, of hydrothermal origin. Martite occurs largely in the biotitic and sillimanitic microcline granite gneisses, in association with ilmenorutilohematite where oxidation has been intense. Martite was found in only one area of normal hornblende granite gneiss, southeast of Catherineville School, east-central rectangle, Potsdam quadrangle.

The magnetite, ilmenomagnetite, ilmenohematite, ilmenorutilohematite, titanhematite, rutilohematite, and rutile almost everywhere occur as independent grains coordinate with the silicates, or as composite grains of two of the minerals. In the latter case, the junction is always a plane surface. The only evidence of alteration of one mineral to another is the partial change of magnetite to martite, or of ilmenite to a fine secondary aggregate, as described above.

CORRELATION WITH AEROMAGNETIC ANOMALIES

The following empirical relations between magnetic anomalies and mineralogy of the accessory oxides have been found to hold for rocks of the northwest Adirondacks (Balsley, Buddington, and Fahey, 1952; Balsley and Buddington, 1954 and 1958). Magnetic anomalies given by ore deposits are excluded from consideration here.

All *negative* magnetic anomalies are correlated with one of the following assemblages of oxides:

1. The oxides are a member or members of the hematite-titanhematite-ilmenohematite-rutilohematite series containing 3–18 percent TiO_2. Rocks with these minerals may give rise to *relatively intense* negative anomalies.
2. Both magnetite or ilmenomagnetite and members of the titanhematite-ilmenohematite series are present, but the magnetite or ilmenomagnetite is subordinate. Rocks with this assemblage may yield *mild* negative anomalies.
3. Both ilmenomagnetite and ilmenite (containing more than 6 percent Fe_2O_3 in solid solution) are present, but ilmenite forms more than one-third of the oxides. Rocks with this assemblage often give rise to *mild* negative anomalies. Where ilmenite is present as the only oxide in granite or granite gneisses, the rocks give rise to a negative magnetic anomaly. Where ilmenite is the only oxide in amphibolite, there may be either a very low negative or a very low positive magnetic anomaly.
4. Ilmenohematite and hemoilmenite occur together as the sole oxides only locally. They yield a negative magnetic anomaly.

Magnetite or ilmenomagnetite, occurring alone or forming more than two-thirds of the oxides, always gives rise to *positive* magnetic anomalies.

TITANIFEROUS MAGNETITE AS A GEOLOGIC THERMOMETER

The percent of TiO_2 in the magnetite of many of the Adirondack rocks has been determined by chemical analysis and the significance discussed by Buddington, Fahey, and Vlisidis (1955), and Buddington (1956). A summary of their work is given here (table 34).

TABLE 34.—TiO_2 *content of accessory magnetite in certain rock types of the northwest Adirondacks*

	Percent TiO_2 in magnetite
Igneous rocks:	
Pyroxene-microperthite granite	9.6–8
Hornblende-microperthite granite	6.7–5
Microperthite alaskite	4.3–3
Microperthitic microcline alaskite	3.8–3.1
Granite pegmatite	5.7–1.0
Ilmenite-magnetite-apatite-pyrite veins (segregations) in alaskite. (Magnetite here is not accessory.)	4.3–3.9
Metasomatic rocks:	
K_2O-rich microcline granite gneiss formed by granitization	3.1–2.2
Metamorphic rocks:	
Gneisses reconstituted in granulite facies and upper temperature range of amphibolite facies	4.1–3.1
Gneisses reconstituted in middle and lower temperature range of amphibolite facies	2 –0.65

Where ilmenite is present as separate, discrete grains and the TiO_2 in magnetite is present as ilmenite, the percent of TiO_2 in the magnetite may be inferred to vary systematically with the temperature—the higher the temperature, the higher the TiO_2. The percent of TiO_2 in magnetite of the granites is similar to that in magnetite phenocrysts from dacite and rhyolite. The magnetite of the granites is therefore assumed to have formed at magmatic temperatures, perhaps between 700° and 800°C.

The garnetiferous facies of the pyroxene syenite and pyroxene granite gneisses are inferred to have formed by metamorphism at temperatures between 600° and 700° C, and the gneisses of the amphibolite facies between 500° and 600°C.

The alaskite and granite pegmatite could, in consequence of their inferred content of hyperfusibles, have formed from magma whose temperature range was similar to that of the garnetiferous gneisses—600° to 700°C.

Microcline granite gneisses of metasomatic origin are inferred to have formed at temperatures comparable to those of granite pegmatite and of rocks reconstituted in the intermediate and upper ranges of the amphibolite facies.

A variety of hornblende-microcline granite gneiss that occurs in the Childwold quadrangle was earlier thought to be in substantial part metasomatic. However, the TiO_2 content of its accessory magnetite is similar to that of magnetites from microperthite alaskite, and the temperature of formation of this hornblende-microcline granite gneiss is therefore inferred to be similar.

PETROLOGY

EMPLACEMENT: MAGMATIC INTRUSION OR GRANITIZATION?

The discussions of the Precambrian rocks have been based on the assumption that the members of the quartz syenite series and the granite series, with the possible exception of a large part of the microcline granite gneisses, were all derived from magma intruded from depth.

Rocks similar to those under consideration, however, have elsewhere in many cases been interpreted as a product of granitization of metasedimentary rocks or locally of other rocks through the agency of migrating ions or atoms, or the flow of solutions (hydrothermal or pneumatolytic). On this latter hypothesis, any magma involved is dependent upon the granitizing materials and develops in place consequent upon their activity. There is little doubt that many geologists would likewise interpret the quartz syenite complexes and the granitic rocks of the Adirondacks as syenitized or granitized metasedimentary rocks, as the case may be. Indeed the layered structure and variation in composition of the quartz syenite complex led the early geologists in this district to interpret the rocks as metamorphosed sediments. The sillimanitic character of one facies of the microcline granite gneisses immediately suggests the possibility that this gneiss represents a granitized or reconstituted aluminous metasedimentary rock. The large and small schlieren or ghosts of metasedimentary rocks, which occur in many places in the quartz syenite complexes and in certain of the granite gneisses, also raise the question as to whether all the gneisses are not actually modified metasedimentary rocks. Finally the fact that for most practical purposes the planar structure can be used as though it were bedding planes in working out structure and in exploration programs for ore could be taken to be consistent with a hypothesis that it actually does represent a planar structure inherited from metasedimentary rocks which have been granitized. It therefore seems necessary to present the evidence for what is believed to be the magmatic origin of most of these rocks. The discussion to a considerable extent follows that of a previous publication (Buddington, 1948).

One postulated mechanism whereby granitization might be effective is a diffusion of atoms or ions through solid rock. (See Backlund, 1938, p. 339–396; Ramberg, 1944, p. 98–111; 1945, p. 307–326; Bugge, 1946, p. 5–59). There are several arguments against this as it would apply in the Adirondacks. Much of the rock in the border facies of the anorthositic massif shows a distinct difference between the composition of the larger crystals and those of the smaller (Buddington, 1939, p. 31

and 104; Stephenson, 1945, p. 20–21). Similarly the cores of the larger plagioclases in the pyroxene syenites and quartz syenites are more calcic than their outer portion or than the plagioclase in the groundmass. If diffusion within solid rock was so free that thousands of cubic miles of metasedimentary rocks could be granitized, it is difficult to understand why the plagioclases were not brought into equilibrium, since interdiffusion to the extent of only millimeters would be involved.

It is also a question how such vast quantities of particles could have passed through the widespread quartz syenite sheets to granitize the overlying metasedimentary rocks without modifying or destroying such a characteristic feature of the quartz syenite sheets as the superposed layers of systematically varied composition.

Evidence for the emplacement of a thin granite sheet of the Stark complex by mechanical displacement of the walls is clearly shown in the roof rocks of the Clifton mine. The wall rocks of the granite sheet consist of thin-layered metasedimentary rocks with two well-defined pyroxene skarn layers in light-colored quartz-feldspar gneiss that can be used as horizon markers. Where the granite sheet is only 0.5 foot thick, the two pyroxene skarn layers are 4 feet apart; but where the granite widens to 4.5 feet in thickness, about 40 feet along the strike the skarn layers are spread to 8 feet apart.

The zircons of all the rocks of the Stark, Diana, and Tupper complexes and of all the younger granitic rocks are characteristically euhedral. A reconnaissance inspection also suggests that for a major part of the zircons the ratio of length to breadth is generally greater than 2. Both these criteria are cited by Poldervaart and von Backström (1949, p. 433) as characteristic for rocks of magmatic origin.

MAGMATIC ORIGIN OF GABBRO AND DIORITE

Rocks of gabbroic composition and texture are generally accepted as of magmatic origin. Within the main igneous complex the gabbro and diorite bodies occur almost exclusively as long, narrow layers or lenses that are included in younger granitic masses. Their relations to older rocks are therefore not generally evident. Within the metasedimentary rocks of the Grenville lowlands, however, the gabbro, metagabbro, and equivalent amphibolite bodies occur as roughly conformable sheets ranging in size from small lenses to bodies a score of miles long. They do, however, locally crosscut the bedding of the metasedimentary rocks. Their habit is the same as that within the main igneous complex, and a similar mode of emplacement is assumed. Conformable sheets and lenses of diorite are also found within the metasedimentary rocks. The form of the diorite sheets within the main igneous complex is consistent with the hypothesis that they were similarly emplaced as roughly conformable sills within the metasedimentary rocks.

MAGMATIC ORIGIN OF QUARTZ SYENITE SERIES

The clearest evidence for at least the local magmatic nature of the quartz syenite gneiss is the occurrence of a conspicuous igneous breccia of quartz syenite, quartz-feldspar granulite, and metasedimentary pyroxene gneiss, which extends in a belt for 20 miles from just southwest of Harrisville to about 3 miles west of South Russell. Commonly, the included blocks in this breccia are relatively thin tabular layers several feet to several hundred or several thousand feet long; angular fragments as small as a few inches in size are also found. The belt of breccia is one-half mile or more in width. The nature of the igneous breccia is especially well shown for several miles north and south of Portaferry Lake (Oswegatchie quadrangle). In the drill cores of the Clifton mine the coarse phacoidal hornblende granite gneiss also alternates with included layers of fine-grained quartz-feldspar granulite. The contact between the two rocks is knife-edge sharp. There is no evidence of gradation. Syenite sills with sharp contacts occur in both crystalline limestone and gneisses northwest of the outer border of the quartz syenite complex. They are characteristically of finer grain than the main mass.

There are also many mafic schlieren here and there throughout the quartz syenite mass, which are interpreted as relics of modified metasedimentary rocks.

There is in general no correlation between the nature of the igneous rock and the nature of its inclusions of country rock. Inclusions of pyroxene skarn and mixed rock arising from disintegration of pyroxene skarn occur indiscriminately in mafic pyroxene syenite, pyroxene-quartz syenite, and hornblende granite. Similarly, layers of quartz-feldspar granulite occur in the same igneous rocks.

There are many examples where the syenite bodies contain inclusions of rocks that must have been brought in from a distance. Travel by magma would be the simplest explanation.

Within the anorthosite massif, many dikes of composition ranging from mafic syenite to quartz syenite cut the anorthosite and its planar structure. Occasionally these dikes contain angular fragments of anorthosite. One outstanding example, interpretable only as a magma intrusion, is the polymict breccia dike of Jenkins Mountain (Buddington, 1939, p. 123–124). This dike is 150–200 feet wide and more than 2 miles long. It contains an abundance of fragments that are in part inclusions from the anorthosite wall rock, though most are fragments of metasedimentary rocks that have not been observed within a mile of the dike. The evidence therefore indicates that the syenite must have been derived from a magma that must have travelled a substantial distance, in order to have brought the fragments of foreign metasedimentary rocks into the anorthosite massif.

There is also evidence, already described, that the Diana, Stark, and Tupper complexes locally contain xenocrysts of labradorite, which were probably carried from a distance as there is no anorthosite in place locally.

The feldspar crystals of the pyroxene syenite and pyroxene-quartz syenite are commonly zoned, having plagioclase cores and microperthite borders. This, again, is a feature consistent with crystallization from a magma.

It has also been shown that the thick Diana sheet of the quartz syenite series is layered in such a fashion that the rocks range asymmetrically from mafic syenite to quartz syenite, independent of the kind of country rock, and that when the complex structure is taken into account the arrangement is such as to be consistent with the hypothesis that we are dealing with a thick gravity-differentiated sheet which has been closely folded. The more felsic facies is at or near the top of the sheet, and the more mafic are at or near the base. Such a type of variation is consistent with what would be expected and is known of differentiation in sheets of magma. It has also been shown that the degree of diversity and contrast in the nature of the differentiated layers varies with the thickness of the sheet, a result again consistent with and to be expected in a gravity-differentiated sheet of magma. The border (basal) facies of the Diana complex has a chemical composition approximately equivalent to the average composition of the sheet as a whole and may represent the relatively undifferentiated magma.

The local concentration of ilmenite with pyroxene in the feldspathic ultramafic gneiss and shonkinite layers of the Diana complex is also consistent with explanation as a gravity concentrate of early crystals from a magma. The combination of much ilmenite with pyroxene has not been demonstrated anywhere to have formed from sediments. Zircon is also locally concentrated with the pyroxene and iron oxides, and this is again consistent with gravity accumulation.

There is locally a variation in the Diana complex such that layers consisting almost wholly of feldspar crystal aggregates alternate with similar layers carrying a little more mafic material. There is locally also a similar alternation between feldspathic ultramafic gneiss and shonkinite or syenite layers. Such rhythmic layering is known to be characteristic of stratiform noritic sheets formed from magma.

MAGMATIC ORIGIN OF HORNBLENDE-MICROPERTHITE GRANITE

The structural relation of the hornblende-microperthite granite and equivalent granite gneiss to the older rocks is predominantly one of conformity, but evidence of magmatic intrusion is found in local and regional transgressive relations.

The borders of the main masses of hornblende granite gneiss are for long distances distinctly transgressive across the structures of the metasedimentary rocks and the quartz syenite complexes, as has been previously described (Buddington, 1939, p. 49–152). Locally, though rarely, the granite has developed an igneous breccia or agmatite with angular inclusions of the country rock. One such breccia of granite with angular inclusions of metagabbro is well shown in the area a little west of a point 1½ miles south of Brandon (Malone quadrangle). At another locality (Buddington, 1939, p. 212) the hornblende granite contains angular inclusions of anorthosite, and at several localities it has been observed to have xenocrysts of labradorite of a character identical with that of the anorthosite. At none of these localities has anorthosite been observed in place, and it is thought that the inclusions and xenocrysts have been carried substantial distances by magmatic movement (Buddington, 1939, p. 149–152).

The evidence thus leads to the interpretation that the metasedimentary rocks, together with the quartz syenite sheets, were folded and deformed before the intrusion of the hornblende granite. The hornblende granite has been shown locally to cut across the folded structures, yet in the main the dominating feature is a conformity between the internal foliation of the granite and its surfaces of contact with the country rock. It is thought that the granite magma entered the folded structures during a period of deformation, in the form of sheets and phacoliths more or less conformable with the structure of the country rocks. The conformability of the structure of the country rock with that of the granite, however, has certainly been accentuated in varied degree by strong deformation of the whole complex subsequent to the emplacement of the granite magma. The granite may originally have shown much more extensive crosscutting relations than is now clearly observed.

The essentially phacolithic sheetlike structure of the hornblende granite is well shown on the flanks of the Stark anticline, where the older phacoidal granite gneiss of the core of the anticline forms a floor for the younger granite sheet. The roof is formed by the metasedimentary rocks and associated biotite granite gneiss sheets of the South Russell and Clare-Clifton-Colton synclinoria, to the northwest and southeast respectively, of the Stark anticline. The wide belt of granite between the Jarvis Bridge and Brandy Brook synclines on the north and the Bog River syncline on the south is interpreted as essentially a complexly deformed compound sheet, with the metasediments of the synclines forming the roof and the main floor not exposed. The quartz syenitic rocks are interpreted as an older sheet of igneous rock with anticlinal structure, which is included as a bent plate within a great phacolithic granite sheet. There has, however, certainly been extensive crosscutting of the structural units involving the metasedimentary rocks, as well as of the quartz syenite complex which has been previously referred to. The rocks of the Darning Needle syncline are very different from those of the Dead Creek syncline and must represent different stratigraphic horizons.

The great thickening in width of the hornblende granite gneiss across the Russell quadrangle is much more reasonably interpreted as the product of intrusive magma bowing out the metasedimentary rocks than as the result of granitization of sediments, even if some plastic flow of the latter is postulated.

It is possible that some of the largest granite masses having domical foliation on the core and steep dips on the flanks, such as that in the extreme southern parts of the Oswegatchie and Cranberry Lake quadrangles, may be of such thickness as to be of batholithic or diapiric rather than phacolithic nature.

Clean-cut hornblende granite dikes have been observed within each of the older igneous rocks, such as the anorthosite and metagabbro, and each of the members of the quartz syenite series.

Dikes of hornblende granite are common across the structure of the phacoidal hornblende granite gneiss of the Diana complex at Jayville and south of there. Intrusive sheets of hornblende granite have been found in the syenite gneiss of the Tupper complex.

The hornblende granite has an extremely limited range of variation in composition; probably it could not have been formed by replacement and modification of metasedimentary rocks of such highly varied character through diffusion processes in the solid state. The hornblende granite east of Bryants Bridge (Oswegatchie quadrangle) occurs over areas of several square miles in uniform character, with no inclusions and containing only a rare pegmatite vein.

Where least metamorphosed, the feldspars of the hornblende granite occur almost exclusively as microperthite. Tuttle (1952) has emphasized that if the microperthite is interpreted as an exsolution phenomenon, as seems most reasonable, the temperature at which the original homogeneous feldspar formed must have been so high (at least 660° C) that it was near the range of magmatic temperatures for granite. The nature of the ilmenomagnetite in the granite is also consistent with a magmatic origin, for the amount of ilmenite in the ilmenomagnetite (10–12 percent) is similar to that of the ilmenomagnetite of rhyolite lava flows and much greater than that normal for high-temperature vein deposits (Buddington, Fahey, and Vlisidis, 1953).

DIVERSE ORIGIN OF HERMON GRANITE GNEISS

The Hermon granite gneiss has been ascribed (Buddington, 1939, p. 160–161) in part to permeation and modification of Grenville aluminosiliceous strata by solutions, followed by the development of porphyroblastic microcline; in part to processes attendant upon intimate lit-par-lit intrusion of magma into Grenville gneiss, granulite, and amphibolite; and in part to porphyritic crystallization of intruded magma.

The origin of the Hermon granite gneiss has lately been restudied in detail by Prucha.[4] A summary of his conclusions follows: In belts of mixed Hermon granite gneiss and amphibolite, the latter has largely been mechanically injected by the former along planes of foliation. Contacts between the amphibolite layers and granite gneiss layers remain relatively sharp. The Hermon granite gneiss has sharp contacts with marble where intruded in it. By contrast, in mixed gneiss consisting of layers of Hermon granite gneiss and biotite-quartz-plagioclase gneiss, contacts are gradational and the latter is more or less granitized and thoroughly permeated by granitic and pegmatitic material. The introduction of granitic material into the gneiss took place, not along well-defined, well-spaced structural planes—as in the case of granite gneiss-amphibolite association—but rather by a gradual intimate permeation along ill-defined, closely spaced foliation planes.

Locally, however, the amphibolite has been granitized in much the same way as the biotite-quartz-plagioclase gneiss.

A recent detailed study of the problem has been made by Engel and Engel (1953, p. 1065–1078). They also conclude that the "Hermon type" of granite gneiss is in part a hybrid rock evolved by interaction of granitic components and gneisses, and that the processes of injection and granitization involve both magma and

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

fluids from magma, which permeated the gneiss, interacting with it and replacing it. The Engels stressed a somewhat more widespread granitization of amphibolite than is suggested by Prucha.

The Hermon granite gneiss is thus in part interpreted as the product of injected magma, in part as a product of injection and granitization of biotite-quartz-feldspar gneiss and, locally, of amphibolite.

MAGMATIC ORIGIN OF ALASKITE

The alaskite of the main igneous complex has its major development as a facies of the hornblende granite masses—at or near their roof, and as sheets within the included belts of metasedimentary rocks. In a discussion of the age relations of the igneous rocks, the occurrence of alaskite dikes at numerous localities has also been cited. No evidence has been noted that these dikes were emplaced otherwise than as intrusive masses.

The hornblende granite has been proved to grade locally into a fluorite alaskite on the noses of anticlines, and at or near the roof of the granite masses. A similar relation is known in the case of several examples of unquestioned magma intrusions (Daly, 1914; van Biljon, 1940; Söhnge, 1945; and Strauss and Truter, 1945).

It is noteworthy that hornblende granite similar to the normal hornblende granite of the main igneous complex occurs only meagerly within the metasedimentary rocks of the Grenville lowlands. The granite masses within this broad belt of metasedimentary rocks are predominantly an alaskite called the Alexandria granite, and a coarse biotitic, locally hornblendic, augen gneiss called the Hermon granite gneiss (Buddington, 1929). The latter also has a medium-grained facies.

The alaskite masses occur in large part as phacoliths on the crest of anticlinal folds in marble. They are usually a few miles long but in some places are as much as 15 miles. Thirteen such phacoliths have been found in the marble, and in all but one a characteristic and peculiar metamorphic aureole of amphibolite and almandite-quartz-feldspar gneiss has been developed at least locally in the contact zone. These have been previously described (Buddington, 1929). The development of these aureoles, the common occurrence of tourmaline veins in the alaskite, and the presence of fluorite in the alaskite of the main igneous complex have together been considered to indicate that the alaskite magma was highly charged with volatiles and, therefore, highly mobile. A high mobility would give a reasonable explanation why alaskite magma and not the normal hornblende granite magma found its way into the metasedimentary rocks.

The emplacement of the alaskite masses into potential low-pressure areas within the folds of the metasedimentary rocks during deformation is well shown on the Potsdam and Russell quadrangles. Several alaskite masses were intruded into the marble in the great bend in structure southeast of High Flats. The crescent-shaped body of alaskite gneiss of Hawk Ledge is in effect a phacolith contained in a synclinal cross fold within the metasedimentary rocks. The alaskite gneiss mass west of Colton has a domical structure and is interpreted as an anticlinal phacolith, as is the mass from 1 to 3 miles northeast of Stellaville. The alaskite gneiss west of Stalbird occupies the crest of a cross fold at the contact between marble and migmatite gneiss, where there has been slippage of the migmatite gneiss relative to the marble.

The hypothesis that the alaskite masses are replacements of thickened and crumpled metasedimentary rocks in the crests (or troughs) of folds must be considered. However, most of the alaskite bodies are in marble; and most of the alaskite phacoliths contain amphibolite layers which, with a few exceptions, are interpreted as metasomatized marble layers. The development of the granite might then be assumed to have proceeded by the development of amphibolite as a first stage, and then by granitization of the amphibolite as a second stage. Most of the contacts of granite and amphibolite, however, are quite sharp, though the amphibolite layers themselves may be modified by impregnation with quartz and alkalic feldspars. The relation is as though the amphibolite was mechanically intruded and broken up by an alaskite magma, rather than as though alternate layers of amphibolite were completely replaced with little or no effect on the intervening layers. It might also be predicated that it was biotite-quartz-feldspar gneiss layers in the marble that were transformed into alaskite. It can only be said that no evidence of this has been found.

ORIGIN OF MICROCLINE GRANITE GNEISSES

All the microcline granite gneisses show essentially the same granulose structure throughout the area. In part, they also show many phenomena that—in contrast to those of the hornblende granite and alaskite—can be cited as evidence of origin by granitization of metasedimentary rocks. Granitic rocks which are thought to be the product of replacement, modification, or recrystallization of metasedimentary rocks, and which have inherited a distinct gneissic or foliated structure, will here be called gneiss. There is then the very difficult problem of trying to distinguish between (a) a contaminated granite that has a primary gneissoid flow structure, (b) a granite gneiss that is the secondary product of deformation, recrystallization, and plastic flowage of a contaminated granite, and (c) a granite gneiss that was derived in part from metasedimentary rocks by processes of granitization with inheritance of a layered or foliated structure, by migmatitic injection, or by some combination of these processes.

If we assume that at least part of the microcline-bearing granitic rocks are gneisses due to granitization of metasedimentary rocks, there is then the additional problem whether the granitization was effected largely by recrystallization, by emanations accompanying or preceding the hornblende granite or alaskite magma, by a volatile-rich microcline granite magma itself, or by emanations independent of any related source magma and having their source at great depths. There is the further problem whether the emanations are hydrothermal solutions, pneumatolytic solutions, migrating ions, or migrating atoms. There is also the problem whether such emanations have generated a secondary granitic magma from the metasedimentary rocks.

The general heterogeneity of the microcline granite gneiss sheets; their general conformability with the associated metasedimentary layers; the great variability in the nature of the varietal minerals, systematically, in accord with the nature of the associated metasedimentary materials; and the chemical composition of the gneiss—all would permit it to be reasonably interpreted as a metasedimentary rock, a reconstituted member of the Grenville series. Indeed it has not been really conclusively proved that this is not the case.

There are, however, no comparable rocks known within the great thick and varied series of metasedimentary rocks of the Grenville lowlands. On the contrary, the microcline granite gneisses are almost wholly confined to the area of the dominantly igneous complex. The percentage of quartz in the gneiss in general has a more restricted and lower range than might be expected for a thick sedimentary bed. There are gradations between enclosed schlieren and layers of biotite-quartz-plagioclase gneiss and the sillimanite-microcline granite gneiss similar to the gradations one would expect if metasomatism were involved. The almost universal presence of rutiloilmenohematite or ilmenohematite in the gneiss, and the occurrence of hemoilmenite in the hornblende- and some of the pyroxene-microcline granite gneisses, in contrast to their absence in normal metasedimentary rocks, is reasonably interpreted as the effect of the activity of high-temperature (perhaps $\pm 350°$–$600°C$) oxidizing fluids. The percent TiO_2 dissolved in the accompanying magnetite is consistent with such a high temperature. And finally there is evidence of local tourmalinization of adjoining country rock and local introduction of tourmaline pegmatite veinlets in the gneiss. All the foregoing features individually and collectively are more consistent with the hypothesis that high-temperature fluids have been active and have introduced and removed substantial quantities of material than with a hypothesis that reconstitution of a sediment alone is involved.

MICROCLINE GRANITE GNEISS—MAGMATIC PART

The possibility that a volatile-rich microcline granite magma was intruded from below and either directly or through the activity of emanations produced much granitization of the metasedimentary rocks yielding gneisses, will be discussed first.

In general, considering the granites of the world, granite as rich in microcline (or potassic feldspar) as that under discussion forms a relatively small percentage of the total. Furthermore, the microcline-rich granite gneisses do not approximate the cotectic ratio for orthoclase-albite-anorthite, such as is indicated by existent experimental data and as would be expected of a late magmatic residuum. In the light of our present knowledge of the possible theories for the origin of variations in the composition of granite, such K_2O-rich granites demand a special explanation.

Vogt (1931) has shown that the percentage of normative orthoclase in the total normative feldspar in most granitic rocks is between 30 and 50 percent, and in most granite pegmatites between 50 and 70 percent. Many pegmatites have a normative ratio of 60–70 percent orthoclase to total feldspar. Five analyses of the microcline granite gneisses of this area show that the ratio of normative orthoclase to total feldspar ranges between 63 and 74 and averages 69. Vogt has suggested (1931, p. 240) that the processes which give rise to K_2O-rich granite pegmatite on a small scale, may under certain conditions give rise on a large scale to K_2O-rich granites. One current hypothesis as to the origin of granite pegmatites assumes that they are the products of a residual magma in which there has been a substantial concentration of volatiles (largely water) and other hyperfusibles.

White (1940, p. 981) has described a hypabyssal, strongly miarolitic aplite in which quartz is 37.5 percent of the rock and orthoclase 73.8 percent of the total feldspar. The miarolitic structure is indicative of the activity of volatiles, which, together with the nature of the rock, indicates the existence of a volatile-rich, K_2O-rich magma.

K_2O-rich felsic dikes thought to be derived from magma have also been described by Noble (1948).

There are many K_2O-rich aphanitic rhyolites. If these are flows or hypabyssal intrusions which show little or no alteration, they are reasonably interpreted as evidence of origin from a K_2O-rich magma. A number of chemical analyses of such rocks are given in

TABLE 35.—*Comparison of normative composition of K_2O-rich rocks and of microcline granite gneiss of the northwest Adirondacks*

Rock type and locality	Specimens averaged	Quartz	Orthoclase	Albite	Anorthite	Corundum	Iron oxides	Norm ratio, orthoclase: total feldspar
Rhyolite (magdeburgose)	13	41.6	37.6	13.8	1.4	1.3	1.66	71
Rhyolite and felsite (omeose) with normative corundum	5	25.15	45.92	17.5	1.0	2.22	3.27	71
Rhyolite and trachyte (dellenose)	4	28.91	40.87	12.32	8.69	1.87	2.95	66
Granite pegmatite and graphic granite (omeose)	13	23.87	53.85	19.03	1.22			73
Vitric tuff with $K_2O>5.5$ and $SiO_2>70$ percent [1]	26	41.2	38.0	12.8	4.7	.79	2.53	69
Microcline granite gneiss, northwest Adirondacks	5	28.6	42.4	17.0	2.47	1.56	4.8	69
Microcline granite, late-kinematic in Svecofennides of southwest Finland [2]	11	30.1	35.6	24.1	5.0	1.12	1.16	52
Granite determined by mode [3]	31	26.43	31.69	25.15	7.92	.77	2.00	49
Granite of the world [4]	546	28.38	28.46	29.34	8.90	1.20	3.08	39

[1] Swineford, Frye, and Leonard (1955).
[2] Simonen (1948).
[3] Johannsen (1932, p. 193).
[4] Daly (1933).

