The Precambrian geology and talc deposits of the Balmat--Edwards district, northwest Adirondack Mountains, New York

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31 plates (separate, tube no.2)
22 figures (separate)
33 tables (included)

by

A. E. J. Engel

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by A. E. J. Engel

Abstract

The Balmat-Edwards, or Gouverneur, mining district is in the Grenville Lowlands of the northwest Adirondacks, St. Lawrence County, New York. In the district, important deposits of talc and zinc and minor deposits of lead are mined from a northeast-trending belt of marble. Deposits of pyrite that were prospected and mined during and before World War I occur in schists in the vicinity of Pyrites and Antwerp villages. Fifteen to 50 miles east-southeast is the St. Lawrence County magnetite district, which contains commercially important magnetite deposits.

Most of the bedrock of the Balmat-Edwards district and adjoining areas consists of metasedimentary and metasomatic rocks commonly referred to the Precambrian Grenville series. The marble that encloses both the talc and the zinc deposits lies conformably, or nearly so, upon a very thick mass of quartz-biotite-oligoclase-garnet gneiss (metagraywacke?). The gneiss is extensively veined and replaced by quartz-perthite pegmatite and inequigranular granitic gneiss. Several thinner paragneisses of more potassic composition are interbedded with the marble, and there are thin sheets and lenses of amphibolite in both the marble and the underlying gneiss. The amphibolite is of ambiguous origin; some of it probably represents pre-granitic, semi-concordant intrusions of basaltic and gabbroic magma, but much of it may have been derived from beds of tuff and impure carbonate. The metasedimentary rocks exposed in the Balmat-Edwards district are about 4,000 feet thick.

Metamorphism in the region is pervasive and complicated. All the rocks are complexly folded and thoroughly recrystallized to an intermediate or slightly higher grade of metamorphism (upper amphibolite facies). Many large masses of marble, gneiss and schist behaved as plastic solids during the more intense stages of deformation and flowed extensively. In contrast, scattered segments of many silicate-rich beds remained relatively brittle; these were ruptured and complexly dispersed in the more mobile associated rocks, especially the marble.

Granitic rocks have formed by mechanical injection, replacement, and probably by partial melting of metasedimentary rock, although the relative extent of these processes is uncertain. Gradation in composition between paragneisses and paraschists on the one hand and granite on the other are common and widespread. Amphibolite is less widely and more irregularly granitized and injected, and marble is only locally injected and replaced by granite and pegmatite. The formation of most of the existing minerals in the metasedimentary rocks was facilitated by pore fluids, partly of magmatic origin, but also partly derived from the dehydration and decarbonation of the metasedimentary rocks.

Granitic rocks constitute about 25 percent of the exposed bedrock, amphibolite about 5 percent, and metasedimentary rocks about 60 percent. The remaining 10 percent consists of patches of quartzite and quartz breccia in the Potsdam sandstone of Cambrian age.

Age determinations based on analyses of uraninite in pegmatites near the district, helium content of magnetites of the district, and A^{40}/K^{40} ratios in biotites and feldspars indicate that granitic intrusion and deformation culminated about 1 billion years ago.

The age of the deposition of the present sedimentary rocks is unknown, but all are older than the intrusive activity and orogeny.

The marble appears to have been chiefly derived from almost pure dolomite and dolomite with cherty and quartzose interlayers. Primary calcite seems to have been rare, and only one of the many major carbonate-bearing zones appears to have been appreciably argillaceous. All the marble is slightly pyritic, containing traces of (metamorphic?) pyrite, and many beds are slightly graphitic.

Metamorphic reactions between silica and dolomite produced various magnesium- and calcium-bearing silicates, especially diopside, and, in the talc-bearing zone, tremolite. Forsterite occurs rarely. The diopside, forsterite and tremolite are partly replaced by anthophyllite, serpentine, and talc. The paragenesis appears to be diopside (fosterite)-tremolite-anthophyllite-serpentine-talc, but with considerable overlap. The earliest metamorphic mineral, diopside, was formed at the highest temperature, and the ensuing mineral transformations were accompanied by falling temperature, and increasing participation of water. Much of the diopside must have formed through reaction of dolomite with sedimentary silica. At many places the formation of diopside probably involved little change in composition, except for loss of CO_2. Elsewhere the formation of diopside, as well as of tremolite, anthophyllite, serpentine, and talc, was partly due to the introduction of silica and water. Calcite is a common by-product of the metamorphic reactions that produced tremolite, anthophyllite, serpentine and talc.

The marble is divisible into at least seven major metasedimentary units, although in the southwest part of the district many more marble zones can be identified. In most places the beds are overturned, so that the unit regarded as the oldest (the one next to the quartz-biotite-oligoclase gneiss) appears to be the highest in the sequence. This unit, called Zone 1, is an almost pure dolomite. It is in places as much as 250 feet thick. Zone 2, the next-younger unit, is a rusty, pyritic quartz-feldspar-chlorite-mica schist, called the pyritic schist. This unit is present only locally, although prior to metamorphism it probably was more widespread. Like Zone 1, it has a maximum thickness of about 250 feet.

Zone 3 is a dolomite, like Zone 1, and the two cannot be distinguished where the pyritic schist is not exposed. Zone 3 ranges up to 200 feet thick. Adjacent to Zone 3 is a thick sequence of siliceous diopsidic dolomite, interlayered with dolomite and an anhydrite-bearing stratum. This sequence, described as the upper siliceous dolomite, is as much as 1,200 feet thick in the Sylvia Lake area. There it comprises eight distinguishable zones, numbered 4 to 11. Zones 5, 7, and 9 are dolomites much like Zones 1 and 3. Zones 4, 6, 8, and 11 consist of highly siliceous dolomite. Zone 10 includes layers of anhydrite and of tremolite mixed with calcite. Zone 12 is a nearly pure dolomite, as much as 200 feet thick, which is succeeded by the rock mined for talc, Zone 13.

Zone 14 is the so-called "footwall marble." Near the talc deposits this unit consists largely of siliceous calcite, but lateral extensions of it, in areas where talc rock is poorly developed or absent, are dolomitic. It has a maximum thickness of 350 feet. The succeeding and youngest unit (Zone 15) distinguished is a rusty, pyritic calcite-diopside-quartz-feldspar-tremolite-phlogopite marble, called the rusty marble, with a maximum thickness of about 250 feet.

All the above units are believed to represent stratigraphic zones within the marble. The dolomite and at least half the quartz appear to be sedimentary or diagenic in origin. They have distinct relict bedding, and most of the major variations in the thickness and form of these units appear to be caused by deformation accompanying metamorphism. The anhydrite in the upper siliceous dolomite may be a true sedimentary bed, although it could have been formed by remarkably uniform replacement of a stratum of marble. The rusty marble appears to be derived from an argillaceous dolomite, but some silica and alkalis were introduced into it during metamorphism.

A weighted average composition of all the marbles, in volume percent, is roughly: dolomite 42, quartz 16, calcite 18, diopside 8, tremolite 7, other minerals 9.