Washington's tables (1917). Some of these rocks may be in part K_2O-rich as a result of alteration, but it seems reasonable that some are unaltered. All within a given group have a close range in composition. Swineford, Frye, and Leonard (1955) have given many analyses of late Tertiary fresh vitric tuffs which are rich in K_2O relative to Na_2O and which appear to be assuredly of magmatic origin. The average normative compositions of several groups and of granite pegmatites based on Washington's data are given in table 35 for comparison with the average normative composition of the K_2O-rich granite and granite gneiss of the St. Lawrence County district. The average normative composition of several other granite families is also given.

Evidence that a K_2O-rich, volatile-rich magma may exist at certain times and places has also been emphasized by Bowen (1928, p. 128–131, 227).

It may also be noted that in the belt through Windfall Pond on the Childwold quadrangle there is a K_2O-rich ferrohastingsite-microperthite granite (table 30, No. 13) that is transitional in composition between the normal microperthite granite and the microcline granite gneisses. Its character and relations are such as to warrant the interpretation that it formed from a K_2O-rich magma. The ilmenomagnetite that it contains holds 8–9 percent $FeO \cdot TiO_2$ in solid solution and as exsolved intergrowths. This is consistent with magmatic temperatures of origin (Buddington, Fahey, and Vlisidis, 1955). The normative ratio of potassic feldspar in the normal hornblende-microperthite granite and granite gneiss is 52, in the transitional K_2O-rich hornblende-microperthite granite 61, and in the microcline granite gneisses, 69.

The foregoing discussion gives good bases for the conclusion that a volatile-rich, K_2O-rich magma may be formed in the earth's crust under appropriate conditions.

Such a magma would have high fluidity and would be expected to have exceptional capacity for penetration and permeation of country rocks, with extensive development of pegmatitic facies. It would also serve to explain the occurrence of a more-than-normal percentage of the iron in the form of oxide rather than as complex iron-bearing silicates. The high percentage of water could serve to split off the iron and titanium oxides from the silicates with accompanying oxidation. The percentage of iron in the biotitic and sillimanitic microcline granite gneiss is about the same as that in the biotite-quartz-plagioclase gneiss. The iron in the latter, however, is largely ferrous iron present in the biotite, whereas in the microcline granite gneiss it is largely ferric iron present in hematite and magnetite.

The microcline granite gneiss locally contains layers and schlieren of each kind of metasedimentary rock of the Grenville series and of the amphibolite, as though it were intrusive into them. In part, also, the biotite-microcline granite gneiss is uniform and homogeneous, without evidence of any inheritance of minerals or structure from the metasedimentary rocks, and of the character of granite crystallized from magma. These relations would be consistent with a hypothesis such as the following. It would seem necessary that some substantial mass of relatively uniform composition be involved in the development of these microcline granite gneisses because of their general gross homogeneity, even through they are heterogeneous in detail. The effective agent in the development of the microcline granite gneisses is then postulated as a highly mobile fluid of the nature of a granitic magma exceptionally rich in H_2O and other volatiles, and exceptionally rich in K_2O relative to Na_2O. This fluid invaded a series of metasedimentary rocks and amphibolite. In the biotite-quartz-plagioclase gneiss the fluid moved in small part along the foliation planes, in large part as films along intergrain boundaries. The pervasive ingress along the intergrain boundaries can be visualized in terms of the mechanism proposed by Fenner (1914). The volatile-rich portion proceeded ahead and fluxed out much of the material of the country rock, permitting the less mobile part to follow, with consequent modification and disintegration of the invaded rock. This would account for the extensive leaching which most of the country rock must have undergone, and for the relatively homogeneous distribution of relict minerals of the modified country rock within the invading material. Some expansion in volume of the composite residual mass would normally be involved, though replacement would be by far the predominant factor. The granite pegmatite veins, in part, and also the schorl-bearing quartz veins are consistently explained as formed by late-stage residual solutions.

The term "migmatization" will here be used for all the results of mixing of the postulated invading fluid and modified country rock, whether the intrusion is as sheets along foliation planes of country rock or as a pervasive intergranular film. The term "granitization," according to Grout (1941, p. 1540), "includes a group of processes by which a solid rock (without enough liquidity at any time to make it mobile or rheomorphic) is made more like granite than it was before, in minerals, or in texture and structure, or in both". Read (1948, p. 9) also defines granitization as "the process by which solid rocks are converted to rocks of granitic character without passing through a magmatic stage." The idea of replacement or metasomatism looms large in the use of this term, but the mechanism for development of the microcline granite gneisses postulated here envisages all gradations between microcline granite gneiss formed strictly by granitization as defined above, and microcline granite gneiss formed directly from magma that is essentially of its own composition.

In the absence of definite indications as to the age relations of the microcline granite gneiss to the hornblende-microperthite granite and alaskite, discussion of possible genetic relationships between them is unwarranted.

SILLIMANITE-MICROCLINE GRANITE GNEISS

REVIEW OF LITERATURE

Many examples, from Precambrian terranes, of granitic gneiss exceptionally rich in potassic feldspar and carrying sillimanite-quartz nodules or aggregates have been described in the literature.

Adams and Barlow (1910, p. 127–139) have described a body of granite exceptionally rich in quartz and potassic feldspar (microcline and orthoclase), which locally carries abundant nodules, in part spherical, consisting largely of quartz and subordinate sillimanite and muscovite. Feldspar is an accessory mineral in the nodules but is varied in amount. The authors interpret the nodules as resulting from crystallization of globules, composed essentially of silica, alumina, and a small amount of boracic acid, which separated as immiscible globules from the main magma.

One typical nodule consists of 63.90 percent quartz, 18.11 percent feldspar, 15.71 percent sillimanite, and 2.60 percent ferromagnesian constituents and water.

Nodular sillimanite-microcline granite gneisses that are similar in every way to those of the Adirondacks have been described as locally occupying extensive belts in the Precambrian of Norway (Brøgger, 1934; Bugge, 1943; and Hofseth, 1942). The nodular granite gneiss is described as closely connected with sillimanitic quartzites and gneisses and also with normal granites whose border zones they often follow. The nodular rocks, however, never show intrusive relations to the country rock or cross the schistosity. Brøgger (1934) interpreted the sillimanite-quartz nodules as originating from a liquid magma that had been highly enriched in SiO_2 and to a lesser extent in Al_2O_3, as the result of the solution of quartzite and perhaps also of aluminous rocks by pneumatolytic influences. The apparent immiscible relation of the SiO_2-rich melt he attributed to its development from solution of the quartzite, and inhomogeneous mixing. The nodular character he thought to be due to the breaking up of the SiO_2-rich magma into minor schlieren and drops during intrusion.

Bugge (1943), on the other hand, explained the nodular granite gneisses as having been formed from a series of aluminous sediments through their penetration by a K_2O-rich ichor, which transformed them metasomatically—in other words, through a process of granitization in connection with a metamorphic differentiation. The latter is indicated, according to him, since the nodular gneiss has often acquired a partial mobility, which is shown as a plastic deformation of the rocks.

Coetzee (1942) has described an aplogranite with sillimanite-quartz nodules bordering an inclusion of biotite-quartz-plagioclase gneiss. The nodules consist chiefly of about 61 percent quartz and 37 percent sillimanite. They occur in the granite for a width of 19 feet along one side of the included layer of biotite-quartz-plagioclase gneiss and decrease in diameter and eventually disappear away from the contact. The percentage of biotite also decreases to normal away from the contact. The granite is a facies of a normal aplogranite, and the composition of the contaminated granite, exclusive of the sillimanite-quartz nodules, is varied, consisting of 42–45 percent quartz, 33–43 percent microcline, 7.5–16.4 percent oligoclase, and 4–9 percent biotite and accessories. In the contaminated facies the ratio of microcline to oligoclase is much higher than normal. The normal granite carries 38 percent microcline and 18.5 percent oligoclase. Coetzee interprets the nodules as the result of incorporation of an aluminous argillaceous sedimentary bed. He thinks, however, that the contaminated rock is not the result of incorporation of a bed like the biotite-quartz-plagioclase gneiss, which is composed in part of 45 percent oligoclase and 29 percent biotite and accessories, because if this were the case the aplogranite should show a higher percentage of plagioclase and biotite.

Another occurrence of sillimanite-bearing microcline-rich granite gneiss has been described by Harry von Eckermann (1922). The Mansjö granite gneiss consists of 36.9 percent quartz, 40.3 percent microcline, 8.4 percent plagioclase, 5.5 percent mica, 4.5 percent almandite, 4.3 percent andalusite and sillimanite. The andalusite, sillimanite, and garnet are interpreted as the product of assimilation of a garnetiferous gneiss. The granite is very rich in pegmatitic segregations and veins that locally give it a gneissic appearance. Von Eckermann interprets it as the last portion of a salic differentiated magma, enriched in volatile components.

Miller (1922) has previously described the occurrence of the nodular sillimanite-quartz-bearing granitic gneiss of the Russell belt and interpreted it as fragments of sedimentary quartzite or quartz schist of the Grenville, included in granite. He writes,

> the granite magma entered two long belts of Grenville well bedded quartzite, forced its way between, and wedged apart the layers, and broke them up into myriads of lenses, all of which became arranged more or less perfectly parallel to the magmatic currents. * * * The magma appears to have been totally unable to assimilate any of the quartzite. * * * Since sillimanite needles commonly lie in the granite several millimeters out from the lens contacts, where they show exactly the same relation to the granite that they do to the lens material—it is probable that the sillimanite did not develop until the intrusion of the magma. * * * The sillimanite * * * probably developed as a result of contact metamorphism, the alumina possibly having been furnished by the granite magma and the silica by the quartzite.

Boos (1935, and Boos and Boos, 1934) has described sillimanite granite from the Front Range of Colorado, where the roofs and sidewalls of the youngest batholiths are locally full of felted masses of sillimanite needles, and the proximity of roofs and sidewalls removed by erosion may be judged from the concentration of sillimanite in exposed granite masses. Some of the granite of these batholiths is rich in potassic feldspar.

Sillimanite granites composed chiefly of oligoclase, less abundant potassic feldspar (microcline or orthoclase), and quartz have been described by Read (1931). They occur as small sheets in sillimanitic metasedimentary gneisses. Read states:

> the presence of sillimanite in these leucocratic granites and aplites is not, in the writer's opinion, to be attributed to an excess of alumina arising by assimilation of pelitic country-rock. These sillimanite-bearing granites have none of the characters of mixed or contaminated granitic rocks. These latter are invariably rich in biotite. The sillimanite and the replacing or late muscovite of the Kildonan rocks are genetically connected, and both most likely arise by the break-up of aluminum halides, alkali aluminates, and similar compounds contained in residual portions of the granitic magma. In the writer's opinion, the operation of similar solutions in the formation of certain types of sillimanite-bearing country-rocks is not impossible.

These rocks have been further studied by Watson (1948), who concludes that the sillimanite was formed under the influence of pegmatitic solutions, late juices of granites, and migmatites at a late stage of migmatization. The sillimanite nodules are interpreted as a metasomatic product later than the process of feldspathization of the metasedimentary rocks.

Geijer (1931) has described narrow pegmatite veinlets containing much tourmaline, which have produced veined gneisses and caused the development of sillimanite in leptites.

Nodular (nodules of quartz and subordinate sillimanite) microcline-rich and quartz-rich granitelike gneisses have thus been diversely interpreted as—

1. The product of immiscible segregation from an originally homogeneous magma (Adams).
2. Originating from an apparently but not actually immiscible ultrarich SiO_2 magma which has resulted from the fluxing of quartzitic beds by pneumatolytic influences related to an intrusive granite magma (Brogger).
3. The product of disintegration of quartzitic beds and modification of the fragments by intrusive granitic magma (Miller).
4. Primary pyrogenic deposits from the break-up of aluminum halides, alkali aluminates, and other compounds in residual portions of a granitic magma (Read).
5. Result of early crystallization of quartz (and sillimanite) in consequence of an ultra-acid composition of the granitic magma (J. Vogt).
6. A peculiar development of the last mother liquor of the rock magma in connection with pneumatolytic agents (Geijer).
7. Product of assimilation of aluminous argillaceous beds (Coetzee) or of a garnetiferous gneiss by a volatile-rich magma (von Eckermann).
8. Result of granitization of aluminous sediments by a K_2O-rich ichor with contemporaneous metamorphic differentiation (J. A. W. Bugge).
9. A metasomatic replacement of feldspathized pelitic and semipelitic schists and granulites effected by solutions related to late juices of granite and granite pegmatites at a late stage of migmatization (Watson).

In general, the groundmass of the nodular sillimanite-quartz-bearing gneisses is a microcline-rich granitic

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material, but there are also associated biotite-quartz-feldspar gneisses commonly rich in plagioclase.

GNEISS OF ST. LAWRENCE COUNTY

The following facts must be taken into account in any hypothesis for the origin of the St. Lawrence County sillimanite-microcline granite gneiss. The country rock with which the sillimanite-microcline granite gneiss is always associated is, in effect, almost exclusively a biotite-quartz-plagioclase gneiss of the Grenville series. The sillimanite-microcline granite gneiss at many places shows all gradations between layers, schlieren, and ghost structures of biotite-quartz-plagioclase gneiss in microcline granite gneiss. Local facies of the sillimanite-microcline granite gneiss may show no inclusions of metasedimentary gneisses over large areas, as between Worden Pond and Sweet Pond (Russell quadrangle). Microcline granite gneiss similar to the sillimanite facies—except for the substitution of hornblende, andradite, or pyroxene for sillimanite—occurs associated with amphibolite or skarn, as in the Childwold area. At many places the sillimanite-microcline granite gneiss carries a little almandite garnet and in the Dead Creek and Loon Pond synclines locally passes into a facies carrying several percent of almandite (with or without biotite) but little or no sillimanite. At many localities the quantity of sillimanite-quartz aggregates decreases away from contact with a layer of biotite-quartz-plagioclase gneiss. Some thin sills of granitic gneiss within the biotite-quartz-plagioclase gneiss are especially rich in sillimanite-quartz discs and nodules throughout, as east of Red School (Russell quadrangle). The sillimanitic facies of the granite gneiss are richer in quartz than the non-sillimanitic facies. The sillimanite-quartz aggregates are not homogeneously distributed but vary widely in quantity from layer to layer. They show the results of development during a period of deformation. Accessory ilmenohematite (locally with magnetite) is commonly associated with the nodules. Granite pegmatite veins are present but not necessarily abundant. The biotite-quartz-plagioclase gneiss at very few localities is itself sillimanitic. Several facies of rocks related to the development of the sillimanite-microcline granite gneiss are shown in figures 7–9.

Sillimanite-quartz nodules have been found in three different kinds of rock in the Grenville series of the Grenville lowlands province: (a) in biotite-rich quartz-plagioclase schist, as one mile northeast of Elmdale (Hammond quadrangle), (b) in biotite-rich microperthite-quartz gneiss at many places in a belt southwest of Black Lake (Hammond quadrangle, Buddington, 1934, p. 127–129), and (c) in feldspathic (microcline) quartzites (west of Bear Pond, Tupper Lake quadrangle, and 0.6 mile north of Pine Grove, Lowville quadrangle). In all these cases the sillimanite-quartz nodules and eyes have a relationship satisfactorily explained as a product chiefly of metamorphic differentiation and segregation in place. The formation of the sillimanite has been at the expense of the aluminous minerals with some associated leaching.

A type of rock which may be closely related to the development of the nodular sillimanite-microcline granite gneiss has been described as occurring at many localities in the Hammond quadrangle (Buddington, 1934, p. 127–129). These rocks are biotite-rich feldspar-quartz schists with a more or less pronounced development of elongate eyes or nodules of sillimanite-quartz aggregates. The nodules are ¼–1½ inches long. In many nodules sparse grains of an iron oxide mineral occur as an accessory. These beds of nodular schist are interbedded with more quartzitic layers and with garnetiferous migmatite. The feldspar of the groundmass is most commonly a microperthite, but in the belt of aluminous metasedimentary rocks through Elmdale there is a bed in which plagioclase is the major feldspar.

Within the main igneous complex, however, rocks analogous to the nodular sillimanite-bearing schists of the Grenville lowlands have not been found. The nodular sillimanite-quartz rocks of the main igneous complex are all microcline-rich granite gneiss.

An example of sillimanite-quartz eyes developed in biotite-quartz-plagioclase gneiss in conjunction with the emplacement of schorl-bearing granitic veinings is shown in figure 8. Other facies related to the development of the sillimanite-microcline granite gneiss are shown in figure 9.

An excellent example of interlayered microcline granite gneiss, sillimanite-microcline granite gneiss, and biotite-quartz-plagioclase gneiss is shown in exposures along Plumb Brook just east of Derby Corners (Russell quadrangle). A rough section is given in table 36.

The sillimanite-microcline granite gneiss always appears to have conformable relations with the enclosing rocks. No transgressive relations were seen, except that the gneiss contains relics of the biotite-quartz-plagioclase gneiss of the country rock.

Most of the microcline granite gneiss, including the sillimanite facies, has a distinctly thin-layered or thin-veined appearance, owing to the development of thin, coarser grained (pegmatitic) facies parallel to the foliation. These may in large part be secretion pegmatites. At many localities there are pegmatitic veinlets carrying schorl, and also at many localities there are numerous veinlets of white pegmatite, in which the

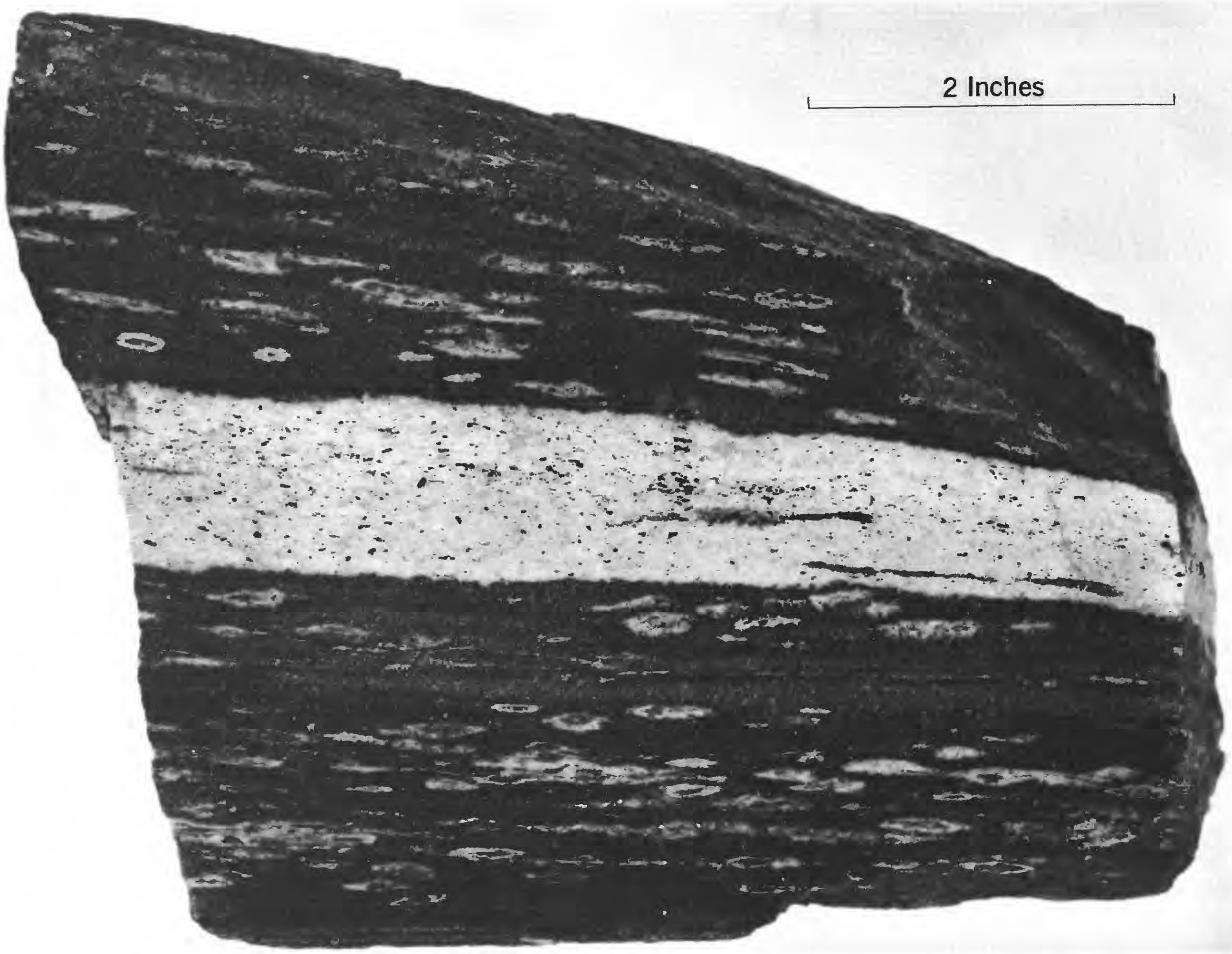

FIGURE 8.—Biotite-quartz-plagioclase gneiss with sillimanitic eyes developed in conjunction with emplacement of schorl-bearing granitic veinings. Wells Island, St. Lawrence River.

TABLE 36.—*Section and modes of gneisses along Plumb Brook, east of Derby Corners, Russell quadrangle*

[Volume percent]

Thickness (feet)	Rock type	Quartz	Microcline	Plagioclase	Biotite (and chlorite)	Iron and iron-titanium oxides	Apatite	Sillimanite	Garnet	Muscovite
	Exposed:									
	Biotite-quartz-plagioclase gneiss with plagioclase almost wholly altered to scapolite	31.3	-----	45.3	10.8	10.9	1.0	-----	0.7	-----
20	Sillimanite-microcline granite gneiss	47.4	28.0	6.4	-----	4.3	.5	13.4	-----	-----
	Covered:									
25	Microcline granite gneiss	39.4	59.6	-----	-----	.4	-----	-----	-----	0.6
7	Sillimanite-microcline granite gneiss	31.6	50.8	15.0	-----	1.7	-----	.5	-----	.4
7	Biotite-quartz-plagioclase gneiss with thin microcline layers	27.1	20.6	43.0	3.5	5.4	.4	-----	-----	-----
15	Microcline granite gneiss	25.1	57.5	11.2	.8	3.5	.5	-----	-----	1.4
25	Sillimanite-microcline granite gneiss	31.9	35.6	24.9	2.2	3.7	.7	1.0	-----	-----

predominant feldspar is plagioclase as distinct from the usual microcline. These phenomena are interpreted as indicative of formation by solutions rich in volatile compounds.

The sillimanite-quartz aggregates can in many places on a small scale be proved to have formed later than the microcline granite gneiss, for they are found developed on shear planes that cut either parallel to or across the foliation of the country rock, and rarely there are crosscutting veinlets of quartz with sillimanite and with or without muscovite. The development of sillimanite-quartz aggregates in veinlike form at 20°–30° to the foliation is well shown in the microcline granite gneiss at the Lyonsdale Bridge (Port Leyden quadrangle), where there are also concentrations rich in sillimanite and iron oxides parallel to the borders of a few pegmatite veins.

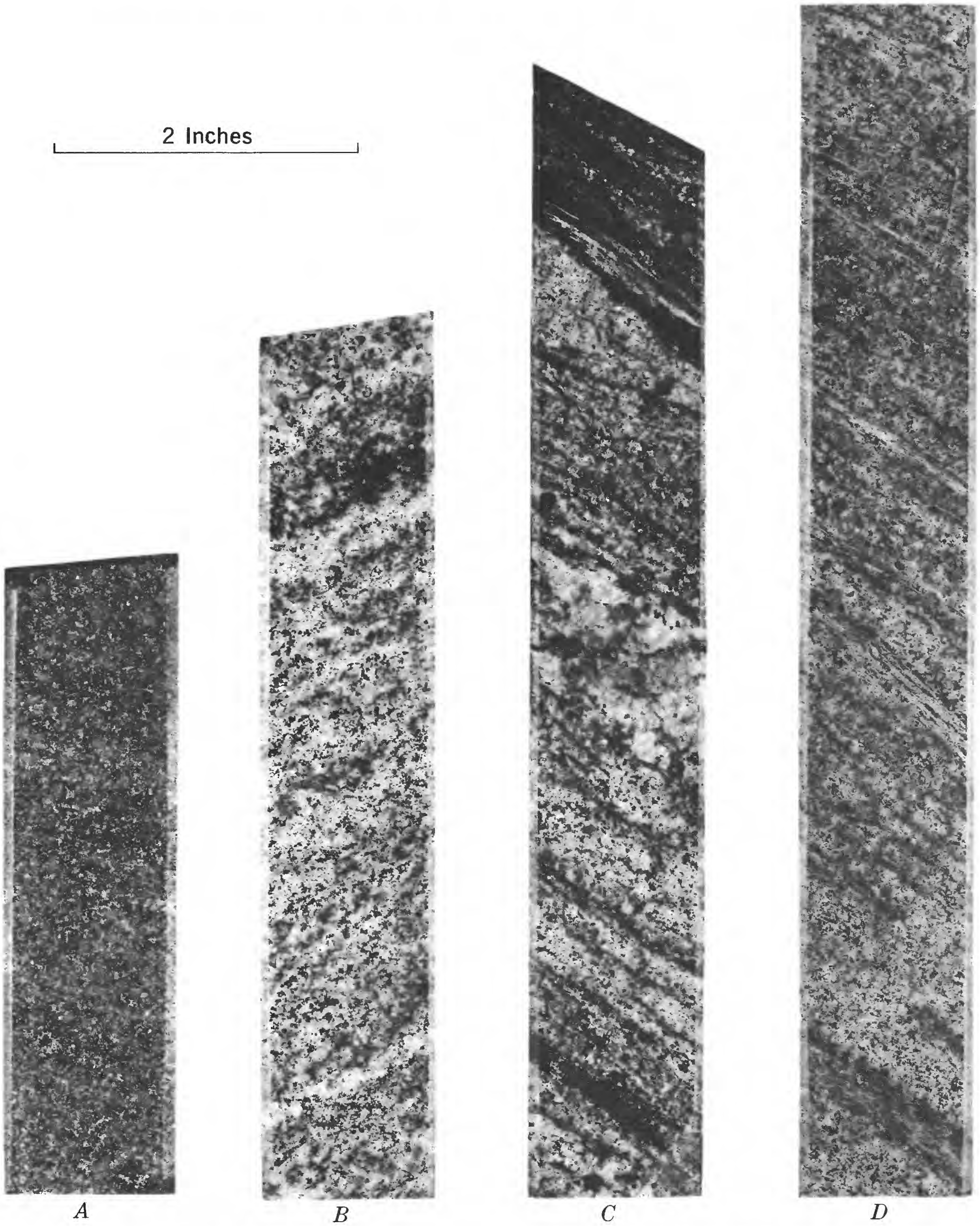

FIGURE 9.—Drill core, Skate Creek magnetite prospect, Oswegatchie quadrangle. A, Biotite-quartz-plagioclase gneiss; B, microcline granite gneiss contaminated with biotite; C, sillimanite-microcline granite gneiss with granitic pegmatite veins and dark schlieren of modified biotite-quartz-plagioclase gneiss; D, uniform sillimanite-microcline granite gneiss. B, C, and D show various aspects of granitization of the biotite-quartz-plagioclase gneiss.

Alaskite gneiss and its contaminated facies are found at least locally in all belts or areas of the sillimanite-microcline granite gneiss. However, we have never found one of these granitic types cutting the other, nor have we found any evidence of one grading into the other. They appear to be quite distinct.

The relation of the microcline granite gneisses to the hornblende granite gneiss has also nowhere been satisfactorily determined. The succession of rocks as found in two localities is given in tables 37 and 38.

TABLE 37.—*Section and modes of rocks in junction zone of hornblende granite gneiss and microcline gneiss, Skate Creek deposit, diamond-drill hole 5*

[Volume percent]

Thickness (feet)	Rock type	Quartz	Microcline	Plagioclase	Hornblende	Biotite	Iron oxides	Apatite	Garnet	Allanite
2	Garnetiferous biotite-microcline gneiss impregnated with iron oxides	5.6	80.2	3.5	-----	1.9	6.2	-----	2.4	0.2
42	Thin-layered pseudomigmatite	11.5	38.2	44.0	-----	4.4	1.5	0.3	-----	-----
>25	Uniform hornblende granite gneiss	9.6	[1]46.5	38.2	3.8	.8	.7	.4	-----	-----

[1] Microperthitic.

TABLE 38.—*Section and modes of gneisses 1.2 miles north of South Russell*

[Volume percent]

Thickness (feet)	Rock type	Plagioclase	Microcline	Quartz	Hornblende	Biotite	Iron oxides	Apatite	Sillimanite	Sphene
	Exposed:									
	Hornblende granite gneiss	31.3	30.1	28.8	7.0	0.4	2.0	0.2	-----	0.2
	Hornblende-biotite granite gneiss	25.2	38.8	25.5	5.0	3.0	1.4	.2	-----	.9
35	Biotite granite gneiss	33.2	30.8	29.0	-----	4.4	1.4	.2	-----	1.0
	Covered:									
40	Sillimanite-microcline granite gneiss	10.4	56.0	30.9	-----	2.7	-----	-----	[1]	-----
25	Biotite-quartz-plagioclase gneisses	48.7	27.0	21.4	-----	.8	1.6	0.5	-----	-----
		60.7	10.5	19.5	-----	4.9	4.1	.3	-----	-----
>20	Sillimanite-microcline granite gneiss	-----	-----	-----	-----	-----	-----	-----	-----	-----

[1] Sillimanite-quartz nodules abundant in rock. Section is of matrix.

The sillimanite-microcline granite gneiss of the South Russell, Granshue, and Dead Creek synclines could all be interpreted as isolated erosion remnants of a single great sheet of gneiss that has been isoclinally folded and whose parts are now preserved only in deep troughs of the synclines, or they could equally well be interpreted as a single biotite-quartz-plagioclase gneiss zone that had been intruded and granitized subsequent to folding.