The above sequence of marble, together with several large associated lenses of quartz-diopside rock of unknown stratigraphic affinities, is mapped as the Sylvia Lake-Cedar Lake marble belt. It is "underlain" by a younger potassic migmatite, the median gneiss. This unit appears to represent a granitized and injected argillaceous sandstone. It is as much as 900 feet thick in some places, but in the thicker parts it may be doubled by isoclinal folding. The median gneiss separates the marble described above from younger marble in which no subdivisions were distinguished, and whose exact form and thickness are obscured by deformation and masses of amphibolite and granite.

The Balmat-Edwards district lies on the southeast flank of an anticlinorium. To the northwest of the district, constituent folds in the anticlinorium plunge gently northeast and southwest like the axis of the anticlinorium. Within the district and southeast of it, the folds and other linear elements are refolded into forms that plunge northwest at moderate angles, athwart the trend of the anticlinorium. This refolding in the anticlinorium takes place in a zone where rock flow and shearing are more extensive than they are in the northwest. The grade of metamorphism also increases to the east and southeast, from typical almandite-amphibolite facies in the Balmat-Edwards district to a lower granulite facies in the adjacent margin of the Adirondack massif. This increase in grade of metamorphism is obviously correlated with the increasing temperature, pressure, and abundance of magmatic rock. The refolding may have occurred either in one complex epoch or during a second, separated period of deformation.

The talc deposits are zones rich in tremolite, anthophyllite, talc, and serpentine. They are interlayered with the marble and conform in general with the bedding. Of the five that have been mined, the two largest, the Fowler and Talcville belts, and a small one, the Edwards belt, appear to be all at about the same stratigraphic horizon. The other two, the American and Balmat talc belts, may also be in the same stratum, duplicated by folding. The total length of these five talc belts is a little more than 6 miles. They are as much as 425 feet thick, and have been traced down dip to the northwest for about 2,500 feet. In places the commercial talc rock makes up much or all of a particular belt. Elsewhere, zones composed of these minerals are interlayered with silicated calcitic marble and manganiferous tremolite, or locally with anhydrite and micaceous amphibolite.

The talc belts were formed by partial to complete replacement of dolomite and siliceous dolomite. Silica probably was introduced during the process. Most of the constituent magnesia was probably leached from the adjoining footwall marble, which is now calcite but probably was originally dolomite. Most of the silica that is believed to have been introduced into both the talc zones and the underlying marble may have been sweated out of subjacent metasedimentary rocks.

The sequence of processes which resulted in the formation of talc is inferred to have been as follows: Dilute silicate solutions saturated with carbon dioxide migrated into the marble along shear zones in adjacent migmatites, especially from the median gneiss, dissolving the nearby dolomite either congruently or incongruently, especially along its contact with the gneiss. Calcite and quartz were precipitated, while the magnesia from the dolomite, together with some silica and water, were moved into the overlying marble along a zone of profound shearing that tended to follow bedding. There the magnesian silicates were formed in the sequence: tremolite-anthophyllite-serpentine-talc. Carbon dioxide and calcite, formed as byproducts of these reactions, were largely expelled from the marble. The associated zinc deposits replace tremolite and most of the anthophyllite, but are locally replaced by some of the last-formed serpentine and talc. Some of the talc rock that is mined selectively contains much tremolite, and some of it contains mixtures of tremolite and all the other silicates listed above. Masses of commercial size rarely contain more than 60 percent of the mineral talc and much commercial "talc" is 80 percent or more tremolite.

The weighted average mineral composition of all the rock in the talc belts of the district, commercial and non-commercial, in weight percent, is approximately: tremolite 49, talc 18, calcite 11, serpentine 10, anthophyllite 7, quartz 3, other minerals 2.

In 1946, seven talc mines and six mills were in operation. Production in the decade 1936-1946 averaged slightly over 100,000 tons per year. Known reserves in 1946, as calculated to depths of 2,000 feet below the surface, were about 10 billion tons of somewhat more tremolitic talc than that already mined.

Introduction

Location and importance of the district

The Balmat-Edwards mining district, which has also been called the Gouverneur district, is in northern New York State, in the northwestern part of St. Lawrence County. It lies about 100 miles north of Syracuse and 25 miles southeast of the St. Lawrence River (Fig. 1).

Figure 1. in ~~near here.~~ pocket

The village of Balmat lies at the southwest end, and Edwards at the northeast end of the district; they are about 12 miles apart. The name Gouverneur district evolved because Gouverneur is the largest nearby business center, but the town is about 7 miles northwest of the nearest point in the district. The district extends along the southeast margin of the lowlands of the St. Lawrence valley, which merge into the northwest foothills of the Adirondack Mountains.

The talc deposits in the district are the largest of the tremolitic type known in the world. Their annual production has been between 100,000 and 150,000 tons. Seven talc mines were operated in 1949. The mines are owned and operated by three companies: the Gouverneur Talc Company with offices at Balmat, N. Y.; the International Talc Company of Hailesboro, N. Y.; and the Loomis Talc Company of Gouverneur, N. Y.

Figure 1. Index map showing location of the Balmat-Edwards (Gouverneur) mining district in the northwestern Adirondack Mountains, New York.

The district also contains important deposits of zinc ore. The zinc mined from the district made up about 10 percent of the total U. S. production in 1950. Most of it came from the Balmat and Edwards mines of the St. Joseph Lead Company (Brown, 1936a; 1942; 1947a). Small amounts of lead are also produced from these mines. The Hyatt mine, until recently operated by the Universal Exploration Co., has produced some zinc and a little lead.

Sizable deposits of pyrite and pyrrhotite (Smyth, 1912; Buddington, 1917, 1934, p. 209-214; Prucha, 1953) also occur in and around the area that contains deposits of zinc and talc (fig. 2). Several

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Figure 2. in pocket.

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of the pyrite and pyrrhotite deposits were highly productive sources of sulfur before World War I, and economic interest in these deposits was revived in the 1950's (Purcha, 1953).

Figure 2. Generalized geologic map of the northwestern Adirondacks showing distribution of the major metasedimentary, igneous, and metasomatic rocks, and mineral deposits.

Fissure veins of galena and calcite occur at several places northwest of the district, especially near Rossie, N. Y. Buddington (1934, p. 203-209) has summarized information on these deposits, and notes that they must have produced several thousand tons of galena. Diamond drilling along two of the veins was undertaken by the U. S. Bureau of Mines in 1950, but with negative results.