A comparison of the structure and stratigraphy of the rocks of the Loon Pond syncline with those of the Edwards-Balmat syncline suggests that the microcline granite gneiss with sillimanite and almandite of the Loon Pond structure corresponds to the veined garnet-biotite-quartz-plagioclase gneiss of the Edwards-Balmat structure. In each syncline these respective rocks are overlain by limestone, with interbeds of quartzite and lime-magnesian silicate rocks which are so similar in character as to warrant a tentative hypothesis that they are equivalent formations. The microcline granite gneiss of the Loon Pond syncline contains beds and schlieren of biotite-quartz-plagioclase gneiss similar in all respects to the rocks beneath the limestone of the Edwards-Balmat syncline.

PREFERRED HYPOTHESIS

The sillimanite-microcline granite gneiss, as has been previously noted, is here interpreted as the product of a complex series of processes involving the migmatization and granitization of biotite-quartz plagioclase gneiss, yielding a microcline-rich granite gneiss, which were accompanied and followed by the development of sillimanite-quartz nodules and pegmatitic veinings. Sillimanite rosettes locally replace portions of microcline granite pegmatites, and some of the sillimanitic laminae that developed on shear planes of the gneisses indicate a late origin. There has been much recrystallization of postdeformation age.

In detail, a K_2O-rich, Na_2O-poor salic magma with a meagre percentage of iron and titanium oxide and strongly enriched in volatiles—predominantly water but with subordinate boron, fluorine, chlorine, phosphorus—is assumed to have been intruded. Without contamination it would have yielded a microcline granite low in plagioclase, with local veinings of tourmaline and quartz, as in the Palmer Hill area. The volatile-rich and K_2O-rich magma was intruded in part along foliation planes of the biotite-quartz-plagioclase gneiss. In small part the country rock was incorporated in the granite magma, giving rise to a plagioclase-microcline granite gneiss. In part layers and schlieren of country rock were so modified by solutions or migration of ions from the magma, or by metasomatizing magma itself, as to yield a microcline granite gneiss closely resembling or virtually identical with the microcline granite of direct magmatic origin. In part the inclusions were decomposed and partially leached, yielding a residue of sillimanite and quartz, which was concentrated into flattened and recrystallized boudinagelike aggregates or nodules accompanying contemporaneous deformation. The problem arises whether the nodules developed from materials more or less solid throughout their history, or whether they were fluxed to form a second liquid immiscible with the residual liquid of the late stages of consolidation of the microcline granite magma.

It is here thought that the sillimanite-quartz aggregates were for the most part developed from included layers of biotite-quartz-plagioclase gneiss by a very volatile-rich fluid fraction of a K_2O-rich magma. Such a volatile-rich fluid fraction could have been either a gas phase, an immiscible late segregate, a late-stage residuum, or a combination of these. The data are not adequate to distinguish between the possibilities.

The process of granitization of a biotite-quartz-plagioclase gneiss to yield a microcline-rich rock would necessarily involve the driving off of Na-rich solutions. The mechanics of this, however, are quite speculative. One can think of interchange of ions whereby Na ions move out of the plagioclase of the country rock into magma and K ions move out of the moving films of intergranular magma and into the positions formerly occupied by the Na ions in the plagioclase. The result would be an Na-rich magma, which could then move out as such. Again, it is conceivable that emanations from the microcline granite magma fluxed the quartz-plagioclase gneiss, yielding a solution which moved ahead and out as the K-rich magma moved in and crystallized.

In the process of decomposition and partial leaching of the biotite-quartz-plagioclase gneiss—yielding the sillimanite-quartz nodules—Na, Mg, and probably some Si must be removed.

Locally, the plagioclase of the biotite-quartz-plagioclase gneiss inclusions is partially to wholly altered to sericite. This too would yield Na-rich solutions as a byproduct.

Either an Na-rich magma, such as would be developed in the first case, or an Na-rich hydrothermal solution, such as is developed by the other processes, could give rise to the directly associated sodic pegmatite veins and to the sodic granite masses.

Late-stage aftereffects of magma consolidation include some introduction of iron oxides into the nodules as replacements of sillimanite and, locally, alteration of sillimanite or biotite to muscovite. Also at a few localities there are crosscutting veinlets of quartz carrying muscovite or sillimanite.

The nodular structure of the granite gneiss in the Russell belt has certainly been accentuated during a period of orogenic deformation that induced plastic flowage. In this belt the nodules are in part concentrations on the apices of folds that have been sheared off from the limbs. Such structure is well developed in the gneiss 1 mile north of South Russell and 1.5 miles north-northwest of Benson Mines. Some of the cross-cutting veinlets were also formed at this period.

W. J. Miller (1922) assumed that the magma came into a series of quartzite or quartz schist beds, basing his conclusions primarily on the occurrence of a lens one-eighth mile long at a spot along the western border of the nodular gneiss. Sillimanite-quartz schists, however, have been observed at only two localities in this district and are thought to be of minor significance. The dominant country rock of the sillimanite-microcline granite gneiss is overwhelmingly biotite-quartz-plagioclase gneiss, as are the inclusions. Probably to a quite minor degree, the nodular granite gneiss is the product of the intimate intrusion of magma into sillimanite-quartz schist, as advocated by Miller.

PYROXENE-ANDRADITE AND HORNBLENDE-MICROCLINE GRANITE GNEISS

The microcline granite gneiss with knots and disseminated accessory grains of pyroxene and andradite presents another aspect of the problem of origin of these gneisses. Its nature is such as to make it very improbable that the rock could represent simply reconstituted sediment, for sediment of appropriate composition is very rare. Furthermore, the presence of local lenses of pyroxene and andradite skarn make it probable that limestone was the original country rock. It is again difficult to believe that the granite gneiss could be exclusively the product of replacement of skarn and still yield such a homogeneous distribution of accessory andradite and pyroxene. A few thin amphibolite layers are included in the granite gneiss east of the Raquette River crossing of the road from Piercefield to Tupper Lake. Sodic granite pegmatite lenses occur in these amphibolites and as lenses in the microcline granite gneiss, where they are thought to be inherited. The relations as a whole seem to be best interpreted in terms of intrusion of a K_2O-rich magma with inclusion of some relict lenses, some homogeneous incorporation of suitable country rock, and subordinate concomitant replacement.

A similar interpretation is called for in the case of the hornblende-microcline granite gneiss. The hornblende-microcline granite gneiss characteristically has hemoilmenite or hemoilmenite and ilmenohematite associated with magnetite as accessory oxides. Magnetite is commonly predominant. The oxides of the hornblende facies do not show as intense a degree of oxidation as do the oxides in much of the biotite- and sillimanite-microcline granite gneiss. The hornblende-microcline granite gneiss is always associated with amphibolite. The amphibolite in part occurs in pyroxene-microcline granite gneiss and could have been derived from metasedimentary layers of pyroxenic gneiss or skarn or from para-amphibolite, and in part the amphibolite is probably derived by alteration of gabbro.

The biotite- and sillimanite-microcline granite gneisses are satisfactorily interpreted as largely the product of metasomatism of metasedimentary biotite-quartz-plagioclase gneisses, with contemporaneous deformation and metamorphic differentiation modified by late stage recrystallization. The hornblende-microcline granite gneiss, however, has some characters which are more consistent with its being largely of magmatic origin and only to a subordinate extent the product of replacement of amphibolite and contamination by disintegration of amphibolite. The oxides are thus in part those characteristic of the allied hornblende-microperthite granite and in part slightly more oxidized, but in no case as much oxidized as the oxides of the sillimanite-microcline granite gneiss. The original amphibolite contains much more FeO than does the biotite-quartz-plagioclase gneiss. Similar fluids acting on both rocks might therefore be expected to yield less-oxidized minerals in the replacement of amphibolite than of the biotite-quartz-plagioclase gneiss. Positive evidence of replacement of amphibolite by microcline granite gneiss, however, is not common, the hornblende-microcline granite gneiss appears to be transitional to the hornblende-microperthite granite, and an origin of the hornblende-microcline granite gneiss by replacement of amphibolite would require the removal of substantial amounts of iron and titanium, a phenomenon that did not occur in the formation of the sillimanite-microcline granite gneiss. For these reasons the hornblende-microcline granite gneiss is thought to have formed largely from a magma contaminated by reaction with a little amphibolite. The common occurrence of sphene is in accord with this hypothesis.

REGIONAL DYNAMOTHERMAL METAMORPHISM

Two major types of metamorphism on a regional scale are exemplified in the northwest Adirondacks. One type involves a regional dynamothermal metamorphism that is consequent on intense deformation under relatively high temperature conditions accompanied by reconstitution and recrystallization, under conditions of moderate to great depths and relatively high temperatures and in the presence of a slight amount of permeating fluids. The other type is connected more intimately with emplacement of the younger granites and involves various degrees of intense granitization and the development of veined gneisses. There were at least two periods of dynamothermal metamorphism, one preceding the emplacement of the younger granites, and the other contemporaneous with and following the emplacement, and most intensely localized northwest of the Diana and Stark complexes. In general, the products of metamorphism directly related to granitization, migmatization, and the emplacement of the younger granites involve more substantial changes in composition than do the products of regional dynamothermal metamorphism, although in the latter case there may be some change, even though slight.

Regional metamorphism with substantial changes in composition of the country rock is well exemplified in certain rocks developed from the biotite-quartz-plagioclase gneiss and from marble.

Biotite-quartz-plagioclase gneiss constitutes one of the major country rocks, the raw material which has been involved in migmatization and granitization. It has yielded three major kinds of rock. In the Grenville lowlands it has yielded on the one hand a veined rock, in many places garnetiferous, as a result of arteritic pegmatite injection and pegmatitic vein replacement along foliation planes; and on the other hand two kinds of granite gneiss, as a result of uniform permeation. One type of granite gneiss is usually porphyritic, because of the coarse development of porphyroblastic feldspars, but locally may be even grained. It constitutes a large part of the Hermon granite gneiss. The third product, a result of granitization of the biotite-quartz-plagioclase gneiss, is the microcline granite gneiss with sillimanite-quartz discs and nodules of the main igneous complex. These rocks also show the results of contemporaneous dynamothermal metamorphism as well as change of composition as a whole. They are commonly layered, with alternating coarser and finer texture and the development of pegmatitic veinings. This textural layering is interpreted as related to contemporaneous recrystallization and metamorphic differentiation during the formation of the gneiss.

The widespread development of silicates and silicate nodules in the marble is in part related to materials introduced in connection with the emplacement of the younger granites, as is the local formation of some quartz-mesh marble through silica replacement. The skarn deposits are also metasomatic replacements by magmatic solutions.

There is also a regional variation in the nature and grade of dynamothermal metamorphism of the rocks. For some of the rocks, there is a systematic regional variation across the strike—from northwest to southeast; and for others, a systematic regional variation along the strike—from southwest to northeast. The first part of the discussion will be based on the nature of the rocks as they now occur, and the possible relation to period of deformation will be taken up later.

METAMORPHISM OF SEDIMENTARY ROCKS

Studies of the mineralogy of the metasedimentary rocks of the district have not been sufficiently detailed to yield the data necessary to prove whether there is or is not a regional variation in the intensity of their metamorphism essentially without a corresponding change in chemical composition. Such data as are available seem to indicate a similar mineralogy in similar rocks throughout the region.

The biotite-quartz-plagioclase gneiss has in general a similar mineralogy throughout the main igneous complex. Locally, however, sillimanite has developed in zones of intense metamorphism. The Engels (1953, p. 1068–1071) have shown that in the Grenville lowlands the mica in the least metamorphosed facies of the biotite-quartz-plagioclase gneiss is green, and that with increasing degrees of metamorphism, related to approach to granitic materials or masses, the mica has changed to a brown variety. They have also shown that in the development of sillimanite the relations are similar. Their detailed studies of the mineralogy and metamorphism of the biotite-quartz-plagioclase gneiss (Engel and Engel, 1958; 1960) were published while our report was being prepared for publication.

The combination tremolite–diopside–calcite is found in the marble belts of the Grenville lowlands, as in the Balmat-Edwards area on the Gouverneur quadrangle; in the marble belt on the northwest flank of the main igneous complex; and in marble belts within the main igneous complex just north of Fine, a mile west-southwest of Lower District School, and in the Loon Pond syncline (Tupper Lake quadrangle). Also diopside granulites are found in the marble belts throughout the area. These mineral combinations may form under physicochemical conditions that yield the albite-epidote amphibolite facies or the amphibolite facies (Turner, 1948, p. 89–90). The combination albite-epidote-hornblende is not found in any of the gabbros of the region, so that in the Grenville lowlands the conditions of formation for the reconstitution of the marbles must correspond to those which yield the normal amphibolite facies in reconstitution of gabbro. Somewhat more intense conditions under higher pressures at greater depths must have obtained within the area of the main igneous complex, where garnet amphibolites and garnet-pyroxene-plagioclase granulites were formed from metagabbro or metadiabase.

METAMORPHISM OF GABBRO AND DIABASE

No garnet has been found in any of the numerous metagabbro or equivalent amphibolite bodies west of the Cranberry Lake and Russell quadrangles or north of the Russell and Stark quadrangles, with the single exception of a small mass on the Oswegatchie quadrangle southwest of Little Otter Pond. Traces of garnet occur sporadically in the anorthositic metagabbro mass on the Russell quadrangle, and locally there is some garnet in the metagabbro sheet in the northwest corner of the Russell quadrangle. By contrast, coarse garnets in substantial volume are found in many metagabbro and amphibolite masses east of the Stark anticline on the Stark quadrangle, and on the Cranberry Lake and Tupper Lake quadrangles. The hornblende of the metagabbro masses within the metasedimentary belt of the northwest is also different from the hornblende of the amphibolite bodies within the granite and granite gneiss masses of the main igneous complex to the southeast. The hornblende of the amphibolite in the northwest is a normal green hornblende consisting of 4 percent or more Fe_2O_3 and 10–12 percent FeO, whereas the usual hornblende in the amphibolite of the main igneous complex is brown to olive green in thin section and carries less than 3.5 percent Fe_2O_3 and 9–13 percent FeO.

Metadiabase dikes are common in the Diana complex on the Lake Bonaparte and the northwest corner of the Oswegatchie quadrangles. No garnet has been found in these rocks. On the Cranberry Lake and Tupper Lake quadrangles, equivalent metadiabase dikes are also present in the Tupper complex, but nearly all are here reconstituted to a garnet-augite-plagioclase granulite. Similar garnet-augite-plagioclase granulite dikes are also found at the Clifton mine on the east side of the Stark anticline.

The garnetiferous amphibolites occur as lenses and belts within the granite masses of the main igneous complex, whereas the garnet-augite-plagioclase granulites occur exclusively within the quartz syenitic complexes. The amphibolites are thought to have been derived from gabbro by deformation and reconstitution in the presence of vapors or hydrothermal solutions moving through them, whereas the granulites are thought to have developed by deformation and reconstitution under relatively dry conditions. The garnet of the amphibolite, however, is not an effect of metamorphism in contact zones with the granite, for it has been previously shown that in such zones garnet is destroyed (Buddington, 1939, p. 184–187).

The strongly hornblendic facies of the amphibolites have been reconstituted in the presence of such volatile materials as H_2O, F, Cl, and P, for OH, F, and locally Cl enter into the composition of the hornblende in small amounts, and apatite is commonly present in larger amount in the amphibolite than in the primary gabbro.

The manner in which the metamorphism of the gabbroic masses differs from the northwest to the southeast has been previously discussed (Buddington, 1939, p. 267–282; 1953). The predominant metagabbro in the Grenville lowlands is a hornblende-plagioclase or hornblende-hypersthene-augite amphibolite, and locally a hornblende-hypersthene amphibolite. Both types may be biotitic and may include associated or relict augite. Within the main igneous complex both types are common; in addition, there is a garnetiferous facies.

METAMORPHISM OF QUARTZ SYENITE SERIES

DIANA COMPLEX

The rocks of the Diana complex show a marked variation and increase in the intensity of metamorphism from the northwest to the southeast, across the strike. The rocks of the Stark complex show an increase in intensity of metamorphism from southwest to northeast along the strike; the Tupper complex is garnetiferous only in the easternmost part. In general, the intensity of metamorphism increases toward the east.

The rock along the northwest border of the Diana complex in the Oswegatchie quadrangle is an augen gneiss with a mortar of mylonitic grains averaging less than 0.2 mm in size, commonly less than 0.1 mm, and locally 0.01–0.03 mm. Locally, there are sphene porphyroblasts. This belt is about 0.5–1 mile wide. North along the strike the pyroxene syenite facies thins out on the Russell quadrangle, and the augen gneiss is there replaced by hornblende-quartz syenite flaser gneiss.

The augen gneiss grades also southeastward across the strike into flaser gneiss. The flaser gneiss has a width of about 3 miles where it passes southwestward off the Oswegatchie quadrangle but is only about 1 mile wide north of Kalurah. The flaser gneiss on both the Oswegatchie and Russell quadrangles consists predominantly of a granular recrystallized groundmass, with the feldspar in polyhedral grains averaging 0.2–0.5 mm in a mosaic fabric. The quartz is commonly in flat leaves of uniform extinction or an aggregate composed of only a few grains. There is usually a small percentage of porphyroclasts of feldspar, but in part the flaser gneiss is a wholly granoblastic facies with porphyroblasts of hornblende and locally of sphene.

Farther southeast, the phacoidal granite gneiss of the Diana complex north of Jayville is a granoblastic gneiss in which the polyhedral grains average 0.8–1.5 mm, though some are as large as 4 mm. Locally, as on the trail 0.6 mile north of Jayville, the phacoidal granite gneiss is sheared out to an even-grained uniform gneiss with complete loss of phacoidal structure, but without diminution of average grain size. Some of the feldspar in the gneiss with phacoidal structure, however, is more perthitic than any in the more recrystallized, evenly foliated gneiss.

There is considerable variation in grain size within a single thin section and from layer to layer throughout the quartz syenite complex. The granular material of some phacoids may be coarser than the groundmass. In general, however, there is at least a tenfold increase in grain size of the recrystallized feldspars from the northwest to southeast on the Oswegatchie quadrangle within a distance of 3 miles across the strike.

Coincident with the increase in grain size from northwest to southeast, there is also an increase in the degree of recrystallization. In the mylonitic material, where the grain size of the feldspars is less than 0.1 mm, the quartz is uniformly coarser than the feldspar, being about 0.2 mm where the feldspar grains are 0.03 mm, and 0.1 mm where the granulated feldspar is 0.05 mm. Where the grain of the recrystallized feldspar is 0.1 mm or more, the leaves of quartz are commonly single grains with uniform optical extinction. In the flaser and granoblastic gneisses, all but the porphyroclasts are completely recrystallized, so that the perthite of the primary rock has unmixed to yield individual grains of microcline or untwinned potassic feldspar on the one hand and plagioclase on the other. In the phacoidal granite gneiss, all gradations may be seen, from microperthitic feldspar clasts with about equal amounts of plagioclase and orthoclase, to clean microcline without any plagioclase intergrowth.

There is also a change in the new minerals developed from northwest to southeast. In a belt along the northwest, sphene is the characteristic new mineral developed and in part occurs as porphyroblasts; in the southeast part, hornblende is developed as porphyroblasts, and secondary sphene is absent.

STARK COMPLEX

The least deformed facies (fig. 10) of the Stark complex forms a lens about 0.5 mile wide and 2.0 miles long near the north border of the Stark quadrangle, for more than 1 mile on each side of Cold Brook. The rock of this belt is predominantly a hornblende granite ranging from the normal hornblende facies to an alaskitic type. The rock is very coarse, the feldspars varying from 0.5 inch to more than 1 inch in diameter. Much of the rock appears to the eye to be wholly undeformed, but examination with the microscope shows that the feldspars are cracked and veined, and that the borders are locally granulated, so that the rock is actually a mortar gneiss. There are a few

FIGURE 10.—Coarse hornblende granite, practically undeformed, from the Stark complex. Locality is 1.5 miles west-northwest of Cold Brook School, Stark quadrangle.

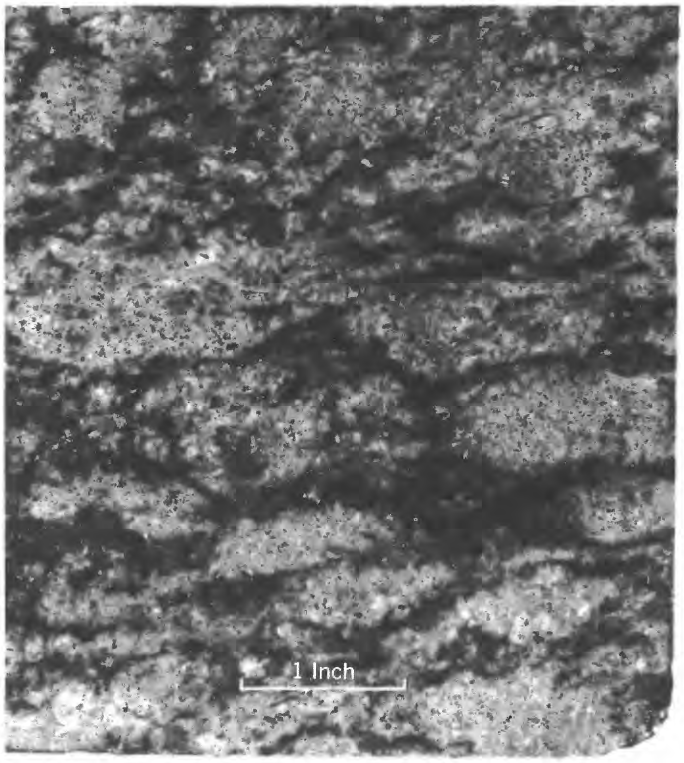

FIGURE 11.—Phacoidal hornblende granite gneiss, deformed and recrystallized, showing granoblastic feldspars and recrystallized quartz leaves; original rock similar to that of figure 10. Clifton mine, Russell quadrangle.

euhedral plagioclase crystals of similar size to the microperthite. The rock 1.1 miles west-northwest of Cold Brook School is also almost undeformed and equally coarse. Here, however, the plagioclase forms the core of the feldspars, and microperthite forms the rim. The mafic minerals include both augite and hornblende, the latter in part secondary after the pyroxene.

The lens of relatively massive primary quartz syenite and granite just described is but a local relic of the original rock. All the rest of the mass is almost completely granulated and recrystallized (fig. 11). The rock of the Stark complex west and southwest of the highway through Cold Brook School is a granoblastic gneiss in which polyhedral grains average 0.8–1.5 mm in size, but larger ones are up to 4 mm. The pyroxene syenite gneiss in the local facies on the flanks of the Stark anticline has a grain size of 0.4–0.7 mm.

It has been previously described how the members of the quartz syenite series in the northeast part of the Stark complex, northeast of the main highway through Cold Brook School (Stark quadrangle), almost everywhere contain a little garnet, predominantly as coronas around the mafic minerals but in part as small porphyroblasts.

The garnetiferous quartz syenite facies of the Stark complex are composed of a mosaic of feldspar grains of somewhat varied diameter and leaves of quartz of uniform extinction. The larger recrystallized feldspar grains average around 0.8 mm in size, and the untwinned potassic feldspar and microcline carry from considerable to a little microperthitic intergrowth. The smaller potassic feldspar grains (0.2–0.5 mm) are microcline and untwinned potassic feldspar, usually free of perthitic intergrowths. A little myrmekite is present.

The development of garnet in the eastern and northeastern facies of the quartz syenite series is independent of the composition of the rock. It seems best interpreted as a facies of metamorphism under deeper seated and warmer conditions.

INLET SHEET

The rock of the southern and southeastern part of the Inlet sheet is strongly deformed and has a well defined foliation and a granoblastic structure. The size of grain is different from one thin layer of the foliation to another; the average is 0.3–0.8 mm, though a few porphyroclasts are as large as 3 mm. The coarser grained layers are more common. The most deformed rock is at the east and west ends of the sheet. The smaller grains of potassic feldspar in the deformed rock are in general less microperthitic than the larger grains; and there are more discrete plagioclase grains, indicating some unmixing and recrystallization of the primary perthite during metamorphism.

ARAB MOUNTAIN SHEET

The pyroxene syenite gneiss of the Arab Mountain sheet consists of a granoblastic aggregate, averaging about 0.3–0.4 mm in grain size, but containing a varied percentage of porphyroclastic grains, commonly 1–3 mm in size but locally as large as 4 mm. The potassic feldspar of the groundmass, in general, carries less plagioclase in microperthitic intergrowth than the larger grains. There is also more plagioclase in the groundmass than in the normal undeformed rock of similar type. These phenomena indicate that during the deformation there has been a recrystallization of a strongly microperthitic feldspar, yielding a less microperthitic facies and discrete plagioclase grains.

Ramberg (1949, p. 35) writes that rocks of charnockitic affinities (norites, enderbites, charnockites) were formed not by magmatic differentiation but by high-grade regional metamorphism yielding a "granulite facies." He also believes (1948, p. 553) that anorthosites are recrystallized under "granulite facies" conditions and (1948, p. 566) have received their present structure and mineralogical character by regional metamorphic recrystallization in the solid state. He further (1948, p. 567–568) implies that the occurrence in the Adirondacks of charnockitic gneisses in a location above the anorthosite massif is due to a differential diffusion of elements and molecules, whereby O, K, Si, and Na tend to diffuse upward and elements of the heavy minerals migrate downward, so that only plagioclase (anorthosite) of intermediate composition remains stable, together with pyroxenes, garnet, olivine, and ore minerals.

At least part of the Adirondack charnockitic gneisses (especially those carrying garnet), part of the gabbros, and the border facies of the anorthositic rocks have certainly been recrystallized under the physical conditions appropriate to the "granulite facies." It is equally certain, however, that the systematic stratiform arrangement of the variation of composition in the Tupper complex is due to magmatic differentiation—largely by fractional crystallization—*before* metamorphism, and the charnockitic character of its mineralogical composition is in part inherited from a primary feature.

The suggestion that differential diffusion is the explanation of the origin of the charnockitic rocks overlying anorthosite is too simple and general a proposal to warrant discussion until it has been shown how it could apply in detail to the great and complex variety of rock relations which obtain and which appear to be in complete conflict with the idea. One such relation is the presence and sharp contacts of metadiabase dikes and of charnockitic dikes within anorthosite, all of premetamorphic age. Diffusion of the major elements has taken place locally to the extent of feet, but no evidence that it has taken place to the extent of hundreds of feet or of miles has been seen. However, such materials as OH, H_2O, Cl, and F have diffused or migrated extensively in certain zones.

There also appears to be the implication (Ramberg, 1949, p. 30, 35–47) that the charnockitic rocks formed from hornblendic rocks by metamorphism. This hypothesis is quite inapplicable to the Adirondack charnockites, for hornblende granites of similar composition to the charnockite, and associated with them, have been recrystallized in the "granulite facies" with the hornblende still preserved.

METAMORPHISM OF GRANITE SERIES

The metamorphism of the hornblende granite and the alaskite may be described in terms of three belts, which differ from each other in the intensity of deformation and recrystallization. The granites of the belt that lies north, northwest, and west of the Stark anticline are most intensely deformed and recrystallized; those of the southern two-thirds of the Oswegatchie quadrangle and of the Cranberry Lake and Tupper Lake quadrangles are the least deformed and recrystallized. The hornblende granite in the belt between the Stark anticline and the belt of metasedimentary rocks to the east has characters transitional between those of the northwestern and the southern belts.

The least deformed facies of the hornblende granite occurs in the main mass of the southern belt. The bulk of the granite consists of more or less the same amounts of porphyroclastic(?) microperthite grains 1.5–3.5 mm in diameter, and a granular groundmass in which the same kind of microperthitic feldspars are 0.2–0.4 mm in diameter. Locally the mortar may form as little as 10 percent of the mass. To the north, on the east flank of the Stark anticline and between the phacoidal granite gneiss and the metasedimentary rocks, the hornblende granite is more deformed (fig. 11) and has an average grain size of 0.6–0.7 mm, locally with clasts 1.5–3 mm in diameter. To the west of the Stark anticline, between it and the northwest border of the igneous complex and north of Vroomans Ridge, the hornblende granite gneiss is completely recrystallized. Here its average grain commonly ranges from 0.5 to 0.8 mm, but locally is as large as 1.2 mm.

Concomitant with these changes in the size and uniformity of the grain there is also a change in the nature of the mineralogy. In the least deformed hornblende granite the feldspar is almost wholly microperthite, and both the primary grains and that in the granular groundmass are the same kind of microperthite. A few percent of plagioclase is present in the groundmass. Table 25 shows that there is a successively increasing recrystallization and unmixing of the microperthite as we pass from the border facies of the main mass through the facies on the east flank of the Stark anticline to the gneiss of the South Russell synclinorium. The amount of plagioclase increases at the expense of the microperthite, until there is only a little microperthitic intergrowth left in a few grains of that facies of the gneiss which has been most intensely recrystallized. The microperthite is not unmixed by an abrupt recrystallization into clean plagioclase and clean potassic feldspar. On the contrary, the unmixing of the plagioclase takes place through the development of clean plagioclase at the expense of the microperthite, in such a fashion that the new microperthite grains gradually come to have less and less plagioclase intergrowth. The nature of the potassic feldspar also changes. In the granites on the core of the Cranberry Lake anticline and on the east flank of the Stark anticline it is almost wholly orthoclase, whereas in the gneiss of the South Russell syncline it is almost wholly microcline. Another noteworthy change is the development of sphene and an increase in the percentage of magnetite in the gneiss of the South Russell syncline. The hypothesis that this might be a phenomenon of contamination from the metasedimentary rocks was considered but abandoned, for there has been as much opportunity for contamination of the granite gneiss on the east flank of the Stark anticline as on the west. The titanium and iron have presumably been in part derived from the primary ferromagnesian minerals during their recrystallization.

The hornblende granite gneiss differs from the equivalent gneissoid hornblende granite, however, in having a slightly higher percentage of Fe_2O_3 and TiO_2. This is brought out by the data in table 39. It is possible, though not necessary, that there may have been introduction of a slight amount of Fe_2O_3 and TiO_2 during the regional metamorphism.