The region that includes the Balmat-Edwards district is also of keen interest to geologists and mineral collectors as a source of fine specimens of many minerals other than ore minerals. Handsome crystals of apatite, danburite, hexagonite, tremolite, pyroxene, phlogopite, scapolite, tourmaline and other minerals collected there are described in the writings of European as well as American mineralogists. (See, for example, Kunitz, 1930; Dana, 1893, p. 1062-1065). These mineral specimens were most eagerly sought during the period 1850 to 1890, but many of the more famous localities (Agar, 1923; Buddington, 1934, p. 221-222; Cushing and Newland, 1925, p. 80-89), are now obscured by weathering and plant growth. Interest in beautiful and showy crystals has gradually been superseded by a deeper, widespread interest in the origin and evolution of the Grenville-like rocks. An attempt has recently been made to review and summarize most of the work that has been done on these problems (Engel and Engel, 1953a)

Purpose and scope

Despite the size and value of the talc deposits in and near the Balmat-Edwards district, comparatively little geological information about them has been published. This report, which is a phase of the strategic and critical minerals investigations of the U. S. Geological Survey, is a study of the occurrence and origin of the talc deposits and the associated Precambrian rocks. As the zinc deposits are well known, especially through the long-continued and thoughtful studies of Dr. J. S. Brown, chief geologist for St. Joseph Lead Company (Brown, 1936a, 1936b; 1941; 1942; 1947a; 1947b), no attempt is made to treat them in detail. The zinc and talc deposits are so closely interrelated, however, that an understanding of either requires some attention to the other. This fact was recognized by the zinc producers, especially by Dr. Brown and other officials of the St. Joseph Lead Company and by Mr. Edward Jennings, general superintendent of operations for the Universal Exploration Company. I was therefore cordially invited to study both talc and zinc in the deposits being mined by those companies, and to examine many of the drill cores obtained in exploration for zinc. One result has been a joint paper on the stratigraphy and structural features of the Grenville series in the Balmat-Edwards district (Brown and Engel, 1956).

In conjunction with our studies of talc and zinc deposits, mapping was extended outward to the limits of the area shown in plate 1 to determine the distribution and geologic setting of the deposits. Field work and mine mapping were begun in October 1944 and carried on continuously until March 1946. Assistance was given throughout the investigation by my wife, Celeste G. Engel, and by C. G. Johnson, Karl Stafansson, and James J. Page; C. N. Bozion also assisted for about six weeks in mine and field mapping. Geologic maps of all accessible mine workings were prepared largely on a scale of 50 feet to one inch. About 75,000 feet of drill core was logged in varying degrees of detail. An outcrop map prepared on a scale of 800 feet to the inch, forms the basis of plate 1. Between May 1946 and August 1950, new or critical mine openings and some surface exposures in the district were revisited and studied by me and by James J. Page and C. N. Bozion. Petrographic, chemical, and X-ray studies were made by me and my wife.

This report has been written by A. E. J. Engel, who is responsible for all descriptions, comments, and conclusions. The illustrations were prepared jointly by A. E. J. Engel, C. N. Bozion, Celeste G. Engel and James J. Page. The report describes and discusses the talc deposits, the associated Precambrian rocks, and the regional geologic setting. Readers interested solely in the talc deposits may wish to limit their reading to the latter part of this paper. ~~(p. 197-286).~~

Acknowledgments

Specific help in the form of information and ideas is acknowledged, insofar as possible, at appropriate places in the text, but some general thanks must be given here. Help was continually asked and received from both officials and employees of the International Talc Company, the Loomis Talc Company, and the Gouverneur Talc Company. Among those who were particularly helpful were Fred Kuehl, George Hurly, and Max Tessmer of the International Talc Company; Eugene Magnus, Prescott Loomis, Frank Atkinson, David Ryder, and Orton Smally of the Loomis Talc Company; and Robert MacClellan and John Bolger of the Gouverneur Talc Company.

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Residents of the region who befriended and helped us in many ways include Stella Ackerman, Roy Brown, Harry Rogers, Ward Howard, Charles Kinney, and Mrs. Harold Story.

The scientific writings of many geologists mentioned in the body of this report have contributed much to the author's thinking. Association in the field with John S. Brown, Arthur Buddington, Norman Donald, and Benjamin F. Leonard III, was stimulating and enlightening.

Geologic setting

Grenville Lowlands

The Balmat-Edwards district lies in what is commonly called the Grenville Lowlands, an arcuate belt some 25 miles wide at the northwest side of the main Adirondack massif (fig. 2). The district consists largely of the metasedimentary rocks referred to as the Grenville series (Engel and Engel, 1953b; Engel, 1956).

Metasedimentary rocks

In the Grenville Lowlands, Precambrian metasedimentary and metasomatic rocks predominate by far over igneous rocks. About 5 or 10 percent of the exposed bedrock is amphibolite, and roughly 20 percent is granite pegmatite and granitic gneiss, partly of igneous and partly of metasomatic origin.

The metasedimentary rocks occur as a complicated sequence of lithologic zones that converge to the northeast at a very low angle with the convex margin of the central Adirondack massif to the southeast. These relations are generalized in figure 2. The dominant strike of the metasedimentary rocks is northeast, and their general dip northwest.

Stratigraphy

The metasedimentary rocks in the Grenville Lowlands may be regarded as comprising either three or five major units, depending on which of two differing interpretations of structure one adopts. Assuming, as was previously done (Engel and Engel, 1953b, figs. 1 and 2, table 1), that the series includes five units, all overturned to the southeast, the sequence from the central massif northwest to the St. Lawrence River (youngest to oldest beds) is as follows (fig. 3): (a) Zone 5,

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Figure 3. ~~near here.~~ in pocket

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feldspathic gneiss or granulite, (b) Zone 4, younger marble belt, which forms the host rock to the talc and zinc deposits of the district, (c) Zone 3, quartz-biotite-oligoclase gneiss or major gneiss complex, parts of which form the southeastern, northwestern, and southern boundaries of the district (pl. 1), (d) Zone 2, central marble belt, and (e) Zone 1, metasedimentary belt of the Black Lake area, which consists of calcitic marble and quartzose gneiss. These are briefly described in table 1, in the inferred order of age, with estimates of their thickness

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Table 1. General stratigraphic sequence (in order of increasing age) in the Grenville series, Grenville Lowlands, northwestern Adirondacks, New York. (p.19a)

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before metamorphism. As may be seen from figure 3, the metasedimentary sequence is interpreted as having its uppermost, youngest exposed member, the feldspathic gneiss or granulite, in contact with and partly obliterated by igneous and metamorphic rocks of the Adirondack core.

Table 1. General stratigraphic sequence (in order of increasing age) in the Grenville series,

Grenville Lowlands, northwestern Adirondacks, New York

Major rock unit		General lithology	Approximate thickness ~~after~~ before metamorphism
Geographic designation	Common designation		
Feldspathic gneiss of Harrisville-Russell region Zone 5, Pl. 1 and Fig. ~~1~~ 3	Upper feldspathic gneiss or feldspathic granulite	Quartz-microline gneiss, in part pyroxenic and biotitic, and locally hornblendic	500-1000'
* Marble complex of Natural Bridge- Balmat-Edwards-West Pierrepont regon Zone 4, Pl. 1 and Fig. ~~1~~ 3	Upper marble belt	Dominantly dolomitic marble with interlayers of pyritic graphitic schist, feldspathic, biotitic and pyroxenic gneiss	3000'
Gneiss complex of Lewisburg-Antwerp-Hermon region Zone 3, Pl. 1, and Fig. ~~1~~ 3	Quartz-biotite gneiss or major gneiss complex	Quartz-biotite-feldspar migmatite in part garnetiferous and silli-manitic. Includes several marble zones and numerous thin inter-layers of amphibolite	2500' (exclusive of most granite)
* Marble complex of Gouverneur-Pyrites region Zone 2, Pl. 1, and Fig. ~~1~~ 3	Gouverneur (central) marble belt	Dominantly marble with scattered interlayers of quartzite and feldspathic and pyroxeni~~t~~c gneisses	3000-4000'
* Metasedimentary rocks of Rossie-Black Lake belt Zone 1, Pl. 1 and Fig. ~~1~~ 3	Black Lake metasedimentary belt	Complexly interlayered marble, quartzite and biotitic, felds-pathic, and pyroxenic gneisses	3000-4000'

* Possibly segments of the same carbonate-rich zone.

19a

1-19a

Figure 3. Generalized plan and section showing the major structural features in the Grenville series of the northwestern Adirondacks, New York.

Possibly, however, the marble units designated Zones 4, 2, and 1 above, in table 1 and on figure 3, are but parts of a single unit flanking opposite sides of an obscure isoclinal fold in the quartz-biotite gneiss. This interpretation, which reduces the number of major units to three, is considered less probable than the one requiring five major units (Engel and Engel, 1953b, p. 1058).

Lithologic features and thickness

The general lithologic features of each unit and its estimated thickness before metamorphism are shown in table 1. These data, if compared with figure 3, indicate the dominance of magnesian marbles in what is believed to be the central and upper half of the exposed sequence, and of thick layers of siliceous calcitic marbles near the base. Gneissic, schistose, and quartzitic interlayers are common in all three of the carbonate units, being most abundant in the oldest (most northwesterly) unit. These were probably derived from shales, sandstones, and tuffs. The largest body of gneiss, regarded as separating the central marble belt (Zone 2) from the younger marble belt (Zone 4), was presumably derived from a clastic unit about 2,500 feet thick prior to metamorphism (table 1). The bulk composition of the least metamorphosed layers of this gneiss is very close to that of many geosynclinal graywackes and of certain dacitic tuffs. Other considerations, however, especially some aspects of lithology and the association of the gneiss with thick strata of clean carbonate rock and fine-grained, well-sorted clastics (especially orthoquartzites), suggest that the unit could have been derived from sodic clays, or possibly from clays mixed with sodium-rich zeolite-like minerals such as analcite or clinoptilolite (Engel and Engel, 1953b).

The feldspathic gneiss or granulite (Zone 5), which adjoins the granite-syenite complex of the ~~Adirondacks~~ Adirondack core, has been studied and mapped by Buddington and Leonard (in press), who believe that this rock was probably derived from an illitic sandstone. Conglomerates and other coarse-grained clastic sediments are unknown in the sequence.

The marbles in the central and upper part of the sequence tend to be low in sedimentary alumina and iron, but high in silica and magnesia, and being intimately associated with pegmatite and granite, they include much secondarily introduced alkali and silica.

Amphibolite sheets of more or less obscure origin are also abundant throughout the section. Some of these may be sills of basalt and diabase, but most appear to be derived from sedimentary beds, perhaps tuffs or marls. There are a few scattered bodies of amphibolite clearly formed by skarn-like replacement of carbonate layers (Engel and Engel, 1953a; Engel, 1956). These metasomatic amphibolites are almost identical with amphibolites known to be derived from gabbroic rocks. Obviously an awareness of the exact rocks from which the amphibolites have been derived would be invaluable in stratigraphic reconstructions of the metasedimentary rock series.

The premetamorphic thickness of the metasedimentary rocks exposed in the Lowlands is estimated to have been 12,000 feet or more (see table 1). At least half of this consisted of siliceous limestone and dolomite. Sandstones, which are the coarsest clastic rocks believed to have been present, appear to have formed less than a third (or at most a half) of the section. The remainder of the clastics, perhaps as much as 25 percent, are thought to have included shale, silt, chert, and tuff.

Age

The metasedimentary rocks are the oldest rocks exposed in the Grenville Lowlands. No exposure of floor upon which they were deposited has been recognized; probably this floor was largely or wholly replaced by, or mobilized into, the great masses of granitic rock which have invaded the Grenville series.

The age of the metasedimentary rocks must be more than a billion years, for this is approximately the age of lead- and uranium-bearing minerals--zircon, allanite, uraninite, and magnetite--in the granitic rocks and pegmatites emplaced in the metasedimentary rocks (Brown, and others, 1951; Shaub, 1940; Marble, 1943; Tilton and Davis, 1959; and the summary in Engel and Engel, 1953a). The lengths of the period of regional metamorphism preceding intrusion, the interval between the beginning of metamorphism and the ending of sedimentation, and the period of sedimentation itself, are all unknown. If these processes proceeded at about the same rates as they are estimated to have done in more recent time (Kay, 1951, p. 96), the basal layers in the Grenville series were deposited at least 1.1 billion and possibly over 1.3 billion years ago.

Structural features

The metasedimentary rocks were complexly folded and otherwise deformed during the regional metamorphism and igneous intrusion. The dominant structural feature in the Grenville Lowlands is believed to be an anticlinorium, roughly parallel to the northeasterly course of the St. Lawrence River. A highly simplified cross section of this anticlinorium is shown in figure 3. If this reconstruction is correct, the Black Lake metasedimentary belt (Zone 1), forms the core of the anticlinorium. The southeast flank of the structure is visualized as overturned to the southeast, against the northwest margin of the central Adirondack igneous massif. Accordingly, the quartz-biotite-oligoclase gneiss (Zone 3), the upper marble belt (Zone 4), and the feldspathic gneiss (Zone 5), which are believed to constitute the upper half of the Grenville series, are largely overturned throughout their extent, as shown in figure 3. Consequently they dip northwest, toward the core of the anticlinorium, and away from the Adirondack highlands.

According to this interpretation, the Balmat-Edwards district lies on the overturned southeast flank of the anticlinorium, largely within the upper marble belt (Zone 4). The gross structure of the marble is synclinal. This syncline, one of many northeast-trending second-order folds on the southeast flank of the anticlinorium, has been very much distorted, however, by a combination of bowing and refolding that formed folds plunging north to northwest. The resulting structure is far more complicated than the folded structures farther northwest, in the central part of the Lowlands belt. There most of the folds, irrespective of their magnitude, trend northeast-southwest, parrallel to the general strike of the anticlinorium (fig. 3). Such folds are referred to hereafter as accordant. The "refolds" in the accordant Balmat-Edwards syncline, and in many other accordant folds near and in the massif, are referred to as cross folds, because their trend is across that of the parent anticlinorium.

The cross folds and cross lineations may be genetically related to, and synchronous with, the accordant folds trending northeast and southwest in the central part of the Lowlands; but more probably they have resulted from refolding of accordant folds, as discussed under the heading of "Structural features in the district" (p. 132). The zone of transition between the little-deformed accordant folds of the central lowlands and the folded folds (cross folds) of the southeastern border zone, trends diagonally across both the arbitrary outlines of the Balmat-Edwards district and the syncline that forms its principal structural feature.

The above interpretations have been discussed in detail by Brown and Engel (1956) and are consistent with the tentative stratigraphic synthesis of Buddington (1934). They also conform in general to Buddington's structural interpretations of the Lowlands as the northwest flank of a major synclinorium (1939, p. 243). They are at considerable variance, however, with views of Cushing and Newland (1925, p. 17) and with an earlier interpretation of Brown (1936).