TABLE 39.—*Weight percent* Fe_2O_3, FeO, *and* TiO_2 *in hornblende granites and hornblende granite gneisses, northwest Adirondacks*

Rock type and locality	Fe_2O_3	FeO	TiO_2
Ferrohastingsite granite, Cranberry Lake quadrangle	1.14	2.62	0.42
Femaghastingsite granite, Lowville quadrangle	1.85	1.97	.41
Hornblende granite, Cranberry Lake quadrangle	1.02	1.92	.43
Composite grab sample of hornblende granite for 0.4 mile northeast of Scanlons Camp, Oswegatchie quadrangle [1]	1.60	1.40	.37
Composite grab sample of hornblende granite for 1 mile west of Briggs, Oswegatchie quadrangle [1]	1.32	1.93	.32
Hornblende granite gneiss, Nicholville quadrangle	2.90	2.07	.51
Ferrohastingsite granite gneiss, Russell quadrangle	2.07	2.16	.57
Composite grab sample of hornblende granite gneiss, 0.6–1 mile southeast of Stone School, Russell quadrangle [1]	3.19	1.96	.60

[1] Analyst, J. J. Fahey, U.S. Geological Survey.

The alaskite similarly shows the same type of concomitant variation in texture and feldspar mineralogy in successive belts northwest from the belt east of Cranberry Lake.

The least metamorphosed alaskite seen is in the Darning Needle syncline. The alaskite here consists of grains of microperthite 1.5–2.5 mm in size, in a granoblastic mortar averaging about 0.3 mm. The quartz is in much flattened, elongate leaves, except for a few rounded granules contained in the mortar or enclosed in microperthite. Thus the rock shows evidence of deformation, though not of unmixing of the feldspars. In some of the alaskite the microperthite grains are elongate parallel to the foliation. The sheet of alaskite in the Dead Creek syncline north of Roundtop Mountain is fine grained (0.8–1.0 mm). Here the microperthite is in granular aggregates, as the product of breakup of larger grains. An albite rim surrounds many of the perthite grains. The quartz is in elongate ameboid forms. At the southwest end of the Jayville syncline the larger microperthite grains are 2.5–4.0 mm in diameter, though most are about 1 mm. In some layers the microperthite grains average about 0.9 mm in diameter; well-defined quartz leaves are present.

The alaskite gneiss sheet in the Brandy Brook belt east of Cranberry Lake is in general a fine-grained, largely granoblastic rock. The texture is inequigranular, the larger grains being about 2.0 mm in diameter and the groundmass about 0.4–0.5 mm. The proportion of larger to smaller grains is varied, though the fine material usually predominates. The quartz is almost wholly in the form of irregular granular aggregates or is present as grains intermingled and intergrown with the feldspar.

None of the alaskite or gneissic alaskite south of the Clare-Clifton-Colton and South Russell belts of metasedimentary rocks has a texture in which the broken grains have a typical polyhedral character in the groundmass. Instead, the borders of the coarser grained primary aggregates are scalloped or somewhat interpenetrated.

The alaskite gneiss around the south end of the South Russell synclinorium, south of Benson Mines and south and southwest of Fine, is strongly deformed. There are a few grains 1 mm or so in diameter in a groundmass averaging 0.5 mm. The feldspar granules of the latter are polyhedral. The quartz is in small rounded grains intermingled with the feldspar or in larger, elongate, ameboid grains. The feldspar has in part been recrystallized to microcline and plagioclase.

The alaskite gneiss of the South Russell synclinorium has been completely recrystallized; the granoblastic feldspar has an average diameter of about 1 mm, though locally it is coarser. The grains commonly have nearly straight borders and range from 0.5 to 2.5 mm in diameter.

Table 27 shows that there has been an unmixing of the microperthite feldspar of the alaskite comparable to that in the hornblende granite in successive belts to the northwest from Cranberry Lake. Similarly, the potassic feldspar ranges from orthoclase, in the least deformed facies, to microcline, in the alaskite gneiss of the South Russell synclinorium. A trace of sphene appears in a few specimens of the alaskite gneiss in the South Russell syncline, paralleling the development of sphene in the hornblende granite gneiss.

MICROCLINE GRANITE GNEISS

The sillimanite-microcline granite gneiss, as has been previously noted, is here interpreted as the product of permeation, migmatization, and partial granitization of biotite-quartz-plagioclase gneiss, accompanied by contemporaneous metamorphic differentiation. The development of the sillimanite-quartz nodules and, in large part, of the thin, coarser, recrystallized layers that give the rock its veined gneissic character, is due to metamorphic recrystallization and differentiation. The coarser grained layers of the granitic gneiss carry fewer but coarser grains of iron oxide minerals than do the finer grained layers; but the nature of the iron oxides is the same in both. The ilmenohematite and titanhematite grains are of similar size and coordinate in development with the other minerals with which they occur. The ilmenohematite grains in the coarser grained and pegmatitic layers of the gneiss are coarser than those in the finer grained gneiss, and the ilmenite intergrowths are also coarser. There is commonly less metallic mineral in the pegmatitic and nodular sillimanite-quartz layers than in the normal gneiss, and in many places there is a slight concentration of the oxides on the borders of the pegmatitic layers and the sillimanite-quartz nodules. Sillimanite continued to form at a late stage of the differentiation, for it replaces pegmatite locally and occurs in late shears. In some of the vein deposits it is contemporaneous with the magnetite of the late-stage ore-forming solutions. The intensity of development of the sillimanite-quartz nodules and lenses also correlates directly with the degree of deformation.

QUARTZ SYENITE SERIES AND YOUNGER GRANITE: CONTRAST IN METAMORPHISM

There are some resemblances and some conspicuous inconsistencies in the manner of variation in the metamorphism of the granite as compared with the nature of variation in metamorphism of the quartz syenitic rocks, from one belt to another.

The metamorphism of the granite and the quartz syenite is similar in that the quartz syenite gneisses of the Diana complex and the younger hornblende granite gneiss west of the Stark anticline carry a little sphene originating through metamorphic reconstitution. Also, members of the quartz syenite series near the northwest border of the igneous complex, on the Lake Bonaparte quadrangle, are locally mylonitized in broad belts (Buddington, 1939, p. 283–284), as are the phacoliths of younger granite in the adjoining metasedimentary rocks (Martin, 1916, p. 70–71; Buddington, 1939, p. 292–294). All this mylonitic rock has been somewhat recrystallized, the development of the microcline being strong in the granite.

In places, however, there are also contrasts in the degree of metamorphism of the quartz syenitic rocks and the younger granite within the same geographic belt The granite gneiss lying to the east and west of the belt of quartz syenite gneiss through Stammerville School (Russell quadrangle) is, in general, coarser grained than the quartz syenite gneiss, even though both are completely recrystallized. The syenite gneiss on the core of the Arab Mountain anticline is also more thoroughly recrystallized and granulated than the associated hornblende granite.

The difference in degree of deformation between the pyroxene-quartz syenite gneiss and the younger granite is well shown 0.5 mile north of the State Ranger School (Cranberry Lake quadrangle), where granite is intrusive into the gneiss of the Inlet sheet. The quartz syenite gneiss is almost wholly granulated, with the perthite grains averaging 0.6 mm and a few porphyroclasts 3 mm in diameter. The granite does not contain more than 10 percent granulated material; the dominant perthite grains are 2–3.4 mm in diameter and average 2.7 mm.

The inconsistencies between the nature of metamorphism of the quartz syenitic rocks and the younger granite series in the same geographic belts is in accord with the hypothesis, based on structural relations, that the quartz syenite complexes and older formations were folded and metamorphosed before the intrusion of the younger granitic rocks. Because the latter are in part also metamorphosed, there must have been in effect at least two major periods of deformation and metamorphism.

SUMMARY OF REGIONAL VARIATION IN METAMORPHISM

In the preceding pages, details have been given of the manner in which the intensity of metamorphism of the rocks older than the younger granites has varied regionally. A more thorough discussion as related to a larger area has been given in another publication (Buddington, 1952), but a brief summary of certain points of interest is given here. Three belts in which there is variation of intensity of metamorphism in rocks older than the younger granites are shown in figure 12. The intensity of metamorphism varied in such a way that in the northwest (northwest of line $A–A'$) the rocks were metamorphosed under the conditions of the amphibolite facies, and there is no garnet in any of the rocks of the anorthosite series, gabbro sheets, pyroxene syenite, quartz syenite series, or metadiabase. Within an intermediate belt to the southeast (area between lines $A–A'$ and $B–B'$), garnet did develop locally in reconstituted gabbro and diabase masses but not in the pyroxene and quartz syenite gneisses. Southeast, east, and northeast of the intermediate belt (northeast, east, and southeast of line $B–B'$) garnet developed not only in the gabbro and metadiabase but also in the border facies of the anorthositic rocks and in the pyroxene syenite, quartz syenite, and hornblende granite gneiss of the Stark and Tupper complexes.

Porphyroblastic and corona sphene occurs in the syenite gneisses of the cooler, northwestern part of the Diana complex developed in the amphibolite facies, whereas porphyroblastic hornblende is developed in the quartz syenite of the warmer, southeastern part of the belt of the amphibolite facies. Porphyroblastic hornblende may accompany recrystallized augite, hypersthene, and porphyroblastic garnet in the syenite and quartz syenite gneisses of the granulite facies.

In addition to the development of garnet successively in diabase and gabbro gneiss, and then in syenite, quartz syenite, and related older granite gneiss toward the southeast, there is also a variation in the amount of perthitic intergrowth of plagioclase in potassic feldspar and of ilmenite in magnetite. The percentage of intergrowth (where present and not completely segregated) for both these minerals increases toward the southeast.

The amount of plagioclase in perthitic intergrowth with the potassic feldspar ranges from about 50 percent to zero, depending on the intensity of conditions during metamorphic recrystallization. The hypothesis is adopted here that the primary mineral in all cases was a feldspar of magmatic crystallization occurring as a homogeneous solid solution of all the potassic and plagioclase components, and that its perthitic character is due to subsequent unmixing (Tuttle, 1952). This interpretation is consistent with the following facts. Lenses of practically unmetamorphosed pyroxene-quartz syenite occur locally in marble throughout the Adirondacks. In all such cases the feldspars occur almost exclusively as microperthite. Relict porphyroclasts are common in much of the gneiss; they are microperthite comparable in composition to the total feldspar present.

During subsequent regional metamorphism the microperthite was granulated and recrystallized. The new potassic feldspars redissolved part of the exsolved perthitic intergrowth, and part was segregated, to form discrete plagioclase grains. On later cooling, the new homogeneous potassic feldspar solid solution again partly unmixed to yield the present microperthite of the gneisses. The amount of plagioclase intergrowth in the potassic feldspar should vary with the original temperature of metamorphic recrystallization (Bowen and Tuttle, 1950).

The normative composition of the total feldspar in the hornblende- and pyroxene-quartz syenite and granite gneisses quite uniformly shows about 46 percent potassic feldspar molecule and 54 percent plagioclase molecule. In such gneisses the microperthite forms about 100 percent of the total feldspar in the primary unmetamorphosed rocks, 65 percent (locally 70 percent) of the total feldspar in the rocks recrystallized in the granulite facies, and 60 percent of the total feldspar in the felsic gneisses recrystallized in the amphibolite facies. The plagioclase might also be expected to be perthitic but it is usually clear, hence it must be assumed that the exsolved potassic feldspar has migrated outside the borders of the plagioclase grains and been counted as part of the microperthite portion. Bowen and Tuttle (1950, p. 497) give a diagram of the "solvus" for $KAlSi_3O_8$ and $NaAlSi_3O_8$ that can be used to obtain a rough approximation for the temperatures of metamorphism of the Adirondack gneisses. The original composition of the total feldspar aggregate of the gneis-

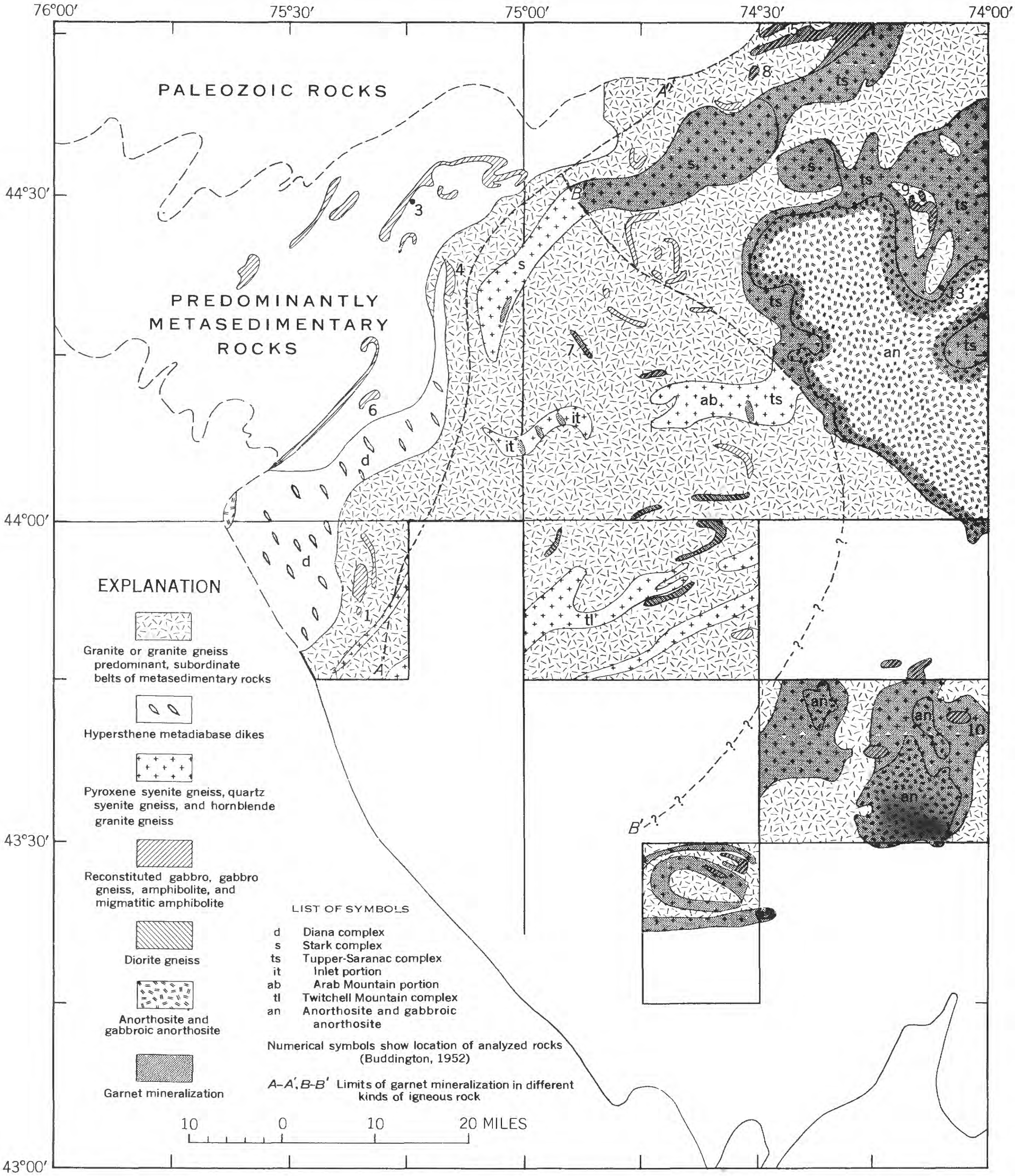

FIGURE 12.—Generalized map showing distribution of garnetiferous facies of rocks. Northwest of line *A–A′*, anorthosite, gabbro, diorite, and quartz, syenite series of rocks are nongarnetiferous, between lines *A–A′* and *B–B′* gabbro, diabase, and diorite are in part garnetiferous, whereas quartz syenite gneisses are nongarnetiferous; northeast, east, and southeast of the line *B–B′*, gabbroic anorthosite, gabbro, diorite, and quartz syenite gneisses are at least locally garnetiferous.

ses, the various phenomena of exsolution exhibited by the rocks, and the ratios of occurrence of the different feldspars have been taken into account in the following estimates. The effect of the anorthite molecule has not been considered but would presumably be small. The mineralogical relations of the two feldspars in the granulite facies (microperthite composed of about two-thirds potassic feldspar, one-third oligoclase) would roughly correspond to a temperature of formation of about 625° C, and of the two feldspars in the amphibolite facies (microperthite composed of about three-fourths potassic feldspar, one-fourth oligoclase), to a temperature of formation of about 500°C. The temperature of the original primary magmatic crystallization would have been above 660°C.

There is a concomitant variation in the percentage of ilmenite intergrowths in magnetite, as shown by the data in table 40. The amount of ilmenite in the ilmenomagnetite of the primary younger hornblende granite, including that present in solid solution and that present as intergrowths, is 7–10 percent, whereas the magnetite or ilmenomagnetite in similar granite recrystallized under conditions of the amphibolite facies carries only 1–4 percent ilmenite, and the ilmenomagnetite of the phacoidal granite gneiss of the Stark complex recrystallized in the granulite facies carries an intermediate content of 5–7.7 percent ilmenite.

The younger granites are associated as closely with the rocks of the amphibolite facies as with those of the granulite facies, hence the temperature differential cannot be related directly to the emplacement of the younger granite bodies. The eastward increase in the temperature of metamorphism of the older rocks seems best interpreted in terms of their development at greater depth, with concomitant rise of the isogeotherms. The belt of the granulite facies now has an arclike shape, which, insofar as our present data go, could be interpreted in terms of a postmetamorphic synclinal warp that brought the originally deeper rocks of the east to the present same topographic level as the originally shallower rocks of the west, or higher.

TABLE 40.—*Decrease in weight percent ilmenite in magnetite of igneous gneisses with recrystallization at lower temperatures*

[Analyses are of grab samples]

Rock type and locality	*Weight percent ilmenite in ilmenomagnetite*
Ilmenite of primary hornblende granite:	
Along trail for 2.5 miles southeast of High Rock, Cranberry Lake quadrangle	8.53
Along railroad for 1 mile west of Briggs, Oswegatchie quadrangle	9.39
Along road, from 0.4 to 1.0 mile west of Strifts School, Lowville quadrangle	9.80
Along trail for 0.4 mile northeast from Scanlons Camp, Oswegatchie quadrangle	7.30
Ilmenomagnetite of phacoidal hornblende granite gneiss (recrystallized in amphibolite facies):	
Along road for 0.8 mile from house west of center of Ormsbee Pond, Stark quadrangle	7.66
Along 0.5 mile of a traverse, 1.8 miles west to southwest of Cold Brook School, Stark quadrangle	5.05
Along road, from 0.6 to 1.0 mile south from Littlejohn School, Stark quadrangle	5.78
Magnetite of younger hornblende granite gneiss recrystallized in amphibolite facies:	
Along road, from 0.6 mile northeast to 0.6 mile southwest of Stone School, Russell quadrangle	1.14
Across outcrops from 0.6 to 1 mile southeast of Stone School, Russell quadrangle	1.90
From Blanchard Hill, 1 mile east of Whippoorwill Corners, Russell quadrangle [1]	3.78

[1] Is of single specimen.

The field and age relations of the younger granites are such that the regional dynamothermal metamorphism could have been effected at a time when the younger granite magma was beneath, but on its way up to, its present level of exposure. The metamorphism could therefore have been effected in the presence of fluids escaping from the magma and permeating the rocks locally by intergranular creep. The development of hornblende from pyroxene, and scapolite from plagioclase, in some of the mafic rocks requires such access of fluid carrying OH, F, Cl, and CO_2.

All the rocks northwest of the Diana and Stark complexes have been subjected to two periods of metamorphism. There is no evidence, however, that the earlier metamorphism was at a higher or lower temperature than the later. The temperature and depth are thought to have been similar for both periods of metamorphism. However, in the area southeast of the Diana and Stark complexes the possibility is not precluded that the younger granites were emplaced with accompanying local development of the amphibolite facies in the metasedimentary rocks, especially the marble, while at the same time the orthogneisses retained the mineralogy of the older developed granulite facies.

STRUCTURAL GEOLOGY

ORIGIN AND SIGNIFICANCE OF FOLIATION

The interpretation of gross structural fabric in this area is based largely on the study of the orientation of foliation planes in the rocks, and to a subordinate extent on bedding in the metasedimentary rocks and primary layering in certain of the igneous rocks. A discussion of the origin and significance of these planes is therefore in order.

FOLIATION OF METASEDIMENTARY ROCKS

Foliation of the metasedimentary rocks of the Grenville lowlands has been studied in great detail by Engel, who has published (1949b, p. 767–783) an excellent discussion from which the following summary is taken.

The lithologic trend and average dips are patterns imparted by the metasedimentary formations and rudely accordant orthogneisses. * * * In a few scattered beds in all rock types, and especially in these quartzose and silicated beds in the marble, the surfaces defined by sedimentary variations in texture and composition seem to have been little disturbed by recrystallization and reaction of components. * * * In other layers, less confidently interpreted as relict beds, the layering tends toward gneissic banding * * *. [Rather than representing a rigorous bedding foliation] it seems more reasonable to me that most of these gneissic layers represent beds in part sheared out and refoliated during severe stages of metamorphism * * *. Commonly, the well-defined folds in the siliceous and harder silicated beds, where opposite flanks of the folds subtend angles somewhat less than 80° or 90°, are accompanied, at least in the axial regions of the folds, by incipient to clearly visible surfaces of fracture and slip. These surfaces * * * tend to parallel the axial planes of folds to which they are genetically related * * *. Commonly, the more closely compressed the fold [of the bedding] the more prominent are these cross-cutting surfaces, and the less definite the outlines of the bedding and the fold * * *. As the folds become more closely compressed, their axial planes, and the essentially accordant [with axial plane] shear surfaces, tend to form successively smaller angles with the average lithologic dip and trend of the Grenville Belt * * *. In many places where the shearing-out is profound, * * * no fragments indicative of true beds remain at any position in the "ghost" of what was originally a fold * * *. Almost invariably a pronounced foliation, divergent from the initial horizons, permeates the relict bed fragments * * *. [One may estimate] that in about one-fourth to one-third of the outcrops of hard, siliceous and silicated gneisses, the dominant foliation is of secondary shear origin * * *. Probably less than one-third of the prominent layering in the marble represents bedding. These more mobile layers have been deformed "ahead" of the gneisses * * *. [A marked constriction and thinning of marble zones is especially prominent] on flanks of major folds in the Grenville series, whereas convergence, thickening, and duplication of marble are characteristic features of axial areas of folds * * *. Large parts of the thicker siliceous and silicated para-gneisses, as well as thinner, harder zones and beds clearly remained as [fairly] continuous sheets throughout the various stages of deformation. These sheets acted as guides to the direction of flow in the more mobile interlayers and zones * * *. [The] more mobile rocks * * * flowed not only in the directions of regional dips, but in the zones now occupied by cross folds, * * * laterally * * * along the surfaces and within the folded structures defined by harder lithologic units. * * * *There is little doubt that in many areas of the northwest Adirondacks a map drawn with the strikes and dips of the diverse, major rock surfaces and layers plotted under a single symbol does little practical violence to the facts, or to gross structural and stratigraphic reconstructions. The thick units of gneiss and marble remain as distinctive masses which outline roughly most major structures* * * *.[5] Some of the most serious errors of interpretation and generalization involving even gross regional features are possible where two or more major lithologic units are closely interfolded and sheared out to the extent that the original forms of these units are repeated and blended together.

It is highly probable that Engel's conclusions regarding secondary foliation are equally applicable to the marble and associated gneisses and harder beds of the metasedimentary rocks within the main igneous complex. The foliation plane within relatively mobile rocks may be parallel to the axial plane of a fold instead of to the bedding. The secondary foliation planes may also in turn have been folded.

The more apparent foliation of the major relatively resistant units of the metasedimentary rocks, however, is thought to be generally conformable with the bedding around major fold structures, except in the most intensely deformed zones. In all diamond drilling so far completed, the working hypothesis that the foliation is in general conformable with the primary layering of the metasedimentary rocks has proved generally satisfactory.

FOLIATION OF IGNEOUS ROCKS AND ORTHOGNEISSES

Where unmetamorphosed, the anorthositic gabbro mass shows a primary flow foliation and locally a layering thought to be due to gravity sorting of minerals during crystallization of the magma. Where intensely metamorphosed, the anorthositic gabbro mass shows a wholly secondary foliation parallel to the primary foliation and to the layering. Similar phenomena are shown in the metagabbro mass northeast of Stellaville.

Primary layering is found locally throughout the Diana complex. Secondary foliation is intense throughout the rocks of the Diana complex and is consistently parallel with the primary layering. Layering is present but not so obvious in the rocks of the Stark and Tupper complexes. At one locality, a primary flow structure without any later superimposed secondary foliation is present in the Stark complex. A secondary foliation, however, is intense throughout almost all the rocks of the Stark and Tupper complexes, where it is parallel to the primary layering and to the primary flow foliation.

The hornblende granite and alaskite everywhere have in general a foliation that is conformable with the schlieren and layers of country rock which they contain, and with the contact with the country rock. The foliation of the granites is also in general conformable with the foliation of the adjoining country rock. This is true both where the foliation is a primary flow structure, as in much of the southern tier of quadrangles, and where it is the product of intense deformation and recrystallization. The primary foliation is thought to have been largely controlled by the form of the walls. The secondary foliation is superimposed upon this primary foliation and conformable with it, insofar as observed, within the main igneous complex of this area. A secondary foliation parallel to the axial plane of the anticlinal structure, however, has been observed to cross the primary foliation on the plunging noses of some of the alaskite phacoliths in the Gren-

[5] Italics are by A. F. Buddington.

ville lowlands belt of metasedimentary rocks, where they have been most intensely deformed.

The sillimanite-microcline granite gneiss of the intensely deformed South Russell isoclinal synclinorium shows many sheared-out isoclinal puckers or chevron folds, a uniformity of orientation of planar structure, thorough recrystallization, and some evidence for large-scale fold structures—obscure so far as the foliation is concerned. There are two planes of foliation at an angle to each other in the microcline granite gneiss at the blunt ends of the belt north and west of Irish Hill School (Russell quadrangle). One foliation is indistinct and parallel to the curving boundary of the microcline granite gneiss with the wall rock. The other is across this foliation and parallel to the general northward trend of the belt of microcline granite gneiss as a whole.

The metasedimentary rocks of the Grenville series, together with their associated sheets of gabbro and differentiated quartz syenite complexes, were all strongly folded and complexly deformed before the emplacement of the younger granites at their present level of exposure. The intrusion of the younger granite magmas has obscured or obliterated the evidence of much of the older structure; it has also introduced new complexity.

In many places, the foliation within the hornblende and alaskite granite has an anticlinal structure. This does not necessarily mean that layers of granite have been folded into this structure. On the contrary, the structure may be a primary flowage structure consequent on the emplacement of crystallizing magma in preexisting folds with the concomitant development of foliation conformable to the walls. It is possible that, locally, uprise of the magma mass was the dominant factor, resulting in development of foliation parallel to a dome- or anticline-shaped surface of the advancing magma.

PLASTIC THINNING AND THICKENING OF METASEDIMENTARY ROCKS

The metasedimentary rocks have undergone marked plastic flowage as part of the intense deformation to which they have been subjected. The stronger igneous rocks, such as the sheets of quartz syenite gneiss, have undergone less complex large-scale distortion and appear to afford the best means to work out some of the major structural elements. The metasedimentary rocks have been deformed into more folds, and more complex folds, than the igneous sheets.

The plastic flowage that resulted in thinning and thickening of the metasedimentary rocks is particularly well shown on the plunging noses of several cross folds in the Russell quadrangle. Limestone forms thick lenses in the axial zone of a major cross fold at Endersbees Corners and Van House Corners and is thin or missing on the limbs to the east and south. The quartz-feldspar granulite northwest of Van House Corners is similarly much thicker on the nose of the fold than on the limbs to the east and south. Metasedimentary rocks also thicken markedly on the plunging nose of the cross fold north of South Russell. The discontinuity of the garnet-biotite gneiss in the southeast corner of the Canton quadrangle, southeast of Van Rensselaer Creek, appears to be due to pulling apart of a former continuous belt by plastic flow. The foregoing examples are thought to represent the kinds of phenomena that have gone on extensively throughout the metasedimentary rocks within the igneous complex, where the record is less easily deciphered.

MAJOR STRUCTURAL UNITS OF NORTHWEST ADIRONDACKS

There are three major structural units in the northwest Adirondacks. A large mass of granite gneiss and subordinate metasedimentary rocks lies in the extreme northwest, on the Alexandria Bay quadrangle, west of the area shown in figure 4. Adjoining it on the southeast is the Grenville lowlands unit, a broad belt, 25–30 miles wide, underlain by rocks consisting about 70 percent of metasedimentary rocks, with subordinate igneous rocks, largely granitic. The Grenville lowlands unit is in turn adjoined on the southeast by the main igneous complex, which comprises two subunits. The eastern subunit is the great anorthosite massif, around which on the north, west, and southwest is the other subunit, consisting predominantly of granitic and pyroxene-quartz syenitic rocks with an associated but subordinate amount of metasedimentary rocks. Within each of these large-scale elements there are subordinate structural units.

The area covered by this report does not include any of the granite mass on the far northwest or the anorthosite massif on the east. Only the great belt of metasedimentary rocks and the subunit of the main igneous complex that consists largely of felsic intrusives are included within the area of study, and only these will be discussed in detail. However, their structures are in part conditioned by and related to the other structural elements.