Igneous and metasomatic rocks

About a quarter of the granitic rock in the Grenville Lowlands is an equigranular quartz-microcline granitic gneiss, which Buddington (1929; 1939, p. 15~~7~~3-158; 1948, p. 34-36) regards as satellitic to, and a volatile-rich derivative of, the larger alaskitic igneous granite in the core of the Adirondacks. The remaining three-fourths is a widespread complex of gneissic rock, which is correlated with Buddington's Hermon type of granitic gneiss (1939; 1948). It resembles an albite-oligoclase granite in the Adirondack massif, but the relations of the two are far from definite. In the Grenville Lowlands, only a minor part of the albite-oligoclase granite is clearly magmatic. The remainder--possibly as much as three-fourths--may have been formed by partial refusion of metasediment or by interaction of granitic magma and magmatic fluids with sedimentary rocks (Engel and Engel, 1953b, 1958a). Igneous and metamorphic granites appear to be closely interrelated. There are lit-par-lit injections into the paragneisses and schists--parts of which show marked gradations between the paragneiss and the granitic gneiss. There are also well defined bodies, both accordant and transgressive, of granite and pegmatite. Texturally, the Hermon granitic gneiss complex in the Grenville Lowlands ranges from equigranular medium- to fine-grained granite to coarse-grained inequigranular granitic gneiss and pegmatite. The inequigranular rocks, which predominate, contain augen and subhedral crystals of microcline (or rarely of sodic plagioclase) in a matrix of quartz, biotite, microcline, and oligoclase-albite. These rocks are mostly gneissic, but in some areas they show little or no foliation. The evolution of these textures is discussed at length in separate papers (Engel and Engel, 1953b; 1958a).

The emplacement of both the alaskitic granite and of the Hermon complex in the Adirondacks may have been preceded, and was certainly accompanied by, extensive alteration of the enclosing metasedimentary rocks. Much of the alaskite forms phacoliths which were emplaced in the apices of folds in marble and in quartz-biotite-oligoclase gneiss as these folds evolved (Buddington 1929; Buddington and Leonard, in press). The interaction of these granites with the rocks of the Grenville series is manifested by pyrometasomatic sheaths of amphibolite, garnet-sillimanite gneiss, and oligoclase-quartz gneiss which partly encircle some of the phacoliths. Most of the sheaths are less than 200 feet thick. Effects of the interaction of fluids from the phacolithic granites and the metasedimentary rocks at considerable distances from the phacolithic masses are not marked, but many such effects may have been obscured by widespread modifications induced during the evolution of the Hermon granitic gneiss complex. In some places shear zones containing quartz and sillimanite extend outward from the phacolithic bodies into the quartz-biotite-oligoclase gneiss along its dominant foliation.

The impregnation of the metasedimentary rocks of the Northwest Lowlands by the alkali-silicate fluids which produced the granitic complex was widespread although by no means uniform. Igneous rock cannot in general be distinguished from granitized metasedimentary rocks, and some pegmatite "dikes" appear to grade into veins of quartz and pegmatite. Well-defined halos of contact-metamorphic minerals around igneous-looking bodies are of irregular shapes and separated by long stretches of barren contacts (see also Agar, 1923, especially p. 133). On the other hand, nests of tourmaline, scapolite, and phlogopite appear at irregular intervals throughout the marbles. Small, isolated grains and well-defined lenses of what would generally be regarded as contact minerals are also scattered through large bodies of marble that have no visible contacts with clearly intrusive granite. Most of the bodies of gneiss and schist are more or less feldspathized and silicified, and garnet and sillimanite have been formed in some of them by interaction of granitic fluid and metasediment. Although potassium-aluminum silicates and silica predominate as introduced constituents, other elements probably introduced by fluids in appreciable amounts include sulfur, chlorine, fluorine, boron, iron, manganese, lead, zinc, barium, and water. Many or most of the introduced constituents seem to have had their source in more deeply depressed, hotter parts of the same metasedimentary rock sequence.

Paleozoic sedimentary rocks

The sedimentary rocks of Cambrian and Ordovician(?) age which encircle the Adirondack Mountains also appear as scattered patchy remnants in the Grenville Lowlands. By far the most abundant of these is the basal Potsdam sandstone which is Upper Cambrian(?) (see, for example, Buddington, 1934, p. 172-182, and Ruedemann, 1934). Quartzitic and breccia facies of the Potsdam are especially resistant, and many have been protected from erosion as filling in ancient streambeds, solution channels, and sinkholes in the Precambrian marble. In the Grenville Lowlands the Potsdam appears to have been deposited upon a slightly hilly and karsted surface having a relief as great as 200 feet. Irregular dissolution and karsting in the marbles seems to have continued after the Potsdam sandstone was deposited. Slump breccias of Potsdam sandstone in sinkholes in the marble are common, but it is difficult to separate some of these breccias from those that may have formed during the deposition of upper parts of the Potsdam.

In the Balmat-Edwards district, sinkholes and solution channelways as much as 800 or more feet deep are found in the marble where it is explored under cappings of Potsdam sandstone or breccia (pl. 1). One of the largest patches of Potsdam sandstone in the district is immediately northeast of Sylvia Lake, and drills have probed sinkholes and solution channelways in the marble 350 feet below the present surface. These are filled with Potsdam sand and conglomerate. The Precambrian zinc deposits just southeast of Sylvia Lake have undergone remarkable supergene alteration that extends 1,000 feet below the present surface (Brown, 1936b). The Potsdam rocks in these fillings are mostly brecciated, and include some slabs of sandstone that stand vertically. The associated marble is deeply stained and partly replaced, mainly by earthy hematite and supergene chlorite. The Potsdam relicts found as fill and talus breccias in sinkholes consist largely of highly quartzose conglomerate, breccia and sandstone, but include fragments of gneiss and altered granite. The mineralogy and texture of some of these rocks indicate their derivation from nearby Precambrian rocks. At many places in the Balmat-Edwards district, large masses of siliceous marble have been incorporated into the basal Potsdam with little reworking. This seems especially true in the central part of the district, southwest of Talcville, at (1.75 E. 9.30 N.) _/ It is very difficult to distinguish the marble in place from

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_/ Refers to the coordinate system on plate 1.

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the fragments of marble in the Potsdam. Brown has concluded (written communication) that most or all of the solution channels found in the marble of the Grenville series have been enlarged, or possibly formed in their entirety, after the basal beds of the Potsdam were laid down.

The basal part of the Potsdam may be Early or Middle Cambrian but it is lumped arbitrarily with the Upper Cambrian in this report. Except for very rare vein fillings that have been described by Buddington (1934, p. 202-209), the district contains no rocks known to be of Paleozoic age other than the Potsdam sandstone, and no rocks of Mesozoic or Tertiary age. Carbonate rocks of Ordovician age conformably overlie the Potsdam not far outside the district near Watertown and Theresa, N. Y. These Ordovician rocks were probably deposited continuously across the district and subsequently eroded away.

Pleistocene and Quaternary sediments

The Potsdam sandstone and the Precambrian rocks in the district are masked in many places by (a) lenses, sheets, and irregular masses of Pleistocene till; (b) glacio-fluviatile silts, sands and gravels; and (c) still-younger deposits of wind- and water-carried clastic material.

Pleistocene glacial till and water-laid deposits of sand and gravel derived from glacial debris, associated with rounded, fluted, and striated bedrock, are prominent throughout the district. Erratic boulders and rudely stratified sands, silts, and gravels appear along the flanks of and between the knolls of Precambrian rocks. Boulder clay occurs sparingly in this district and throughout the northwestern Adirondacks. Dune sands are widespread and conspicuous in the central part of the district near Fullerville and north of it along the west branch of the Oswegatchie River. Beneath the dunes are stratified sands, silts, and fine gravel, presumably deposited in lakes and ponds dammed behind the melting glaciers. Several drill holes just east of the Ontario mine have penetrated as much as 97 feet of sand, silt, and gravel resting on Precambrian bedrock (Engel, 1948). Extensive flats near Talcville and Edwards are underlain by sand, mud, and boulders to known depths of 100 feet. These sediments appear to have been deposited by streams ancestral to the Oswegatchie River and its tributaries during and after the melting of the Pleistocene glaciers.

Adirondack Uplands

Central massif

The Central Adirondack massif consists chiefly of anorthosite, syenite, and granite, but includes small bodies of anorthositic gabbro and gabbro, which are in part younger than the anorthosite (Balk, 1931, and Buddington, 1939, p. 19-52). The anorthositic rocks lie in a belt about 125 miles long whose apex strikes east-northeast, somewhat divergent from the dominant regional grain (Buddington, 1939, p. 238, fig. 22). The anorthosite has a foliation that appears to be in part primary. Its margins have, in most places, a strong secondary foliation, and are injected by gabbro, syenite, and granite. Some scattered remnants of the Grenville series in contact with anorthosite are converted into skarns and pyrometasomatic rocks (Kemp, 1921; Buddington, 1939, p. 39-46). Genetically associated with the anorthosite in several areas are economically important bodies of ilmenite-magnetite, which are mined at Tahawus, Essex County, New York (Stephenson, 1945).

Peripheral granite-syenite complex

Surrounding this elevated core area, especially on the north, west, and south, is a dominantly igneous(?) complex of syenites and younger granites, which separates the core of the massif from the Lowlands. These rocks have been studied in detail by Buddington (1939, 1948) and by Buddington and Leonard (in press). They seem to form a comagmatic series in which decreasing age is correlated with increasing silica content. The oldest include shonkinite and pyroxene syenite, which were followed by quartz syenite, ferrohastingsite granite, and alaskite. All these rocks appear to have been emplaced after the anorthosite and gabbros, largely as crudely accordant sheets into the Grenville series. Profound folding accompanied and followed their emplacement. Buddington regards the progressive changes in composition of most syenites as due to gravity stratification and the lithologic banding in the syenites seems best explained in this way. The granite series is believed by Buddington and Leonard (in press) to have been derived from a magma that crystallized largely as a mixture of hornblende and microperthite, and that gave rise to alaskite, microcline granite, and albite-oligoclase granite as extreme differentiates. The granites, along with associated alkali-silicate fluids, have produced most of the contact metamorphic changes in the enveloping metasedimentary rocks. Minor contact metamorphic changes seem to have resulted from the intrusion of the syenites and only local contact metamorphism appears to have been produced by the emplacement of the anorthosite. The granitic fluids soaked into scattered remnants of the metasedimentary rocks within the Adirondack core and have reacted extensively with them. Large bodies of hybrid rock have thus been formed--migmatite and metamorphic granite in the gneisses, and the magnetite-bearing skarns in the marbles. Certain of these skarns, together with important bodies of magnetite in granite gneiss and migmatite, make up the well known group of iron ore deposits in St. Lawrence County, and in the Clinton (Chateaugay), Mineville, and Port Henry districts of the Adirondacks.

Zonal relations of mineral deposits and metamorphism in the Adirondacks

The crudely semi-circular zonal pattern of the igneous-looking rock in the Adirondacks, and the enveloping sheaths of dominantly metasedimentary rocks, diagrammed in figure 2, is reflected in the distribution of the mineral deposits and of characteristic rock textures. The implications of this zonal pattern, and especially its bearing on the evolution of the Adirondack massif, were first discussed by Buddington (1939, p. 251-303).

The distribution of the more valuable known mineral deposits is particularly noteworthy. The anorthosite bodies of the Adirondack core contain important segregations of ilmenite-magnetite, or titaniferous iron ore. The known deposits of this type are in fact almost entirely confined to the central massif, which contains almost all of the anorthosite.

On the other hand, the non-titaniferous magnetite deposits--or, more accurately, those with less titanium--are associated with the granites. Some appear to have been formed by segregation from the granitic magmas; others--the skarn deposits--by complex replacement of marble in close association with the granites.

Outside the dominantly igneous massif, in the encircling metasediments, are the pyrite-sphalerite deposits, and the concentrations of talc and serpentine, especially those formed by alteration of tremolite and anthophyllite in and near the Balmat-Edwards district. Buddington has noted that the pattern of distribution of all these ore deposits is consistent with theory and is found in certain other districts: that is, their distribution reflects a differentiation in fluids moving outward from a hot central area into a cooler perimeter. There is a gradation from the older, higher-temperature deposits of the Grenville Lowlands; and this gradation reflects a progressive decrease in temperature of formation from the anorthosite, gabbro, syenite, and granite of the core to the volatile-rich granite and pegmatite of the Lowlands. Decrease of temperature was probably correlated with decrease of confining pressure for, as Buddington has suggested (1952), the anorthosite of the core must have solidified at greater depths in the earth than the rocks of the syenite-granite complex around it, and the scattered bodies of granitic rock in the Grenville Lowlands were doubtless emplaced at still less depth, as well as at lower temperature.

The zonal concept may be extended to the metamorphic mineral facies formed during reconstitution of any kind of rock in the Adirondacks. In rocks of a given composition, the grade of metamorphism rises toward the Adirondack core. In most rocks, for example, the constituent grains become larger and change progressively in properties, especially composition, as the central massif is approached. These metamorphic changes in the major paragneiss and associated amphibolites have been discussed in detail elsewhere (Engel and Engel, 1958a, 1958b, 1960). One highly significant element in the zonal pattern is that the rock layers in the Grenville series tend to curve around the central Adirondack massif. Initial differences in composition of strata are semi-accordant in pattern, therefore, with differences due to igneous intrusion and metamorphism. The present zonal arrangement of rocks and of certain mineral deposits thus has a composite origin; it is due partly to original composition and partly to intrusion and metamorphism, and the relative importance of these factors is often hard to evaluate.

Significant changes in the character of the metamorphism can be observed as one crosses the Balmat-Edwards district from southwest to northeast. These are discussed in considerable detail in succeeding sections, as are some possible parallels between the evolution of the talc deposits and the formation of skarnlike bodies in marble enclosed in the Adirondack core.