GRENVILLE LOWLANDS: GROSS STRUCTURE OF THE METASEDIMENTARY ROCKS

The gross structure of the metasedimentary and associated igneous rocks of the Grenville lowlands has

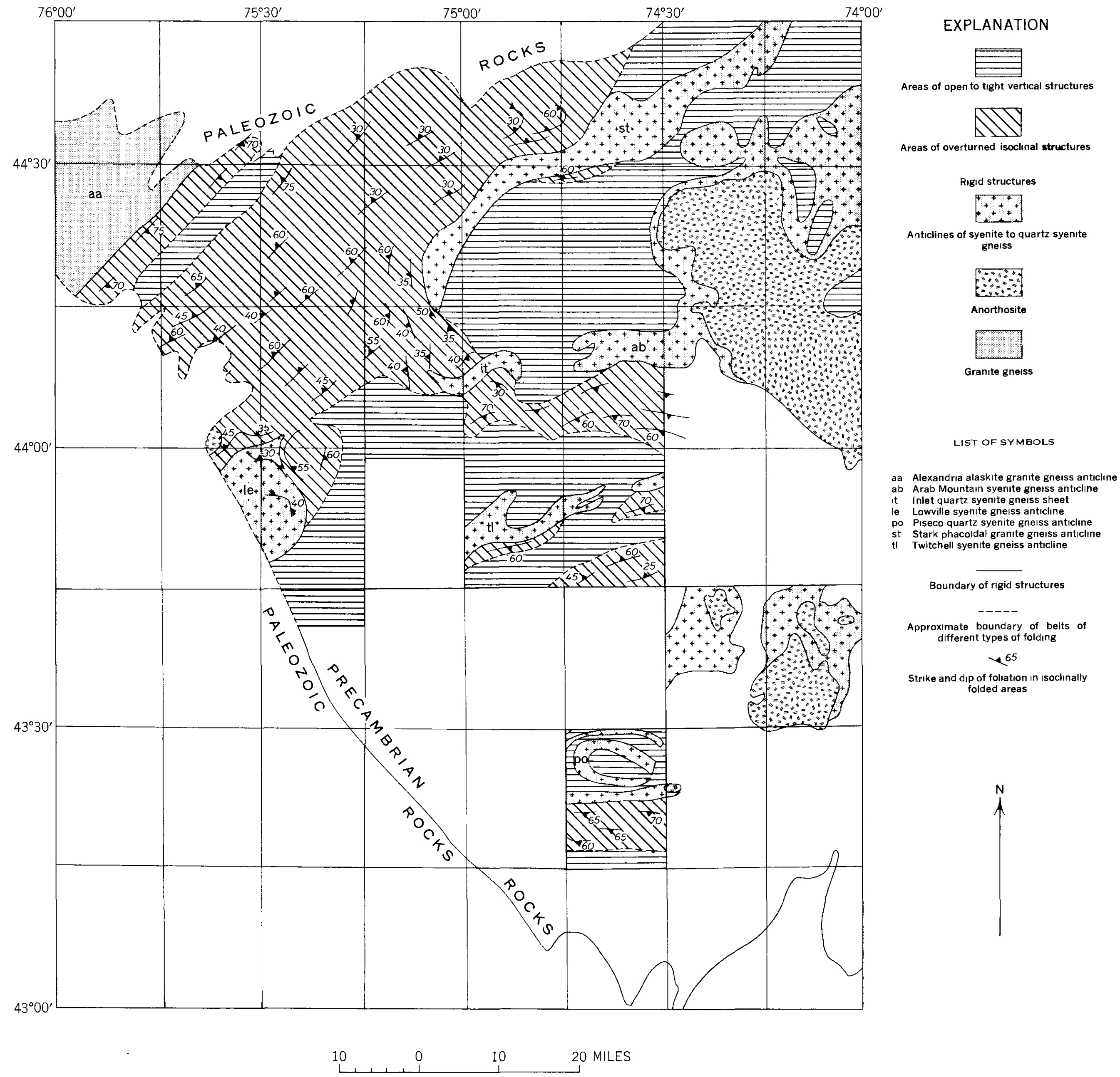

FIGURE 13.—Map showing relations of belts of isoclinal structures overturned toward more rigid elements in the western Adirondack area. Data not available for blank areas.

been previously summarized (Buddington, 1939, p. 237–241).

The general structure within the sedimentary rocks of the Grenville lowlands (figs. 13 and 14) is that of an asymmetric wedge-shaped prism. On the northwest, there is a narrow belt of isoclinal folds whose axial planes dip southeast; adjoining it, there is a narrow central belt of folds with steep axial planes; on the southeast lies a wide belt where isoclinal folds have axial planes dipping moderately northwest. In cross section, the axial planes of folds in the major metasedimentary unit are thus oriented in the pattern of an asymmetric fan. A brief description of these subdivisions follows.

The northwest belt consists of isoclinal folds overturned northwestward against a batholithic mass of granite gneiss; it is 2¼–4 miles wide and parallels the St. Lawrence River. The axial planes dip 60° or more to the southeast, and minor fold axes rake steeply.

The next belt to the southeast is about 6 miles wide;

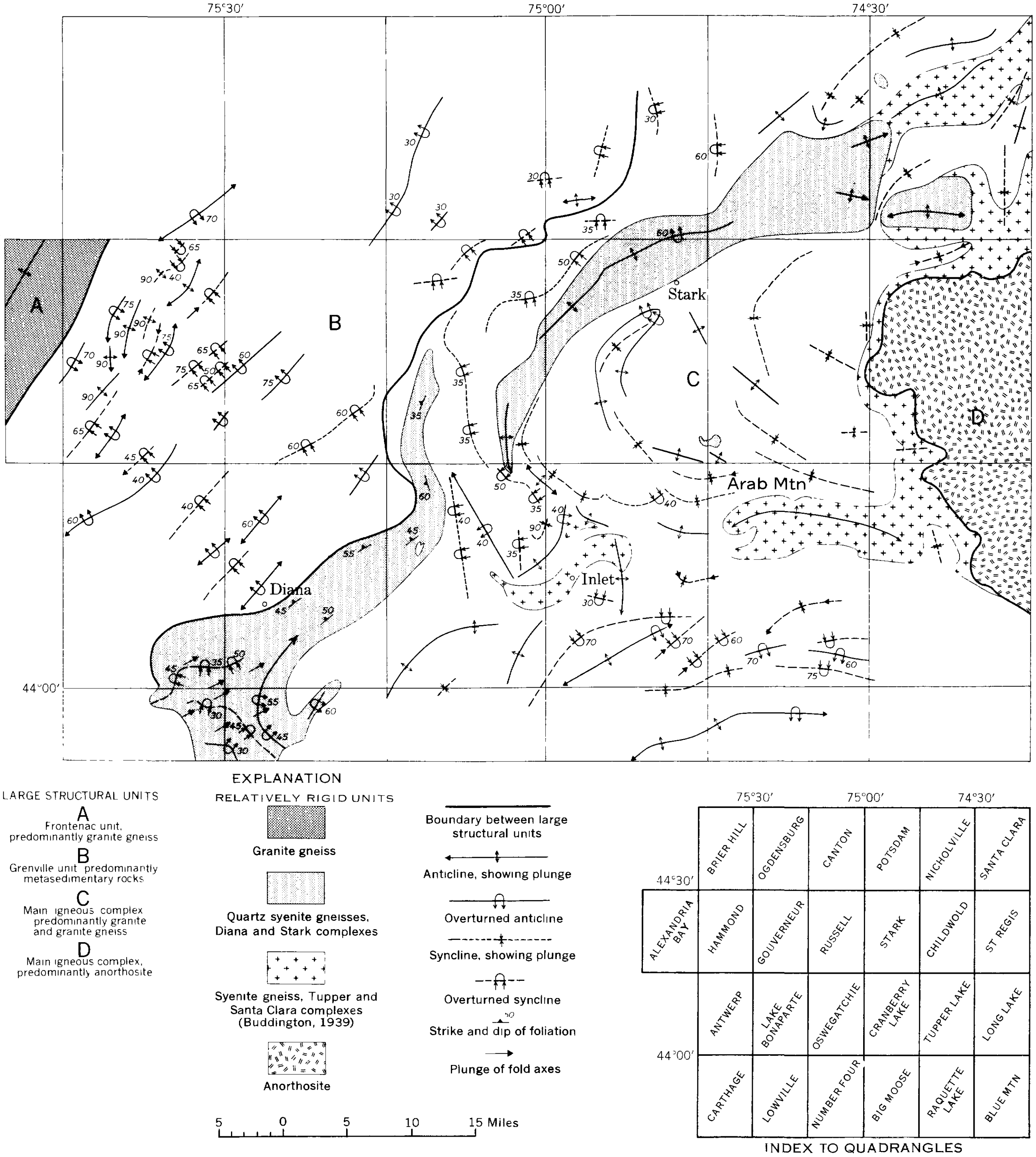

FIGURE 14.—Map of the northwest Adirondack area showing relations of structure axes and isoclinal folds overturned toward more rigid units.

in it the axial planes of the folds dip steeply or are vertical, and the fold axes plunge moderately to steeply at 25°–60°.

The southeasternmost belt is 15–20 miles wide. Within it the axial planes of major isoclinal folds are overturned to the southeast and dip northwest, and most of the minor fold axes rake steeply at 50°–90°. This southeast belt, considered in detail below, flanks the northwest border of the main igneous complex of the St. Lawrence County magnetite district. The belt extends northwestward to a line running from the southwest corner of the Hammond quadrangle approximately through its northeast corner. Across this belt, the degree of dip of the major axial planes in general decreases from vertical in the northwest to as low as 40° NW. in the southwestern part of the Potsdam quadrangle, 25°–30° NW. in the southeast corner of the Canton quadrangle, 30°–40° NW. in the western and northwestern part of the Russell quadrangle, about 60° NW. in the Edwards-Balmat area, and 45°–60° NW. across the central part of the Lake Bonaparte quadrangle (Reed, 1934; Martin, 1916; Brown, 1936, p. 240–243).

A major structural element near the southeast border of the belt of metasedimentary rocks is a great fault, originally noted by Buddington (1926, p. 96), and mapped and described by Brown (1936, p. 237, 243–244). Brown shows the fault as somewhat over 10 miles long, extending to the northeast from about 6 miles southwest of Balmat, and passing just southeast of Balmat Corners and northwest of Fullerville. Brown describes the fault zone as 200–300 feet wide, characterized by either minute shattering and jointing, or strong flow banding and microgranulation of the rocks involved, and by sporadic reticulation by quartz veinlets. He suggests that displacement on the fault is probably horizontal rather than vertical, with a maximum displacement of about 3 miles. Brown puts the time of faulting as later than the porphyritic granite gneiss and related to the instrusion of the nonporphyritic (alaskitic) granite mass that lies to the south of Kellogg Corners (Lake Bonaparte quadrangle). Gilluly (1945) describes one outcrop on the fault plane that strikes N. 40° E. and dips 65° NW. Gilluly differs from Brown in that he dates the age of this fault as later than all the intrusives. He bases this on the fact that the fault zone is characterized by low-grade metamorphic minerals such as albite and chlorite, and by mylonite. This seems to us the more probable interpretation.

BELT ON NORTHWEST FLANK OF MAIN IGNEOUS COMPLEX: STRATIGRAPHY AND STRUCTURE

The belt of metasedimentary rocks on the northwest flank of the main igneous complex lies exclusively within the southeast belt of isoclinal folds overturned to the southeast. Across the Oswegatchie, Russell, and Canton quadrangles, the metasedimentary rocks in general form three parallel zones of quite different lithology.

The first, or inner, zone on the flank of the main igneous complex consists of quartz-feldspar granulites and pyroxene-quartz-feldspar gneisses with interbeds of marble and pyroxene granulite. This zone is in part interleaved with and included in the quartz syenites and granites of the main igneous complex itself.

The second zone outward from the main igneous complex is the marble zone through Snyder Lake, Edwards, Stalbird, Endersbees Corners, and West Pierrepont.

The third zone is the migmatitic biotite-quartz-plagioclase gneiss, in part garnetiferous, which underlies the extreme northwest corner of the Oswegatchie quadrangle and the western edge of the Russell quadrangle, and extends northeastward through Fairbanks Corners, Kimball Hill, and north of Hamiltons Corners, becoming continuous with the sigmoid belt in the southeastern part of the Canton quadrangle. These rocks contain one or more marble beds that have been extremely deformed by plastic flow. The marble may form great bulges at noses of folds and may be pinched out wholly or partially on the flanks of folds. Locally, the gneiss also shows intense thinning and thickening as a result of plastic flow.

Many subordinate isoclinal folds have been identified within the rock of each of these zones. There has however, been so much distortion by flowage, magmatic intrusion, and intense complex deformation throughout the rocks, that the major structure has not been certainly determined. A great sigmoid fold that involves many varieties of rock on the Canton quadrangle has been thoroughly described by Martin (1916). The rocks of this fold are continuous with those on the Russell quadrangle, and the structure as a whole must be even more complex than that described by Martin.

The succession of rocks within the Grenville lowlands adjacent to the area of this report has been previously determined (Buddington, 1934, p. 136; Engel and Engel, 1953, p. 1055) and a comparison of the results is given in table 41. A subsequent study by Brown and Engel (1956) appeared while this report was being prepared for publication. The fundamental assumption for Buddington in working out the succession was that the marble of the Gouverneur belt plunged beneath the migmatitic garnet-quartz-feldspar gneiss near Halls Corners, Antwerp quadrangle, on the nose of an overturned isoclinal anticline. The sequence of rocks as worked out from this starting point has proved to be consistent with the succession inferred from the assumption, for the area of this report, that the sheet of the Diana complex is the overturned limb of an isoclinal anticline. The succession as worked out by Engel and Engel (1953), based on much extremely detailed work in the Gouverneur quadrangle, is similar.

Three kinds of interpretation have been offered for the type of structure in which the metasedimentary rocks of the belt on the Gouverneur quadrangle are

TABLE 41.—*Generalized stratigraphic sequence in the Grenville series, northwest Adirondacks*

Location	Buddington (1934, p. 136)	Engel and Engel (1953, p. 1055)	Buddington (this report)
(6) Boyd Pond and South Russell synclinoria.	----------	----------	Pyroxenic or hornblendic quartz-feldspar gneisses: local masses of pyroxene skarn and marble; biotite-quartz-plagioclase gneiss and its granitized equivalent, microcline granite gneiss.
(5) Northwest flank of the main igneous complex; Hamilton Corners, Russell, Barraford School, Manchester School.	----------	Upper feldspathic gneiss or feldspathic granulite.	Quartz-feldspar granulite, pyroxenic gneisses, and rare lenses of marble.
(4) East Pitcairn, Edwards-Stalbird, Endersbees Corners–West Pierrepont.	Crystalline limestone (in part dolomitic), white quartzite, and siliceous schists—all interbedded.	Upper marble belt. Dominantly dolomitic marble with interlayers of pyritic graphitic schist, feldspathic, biotitic and pyroxenic gneiss.	Marble with local white quartzite beds, diopside granulite, and some tremolite schist and biotite gneiss.
(3) Fairbanks Corners, Kimball Hill, Hamilton Corners, West Parishville, and southwest of Browns Bridge; Hermon.	Garnet-bearing gneiss with intercalated layers of amphibolite and local interbanded quartzite-limestone rock, and beds of schist.	Quartz-biotite-feldspar migmatite, in part garnetiferous and sillimanitic. Includes several marble zones and numerous thin interlayers of amphibolite.	Garnet-biotite migmatitic gneiss and biotite-quartz-feldspar gneiss, each with interlayers of amphibolite. Occasional marble layer.
(2) Somerville-Gouverneur pyrites belt; Gouverneur and Canton quadrangles.	Crystalline limestone, locally with occasional zones of interlayered white quartzite, quartz mesh limestone, and quartzite.	Dominantly marble with scattered interlayers of quartzite and feldspathic and pyroxenic gneisses.	
(1) Belt east of Black Lake, Hammond quadrangle.	Biotitic-feldspar gneisses, white quartzite, limestone, and pyroxenic feldspar gneiss.	Complexly interlayered marble, quartzite, and biotitic, feldspathic, and pyroxenic gneisses.	

involved. Cushing and Newland (1925, p. 50–68) interpreted the structure as a series of isoclinal folds overturned toward the southeast; Gilluly (1945) suggested the possibility that the structure consists of large recumbent folds and thrusts analogous to Alpine nappe structure; and Brown and Engel (1956) have proposed that a series of isoclinal folds overturned toward the southeast have been refolded by a northeastward movement and drag of the lowlands block of metasedimentary rocks relative to the southwest movement of the more rigid main igneous complex of the Adirondack massif. All three interpretations assume that isoclinal folds are overturned to the southeast. The hypothesis of nappe structure does not seem consistent with the intensive plastic flow to which these rocks have been subjected, or with the degree to which formations can be followed around fold structures. The hypothesis of refolded isoclinal folds has much evidence in its favor, but so much of the area of the metasedimentary rocks in the Canton, Potsdam, Russell, and Oswegatchie quadrangles is obscured by drift that only more detailed work will warrant a definite interpretation.

STRUCTURE NORTH OF COLTON

The rocks in the vicinity of Colton (Potsdam quadrangle) occur in a deep easterly embayment in the otherwise generally northeast-trending formations. The pyroxene syenite gneiss and alaskite gneiss sheets south of High Flats appear to be in an isoclinal anticlinal fold overturned to the southeast. The belts of migmatitic garnetiferous quartz-feldspar gneiss and amphibolite that underlie the area south of West Parishville and southwest of Browns Bridge occur in structures for which alternative interpretations have to be considered, and no positive solution is possible with present data. They may be normal isoclinally folded belts with elliptical closures; they may represent a single sheet of gneiss with an original northerly dip compressed into tight U-shaped folds; or they may constitute an original belt of gneiss with a complex isoclinal anticlinal structure overturned to the southeast, which possibly was refolded into tight, U-shaped structures with a northerly strike. The marble between the two belts has an anticlinal structure.

The nature of the structure at the north ends of the two belts of gneiss is critical but indeterminate.

A very thin band of gneiss appears to be continuous around the north end of the isoclinal anticlinal(?) marble zone in the belt of gneiss southwest of Browns Bridge, but this is not positive. A similar situation exists at the north end of an isoclinal anticline(?) of a marble zone in the gneiss 1.5 miles south of Edwards.

STRUCTURE IN NORTHWEST PART OF RUSSELL QUADRANGLE

So much of the bedrock is covered by drift in the northwest part of the Russell quadrangle that structure is difficult to work out.

The southwest-plunging nose of an isoclinal anticline overturned to the southeast is clearly shown southwest of Hermon.

The structure of the belt of gneiss between Mott School and Kimball Hill has not been studied in sufficient detail to warrant discussion as to whether it is actually a single sheet of gneiss or has been duplicated by isoclinal folding.

Northeast of Stellaville there is an anticline several miles across that involves a complex series of rocks. The axis of the anticline strikes east, but there has been intense overturning to the southeast in the southeastern part. The core of the anticline is alaskite gneiss in which the foliation is horizontal or gently dipping on the south, and gently dipping to the north on the Canton quadrangle. Overlying the granite core on the south and west is a sheet of metagabbro. The foliation of the metagabbro appears to have a strike of N. 60° W. to N. 80° E. throughout, and a steep dip. The foliation of the metagabbro on the west end of the anticline thus has a foliation striking almost at right angles to its contact with the granite gneiss. A set of faults strikes north-northeast across the metagabbro. The metagabbro is entirely surrounded by granite gneiss and its relations are those of a xenolith within the granite gneiss. The metagabbro is interleaved with granite in the contact zones and crosscut by granite pegmatite veins.

Between the inner zone of granite gneiss and metagabbro and the outer crescent of granite gneiss there is a narrow belt of rocks of complex character. They consist of a homogeneous almandite-quartz-feldspar rock with associated marble beds, pyroxene gneiss, and a little garnet-mica schist. The almandite-quartz-feldspar rock has a most unusual massive appearance and is a type found to be restricted exclusively to contact zones between alaskite gneiss phacoliths and marble (Buddington, 1929). The garnet-rich rock occurs in thick masses adjacent to the granite and also as sheets a few inches to a few feet thick interlayered in the marble and gneiss. In all other zones in which this type of garnet rock occurs, para-amphibolite is present, but is has not been seen in this belt.

An outer part of the anticline is formed by a crescent-shaped sheet of alaskite gneiss. This gneiss locally carries considerable hornblende. The structure of this sheet resembles half of a hollow cylindrical mass plunging to the southwest.

Marble is thought to surround the complex of igneous rocks completely, as an outer part of the anticline.

All the igneous rocks of this structure are intensely deformed and recrystallized. Garnet has developed locally in the metagabbro, and the granite is intensely sheared, with small relict porphyroclasts.

The discordance between the structure of the foliation of the metagabbro sheet and the granite gneiss in the west part of the anticline is not understood.

FOLDED GABBRO SHEET, PIERREPONT SIGMOID

In the southeast corner of the Canton quadrangle a metagabbro sheet is involved in an isoclinal fold whose axial plane strikes northeast. The fold, overturned strongly to the southeast, plunges to the northwest and is but a part of a great sigmoid isoclinal fold described in detail by Martin (1916) as the Pierrepont sigmoid.

GRANITE PHACOLITHS

There are several granite phacoliths in the belt of metasedimentary rocks along the northwest border of the main igneous complex. Some, such as the alaskite mass southwest of Colton and the one northeast of Stellaville, just described, have a domical foliation and are interpreted as the top part of granite bodies emplaced in the crests of normal anticlines. Others were emplaced in cross folds. These bodies include the mass at Hawk Ledge (Potsdam quadrangle), which was emplaced in a synclinal cross fold, and the body west of Stalbird (Russell quadrangle), which was emplaced on the nose of an anticlinal cross fold. The horseshoe-shaped alaskite mass north of Browns School (Potsdam quadrangle) is also related to emplacement in a fold.

JUNCTION ZONE OF GRENVILLE LOWLANDS UNIT AND MAIN IGNEOUS COMPLEX

The bodies of igneous rock within the marble belt just northwest of the border of the main igneous complex are in small part massive and undeformed, in part granoblastic and gneissic, but in large part intensely deformed, with the development of mortar and augen gneisses with porphyroclasts in a mylonitic groundmass. Locally the rock is mylonitized throughout or has an actual schistose appearance in the field. The average diameter of the feldspar grains in the mortar is less than 0.2 mm. On the Oswegatchie quadrangle the pyroxene-quartz syenite gneiss of the outer part of the main igneous complex itself is a mortar or augen gneiss. The predominant rock of the outer part of the main igneous complex, however, is a granoblastic gneiss with average grain of 0.5 mm or more. Schistose facies of syenite gneiss occurring as sheets in the marble were observed 0.4 mile north-northwest of the cemetery 0.6 mile northwest of Pond Settlement (Russell quadrangle) and just east of the road 1.5–0.7 mile southwest of High Flats (Potsdam quadrangle). Mylonitic and intensely crackled alaskite gneiss sheets are found in marble from a mile east-southeast of Moores Corners (Russell quadrangle) to Sellecks Corners (Stark quadrangle). The alaskite gneiss west of Stalbird is a mortar or augen gneiss consisting of small porphyroclasts of feldspar in a mylonitic groundmass whose grain averages about 0.05 mm.

The hornblende granite gneiss forming part of the main igneous complex at the border, from 0.5 mile to 1.5 miles northeast of Owens Corners, is a much crackled rock with chloritic slickensided surfaces. The border itself here is also wavy.

In the gneisses of this border zone that have a mylonitic mortar or are strongly fractured, the ferromagnesian minerals have very generally undergone retrograde metamorphism. Chlorite, carbonate, and sphene are the secondary minerals.

North of Carthage, especially west of Mount McQuillen, there is a contact zone in which quartz syenite has formed an intrusive breccia with anorthosite. This igneous breccia has been intensely deformed to yield ultramylonite. The anorthosite fragments have in large part been drawn out into bands and changed to a dense greenish rock with porphyroblasts of sphene. The greenish bands were previously mistaken for syenite layers, but new exposures show that they exhibit a complete gradation into little-deformed blocks of anorthosite.

The junction between the main igneous complex and the great wedge of metasedimentary rocks that underlies the Grenville lowlands has been one where movement and deformation have occurred many times.

A belt 5–8 miles wide within the main igneous complex near the northwest border is formed by rocks isoclinally folded and strongly overturned to the southeast. All the major types of rock in this belt—such as metasedimentary rocks, gabbro, syenite, quartz syenite, hornblende granite, alaskite, and microcline granite gneiss—have undergone plastic flow and metamorphism and are gneisses. The southeast border of the belt is formed by younger granite southwest of the Inlet mass of quartz syenite and by the Stark complex on the Russell, Stark, Potsdam, and western part of the Nicholville quadrangles.

MAIN IGNEOUS COMPLEX

The relations of many of the inferred major structures within the main igneous complex are shown in a series of generalized structure sections (pl. 2). These cannot be depended upon for accuracy of detail but are largely for the purpose of presenting the types of structure that are inferred to exist.

The thick sheet of differentiated igneous rock constituting the Diana and Stark complexes has acted as a more competent layer than the metasedimentary beds during folding. The beds of the Grenville series between the Diana and Stark complexes are thought to have been much more complexly folded than the underlying igneous sheet. The younger granite masses must locally have broken through the underlying quartz syenite sheet and entered the complexly folded metasedimentary beds above. A part of the west limb of the igneous sheet forming the Stark anticline shows clear evidence of such disruption by the younger granite.

LOWVILLE ANTICLINE AND DIANA COMPLEX

The Diana complex northeast of Natural Bridge is thought in large part to be the strongly overturned southeast limb of a great anticline. Between Natural Bridge and Lowville the width of exposure of the Diana complex expands greatly in consequence of a series of complex folds.

In the southeast corner of the Antwerp quadrangle and the northeast corner of the Carthage quadrangle, the rocks are isoclinally overturned to the south and southeast, whereas in the Lowville quadrangle they are isoclinally overturned to the west and southwest.

SHEETLIKE FORM OF QUARTZ SYENITE COMPLEXES

The concept that folds, including tight isoclinal folds, may occur in sedimentary formations several miles thick, parts of which are thick, relatively rigid members, is accepted by all geologists without question. However, the concept that thick igneous sheets may also be tightly folded is often treated with much skepticism. One may, nevertheless, cite the deformation of the Stillwater complex, for which there is clear evidence that it represents a stratiform sheet (much more than 11,000 ft thick) turned up on edge and in part overturned (Peoples, 1933).

Certain structural evidence that most of the Adirondack quartz syenitic material has a sheetlike form is herewith presented. In its northern part the Diana complex is clearly a simple sheet, conformably overlain and underlain by metasedimentary beds. To the southwest it is thicker and its structure is more complex. The Stark complex is overlain conformably by metasedimentary beds, and the foliation of each is conformable with the other. Both a floor and a roof for the Tupper complex are clearly demonstrable. On the Tupper and Saranac quadrangles the floor of this complex consists of anorthosite with overlying local relict screens of metasedimentary beds. The foliation and contact planes of all three types of rock are conformable. On the Long Lake quadrangle the Arab Mountain sheet of quartz syenite plunges as an anticlinal nose beneath a conformable roof of metasedimentary beds, and on the Saranac quadrangle (Buddington, 1953) the quartz syenite of the Bloomingdale syncline, and that shown on the southeast corner of the Saranac quadrangle, is overlain conformably by metasedimentary beds. The evidence that the Tupper complex, one of the largest of the quartz syenite complexes in the Adirondacks, is a sheet with a generally conformable floor and roof of metasedimentary beds is thus unequivocal. Again, on the Big Moose quadrangle, the Twitchell Mountain quartz syenite mass has anticlinal structure and a conformable floor and roof of granitized metasedimentary rocks.

Microscopic examination of the rocks of these sheets shows that the degree of granulation and recrystallization is intense and consistent with the tight folds into which these sheets have been deformed.

AREA FROM INLET SHEET TO SOUTH END OF STARK COMPLEX

Northwest of the fayalite-ferrohedenbergite granite and south of the south end of the Stark complex is an area of the most complicated structure in the northwest Adirondacks. The axes of some of the folds here are curved as though the result of crossfolding or refolding of an earlier set of folds, though they may belong to a single period of deformation.

This zone of complicated folds is also the location of the most intense development of magnetite deposits in the northwest Adirondacks.

INLET SHEET

The east end of the Inlet quartz syenite gneiss sheet has the form of the south-plunging nose of an anticlinal fold. The rest of the sheet has a gentle southerly dip flattening on the north. The rock north of the fayalite-ferrohedenbergite granite is hornblende granite that lies on the south side of an anticlinal axis. The original structure of the Inlet sheet can only be guessed at. It may have had an anticlinal structure, but, if so, the north limb has been cut out by the younger granite, except for the lens of pyroxene-quartz syenite gneiss north of Little River.

SCHOOL NO. 15 SYNCLINE

East of the Diana complex, in the northern part of the Oswegatchie quadrangle, is a synclinal structure whose axis is indicated by the belt of metasedimentary rocks through School No. 15 (Oswegatchie quadrangle). The synclinal structure is seen just east of Briggs and north of Mud Creek. The northern part of the syncline is an isoclinal fold strongly overturned toward the east, with a generally westward dip of the foliation at about 40°. At the south end the syncline plunges to the north, the lower part of the fold is exposed, and the overturned isoclinal character is not so obvious.

COLTON HILL AND SUNNY POND ANTICLINES

To the east of the School No. 15 syncline is the Colton Hill anticline, which strikes about N. 25° W., is isoclinally overturned to the east-northeast, and has a general dip of about 40° WSW. In the north, this anticline involves alaskite gneiss and the structure plunges north; in the south, hornblende granite gneiss forms the core.

About 1¼ miles north of Streeter Mountain the axis of the Colton Hill fold strikes south-southeast, directly into another anticline whose axis strikes about N. 55° E., almost at right angles to the Colton Hill anticline. The axis of this fold, the Sunny Pond anticline, can be traced northeastward through Sunny Pond towards Lost Pond (Cranberry Lake quadrangle). Hornblende granite forms the core of the anticline and is overlain on both flanks by alaskite. Between Sunny Pond and Lost Pond the foliation dips uniformly about 40° SE., as though the anticline were strongly overturned to the northwest; whereas at the southwest end the fold plunges southwest as a rather open structure.

The curvature of the Inlet sheet to the south is roughly conformable to the respective trends of the Colton Hill and Sunny Pond anticlines in their southern parts, and all the structural forms are probably related.

NEWTON FALLS ANTICLINE

The axis of an anticlinal dome of foliation in granite strikes northwest from Newton Falls. It plunges to the southeast at Newton Falls and to the northwest about 3½ miles northwest of Newton Falls. This anticlinal structure crosses the northeast corner of the Oswegatchie quadrangle. It appears to be related to a bend in the structure at the south end of the south-plunging nose of the Stark anticline in the older phacoidal granite gneiss. The planar structure is obscure on the nose of the anticline northwest and west of Newton Falls.

The crest of the anticline has a foliation with gentle dips. The north limb, however, steepens to vertical along the valley southwest of Spruce Mountain (Cranberry Lake quadrangle). The south limb dips about 40° S. The axial plane of the fold west of Newton Falls thus dips about 65° S., indicating relative overturning toward the north. The granite in the vicinity of Newton Falls is much contaminated with metasedimentary rocks and contains numerous inclusions of them. An included layer of skarn occurs 1 mile north-northwest of Newton Falls.

The structure south and southeast of Newton Falls is complex, and critical areas are covered. The top of Rocky Mountain is composed of migmatitic pyroxenic and biotitic gneisses. Much of the hill is of biotite gneiss. This locally shows isoclinal plications dipping gently west. The relation of the granite mass east of Rocky Mountain to the Stark anticline is uncertain. One could entertain the hypothesis that it is part of the main anticline overturned toward the east, thus constituting an isoclinal fold with axial plane dipping west, but evidence for this is indecisive. Under such a hypothesis the metasedimentary rocks and associated granite in the vicinity of Newton Falls would occupy a saddle in the crest of the anticline.

STAR LAKE AND BENSON MINES SYNCLINES

The Newton Falls, Sunny Pond, and Colton Hill anticlines in alaskite gneiss and hornblende granite gneiss almost enclose on three sides a complex synclinorial structure that involves metasedimentary beds and sillimanite-microcline granite gneiss with local sheets of alaskite gneiss and hornblende granite gneiss. This synclinorium includes at least two synclines, one north from Benson Mines and one north through Star Lake. The whole area is so obscured by overburden that the structure cannot be ascertained with certainty.

One isoclinal syncline strongly overturned to the east, with foliation dipping about 35° W., includes the basin of Star Lake and extends north for at least 3 miles.

The Benson Mines syncline extends as a fairly well defined structure a little east of north from Benson Mines for a couple of miles. Beyond this the rocks and the foliation turn to the northwest and pass south of the south end of the Stark anticline. Their foliation dips uniformly to the southwest. There are inadequate outcrops to work out the structure. A synclinal structure strongly overturned to the northeast, however, is inferred. The Benson Mines syncline lies between the Star Lake syncline, which is strongly overturned to the east, and the Muskrat Pond anticline, which is strongly overturned to the west and northwest. It thus constitutes a narrow zone toward which the adjoining rocks are strongly overturned from opposite directions. Subordinate folds are present. Refolding of preexisting folds may be involved in the structures in the vicinity of Newton Falls and Benson Mines. Detailed study of the structure at Benson Mines is being made by the geologists of the Jones & Laughlin Ore Co. The results are not published.

SOUTH RUSSELL AND BOYD POND SYNCLINORIA

If the interpretation is correct that the Diana complex on the Russell quadrangle is the east limb of an overturned anticline and the Stark complex is on the core of an anticline, then the belt of rocks through South Russell between the two may be inferred to have a synclinorial structure. The rocks are so closely and isoclinally folded that the synclinorial structure cannot be checked by internal evidence from the rocks themselves. No evidence contrary to the hypothesis, however, has been found, and it seems the best tentative interpretation. The rocks of the belt are continuous with those of the Benson Mines syncline.

There is convincing evidence around Boyd Pond of a complex synclinal structure that is isoclinally overturned to the southeast and plunges to the northeast.

The rocks of both the Boyd Pond and South Russell synclinoria consist of metasedimentary rocks; microcline granite gneiss, in large part sillimanitic; sheets of alaskite gneiss; local sheets of hornblende granite gneiss; and rare lenses of metagabbro gneiss. The metasedimentary rocks consist largely of biotite-quartz-plagioclase gneiss, pyroxene-quartz-feldspar gneisses, local pyroxene skarn masses, marble lenses, and considerable hornblendic quartz-feldspar gneisses and migmatites.

It is inferred that the rocks of the synclinoria are much more complexly folded than the thick quartz syenite sheet that is assumed to lie below and to be continuous between the outcrop of the Diana and Stark complexes.

The South Russell synclinorium is offset from the Boyd Pond structure by a thick sheet of hornblende granite gneiss.

The belt of rocks of the Boyd Pond synclinorium becomes narrow across the Potsdam quadrangle and pinches out on the Nicholville quadrangle. This appears to be due, in part at least, to a southwest plunge of the syncline in this portion of the structure.

In general foliation in the rocks of the South Russell synclinorium dips about 35° W. The south ends of the granite gneiss masses near Carr Pond, a mile north of Partlow Pond and three-fourths mile northwest of Irish Hill School, have an anticlinal structure plunging south relative to the surrounding metasedimentary beds.

Several subordinate folds are distinguishable within the Boyd Pond synclinorium. A syncline is clearly indicated around Boyd Pond. A northward-plunging anticline is shown about 1¼ miles north of the west tip of Boyd Pond.

The belt of metasedimentary rocks from Backus Hill to Shingle Pond appears to have an isoclinal synclinal structure on Backus Hill. Northeast of Shingle Pond the belt of metasedimentary rocks frays out into the granite gneiss northeast through Austins Corners as a series of isolated included layers.

Subordinate folds are also indicated east of Horseshoe Pond. The granite prong northeast of Clare has a synclinal structure relative to the metasedimentary rocks on either side. The tongue of metasedimentary rocks west of Big Swamp is on the crest of an anticline plunging gently southwest.

A well-defined syncline plunging west-southwest is shown about 1¼ miles north of South Colton (Potsdam quadrangle). Alaskite gneiss, an arc of metasedimentary rocks, and sillimanite-microcline granite gneiss are all involved in the structure. The syncline is an isocline overturned toward the south with limbs dipping 30°–40° N. Another overturned isocline with moderate westward dip and a north-northwest pitch is clearly defined about a mile north-northwest of Parishville. A similar group of rocks is involved.

Between South Russell and Sellecks Corners there are in the Boyd Pond synclinorium a number of cross folds striking N. 25°–40° W., a corresponding linear structure about N. 40°–55° W., and some small crumples.

STARK ANTICLINE

The quartz syenite series of rocks forming the belt that extends southwest from Santa Clara to south of Degrasse for more than 60 miles is interpreted in general as having an anticlinal structure, though the structure is complex on the Nicholville quadrangle, where the lineation in part has an easterly plunge.

On the west side of the Nicholville quadrangle and the eastern half of the Stark quadrangle, the axial plane of the anticline is overturned toward the south-southeast, with dips of 40°–60° NNW. on the north limb and dips of about 60° NNW. on the southeast limb. On the western half of the Stark quadrangle the anticline is asymmetrical, with steep to vertical dips on the southeast limb and dips of 40°–60° on the northwest limb. Dips are gentle on much of the broad crest of the anticline across the Stark quadrangle, generally about 20°–30°. Interlayered green pyroxene-quartz syenite gneiss and red hornblende granite gneiss form the core of the anticline in the northeast part of the Stark quadrangle. This is overlain by phacoidal hornblende granite gneiss. The anticlinal structure is also brought out by the presence on the western part of the Stark quadrangle of remnants of a more syenitic border rock forming the outer portion of each limb.

On the Russell quadrangle the northwest limb of the anticline, from Big Swamp south to a point about 1 mile northwest of Long Pond, is cut out by younger intrusive granite. The foliation of the older phacoidal granite gneiss for much of this distance strikes into the contact with the younger granite, instead of being parallel to it. Throughout this portion of the structure the east limb has a general dip of 35°–50° SE., with occasional secondary small open folds or rolls superimposed upon it.

The Stark anticline again has its full development from one-half mile north of Long Pond (Russell quadrangle) for 3 miles south to the termination of the phacoidal granite gneiss on the Oswegatchie quadrangle. This part of the belt constitutes the southward-plunging nose of the anticline. The foliation dips about 30° on each limb. The planar structure is obscure to indistinguishable on the crest or axial zone of this plunging nose of the anticline, and in its place is developed a very strong linear structure which plunges 10°–15° S. There seems to be some kind of discontinuity in the structure along an east-west line about one-half mile north of Long Pond.

The phacoidal granite gneiss on the Russell quadrangle, here and there, contains inclusions of metasedimentary rocks. The ore bodies of the Clifton mine are in one such inclusion. In the vicinity of Clifton, many thin layers of metasedimentary rock are included in the granite in a fashion similar to the igneous breccia southwest of Red School (Oswegatchie quadrangle).

CLARE-CLIFTON-COLTON BELT OF METASEDIMENTARY ROCKS AND GRANITE

A belt of metasedimentary rocks and associated granitic intrusions forms a great arc or crescent which lies almost wholly in the townships of Clare, Clifton, and Colton, and which we termed the CCC belt (Buddington and Leonard, 1944). It lies east of the Stark anticline. Starting north of Blue Pond (central Stark quadrangle), it swings southwest and south between the Clifton mine and Wilson Mountain—where it is about 10 miles wide—and runs southeast through the village of Cranberry Lake and then east to a point about 1 mile east of Childwold Station, where it thins out. However, remnants are found far beyond, along the general line of strike, as on the north shore of Tupper Lake on the east side of the Tupper Lake quadrangle. Remnants of metasedimentary rocks of this belt are also found about 4 miles northeast of Blue Pond (Stark quadrangle) at the Brunner Hill magnetite prospect. The main belt has a total length of a little more than 30 miles. Fourteen magnetite deposits are located in this belt, and it is therefore of much economic significance.

The belt is bordered largely on the west and south by the quartz syenitic rocks exposed in the Stark and Arab Mountain anticlines, and on the east and north by the microcline granite gneisses of the southwest part of the Childwold quadrangle.

Much of the belt appears to have a synclinorial structure, but as it is closely folded and much involved with granitic rocks, the structure has not been thoroughly worked out. Some features and interpretations of the structure will be discussed.

JARVIS BRIDGE SYNCLINE(?)

A belt of metasedimentary rocks and associated sheets of microcline granite gneiss extends from opposite Matilda Island in Cranberry Lake through the valley south of Marble Mountain, across the swamp north of Chaumont Pond, through Jarvis Bridge, and northwest through the valley southwest of Spruce Mountain, to a point about 2 miles north-northwest of the southeast corner of Russell quadrangle. This belt is about ⅓–½ mile wide and 8½ miles long. A clearly defined synclinal structure plunging gently south-southeast is shown near the northwest end, just north of the township line. Southeast of here, along the valley of Spruce Creek, the metasedimentary rocks have a generally vertical dip and are interpreted as involved in an isoclinal synclinal fold, for the dips both to the northeast and southwest gradually flatten and dip towards each other. The structure of the belt from Jarvis Bridge east to the main highway is so obscured by overburden that no definite interpretation can be made. The swamp area north of Chaumont Pond in particular is critical, but without outcrops. The structure of the belt south of Marble Mountain may from some indications be interpreted as a syncline having a steep axial plane. The rocks of State Ridge and Twin Mountain dip gently (15°–25°) north on the south side but become increasingly steeper (60°–90°) toward the valley on the north. The rocks on the north side, in Marble Mountain, are about vertical. East of the highway the foliation dips uniformly south. The foliation within the granite forming Buck and Marble Mountains may give a clue to the nature of the deformation here. From Buck Mountain eastward to Lost Pond, the dip changes successively from moderately northeast, through vertical, to moderately southeast. On the east side of Cranberry Lake, the foliation of the granite dips north. The changes in direction of dip may be interpreted as the result of a local northward movement that has gone so far as to overturn the rocks along the west side of the north arm of Cranberry Lake. Both the sharp bend or drag in strike at this locality and the tight folding northeast of Cranberry Lake village conform to the hypothesis. The metasedimentary rocks and the mass of granite east of State Ridge have been similarly deformed.

BRANDY BROOK BELT

The structure of the belt of metasedimentary rocks and associated intrusive rocks that extends from Silver Pond (Cranberry Lake quadrangle) to beyond Childwold Station has not been positively determined. There are too few outcrops in critical areas. The problem is whether the metasedimentary rocks are involved in a synclinal structure or whether they have a generally northerly dip with local gentle rolls. At the head of Brandy Brook Flow there is definitely a broad gentle syncline. North of this, however, outcrops are inadequate to prove whether the sedimentary rocks again roll over and dip beneath the microcline granite gneiss, or whether they overlie the granite gneiss. Northwest from Brandy Brook Flow the mineralized skarn horizon steepens from 25° NE. to about 60° NE. in the hill east of the road. It was originally thought that the synclinal structure at the head of Brandy Brook Flow indicated a major feature and extended northwest and became more closely compressed. No evidence of this is available, however, and the Brandy Brook syncline may plunge northwest and be only a local warp on structure that dips generally north-northeast. Northwest of Childwold Station there is no evidence of synclinal structure and the rocks dip uniformly north.

Northwestward from Town Line Pond, through Hardwood Island (Tupper Lake quadrangle) to the main road north of Shurtleff, an anticline plunges northwest and involves rocks predominantly of microcline granite gneiss, but much contaminated and including local layers of gneiss of the Grenville series. Most of the included gneiss layers are amphibolite, but locally the microcline granite gneiss is pyroxenic as from skarn, and limestone float at a point one-half mile northwest of Pine Pond may well indicate the presence of limestone in place nearby. The weathered surface of much of the granite gneiss has the appearance of a thin-layered migmatite or a phantom breccia.

PARISH SYNCLINORIUM

A belt of metasedimentary rocks and associated granite with layers of metasediments extends from the Oswegatchie River north through the region of the Parish ore deposit, and then northeast for more than 5 miles.

The granite mass between Spruce Mountain (Cranberry Lake quadrangle) and Newbridge (Stark quadrangle) has an anticlinal structure. On the domical surface of the anticline are several minor folds with gentle dips. Locally some metasedimentary rocks are associated with the granite. The structure between Spruce Mountain and Buck Mountain has not been determined because of inadequate exposures. Northeast of this anticline, the next clearly defined structure is the anticlinal granite mass north of Spruce Mountain (Stark quadrangle). Traced to the northeast, the anticlinal structure of the granite south of Newbridge appears to pass into foliation that dips east to southeast along the flank of the Stark anticline. The belt of metasedimentary rocks and associated granite between the Spruce Mountain (Stark quadrangle) granite anticline on the northeast and the Stark anticline and the anticline south of Newbridge on the west thus has a synclinorial character with generally steep dips. The belt of metasedimentary rocks dies out to the northeast, where the structure plunges southwest.

SPRUCE MOUNTAIN PHACOLITH (?)

The granite mass north of Spruce Mountain (Stark quadrangle) has a domical anticlinal structure between Spruce Mountain and Tunkethandle Hill and a uniform, nearly vertical structure, suggestive of a simple sheet, northeast of Tunkethandle Hill. A question arises whether this structure is to be interpreted as indicating a phacolith or a lenslike sheet with a broad bulge at the south. If the mass is a phacolith, the rocks on opposite sides of the anticline should be similar. The stratigraphy, however, is not well enough known, nor are the rocks well enough exposed to check certainly on this problem. The available evidence is consistent with such a hypothesis. The very steep plunge of the foliation at the south end of the mass and the uniform dip northeast of Tunkethandle Hill are consistent with a hypothesis of a lenslike sheet with a plunge down dip. On this hypothesis, the structure at the south end should be such as to indicate that the granite came in along a foliation plane and that the formations are split, so that formerly adjacent beds pass around opposite sides of the granite mass. We have no evidence of this, but exposures are inadequate to prove that it is not the case. A mass of granite of very similar shape and internal structure occurs within limestone east of Gouverneur. The continuity of exposures here is good, and the evidence has been interpreted (Buddington, 1929, p. 56–61) as indicating a phacolith rather than a lens. The Spruce Mountain mass is similarly interpreted as a phacolith, in which the northeastern part is so closely pinched that evidence of the original anticlinal structure is lost.

GRANSHUE SYNCLINE

East of the Spruce Mountain granite anticline (Stark quadrangle) is a belt of metasedimentary rocks with associated sheets of biotite-microcline granite gneiss. A synclinal structure is clearly shown by the foliation east of Stone Dam, in what was originally known as the Granshue tract. The axial plane of this syncline is about vertical. The rocks in the basin of the syncline have moderate to gentle dips and consist of a sheet of biotite-microcline granite gneiss, usually sillimanitic, which contains many thin pegmatite veins parallel to the foliation. The pegmatite veins are locally garnetiferous, and there are local schlieren of biotite-plagioclase gneiss in the granite. The central granite gneiss basin is surrounded by a belt of metasedimentary rocks comprising biotite-quartz-plagioclase gneiss, hornblende- and pyroxene-quartz-feldspar gneiss, and local skarn beds. The belt along the east side of the Spruce Mountain granite, from Blue Mountain Stream south to Deerlick Rapids, is almost completely obscured by overburden. However, in the hills about 2 miles north-northeast of Deerlick Rapids, there are outcrops of skarn, migmatitic skarn, and pyroxene-quartz-feldspar gneisses. These have a westward dip, indicating a continuation of the Granshue syncline as a tongue to the south through this area. A sill of sillimanite-microcline granite gneiss outcrops just southeast of Deerlick Rapids. Except for this outcrop the area southeast along the Grass River is completely covered and the structure uncertain.

WEBB CREEK ANTICLINE

East of the Granshue syncline, from east of Ormsbee Pond southward to 1 mile northwest of Wilson Mountain, and parallel to the belt of metasedimentary rocks, the granite shows a foliation indicative of an anticlinal structure whose axis is parallel to Webb Creek. The granite contains included layers of amphibolite.

ARAB MOUNTAIN ANTICLINE

The axial zone of a broad anticline extends east through Hedgehog (Cranberry Lake quadrangle), Wheeler, and Arab Mountains (Tupper Lake quadrangle).

East of Center Pond Mountain the rock forming the core of the anticline is the pyroxene syenite gneiss of the Tupper complex, which is flanked on each side by hornblende granite. The anticline is asymmetrical, having a slightly steeper dip on the north. There are several gentle rolls in the axial zone of the anticline. The dip of the limbs is gentle to moderate, rarely exceeding 30°–40°. The fold plunges gently eastward, and the nose is well shown on the shores of the northern part of Tupper Lake. However, the planar structure on the nose is very indistinct and obscure, whereas the linear structure is strongly developed.

Along Burntbridge Outlet, hornblende granite overlies the pyroxene syenite gneiss. On Center Pond Mountain the granite underlies the syenite in the cliff faces. The granite appears to have come in both below and above the syenite sheet, bowed it up, and broken across it. To the west of Center Pond Mountain the rock is all granite, though it has an inherited anticlinal structure. The structure cannot be traced beyond Bear Mountain (Cranberry Lake quadrangle).

CHILDWOLD SYNCLINORIUM OF MICROCLINE GRANITE GNEISS

The microcline granite gneiss of the Childwold area has a foliated structure, with a considerable number of folds that together form a synclinorium. The folds in general have gentle to moderate dips, though locally they are steep to vertical. The microcline granite gneiss clearly overlies the hornblende granite sheet and metasedimentary beds on the north flank of the Arab Mountain anticline, and much less clearly appears to overlie the hornblende granite on the Stark quadrangle. In the northeast corner of the Stark quadrangle and northwest corner of the Childwold quadrangle, the rocks are thought to be overturned to the south. On the northeast border of the microcline granite area the relations are again obscured by drift, but what data there are indicate that the microcline granite gneiss here also overlies the hornblende granite. The southeast end of the Childwold microcline granite gneiss area may be a fault contact.

McCUEN POND SYNCLINE OF METASEDIMENTARY ROCKS

The McCuen Pond syncline is a major structural element found in the east-central third of the Stark quadrangle and the southwestern half of the northern two-thirds of the Childwold quadrangle. The structure is complex. The central element strikes north, probably turning to the northeast, where it is tightly pinched south of Wilson Lake. There is another element, striking N. 75° W., which is reflected in the axis of a subsidiary pinched synclinal fold on each flank of the central syncline, as from Jordan Lake nearly to McCuen Pond, and east-southeast from Mud Pond (Stark quadrangle). The central north-striking part of the syncline is located where the folds swing from west-northwest to north, and thence to northeast.

A major structure of generally synclinal character extends from southwest of Lone Pond (Stark quadrangle) southeast through Jamestown Falls, Moosehead Rapids, south of Smith Island, and north of Sols Island on the Childwold quadrangle, and north of Piercefield Flow in the northeast part of the Tupper Lake quadrangle. The rocks are microcline granite gneisses. The outcrops along the main highway for several miles east of Piercefield Flow are largely andraditic, pyroxenic, or hornblendic; rarely, sillimanitic or biotitic facies are exposed. At the bridge crossing north of Sols Island, the rock on the inlet in the river is a biotite-microcline granite gneiss, but just back from the east bank of the river there are some beds of pyroxene skarn and pyroxenic gneiss. There also appears to be considerable para-amphibolite in this series of rocks. At one locality the planar foliation parallel to the axis of the fold was observed to cross a series of crumples. The major oxide of most of the rock is ilmenohematite, locally accompanied by hemo-ilmenite and magnetite.

An anticlinal structure with a southeast plunge extends southeast from Hollywood. It consists of pyroxene-microcline granite gneiss and separates the syncline described above from a major synclinal structure to the northeast, the McCuen Pond syncline.

A syncline plunges west-southwest through Pitchfork Pond on the Childwold quadrangle; and an anticline plunging southwest centers around Willis Pond and south of Kildare in the northern and northeastern part of Mount Matumbla. The Willis Pond anticline consists of pyroxene-quartz syenite gneiss with included layers and schlieren of amphibolite. Most of these structures are not shown on plate 1.

In the area north and south of Windfall Pond, the rocks are in a series of gentle folds with a general easterly strike. Much of the area is obscured by drift. The central trough appears to consist of metasedimentary rocks and migmatitic gneisses. The hills east of McCuen Pond consist of pyroxene and pyroxene-garnet skarn with local feldspathic and quartzose layers. The hill about 1 mile east of Number 19 Mountain and the exposures with northeast strike 1–1.5 miles northeast of Number 19 Mountain are all pyroxene-quartz-feldspar gneisses with local pyroxene skarn and quartz-feldspar gneiss layers. The outcrop 0.4 mile southeast of Kildare Pond is also a pyroxene-quartz-feldspar gneiss, in part highly migmatitic. The hill about 1 mile long north of Buck Pond is composed of sillimanitic and garnetiferous biotitic pegmatite-seamed quartz-feldspar gneiss. The south end of the syncline is upright, with moderate dips, but the north end is probably overturned to the southeast.

The rest of the McCuen Pond syncline is composed of granite and included lenses of amphibolite ranging from a fraction of an inch to about 2 miles in length. The amphibolite ranges from relatively unmetamorphosed gabbro (northwest of Amber Pond)—through migmatitic hornblende-plagioclase gneiss with granite seams—to schlieren of amphibolite in granite. The granite usually contains many schlieren of amphibolite and is contaminated with hornblende.

DEAD CREEK AND DARNING NEEDLE SYNCLINES

The Dead Creek and Darning Needle synclines lie almost wholly in the central part of the Cranberry Lake quadrangle. The two synclines are separated from each other by the south-plunging part of the Muskrat Pond anticline of hornblende granite.

The Dead Creek syncline is a tight isoclinal fold strongly overturned toward the north-northeast, with its axial plane dipping about 20°–25° south-southwest. The rocks of the structure consist of metasedimentary rocks containing two thick sheets of sillimanitic and locally garnetiferous biotite-microcline granite gneiss and alaskite. This complex nests in a country rock of hornblende-microperthite granite. The structure and its magnetite deposits are described in detail in Professional Paper 377.*

The Darning Needle syncline comprises several concentric belts of rock which are cut off on the south as though by a fault or intrusive contact. The dips of the foliation in the syncline are gentle. The outer part of the synclinal structure is hornblende granite, which

* See footnote on p. 31.

locally includes layers of metasedimentary rocks, as west-southwest of Chair Rock Island and south-southwest of the island, in a belt along Sixmile Creek. The next belt toward the center of the structure is the narrow arc of metasedimentary rocks along Sucker Brook. This extends from south of Ash Pond northward to Cranberry Lake, then (lying south of Sucker Brook and northeast of Rampart Mountain) eastward and southward to Lake Marian. Successively inward are the Indian Mountain belt of granite with local inclusions of amphibolite; a belt of granite with associated layers of amphibolite, biotite-quartz-plagioclase gneiss, and sparse quartzite; and a heart of metasedimentary rocks with subordinate thin layers of granite. The observed rocks of the inner basin are amphibolite, biotite-quartz-plagioclase gneiss, and subordinate pyroxenic migmatite. A relatively thin bed of limestone may possibly lie directly under Darning Needle Pond and the swampy strip encircling the more resistant core of mafic gneisses. Two thin sections of rock of the Indian Mountain granite belt are biotite-microperthite alaskite.

The major structure of the Darning Needle syncline strikes east-northeast, but a strong cross fold results in a north-south trend for a substantial part of the syncline.

South of the Dead Creek–Darning Needle synclinorial belt is a zone of granite about 1 mile wide. East of Horseshoe Lake (Tupper Lake quadrangle), this belt is a continuous part of the south limb of the Arab Mountain anticline. Toward the west, the foliation of the granite between Little Pine Pond and Silver Lake Mountain indicates a gentle anticlinal structure. On the axis a strong linear structure trends N. 70° E. Just north of Silver Lake Mountain, a small trough of granite, partly occupied by the metasedimentary rocks of the east end of the Darning Needle syncline, plunges gently westward. Farther west, however, these structures disappear, and from Silver Lake Mountain to Buck Pond on the west border of the Cranberry Lake quadrangle the granite has a southerly dip throughout, except for the belt of vertical dips along the north side of Graves and Wolf Mountains. The anticlinal axis defined by the foliation of the granite west of Cat Mountain appears to swing westward just south of the Dead Creek syncline, and to pass into a zone where the dip is uniformly south. If this is so, then this southwestern part of the anticline is overturned toward the north. The Wolf Mountain belt of granite merges into this south-dipping structure.

The axis of the Silver Lake Mountain anticline strikes at an angle into what may be a fault plane located at the base of the steep slope on the northwest side of the mountain ridge. This fault may cut out the northwest limb of the anticline southwest of Lake Marian.

BOG RIVER SYNCLINORIUM

The structure of the Bog River synclinorium is intricately complex and only some of the largest features have been worked out. The metasedimentary rocks of the synclinorium extend from Round Lake (Tupper Lake quadrangle) across the southern part of the Cranberry Lake quadrangle into the Number Four quadrangle.

LOON POND SYNCLINE

The eastern part of the synclinorium, in the eastern two-thirds of the Tupper Lake quadrangle, appears in large part to form a syncline whose axis runs northeast through Loon Pond and just north of the outlet of Round Lake. This structure is here called the Loon Pond syncline. Its central core, north of Loon Pond, consists of a sheet of contaminated biotite-microcline granite gneiss, which locally includes layers or schlieren of garnetiferous biotite-quartz-plagioclase gneiss and of amphibolite. The rocks underlying this granite sheet consist of marble interbedded with coarse white quartzite, thin-bedded quartz schist and quartzite-carbonate rock, local tremolite schist, diopside granulites, and subordinate garnetiferous biotitic pegmatitic veined gneiss. Graphite occurs in the marble and pyroxene granulites. The rocks quite closely resemble those found in the trough of the Edwards-Balmat syncline. The rocks that underlie the marble, quartzite, and silicate series of beds consist largely of microcline granite gneisses, which may be biotitic, garnetiferous, or sillimanitic, or all three, often with local layers and schlieren of garnet-biotite-quartz-plagioclase gneiss, in part migmatitic with pegmatite. Locally the microcline granite gneiss is hornblendic by contamination with included layers of amphibolite. Much of the granite gneiss is biotitic and garnetiferous. These granite gneisses are underlain by a series of metasedimentary rocks comprising thin-bedded feldspathic quartzites, silicated marble, calcareous pyroxene granulite, hornblende gneiss, and pegmatite-veined garnet-biotite-quartz-plagioclase gneiss. Underlying the metasedimentary rocks in turn is another series of gneisses comprising garnetiferous (locally sillimanitic) biotite-microcline granite gneiss, contaminated alaskite, amphibolite, and local sheets of hornblende granite gneiss—all with local layers of metasedimentary rocks. In the lowlands along the highway south of Coney Mountain are outcrops of graphitic feldspathic quartz gneiss, amphibolite, and (at the old quarry west of Sperry Pond) calcareous quartz-diopside granulites.

A line of pronounced structural discontinuity runs through the south part of the Loon Pond syncline, passing across the south side of Bear Pond and the head of Round Lake. The structural line is probably a fault, for the southwestward extension of the trough of the Loon Pond syncline is abruptly cut off on the south, just north of Otter Pond. Evidence is not at hand as to the age of the fault, or whether this is a normal fault, a thrust fault, or a fault with predominantly horizontal movement.

BOG LAKE PART OF SYNCLINORIUM

The Bog Lake (locally known as Robin Lake) part of the synclinorium between Tomar Mountain and the Charley Ponds is complexly folded.

The rocks in the belt from just west of the Charley Ponds to a line southeast of the railroad are largely microcline granite gneisses carrying biotite, garnet, or sillimanite, or a combination of these; subordinate hornblende granite contaminated by admixture from amphibolite; and included layers of biotite-quartz-plagioclase gneiss, more or less pegmatite-veined and granitized. Locally there are included layers of amphibolite. Similar rocks form a belt southwest from Sabattis along the east side of Bog Lake, and these are thought to form a closely appressed anticline. A closely folded synclinal structure is inferred for the rocks of the belt extending from Rainer Pond northeastward through Clear Pond to a place about 1½ miles southwest of Sabattis. The rocks consist of pegmatite-veined garnet-biotite-quartz-plagioclase gneiss with amphibolite layers, hornblende-garnet gneiss, and some granite sheets. Locally, there are sillimanitic gneisses. Amphibolite seems to be more common southwest of Clear Pond. It contains granitic veinings and local lenses of feldspathic skarn.