The evolution of many of the above features and of some prominent textures in the rocks of the Adirondacks is closely related to the dynamothermal metamorphism that accompanied the emplacement of the granites. Regional stresses and the consequent shearing of many layers must have facilitated chemical reactions and formed channels along which substances might migrate. In general, the anorthositic core of the Adirondacks shows the least deformation and alteration, although very little of the anorthosite remains completely undeformed or unaltered. There has been a great, though irregular, increase in the degree of mashing and refoliation outward from the anorthosites to the margin of the central massif, where there is a wide zone of pervasive shearing and displacement. Mylonites and mortar gneisses are common there, along with variously recrystallized gneisses, schists, and marbles.

Along the middle part of the Grenville Lowlands the deformation was much less thorough and profound, though all the rocks have been completely recrystallized and reconstituted to the amphibolite facies as defined by Eskola (1920). Temperatures of metamorphism were as high as 500° C (Engel and Engel 1958a).

Rocks of the district

Introduction

The distribution of the rocks in the Balmat-Edwards district is shown on plate 1. The outlines of exposures of bedrock have been

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Plate 1. Geologic outcrop map of the Balmat-Edwards district, northwestern Adirondacks, New York. (in pocket)

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shown, as well as those parts of lithologic boundaries that are accurately located and the parts that are inferred. References to specific rock exposures or localities make use of the coordinates on plate 1. Thus, the cross-roads at Balmat corners is at 0.5 S, 9.45 W. and the small exposure of rock at 6.4 N, ~~6.55~~ 6.55 W. is magnetite-grunerite skarn. The inferred age relations and present thickness of the rock units are summarized in table 2.

Table 2. Inferred stratigraphic relations between the metasedimentary rock units of the Grenville series in the Balmat-Edwards district, New York.

Table 2. Inferred stratigraphic relations between the metasedimentary rock units of the Grenville series in the Balmat-Edwards district, N. Y.

	Rock unit	LITHOLOGIC FEATURES			
		Description of unit	Most abundant mineral constituents given in volume percent where feasible	Inferred pre-metamorphic composition in volume percent	Present thickness in feet
	3/ Marble overlying median gneiss Fowler-Fullerville	Gray medium- to coarse-grained massive to distinctly foliated dolomite, with lenses of foliated quartzitic rock, serpentinized diopside, pegmatite, gneiss, and schist.	Dolomite 65, quartz 10 serpentinized diopside 10, others 15.	Dolomite 80 Quartz 15 Others 5	1000
	Median gneiss (Zone 16)	Migmatite consisting of micaceous and amphibolitic quartz-feldspar gneisses and schists permeated and injected by pegmatite and granite	2/ Quartz, microcline, biotite, muscovite, and plagioclase	Argillaceous sandstone with scattered thin lenses of tuff and dolomite	900 4/
Sylvia Lake - Cedar Lake marble belt	Rusty Marble (Zone 15)	Buff-weathering, medium-grained marble containing silicates and pyrite, largely disseminated, but partly in distinct layers and lenses.	2/ Calcite, potash feldspar, diopside, tremolite, quartz, plagioclase, phlogopite, and pyrite.	Argillaceous dolomite Dolomite 65 Clay mineral20 Quartz 12	250
Sylvia Lake - Cedar Lake marble belt	Footwall marble (Zone 14)	Medium- to coarse-grained marble, partly dolomitic, partly siliceous, calcite containing quartz and diopside layers, lenses, and small clots.	2/ Slightly serpentinous and diopsidic dolomite, and partly siliceous calcite	Dolomite 76 Quartz 18 Others 6	350
Sylvia Lake - Cedar Lake marble belt	Talc belt (Zone 13)	Tremolite schist, largely coarse grained, containing quartz, calcite and talc	2/ Tremolite, talc, calcite, serpentine, anthophyllite, and quartz	Dolomite 84 Quartz 15 Others 1	450
Sylvia Lake - Cedar Lake marble belt	Dolomite 1/ (Zone 12)	White to pale-buff or gray dolomite mostly coarse-grained with local clots and lenses of quartz, serpentinized diopside and other minerals	Dolomite 80; serpentine, diopside, calcite and quartz.	Dolomite 90 Quartz 3.7 Others 6.3	250
Sylvia Lake - Cedar Lake marble belt	Silicated marble (Zones 4 to 11 inclusive in the Sylvia Lake area)	Coarsely crystalline dolomite (Zones 5, 7, 9), intercalated with silicated dolomite Zones 6, 8, 11, of quartz and serpentinized diopside. Zone 10 includes a layer of anhydrite and gypsum	Dolomite 43, and quartz, diopside, calcite, serpentine and tremolite	Dolomite 73 Quartz 22 Others 5	1600
Sylvia Lake - Cedar Lake marble belt	Dolomite (Zone 3)	White to pale buff or gray dolomite mostly coarse-grained with local clots and lenses of serpentinized diopside, quartz, and other minerals	Dolomite 90; calcite, diopside, and serpentine.	Dolomite 90.0 Calcite 2.5 Quartz 1.0 Others 1.5	250
Sylvia Lake - Cedar Lake marble belt	Pyritic schist (Zone 2)	Pyritic migmatite schist, which weathers to mostly yellowish-brown but locally irridescent brownish black.	2/ Potash feldspar, quartz muscovite, biotite, chlorite, and pyrite.	Argillaceous sandstone, possibly pyritic with thin dolomitic lenses	250
Sylvia Lake - Cedar Lake marble belt	Dolomite (Zone 1)	White to pale buff or gray dolomite mostly coarse-grained with local clots and lenses of serpentinized diopside, quartz, and other minerals	Dolomite 90; calcite, diopside, and serpentine.	Dolomite 95.0 Calcite 2.5 Quartz 1.0 Others 1.5	250
	Quartz-biotite-oligoclase gneiss	Migmatite veined and partly replaced by pegmatite and inequigranular granitic gneiss	Least migmatitic layers consisting of quartz-oligoclase-biotite gneiss, injected along dominant foliation by quartz and perthite.	Graywacke or sodic shale	1000

1/ Forms the hanging wall of talc belts in the marble, which is believed to be largely overturned.
2/ Approximately in order of decreasing abundance in the "average sample".
3/ as shown in the text, this may be a highly sheared segment of the Sylvia Lake-Cedar Lake marble belt.
4/ As shown in the text, the gneiss may be duplicated.

The total thickness of the meta~~sedi~~sedimentary rocks exposed in the district is estimated at about 4,500 feet before metamorphism. The two major types of metasedimentary rocks are: (1) quartz-biotite-oligoclase gneiss, and (2) dolomitic marble, in part highly silicated. The gneiss is regarded as the oldest rock in the district, and as having been originally a clastic sediment which was irregularly but extensively permeated by alkali-silicate fluids and completely reconstituted. Only its upper part, perhaps 1,000 feet thick at most, occurs in the area mapped (pl. 1). The marble, which has clastic interlayers designated as the pyritic schist and the median gneiss, overlies the quartz-biotite-oligoclase gneiss and is essentially conformable with it. It is believed to represent quartzose dolomite sediments. Its maximum thickness, including interlayers of gneiss and schist, is estimated to be about 3,500 feet. Within the marble there are layers of quartz schist, often described as quartz-mesh rock (Martin, 1916), as well as zinc-talc deposits which appear to have resulted from marked alteration of certain of the original sedimentary beds. All paragneisses and paraschists are partly altered to granitic-looking rock, but the principal igneous-looking rocks in the district are the inequigranular granitic gneisses referred to as the Hermon complex, which extensively inject lit-par-lit and permeate the quartz-biotite-oligoclase and median gneisses, and locally intrude or replace most other rock in the district.