North and west of Bog Lake—except for a thin border of amphibolite, pyroxene gneiss, and biotitic gneiss—the complex consists of porous-weathering white pyroxene quartzite. Rarely, a few thin interbeds of amphibolite, biotite-quartz-plagioclase gneiss, or feldspathic pyroxene skarn are present. Limestone may be present beneath some of the lowlands. These rocks are similar in nature to those in the trough of the Loon Pond syncline. Also, the structure is similarly inferred to be a syncline with a steep or vertical southeast limb, and a northwest limb that dips 55°–80° southeast.

The metasedimentary rocks in the belt extending from Mud Lake to the western border of the Cranberry Lake quadrangle are rarely exposed. The few outcrops observed consisted of granular pyroxene-feldspar gneisses and, locally, amphibolite, quartzite, and skarn. Here too the structure of the metasedimentary rocks appears to be synclinal relative to the enclosing granite, though dips are steep.

The hornblende granite around the southwest end of the Bog Lake portion of the synclinorium changes dip from southeast north of Lake Lila, through vertical to the west of Lake Lila, and back to steep southeast on the northwest limb. This structure is consistent with what would be expected for a syncline slightly overturned toward the northwest and plunging northeast.

Within the Bog Lake part of the synclinorium two kinds of planar structure are present. The first is the alinement of platy minerals in biotite gneiss, amphibolite, and quartz-feldspar gneisses. The second is relict bedded structure. Many outcrops of amphibolite, for example, contain beds only a few inches thick of garnetiferous quartzite. Or garnet-biotite quartzite and biotite gneiss, interlayered on the scale of one to two feet, may show inch-thick interbeds of amphibolite. In the apparently massive quartzites, slight differences in content of pyroxene and calcite produce layers of slightly differing weathering characteristics. Rarely, the quartzite shows boudinage structure where interbeds of pyroxene skarn, a few inches to half a foot thick, have been pulled apart. It would be extremely difficult to explain these compositional differences as due to agencies other than sedimentation, with accentuation of bedding during deformation.

The planar structures in both granite and Grenville rocks determine the folds. As the map shows, the contacts between the two rock types are virtually parallel.

Linear structures are prominent in the southern third of the southeast rectangle. Within the granite, lineation is given by aggregates of hornblende in very flat lenses (streaks). In the vertical or steep-dipping granite of Webb and Smith Mountains and the hill east of Harrington Pond, the lineation trends uniformly S. 65° W. and plunges 10°–20° SW. Less than 1 mile east of Partlow Lake, it trends about N. 78° E., plunging 20°–22° NE. Curiously enough, no lineation has been observed in the granite block west and northwest of the lake. Absence of this structure may be, in part, more apparent than real: outcrops are poorer in that block; and where dip surfaces are fewer, there is less chance of finding a lineation. In part, however, the lineation may never have been prominently developed in rocks so far from the contact zones.

Linear structures in the metasedimentary rocks are manifest in ribbing by dimensionally-oriented hornblende crystals, and (rarely) by axes of minor folds. None of these structures is found in the quartzite; hence a valuable structural element is missing from almost the whole northwest part of the Robin Lake folded complex. In the vicinity of the lake itself, the lineation trends S. 45°–85° W. and plunges 17°–45° SW. The value most frequently observed is S. 60°–65° W., 40°–45° SW. This is essentially the same trend as that found in granite to the south, but the plunge is steeper.

The only lineation seen in the northwest part of the complex was on an outcrop 0.1 mile southwest of Tomar Pond. Here the trend was S. 5° W. and the plunge 35° S.

The intersection of the various oblique joints with the foliation surface gives a different type of lineation. This has not been mapped.

A highly developed joint system is characteristic. The presence of the same joint systems in granite and metasedimentary rocks, as well as the same patterns of foliation and lineation, strongly suggests that both rock types have been subjected to similar deformation.

SOUTHERNMOST PART OF AREA

ANTICLINE SOUTH OF SCANLONS CAMP

An anticlinal axis in the hornblende granite of the southern third of the Oswegatchie quadrangle lies west of Bald Mountain and Shaws Camp, where it trends north-northeast, then turns east-northeast and passes between Scanlons Camp and Maple Hill. The southwest part of the structure is not shown on plate 1.

CRACKER POND ANTICLINE

The hornblende granite in the southwest corner of the Cranberry Lake quadrangle has an anticlinal structure relative to the two parts of the Bog River synclinorium of metasedimentary rocks. A main axis of this anticline runs from just north of Tomar Mountain through Cracker Pond to the southwest. Linear structure is prominent on the southwest-plunging nose of the anticline southwest of Toad Pond. Northeast of Cracker Pond there is wavy structure on the crest of the fold. A subordinate anticlinal structure occurs southwest of Partlow Mountain. The dips of the foliation are generally steep, but locally there are areas of quite gentle dips, as around Gal and West Ponds. The anticlinal structure is overturned to the north in the Tomar Mountain prong.

SALMON LAKE ANTICLINE

South of the Arab Mountain anticline, the next major anticline that can be clearly determined is the Salmon Lake anticline. This feature, which occupies the whole northern third of the Raquette Lake quadrangle (Rogers, C. L., oral communication), is several miles broad and strikes east. The east-central portion is clearly defined by an eastward-plunging arc of amphibolite lenses. The amphibolite of the north limb of the fold is continuous with the amphibolite belt west of Antediluvian Pond (south border of Tupper Lake quadrangle). The anticline is a tight fold slightly overturned to the north.

SOUTH PART OF TUPPER LAKE QUADRANGLE

The foliation of the south part of the Tupper Lake quadrangle, south of the Loon Pond syncline and southeast of the Little Tupper Lake fault, dips almost uniformly south and the structure is largely indeterminate.

The axis of a tightly pinched syncline is indicated by opposed dips of the granite and associated amphibolite around Lake Lila. This syncline lies on the north flank of the Salmon Lake anticline, which, as has been described, occupies the adjoining part of the Raquette Lake quadrangle.

East of the Little Tupper Lake fault, as has already been noted, the amphibolite west of Antediluvian Pond is definitely on the north limb of the Salmon Lake anticline. The metasedimentary rocks and amphibolite along the general locus of Slim Pond and Stony Pond may occur in a tightly pinched syncline and be the equivalent of the Lake Lila syncline. No other structures have been clearly identified in this area, though isoclinal folds overturned toward the north are suspected.

PLANAR STRUCTURE IN ADIRONDACKS AS A WHOLE

The origin of certain structural features of the St. Lawrence County district can only be understood by considering the structure of the Adirondacks as a whole. A generalized sketch of the orientation of the planar structure of the rocks of the Adirondacks, insofar as now known, is given in plate 4.

A number of anorthosite masses occur in the eastern half of the Adirondacks and one small mass in the northwestern Adirondacks, on the south border of the Antwerp quadrangle. In general, the larger anorthosite masses are anticlinal with respect to the surrounding rocks. They are thought to have acted during deformation as relatively rigid structures. However, they have been reactivated, subsequent to their consolidation, in such fashion as to deform, rise somewhat as domes with diapiric flow on their flanks, and yield granoblastic gneiss on a large scale along their borders, and locally throughout some small domes.

In the eastern Adirondacks, the strike of the foliation, which is to a major extent determined by the primary form of the anorthosite masses, is highly irregular.

In the central and southern half of the Adirondacks, the strike of the foliation is predominantly east.

In the northwestern and extreme western Adirondacks, the strike of the planar structure is predominantly northeast. Similarly, for more than 100 miles northwest of the St. Lawrence River in Canada, the strike is prevailingly northeast ("Tectonic Map of Canada, 1950"). For the northwestern Adirondacks it would thus seem reasonable to assume tentatively that major stresses acted along northwest-southeast lines.

The east-west trend of the foliation and its arcuate curvature to the north in the southern half of the Adirondacks would equally well be interpretable on the basis of major stresses acting along generally north-south lines. The north-south stress could perhaps be a major component of generally prevailing major northwest-southeast forces, but the geologic history of the southern half of the Adirondacks is inadequately known to us to warrant further discussion of this problem.

ANTICLINAL RIGID STRUCTURES AND MOBILE ZONES OF OVERTURNED ISOCLINES

The major zones of overturned isoclines and associated anticlinal bodies of igneous rock have been plotted in figure 13, for those quadrangles for which data are available. The anticlinal masses include the alaskite gneiss (Alexandria batholith) in the extreme northwest, the Lowville and Stark anticlines of quartz syenite and older hornblende granite gneiss, the Inlet sheet and Arab Mountain anticline of syenite and quartz syenite gneiss, the Twitchell anticlinal mass of syenite gneiss (Big Moose quadrangle), the anticlinorial bodies of anorthosite and associated syenite gneiss of the Thirteenth Lake and Indian Lake quadrangles, and the Piseco dome of quartz syenite gneiss in the south. It will be noted that each of these anticlinal masses is bordered on one side by a belt of rocks having overturned isoclinal structure. It is as though the anticlinal bodies had acted as relatively rigid elements in relation to the more mobile gneisses, which have been deformed against them.

Within the zone lying between the Alexandria batholith on the northwest and the Lowville and Stark anticlines on the southeast, there are two belts in which the axial planes of the isoclinal folds dip away from the anticlinal masses and toward each other, and there is an intermediate zone of open to tight folds between them.

The Diana complex northeast of the Lowville anticlinorium has been shoved forward to the southeast, giving a great arcuate curve convex to the southeast (fig. 14). Similarly on the southeast side of the Diana complex, the foliation of the hornblende granite swings to the northwest, as though bent around the anticlinorium as a rigid element. Within the Lowville anticlinorium itself the folds are overturned to the south and southeast on the northwest portion, and to the northwest and southwest on the eastern side.

The linear structure within the belt (belt 2, fig. 15) of strongly overturned isoclinal folds bordering the Diana complex, and including the complex itself northeast of the Lowville anticlinorium, is about normal to the major fold axes.

Between the Lowville anticlinorium and the Inlet sheet and Stark anticline, the younger granite has cut out the older rocks.

The quartz syenite complexes—constituting the flank of the northwest part of the main anorthosite massif, the Arab Mountain anticline, the Inlet sheet, and the Stark anticline—together form a somewhat interrupted, roughly elliptical ring of anticlines enclosing a synclinorial basinlike structure. One possible interpretation of these structures is that a more-or-less rigid sheet of anorthosite continuous with the main massif extends beneath the surface rocks of this entire area in such a fashion that its outer edge is in general marked by the Arab Mountain and Stark anticlines. Such a sheet could have been responsible for localizing the structures where they are. In effect the real outer edge of the main anorthosite massif would be subsurface along the general line of the Arab Mountain mass, Inlet sheet, and Stark complex, and together with them would constitute a relatively rigid structure. A zone of isoclinal folds overturned to the southeast toward the Stark anticline and to the north toward the Arab Mountain anticline forms an almost continuous arc around the outer edge of this structure.

A belt of mixed rocks comprising metasedimentary beds and interlayered sheets of diorite, quartz syenite, and granite forms a broad zone between the anorthosite-syenite complexes of the Thirteenth Lake and Indian Lake quadrangles on the south, and the quartz syenite mass of the Arab Mountain anticline on the north. This belt of layered rocks has zones of isoclinal folds overturned in opposite directions on each flank, and a central zone of open to tight folds. The data for the Raquette Lake quadrangle are based upon the work of C. L. Rogers (1951, personal communication).

On the Cranberry Lake quadrangle, the granites and associated metasedimentary rocks of the Dead Creek syncline have been isoclinally overturned toward the north against the quartz syenite gneiss of the Inlet sheet, which itself may be isoclinally overturned toward the north.

Again, Cannon (1937, p. 64–65) has described a series of metasedimentary rocks and associated granitic gneiss and porphyritic quartz syenite gneiss that lie in a belt about 7 miles wide south of the Piseco dome of quartz syenite gneiss and granite gneiss, and are interpreted

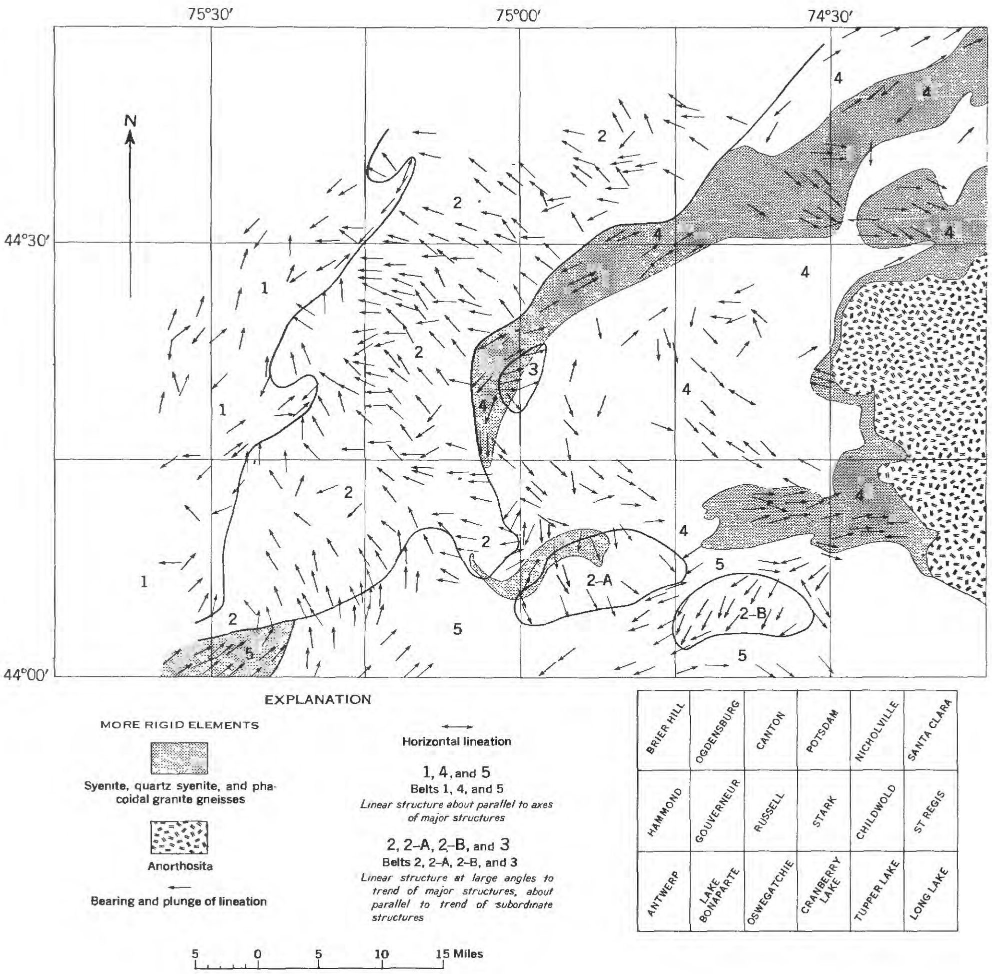

FIGURE 15.—Orientation of linear structure in the northwest Adirondacks and its relations to more rigid structural units.

as probably representing a series of isoclinal folds overturned toward the north against the dome.

Postel (1952, pl. 2) has found that the granite gneisses and associated metasedimentary rocks have been isoclinally overturned from opposite directions against the phacoidal granite gneiss of the Lyon Mountain dome (Lyon Mountain quadrangle), so that along the northwest side the axial planes of isoclinal folds dip northwest, along the south-southeast side they dip south, and along the east-northeast side they dip east-northeast.

Local bodies of anorthosite lie along a belt extending northwest from the north end of the anorthosite prong on the Lake Placid quadrangle, across the northeast part of the Saranac quadrangle, and onto the Loon Lake area. This is taken to indicate the presence of a ridge of anorthosite extending beneath the surface. The quartz syenite complex, which forms an anticline northeast of the ridge, has its southwestern part overturned to the southwest against the anorthosite ridge. The linear structure is about normal to the strike of the fold axis (Buddington, 1953).

The foregoing facts show that there is no uniformity in the direction in which isoclinal folds are overturned throughout the Adirondack region. On the contrary, their localization and direction appear to be correlated with the presence of orthogneiss or anorthosite masses, which have acted as relatively rigid units. The isoclinal folds are uniformly overturned toward these cores, and the direction of the fold axes is determined by the strike of the old anticlinal structures, whatever may have been the direction of the major stress.

ORIENTATION OF LINEATION WITH RESPECT TO FOLD AXES

The relations of the orientation of lineation and planar structure to axes of major folds and foliation structure in the northwest Adirondacks are shown in figures 15 and 16. Linear structure in general may be due to a variety of features, but that upon which the data in the figures is based is a lineation shown by a subparallel orientation of individual elongate mineral grains or mineral aggregates.

In general, the lineation shows two contrasted relations to the major folds, one in which the lineation is subparallel to the axes, and another in which the lineation is about at right angles to the axes. In local belts there are transitions between the two types. Most of the rocks show the relation in which the linear structure is subparallel to the major fold axes. There is, however, a great arcuate belt and two small related areas within which the linear structure is subperpendicular to the axes of major folds.

The term fold is used here to include domical structures in granite masses. These masses are geographically related to folds in the metasedimentary rocks, though the masses themselves may be only incidentally the result of folding.

Two contrasting relations of linear structure to planar foliation in the plunging noses of individual folds are also found. In some examples the planar foliation curves around the end of the fold, and the linear structure is subparallel to the foliation. In other examples there is practically no megascopic planar structure but only an intense pencil or linear structure. In this type the linear structure is always parallel to the axis of the fold and plunges at the same angle as the fold.

In one fold within a mass of quartz syenite gneiss on the Saranac quadrangle the following relations hold. The south end for a length of more than 1 mile plunges south and the limbs have opposite directions of dip; the lineation is parallel to the axis and plunge of the fold. At the north the fold is isoclinally overturned to the southwest, and the major axis trends northwest. The lineation in this part of the fold is about at right angles to the trend of the axis. There is thus a direct correlation here between the subtransverse relation of linear structure and overturned isoclinal deformation. However, subparallel lineation may also occur in zones of overturned isoclinal folds.

The linear structure may also have different orientation in rocks of different ages. The linear structure of the hornblende granite around the areas 2A and 2B (fig. 15) is discordant with that of the older rocks within these zones. The metasedimentary rocks of the two areas were strongly deformed before the intrusion of the hornblende granite. The linear structure of the areas may be related either directly or indirectly to this earlier deformation, whereas the linear structure of the granite is a primary flow structure.

LINEATION PARALLEL TO PLANAR STRUCTURE, CURVING AROUND NOSES OF FOLDS

In folded belts where linear structure is predominantly about parallel to the strike of the major foliation, two different relations have been found to occur at the ends of the folds where the foliation swings around in an arc.

In some examples, as at the southwest end of the Canton granite mass, the linear structure continues to be about parallel to the strike of the local foliation around the blunt anticlinal nose. This is probably in part conditioned by the presence of an original strong primary foliation, for the later deformation of this rock has been intense. Again, the rocks of the McCuen Pond syncline have a lineation which is about parallel to the local strike of the planar structure throughout.

LINEATION PARALLEL TO FOLD AXES, TRANSVERSE TO NOSES OF FOLDS

In many other instances, the linear structure maintains its orientation parallel to the major axes of anticlinal and synclinal foliation structure, even in the rock at the apices of these structural features, and is thus at a large angle to the strike of the foliation in such locations. It is so often the case as to be considered the rule, that on the crests or the noses of tight, plunging anticlines, and occasionally in the plunging troughs of tight synclines, linear structure is developed parallel to the strike of the major axes—almost, if not quite, to the exclusion of planar structure. In the syenite, quartz syenite, and phacoidal granite gneisses of the Diana, Stark, and Tupper complexes, the linear structure is uniformly parallel to the major structure axes where the planar foliation swings around the ends of the folds. Some examples are the northeast end of the Lowville anticline (northwest corner Lowville quadrangle), the south end of the Stark anticline (near

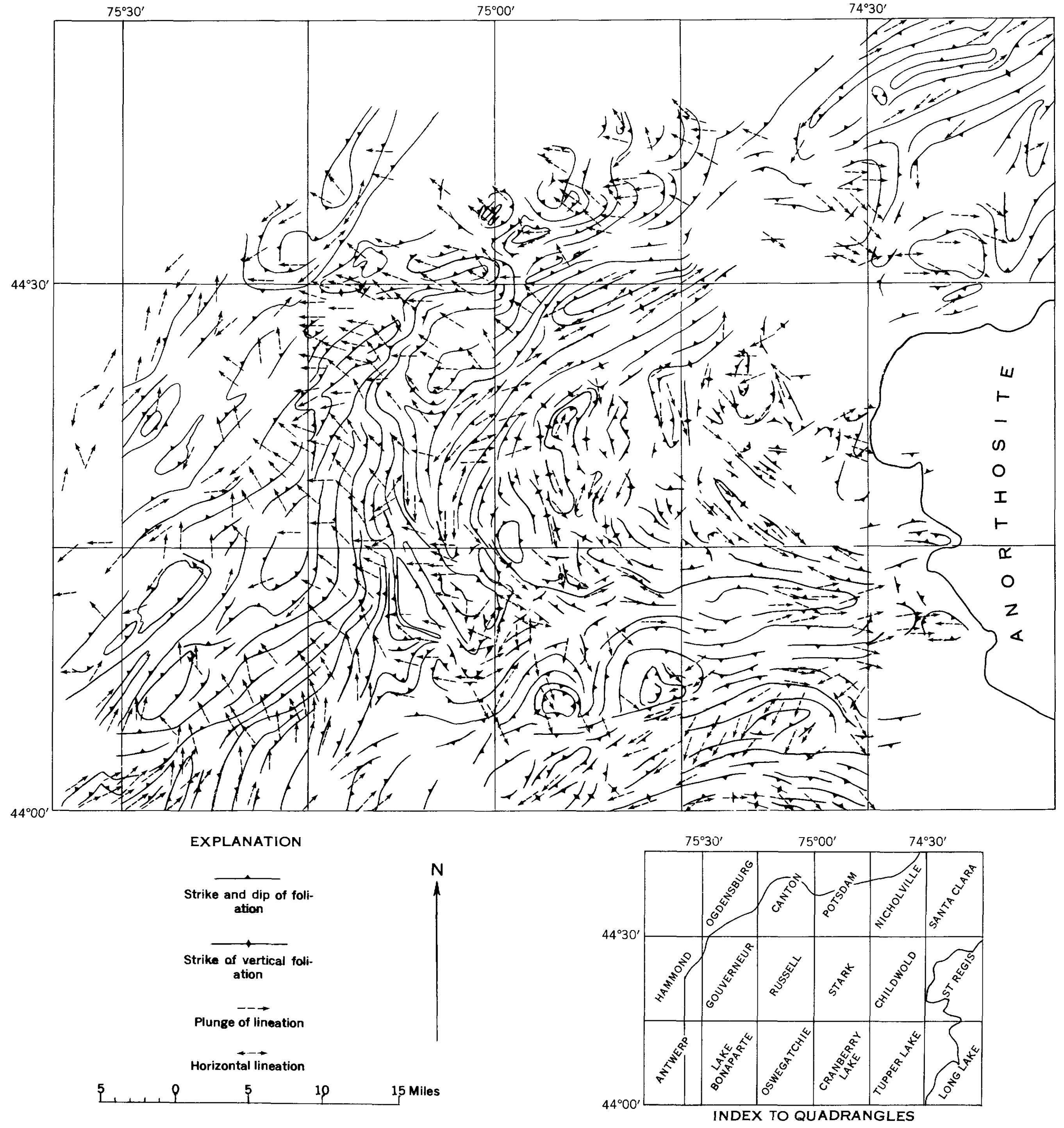

FIGURE 16.—Generalized trends of foliation and linear structure in the northwest Adirondack area.

south border of Russell quadrangle), a local plunging nose of the Stark anticline in the northwest part of the Stark quadrangle, each end of the Jennings Mountain dome (south part of the Santa Clara quadrangle), the east side of the plunging anticline near the southwest border of the Santa Clara quadrangle, and the east end of the Arab Mountain anticline (northwest part of Long Lake quadrangle and northeast part of Tupper Lake quadrangle).

Linear structure parallel to major structure axes and at large angles to the strike of the planar foliation around the ends of synclines is shown in the granite gneiss with associated metasedimentary rocks in the east-central part of the Potsdam quadrangle, in similar rocks of the small Trembley Mountain syncline (Cranberry Lake quadrangle), and at the east end of the Darning Needle syncline. The linear structure is in the plane of foliation.

The plunging noses of some anticlines show a strong lineation as the dominant internal structure. The lineation is parallel to the fold axis. Planar structure is absent or indistinctly developed. This relation has previously been described by Cannon (1937, p. 62) as occurring in the quartz syenite and granite gneisses of the Piseco dome (Piseco quadrangle); by Buddington (1939, p. 308), in the anorthosite gneiss of the Reber anticline (Willsboro and Port Henry quadrangles); and by Balk (1932, p. 51), in metasedimentary rocks of the Newcomb quadrangle.

The linear structure in the broad expanse of the Diana complex at the intersection of the Antwerp, Lake Bonaparte, Lowville, and Carthage quadrangles appears in part to be transverse to the fold axes within the mass. This is not the case however, for the folds here are minor ones superimposed on the broad nose of a major anticlinorium in the Diana complex. The linear structure is generally parallel to the axis of this major anticlinorium, which trends N. 60°–65° E. and plunges east-northeast.

The linear structure in all the rock of the Stark and Tupper complexes is parallel to the major fold axes, with only a few exceptions, such as the mass north of Wanakena (Cranberry Lake quadrangle) and the eastern part of the south end of the Stark complex.

LINEATION ABOUT AT RIGHT ANGLES TO MAJOR FOLD AXES

The larger part of the belt (fig. 15, area 2) in which the lineation is about at right angles to the major structure axes parallels the contact between the main igneous complex and the Grenville belt of metasedimentary rocks and includes a broad belt of rocks on each side. The part of the belt that strikes northeast is more than 60 miles long and 15 or more miles wide. The rocks comprise marble, quartzite, metasedimentary gneiss, and sheets of hornblende granite gneiss, alaskite gneiss, sillimanite-microcline granite gneiss, amphibolite, and syenite and quartz syenite gneiss. All the rocks are isoclinally folded and intensely overturned to the southeast; they are also thoroughly metamorphosed and recrystallized. Much of the igneous rock along the zone bordering the metasedimentary rocks is mylontized.

The subtransverse relation of the linear structure to the major fold axes is somewhat obscured in part (fig. 15, area 2A) of the east-striking belt by the fact that the major structure axes are themselves strongly folded and the linear structure is mostly parallel to the axes of these well-marked but subordinate fold structures. The rock belts as a whole here suggest an east-northeast to east structural trend, upon which folds with north to north-northwest axes have been superimposed. The belt includes the rocks of the Dead Creek and Darning Needle synclines. Farther east (fig. 15, area 2B), the rocks of the Loon Pond syncline (southwest part of area 2B) have a linear structure that is roughly parallel to the major axis. However, the linear structure for much of the rock in the eastern part is at a considerable angle to the strike of the foliation.

The linear structure of the rocks of the Dead Creek, Darning Needle, and Loon Pond synclines is thus in general at a considerable angle to the major fold axes, whereas the linear structure of most of the surrounding hornblende granite is about parallel to the strike of the foliation. This could be interpreted as indicating that the rocks (metasedimentary rocks and microcline granite gneiss) of the synclines had undergone an earlier metamorphism and were later intruded by the hornblende granite, which assumed an independent linear structure consequent on its own magmatic flowage. Another interpretation would be that the rocks in the synclines were weaker and as such had a linear structure induced in consequence of the movement of the enclosing hornblende granite as more or less rigid units. The latter hypothesis seems in part a possible one, as the bordering hornblende granite itself locally has the same structure as the synclinal metasedimentary rocks and microcline granite gneiss.

The rocks within the east-trending part of the belt (2–A, 2–B, fig. 15) are also isoclinally folded and strongly overturned. However, the hornblende granite and alaskite are not thoroughly metamorphosed, as they are in the major part of the belt, but show only a moderate degree of granulation and only a little metamorphic unmixing of the perthite feldspars.

South of Clarksboro (Russell quadrangle), just east of the major belt and in the inner part where the arc curves sharply, there is a small, elongate area (3, fig. 15) in which the linear structure is transverse to the strike of foliation. The area includes the east part of the Stark complex, local belts of metasedimentary rock, and some hornblende granite gneiss. The phacoidal granite gneiss in this area is locally strongly mylonitized and schisted.

Throughout the belts in which lineation is subtransverse to the major structure axes and the major extension of the rock belts, there are subordinate fold axes and local corrugations to which the lineation is parallel.

This relation is excellently shown in the rocks of the great synclinal bends west and north of Cox Meadow

(Potsdam quadrangle), and in the arc northeast of School No. 7 (Nicholville quadrangle). It is equally well shown on the large anticlinal cross fold through Blanchard Hill and Fairbanks Corners (Russell quadrangle).

The correlation of linear structure with the axes of minor cross folds is again well shown in the southwestern part of the Russell quadrangle. West of Irish Hill School the linear structure trends southwest in relation to an anticlinal arc whose axis is about west-southwest; yet offset slightly to the northwest is a cross fold whose axis is about west-northwest and whose linear structure trends correspondingly northwest.

In the northwest part of the Oswegatchie quadrangle is another major bend with a northwest axis and a similar orientation of linear structure.

To cite one final example: on the Cranberry Lake quadrangle, the major cross fold in the vicinity of Wanakena has an axis trending south-southeast, and a corresponding orientation of the linear structure.

All the foregoing cross folds are developed on a relatively large scale. There are numerous minor cross folds on much smaller scales, to which in practically all cases the linear structure is conformable.

The rocks within the Grenville lowlands also often show secondary folds and crumples on the limbs of major folds. The orientation of the secondary folds may be at large angles to the strike of the major structure. The smaller structures appear to indicate subordinate movement about at right angles to the major one (Buddington, 1929, p. 53; 1934, p. 154, 167).

ORIGIN OF LINEAR STRUCTURE

Data have been given to show that in the northwest Adirondacks there is a correlation between particular features of regional deformation and the zones in which the relation of linear structure subtransverse to major fold axes obtains. These features are: intensely compressed folds, for the most part strongly overturned and isoclinal; commonly occurring subordinate to small-scale cross folds whose axes are parallel to the lineation; and plastic flowage of the rocks as evidenced by granulation and deformation, in part of mylonitic type but in large part with recrystallization.

A thorough survey of the general problem of the relation of the orientation of lineation to fold axes and the strike of planar features has been made by Cloos (1946). A summary of several of his conclusions that are pertinent to the Adirondacks is given here. For purposes of description he adopts the following terminology: *b* is the major fold axis; *a* is perpendicular to *b* in the movement plane; and *c* is perpendicular to *ab*. Mineral lineation parallel to *b* is the most frequently observed type. Lineation in *a* has been proved by many observers, though in some instances it is interpreted by authors as the product of a second phase of deformation perpendicular to the first one. The lineation of the *a* type is frequently formed by much more intense deformation than that which is parallel to *b*. Lineation in *a* is in the principal direction of movement in the cleavage plane and in the direction of maximum deformation of a fold. In recumbent or torn overturned folds, the forward movement is in *a* and at a maximum. It is often assumed that folds form perpendicular to "pressure" and that fold axes are normal to the maximum movement. Certain subordinate folds, undulations and wrinkles, however, may form where the principal movements are parallel to their axes. Two fundamentally different directions of movement have been distinguished: the principal direction and a subordinate one, normal to the first. The second direction is a direct consequence of the first. A distinction is essential, because lineations may result from either motion but cannot be distinguished unless a careful analysis is made of a region large enough to establish relations between directions of movement.

Cloos (1946) has further summarized certain results of a study of the relation of the orientation of lineation to other structural features in a large region in parts of eastern Maryland and southern Pennsylvania. He finds that the relations vary in such a way that five different belts are indicated. In one (belt II) he finds that the rocks are schists of the Wissahickon and Peters Creek formations, in a strongly compressed series of folds dominated by intense shortening, with flow cleavage, and with lineation and fold axes parallel; in another (belt IV) the rocks comprise volcanics, quartzites, carbonate rocks, and phyllites, which are characterized by intense overturning of folds and intense forward flowage, as indicated by cleavage planes in which there is a strong lineation normal to fold axes and down dip. The Maryland and Pennsylvania rocks are in general not as intensely metamorphosed (with respect to temperature) as those of the Adirondacks. However, the occurrence of linear structure normal to the major fold axis in a series of strongly overturned folds, interpreted as resulting from intense forward flowage, may still have some pertinence to the problem of similar relations in the Adirondack rocks. It may also be noted that the rocks of one belt in which the linear structure is parallel to the major fold axes are of a higher grade (temperature) of metamorphism than those in the belts in which the lineation is about at right angles to the major fold axes, though in both cases the rocks are intensely deformed (Cloos, 1946, p. 43–45).

Cloos (1946, p. 30) has further pointed out that forward motion of rock masses in folding and thrusting may be unlimited and rocks can flow forward, become elongated, or be stretched to any length in the direction of principal movement, whereas extension normal to this direction is limited (Cloos, 1946, p. 30). In general he believes it can only rarely exceed the amount of arcuation, though exceptions may occur where axes plunge steeply.

Engel (1949b) similarly has shown that in the talc belt of the Gouverneur quadrangle many of the metasedimentary beds have been so deformed as to produce a wavy pattern of outcrop, which indicates that there has been substantial lateral extension and which also indicates the possibility for the development of subordinate forces perpendicular to the direction of major shortening.

In the Gouverneur quadrangle, a portion of the belt whose linear structure is generally about transverse to the major folds has been studied by Engel (1949, written communication) in great detail. He summarized his conclusions as follows:

> The northeast-southwest trend of major folds resulted from pronounced movement along northwest-southeast lines. But the north- and northwest-plunging cross folds, associated foliation, shear surfaces, and lineations are related to pronounced movements in a northeast-southwest direction, essentially along or at low angles to the regional trend lines.
>
> These evidences of two directions of movement do not necessarily imply two periods of deformation. All the features may have originated by major shortening, due to compression, and a couple along northwest-southeast lines, accompanied by lateral elongation and cross folding in a northeast-southwest direction.

Engel (oral communication) later suggested that in the culminating stages of deformation the rocks along the southeast side of the Grenville lowlands were rolled northeastward relative to the adjacent margin of the Adirondack massif, yielding refolded folds within the Grenville series, and developing a lineation subtransverse to the major trend of rock formations.

There is much evidence that all rocks older than the hornblende granite and alaskite were strongly folded and deformed before the intrusion of the younger granite magmas. This deformation was accompanied by plastic flowage of most of the rock of the quartz syenite complexes. For example, the pyroxene syenite gneiss east and west of the Usher Farm road (Tupper Lake quadrangle) is more thoroughly recrystallized than the younger granite to the west on the same anticline. The feldspars of the syenite are 70–80 percent granoblastic, whereas in the granite the granoblastic mortar forms only about one-third of the rock. Similarly, the phacoidal granite gneiss of the Stark anticline in general, and of the Jayville area, is more thoroughly recrystallized and granoblastic than the adjoining hornblende granite. It seems highly probable that this earlier deformation produced a linear structure, but we cannot surely distinguish linear structure formed at this period from that superimposed in a later period of deformation. It is certain that locally there was a later period of deformation, and that it postdated the intrusion and consolidation of the hornblende granite and alaskite magmas. There are large bodies of the younger granitic rocks within the belt where the lineation is subtransverse to the major structure axes, and they have been completely recrystallized and deformed to gneisses with lineation appropriate to the belt in which they are. Some granite dikes in the older syenitic rocks have a foliation and linear structure consistent and continuous with that of the enclosing rock, indicating a contemporaneous deformation of both. The directions of major movement and of elongation of the rock masses appear to have been similar in both deformations.

We have nothing to add toward the solution of the problem—whether in the last period of orogeny both the intensified deformation of the major folds and the development of the cross folds were due to the same major stress orientation, and the cross folds and linear features were due to minor forces oriented at right angles, or whether there were two distinct epochs of deformation in which the principal stresses were about at right angles to each other. The first solution would seem preferable, until definitive evidence to the contrary is found.

FAULTS AND LINEAMENTS

The prevalence of strong block faulting in the eastern Adirondacks and its virtual absence in the northwest Adirondacks has previously been referred to in a discussion of the origin of the topography. The southeastern part of the St. Lawrence County district lies within the faulted area.

At least three major normal faults are thought to occur in the Tupper Lake quadrangle. Perhaps another is present in the Cranberry Lake quadrangle along Dead Creek Flow and the north arm of Cranberry Lake. There may well be others that we have not recognized.

The three faults in the Tupper Lake quadrangle are here called the Pine Pond, Tupper Lake, and Little Tupper Lake faults. All strike north or northeast. In addition, two lines of structural discontinuity that strike about east or east-northeast are marked by the cutting off of the south parts of the Loon Pond syncline (Tupper Lake quadrangle) and of the Darning Needle syncline (Cranberry Lake quadrangle). These may be called the Sabattis Road and Wolf Mountain lineaments. A strong structural line of similar unknown significance strikes northeast along the locus of Dead Creek Flow (Cranberry Lake quadrangle).

Strong topographic lineaments or "trough lines" that cross the foliation structures at an angle are also thought to develop in many places as a result of erosion along close-spaced joint systems, especially along certain zones of close-spaced northeast to north-northeast joints.

Normal faults—each with a length of miles and a throw of hundreds or thousands of feet—have been described from several areas within the Adirondacks, notably from the Piseco Lake (Cannon, 1937) and Lake Pleasant (Miller, 1916) quadrangles, and from the Blue Mountain (Miller, 1917) and Long Lake (Cushing, 1907) quadrangles. Faulting along the southern border of the Adirondacks has been summarized by Megathlin (1938), and in the eastern Adirondacks by Quinn (1933). The description by Kay (1942) of the Ottawa-Bonnechere graben north of Lake Ontario includes much data pertinent to the problem.

A major fault, named by Brown (1936, p. 243) the Balmat fault, occurs on the Lake Bonaparte and Gouverneur quadrangles. This fault causes an offset of more than a mile of the formations on the Lake Bonaparte quadrangle. There is no evidence that this fault has effected any displacement of the Potsdam sandstone or of the pre-Potsdam peneplain surface. It is probably of Precambrian age.

EVIDENCE OF FAULTING

The inquiring geologist cannot reach out his hand and place it on the fault surface of any one of the major faults in the district. Nevertheless, evidence for faulting does exist, though some of it is susceptible of other interpretations. Long, relatively narrow valleys transverse to the axes of major folds are found in several places. Outcrops of fault breccia or mylonite, or both, are preserved in a few places. Marked discontinuities in the foliation of the gneissic rocks cannot always be explained as sharp, local folds. Some rock units have been brought together incongruously, as if by faulting. Others have been offset horizontally for distances of 200 feet to a mile or more. Yet the distinctive topography of block-faulted mountains is lacking in most of the area mapped, and in-faulted outliers of lower Paleozoic sedimentary rocks have not been found. Diabase dikes, in many places associated with other Adirondack faults, have been noted in just one zone of mylonite. The existence of the faults, therefore, is to a substantial extent inferred.

CHARACTER OF THE FAULTS

Because the surfaces of the major faults are not exposed, it is impossible to find out directly what the faults are like. We do know that the principal faults strike northeast or north, that they appear to extend for some miles, and that they cut across the bedrock without regard to its type or its planar structure. The rare exposures of mylonite or fault breccia have a steep to vertical dip. These features are characteristic of many other Adirondack faults, and it seems reasonable to interpret the principal faults of the St. Lawrence County district as high-angle gravity faults that reflect block faulting of the eastern part of the Adirondacks.

When this assumption is made and approximate values for the throw are calculated, we find that vertical displacement along the local faults ranges from 500 feet to about 3,000–4,500 feet. Any horizontal movement along the faults in the eastern part of the Tupper Lake quadrangle would, of course, reduce the vertical component calculated from the horizontal offset. The values for throw are comparable to those estimated for faults in the Lake Pleasant quadrangle (Miller, 1916; 100–2,000 feet), in the Piseco Lake quadrangle (Cannon, 1937, p. 73; minimum throw about 1,200 feet at one place), and in the Champlain Valley (Buddington and Whitcomb, 1941, p. 104; throw of 4,000 feet on Split Rock Point fault, the one showing maximum displacement).

Within the massif, either the east block or the west block may be downthrown relative to its neighbor (Miller, 1916, p. 46, 48; Cannon, 1937, p. 72). Moreover, the throw may differ from point to point along a given fault (Miller, 1916, p. 47). The evidence from the Tupper Lake area is consistent with these earlier observations.

AGE OF FAULTING

There is at present no way of dating the faults described below, nor of determining whether movement along them has been repeated. They appear to be a part of the group of faults which break the country to the east, and which are known definitely to be of post-Ordovician age and are inferred by Kay (1942, p. 1627) to have originated during the Taconic disturbance in early Silurian time. It is possible, however, that some faults such as the Wolf Mountain lineament are much older and of Precambrian age.

PINE POND FAULT

A long, straight valley extends southward from the headwaters of the Grass River, in the northwest quarter of the Tupper Lake quadrangle, to the north end of Hitchins Pond, parallel to the 74°40′ meridian. The valley is deep and narrow, and it served as the channel for a prominent esker and associated kames. It cuts at right angles across the axis of the Arab Mountain anticline of syenite gneiss, which is, at this longitude, broken by prominent vertical north-south joints. A mantle of drift obscures parts of the northern and southern contacts between the syenite anticline and the rocks that flank it. If the contacts are projected parallel to the strike of the foliation and toward the valley, they do not meet; instead they are offset horizontally about one-half mile. The geologic pattern so formed indicates that the western part of the anticline has been faulted downward relative to the eastern part. The apparent throw is about 500 feet at the north contact between quartz syenite and the flanking rocks, and 2,500 feet at the south contact. So far, no outcrops of fault breccia or mylonite have been found along the rock walls of the valley. However, these are at least 1,500 feet apart, and the valley floor is covered with glacial debris.

The marked "trough line" just noted has a length of 6 miles. The fault appears to die out within a mile or two of the ends of this "trough line," though the great thickness of drift around and north of Massawepie Lake does not permit us to fix a northern limit for the trace of the fault.

TUPPER LAKE–COLD BROOK FAULT

Sharp discontinuities in the foliation, as well as horizontal offset of rock units, point to the existence of a major fault along the center line of Tupper Lake. The exact position of the fault is unknown, but its trace must fall just northwest of County Line Island in the northern half of the lake, and close to the boundary between Franklin and St. Lawrence Counties in the southern half. Both segments of the fault are paralleled by a set of prominent vertical joints.

The most obvious effect of faulting is the horizontal offset of the south flank of the Arab Mountain anticline of syenite gneiss. On the west shore of the lake, the trace of the contact between syenite and overlying hornblende granite is at the mouth of Bridge Brook. The corresponding contact on the east shore is at Rock Island Bay, more than 1 mile farther south. The apparent throw is about 3,000–4,500 feet, the figure selected depending on the value of the average dip of the foliation. The west side has moved downward relative to the east side. Slightly steeper dips in the foliation on the east shore suggest that one half was rotated with respect to the other. However, the wavy nature of the foliation in the syenite makes it hazardous to infer rotation from the dips alone.

What of the southern extension of the Tupper Lake fault? It scarcely seems reasonable to suppose that a fault with a throw of several thousand feet ends abruptly at South Bay. Projecting the fault to the south-southeast up the narrow, kame-studded valley of Cold Brook helps to explain why many of the lowermost rock units of the Loon Pond syncline are not recognizable east of Cold Brook. If the block west of the brook had been downthrown relative to the eastern block, and then eroded, the units of the western block would crop out farther north than would those of the eastern block. If the present mapping is correct, that is precisely the relation of the mapped units.

There is a possibility that a major fault extends in a southwesterly direction from Black Bay, but the area is too obscured by drift to determine this.

LITTLE TUPPER LAKE–SPERRY BROOK FAULT

Evidence of faulting was found at three places along the southeast shore of Little Tupper Lake. At a point 1 mile northwest of the southwest end of Stony Pond and 700 feet northeast of a small land-tied island, a zone of calcite-veined, crackled, chloritized fault breccia and a little mylonite is exposed at lake level. The zone, about 5 feet wide, strikes N. 30°–50° E. and dips 60° NW. Pink, medium-grained hornblende granite lies northwest of the fault; pink granite with amphibolite schlieren lies southeast. The foliation in both rocks strikes N. 80°–90° E. and dips 80° S.

Two chloritized shear zones, 2 inches to 1 foot wide, are exposed in migmatites of the Grenville series at the west end of the point 0.8 mile southwest of Little Tupper Lake outlet. They strike about N. 60° E. and dip 65°–90° SE.

Mylonite is present along some joints in migmatitic hornblende gneiss on the point 1.3 miles east of the south end of South Bettner Pond.

A set of vertical joints striking N. 45°–75° E. is present in almost every outcrop along the lake shore. In half the outcrops, this set is dominant. The average strike of N. 50°–55° E. is about parallel to the strike of the observed minor faults.

Though the minor faults noted above may represent all the faulting in the area, it is more likely that they are satellites of a major fault zone extending the length of Little Tupper Lake. Such a fault zone, readily eroded, would account for the position and direction of the lake, which otherwise stands as an anomalous topographic feature transverse to the foliation of the country rock.

Regrettably, one cannot surely infer the type of fault, or the magnitude or direction of relative displacement. The Moonshine Pond belt of steeply-dipping Grenville rocks is apparently offset 1,500–2,000 feet horizontally where it crosses the lake. Were it not for the existence of the minor faults, one might readily join the "loose ends" of Grenville by gently curved lines, giving the belt unbroken continuity. However, if one chooses to accept the offset as additional evidence of faulting, and interprets the fault as a normal fault with essentially vertical dip, then the apparent horizontal displacement indicates a throw of approximately 3,800 feet. Farther southwest, the evidence of horizontal offset is more subjective and less compelling, owing to the difficulty of correlating rock units from the east shore westward under swamps and glacial debris. A roughly estimated offset of 1,000–1,500 feet for the Lake Lila–Slim Pond belt of mixed rocks would indicate a throw of about 2,900 feet on this segment of the fault.

Cushing (1907, p. 488) noted shattering, close-spaced fractures, and secondary quartz veins in syenite along the north shore of Duck Lake (Long Lake quadrangle). The direction of this zone is N. 65° E. Cushing states that the "excessive shattering [was] accompanied in all probability by faulting." This zone of shattering lines up rather well with the northern half of Little Tupper Lake by way of the straight, drift-filled valley now occupied by Sperry Brook and Sperry Pond (Tupper Lake quadrangle). If a major fault zone underlies Little Tupper Lake, the same zone may well continue northeast for 9 or 10 miles to Duck Lake.

C. L. Rogers (oral communication) found a structural discontinuity in the granite mass in the northwest rectangle of the Raquette Lake quadrangle. This discontinuity trends southwest through Frank Pond and Little Salmon Lake. Possibly, it continues southwest through the drift-filled valley of Panther Pond and Shingle Shanty Brook. This discontinuity would project northeastward through a large swamp to the southwest end of Little Tupper Lake.

SABATTIS ROAD LINEAMENT

The south part of the Loon Pond syncline is cut off by a line of marked structural discordance with granite on the south. This lineament is inferred to be a fault plane striking about east, whose relative downthrow is on the north. The structure dies out to the west of Bear Pond and its eastern extension is obscured by the swamp at the south end of Round Pond.

WOLF MOUNTAIN LINEAMENT

The south part of the Darning Needle syncline is sharply cut off from the granite mass of Wolf Mountain along a line of marked structural discordance. Since the granite is younger than the metasedimentary rocks of the syncline, this line could be interpreted as an intrusive contact. The line of structural discontinuity, however, is very straight for an intrusive contact, and in the absence of supporting evidence for the intrusive hypothesis, this east-northeast line appears to be best interpreted as the trace of a fault plane with the relative downthrow on the north. The lineament is at least 10 miles long and appears to die out against the Pine Pond fault at the northeast.

DEAD CREEK FLOW LINEAMENT

A major topographic low or lineament of relatively great length, narrowness, and straightness runs along Dead Creek, Dead Creek Flow, and the north arm of Cranberry Lake. This lineament may mark the site of a fault, though the evidence is inconclusive.

The foliation within the sheet of granite gneiss forming Buck and Marble Mountains gives a clue to the possible nature of the deformation. From Buck Mountain eastward to Lost Pond the dip changes successively from moderately northeast, through vertical, to moderately southeast. On the east side of Cranberry Lake the rocks of the Clare-Clifton-Colton belt of metasedimentary rocks again dip north. The changes in direction of dip may be interpreted as the result of a local northward movement that has gone so far as to overturn the rocks along the west side of the north arm of Cranberry Lake. Both the sharp bend or drag in strike at this locality and the tight folding northeast of Cranberry Lake village conform to the hypothesis. The narrow belt of rocks of the Grenville series west of Matilda Island and the sheet of granite gneiss forming Twin Mountain and State Ridge have been similarly deformed. If a fault exists here, it may be similar to the Balmat fault in nature and origin.

The main magnetic anomaly of the Dead Creek syncline appears to die out in peculiar fashion along the line of Dead Creek. A shear zone or fault might be inferred. A concealed north-south fault with an apparent horizontal displacement of about 200 feet is inferred from drilling data and the pattern of the magnetic anomaly at the northwest end of the Brandy Brook magnetite deposit, northeast of the village of Cranberry Lake. The problem is discussed in detail in the description of the deposit (Prof. Paper 377*). There is the further possibility that the Silver Pond magnetite deposit represents an offset continuation of the Brandy Brook deposit, though the drilling data do not afford any support for this.

There are, however, no positive data available to determine whether this lineament represents a shear zone that formed previous to or after the magnetite mineralization and possibly arising largely from horizontal displacement, or whether some element of normal faulting is involved, or whether the lineament is due solely to a set of strong northeast to northerly joints.

* See footnote on p. 31.

JOINTS

Routine observations on the orientation of the obvious joints were made in connection with the geologic mapping, but no special effort was made to obtain systematically the detailed data requisite for a thorough understanding of the relations of joints to other structural features. The following discussion is therefore of general nature only.

At least two directions of jointing are common in each outcrop, but in many places one is dominant, is the major cliff-former, and may exercise a controlling influence on certain elements of the topography.

The dip of the joints is commonly steep, though locally as moderate as 60°. Sheeting structure is well developed in the larger areas of granite.

In the area of the Bog River synclinorium, southwest of Bog River Flow (Cranberry Lake quadrangle), the foliation varies locally in direction from north to east, though generally it is northeast. Three major sets of joints—strike joints, dip joints (strike parallel to direction of dip of foliation), and "quartering" (oblique) joints—are represented, and at least two are visible in most outcrops. In addition, minor oblique joints are usually present, apparently without fixed relation to the major sets. Of the three major sets, any one may appear best developed in a given outcrop. The shift of dominance of the joint system with change of direction of the foliation is usually accomplished by exchange of the dominant strike joint at one point for a dip joint or at another for a "quartering" joint. Locally, the dip joints or the oblique joints have the relations of true "cross joints," that is, they are normal to the lineation (rodding or mineral elongation). The joints with a N. 65°–90° E. strike are well developed throughout the southern third of the Cranberry Lake quadrangle; they are in part about parallel to the strike of the foliation but also in part maintain their strike independent of the trend of the foliation. There is generally a system of north to north-northwest joints about at right angles to the east-northeast system.

The kinds of relations pictured in the foregoing description are found in many other parts of the area.

Within those parts of the Russell and Oswegatchie quadrangles where the linear structure is about at right angles to the strike of the fold axes, the most common joint systems include a pair or conjugate set oblique to the linear structure, and also a pair or conjugate set respectively about at right angles and about parallel to the linear structure.

In the northeastern part of the McCuen Pond syncline, north-northeast and south-southeast of Buck Pond, the foliation strikes on the average about N. 5° W. There is a conjugate set of joints here, one about parallel to the strike of the foliation but a bit more westerly, the other about at right angles to the foliation, N. 70° E. to N. 70° W.

Similarly, around the Willis Pond anticline (Childwold quadrangle) in pyroxene-quartz syenite gneiss, there is a set of joints about parallel to the strike and one about at right angles.

STRIKE JOINTS

As has been noted, a set of joints is commonly developed about parallel to the strike of the foliation and with a steep dip. Since the foliation itself has such a marked effect in controlling topography, and since the foliation is also steep for much of the area, the strike joints are in many places not as marked or obvious a feature as the cross joints. They are, however, equally well developed in most cases.

CROSS JOINTS

Several of the major structural features, such as the Stark and Arab Mountain anticlines and the arc of hornblende granite in the southern part of the Tupper Lake quadrangle, have a well-developed set of joints about at right angles to the strike of the foliation, or to the general curvature of the part of the structure in which they are located. Since the linear structure in these major structures is about parallel to the fold axes, it follows that the joint system is also about at right angles to the linear structure. However, the right-angled relation, for the data at hand, seems to correspond in part somewhat closer to a right-angled relation to the general curvature of the structure than to the linear structure. There is commonly also a set of joints subparallel to the strike of the foliation. Locally, this latter set is not as well developed as the cross joints.

The relation of joints at right angles to the foliation is especially well shown in the rocks of the Arab Mountain anticline, between Cranberry Lake on the west and the New York Central Railroad on the east. In this belt, two joint systems are dominant, one roughly parallel to the direction of strike (east-northeast to east) of the foliation, the other about parallel to the direction of dip. West of the railroad a joint set strikes about north or N. 10°–20° W., and in a belt east of the railroad a set strikes north to N. 10° E. The strike of the foliation is not constant but swings from east-northeast, west of the railroad, to about east, east of the railroad.

Similarly, the two dominant joint systems in the area between Peavine Creek and Inlet on the southeast and Chaumont Swamp on the northwest are respectively about parallel, and about at right angles, to the direction of strike and to the foliation. Some oblique joints also occur here.

Another arc with a sharp change of curvature occurs in the granite and included metasedimentary rocks of the southwest rectangle of the Cranberry Lake quadrangle. A group of northwest joints is well developed in a belt from Deer Mountain to Partlow Mountain, whereas to the east the equivalent set of joints strikes north to north-northwest.

A body of hornblende granite with the form of an arc of large radius extends from Partlow Lake (Cranberry Lake quadrangle), across the southern part of the Tupper Lake quadrangle, by way of the Charley Ponds, Little Tupper Lake, and Slim Pond. From Partlow Lake to the Charley Ponds there is a very well developed joint system striking N. 10°–30° W. (average, N. 12° W.) about at right angles to the strike of the east-northeast foliation (average, N. 60° E.). Between the Charley Ponds on the west and Round Lake and Little Tupper Lake on the east, the comparable joints strike about north (average, N. 9° E.), again at right angles to the east (average, N. 83° E.) strike of the foliation. In the southwest corner of the Tupper Lake quadrangle, southeast of Little Tupper Lake, the foliation strikes west-northwest (average, N. 70° W.) and a joint system strikes northeast (average, N. 49° E.).

The linear structure in the arc on the Tupper Lake quadrangle plunges generally eastward. The strike of the dominant set of joints is somewhat more easterly than is appropriate for a right-angled relation to the strike of the foliation, but it is more consistent with a right-angled relation to the trend of the linear structure.

The joint system described above is by far the dominant one. However, where the strike of foliation is east-northeast there is also a subordinate set of joints striking northwest (N. 35°–45° W.) and east (N. 70° E.–N. 80° W.); and where the strike of the foliation is west-northwest, there are two subordinate sets of joints striking N. 0°–20° E. and N. 70° E. to N. 80° W. respectively, and a few northwest joints.

Furthermore, in the Dead Creek and Darning Needle synclines there are joint systems whose directions change with marked changes in the direction of strike of the foliation. Also in the Dead Creek syncline there is a well-developed set of joints at right angles to the lineation—some of which are quartering to the direction of strike and dip of the foliation—and also a set of joints parallel to the strike of the linear structure.

Within the Stark complex the strike of the foliation is north at the south, and east-northeast at the northeast. The lineation similarly has a general south trend south of the Middle Branch of the Grass River and an east-northeast trend northeast of the river. Similarly, there is a corresponding sharp swing in the orientation of a system of joints. North and south of Long Pond (Russell quadrangle) the lineation trends north, and there is a prominent set of easterly joints. In a belt about two miles wide from Clarksboro to Degrasse a dominant set of joints strikes about N. 50° W. This set of joints appears to be related to the zone in which the foliation of the complex bends rather sharply from north to east-northeast. Farther northeast, in a belt from Allen Pond to Lower District School, there is a major set of joints with a strike of about N. 25° W. and a corresponding east-northeast lineation. On the Stark quadrangle the lineation trends N. 60°–75° E., and a corresponding system of cross joints strikes north to N. 35° W. Throughout this complex there is also a set of joints about parallel to the linear structure. Oblique joints are common.

BELTS OF NORTHEAST TO NORTH-NORTHEAST JOINTS

In several belts there is a strong development of northeast to north-northeast joints that are, at least in large part, oblique to the direction of the strike and dip of the foliation and to the linear structure.

One such belt with northeast joints as a major system lies on the Oswegatchie quadrangle, between Streeter Camp and Star Lake on the northwest, and Tamarack Creek and Maple Hill on the southeast.

Another belt of joints with a north-northeast to northeast strike, and with a generally oblique relation to other structural features, strikes northeast from the Oswegatchie River through Wanakena, Dead Creek Flow, Cranberry Lake, Brandy Brook Flow, and the southeast corner of the Stark quadrangle.

A third belt of N. 20°–45° E. joints runs across the Tupper Lake quadrangle and includes the southeastern part of that quadrangle in a belt lying southeast of Loon Pond Mountain, Horseshoe Lake, and Gull Pond. Here again the joints are generally oblique to other structural elements.

Joints with a north-northeast to northeast strike are prominent in many areas in the Adirondacks. The eastern half and the southern border of the Adirondacks are broken by a system of normal faults in which those with a north-northeast to northeast strike are exceptionally well developed. The faulting is particularly evident and pronounced near Lake George and Lake Champlain. This north-northeast to northeast fault system has been certainly traced as far west as the Saranac quadrangle (Buddington, 1953), where a strong northeast joint system has been noted in a belt including Middle and Lower Saranac Lakes. The Tupper Lake belt of northeast joints lies along the southwest extension of this belt and is certainly related to this system. The faults are known to be in part of post-Ordovician age, though some may date back to the Precambrian. Faulting of northeast trend has been inferred parallel to the northeast part of Tupper Lake and parallel to Little Tupper Lake. The northeast joint systems through the Tupper Lake and Cranberry Lake belts and the belt southwest of Star Lake may thus be genetically related to the same deformation that produced the faulting of the eastern belt, which was much less intense in the northwest. The northeast joint system of the Tupper Lake belt is correlated with the Tupper Lake elongate basin shown in the generalized contour map of the Adirondacks, and the Cranberry Lake belt is correlated with the east side of the Childwold terrace and the west flank of the Adirondack mountain section.

REGIONAL EAST-NORTHEAST TO EAST JOINTS

Throughout the entire district there are joints whose strike is between N. 65° E. and N. 80° E. These joints are in considerable part oblique to the strike of the foliation and the linear structure. Since the latter structures have such different strikes in different areas, however, it follows that locally the east-northeast joints are more or less parallel to the strike or dip of the foliation, or to the direction of the linear structure. These joints appear to be of regional nature and origin.

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