Amphibolite probably derived from gabbroic intrusives older than the granite occurs northeast of Balmat in the southwest part of the district, and forms thin lenses and layers in all the other rocks. Thin layers of amphibolite in the quartz-biotite-oligoclase gneiss are tentatively regarded as having been derived from sedimentary beds, perhaps tuffs.

Marble

Distribution and outcrop pattern

The marble exposed in the district forms lithologically complex masses, irregular in plan, which generally trend northeastward in accord with the regional strike of all the metasedimentary rocks. The general form of the main marble mass appears to be that of a crumpled and refolded syncline that lacks one flank--the southeastern. A bulbous mass of cross folds appears in the marble on the southwest end of the syncline, which is almost enclosed by quartz-biotite gneiss in the vicinity of Sylvia Lake (2.0 N. - 1.3 S., 11.0 - 13.5 W.). The cross folds plunge northward and produce a major recurvature in the axis of the refolded syncline. Northeastward from the lake the marble may be traced by scattered outcrops, drill holes, and mine openings, to the place where the West Branch of the Oswegatchie River is crossed by New York State Highway 58 (6.0 N., 3.4 W.). For about a mile northeastward along the projected strike, the marble is covered by alluvial and glacio-fluviatile sands and silts of the West Branch sand hills. There, exploratory diamond-drill holes in the sand-covered area just east of the river (7.1 N., 1.4 W.; 6.9 N., 2.35 W.; and 7.45 N., 2.7 W.) cored marble at depths of 34, 47 and 95 feet beneath the alluvium (Engel, 1948). Another major refold in the marble in this area is inferred from this drilling and from exposures on the surface as shown in plate 1.

What is probably the same body of marble is exposed on the wooded slopes of the West Branch of the Oswegatchie at 8.0 - 9.0 N., 0 - 1.0 W., and thence northeastward between the median gneiss on the south and the quartz biotite gneiss and granite on the north as far as Cedar Lake (21.0 N., 11.0 E.), near the northeast boundary of the district. At the east end of Cedar Lake the marble curves abruptly southward in still another large northwest plunging refold (cross fold), and then presumably is refolded northeastward into the Russell quadrangle.__/

footnote on follow

__/. Mapped by Buddington and Leonard (in press) for the U. S. Geological Survey in conjunction with studies of the magnetite deposits of the northwest Adirondacks.

A second large elongate mass of marble, or possibly the same one repeated in a small syncline, lies southeast of the one described above. From the area near the center of the district (8.0 N., 0 ~~E~~) northeastward beyond the Edwards zinc mine, these marble belts continue to be roughly parallel and are separated only by the narrow, sinuous belt of feldspathic gneiss that Brown has designated the median gneiss (1936a, p. 238 and fig. 1). As the marble is traced southwest from 8.0 N., 1.0 E., the infrequent outcrops suggest that the more southerly marble belt is cut into several parts by masses of granite, amphibolite, and quartz-biotite gneiss. In the West Branch sand hills, however, outcrops are very scarce and no subsurface exploration of this marble belt has been undertaken, so that the mapping of the marble in this area involves much conjecture.

In the general area of 1.75 N., 7.0 W., scattered outcrops of marble and the irregular patterns of the intervening lowlands suggest that marble and amphibolite are irregularly intertongued or infolded as shown in plate 1. Other small bodies of marble are infolded or enclosed in the complex of quartz-biotite gneiss and granite in the vicinity of 14 N., 4 E., and 16 N., 6 E. Small lenses and layers of marble, some of them only a few feet wide but from ten feet to several hundred feet long, are also enclosed in various bodies of gneiss, schist, and granite. Typical examples include the interlayers or infolded syncline in the median gneiss where it is abruptly curved north of Edwards (18.3 N., 11.25-11.75 E.), the marble interlayers in the contact zone of the granite mass in the area 7.0 - 9.3 N., 2.0 - 5.0 E., and the marble infolded or interlayered in the quartz-biotite gneiss and granite southwest, west and northwest of Cedar Lake and on the east and southeast shore of Trout Lake (1,000 feet north of the northeast corner of plate 1).

The large elongate area of marble that extends across the district from the vicinity of Sylvia Lake and Kellogg Corners northeastward beyond the strongly curved outcrops northeast of Cedar Lake contains all of the known commercial deposits of talc and of zinc and lead except possibly the zinc deposit developed in the Edwards mine. This area will be referred to as the Sylvia Lake-Cedar Lake marble belt, and the large subparallel marble to the southeast as the Fullerville-Edwards marble belt. The Sylvia Lake-Cedar Lake marble belt contains about 13 mines and 40 exploratory shafts, with lateral drifts 28 miles in aggregate length and at least half a million feet of drill holes driven at large angles to the bedding. Nevertheless, there are large areas of the marble that are essentially unexplored. The West Branch sand hills and other large tracts of Quaternary deposits cover parts of both the Sylvia Lake-Cedar Lake and Fullerville-Edwards marble belts, and may well conceal important mineral deposits. The largest areas of alluvial cover along the Sylvia Lake-Cedar Lake belt include: (a) the flood plain of the Oswegatchie River and related lowlands immediately northeast, north, west, and southeast of the Hyatt zinc mine (11.6 N., 2.4 E.); (b) a continuation of this flood plain between the Wintergreen Hill talc mine (11.9 N., 2.35 E.) and the International No. 2 1/2 talc mine (12.5 N., 3.8 E.); (c) an area immediately to the northeast, in the glacially scoured lowlands between the 2 1/2 mine and the International No. 4 mine (13.75 N., 5.25 E.); and (d) an area near the tailings pond of the Edwards zinc mine (17.25 - 18.50 N., 9.75 - 11.0 E.).

The hope of finding deposits of talc and zinc beneath these covered areas has stimulated exploration by drifts and by diamond drilling, and these data have been used in drawing the tentative boundaries of the marble under major areas of alluvium (pl. 1). Few exploratory data are available, however, to aid in plotting the distribution of the Fullerville-Edwards marble belt, or of other smaller bodies of marble where they are obscured by alluvium. The boundaries shown are based largely upon the outcrops, considered in connection with the fact that the marble tends to be less resistant to glacial scour, subaerial weathering, and erosion than the gneisses and schists and is thus more apt to underlie large areas of alluvium. There is definite proof that most such areas in the district are underlain by marble, but also that some others are underlain by gneiss. Most of the depressions cut in gneiss are elongate parallel to the lithologic strike; the few that cross the strike are obviously related to stream courses, faults, and joint patterns.

Evolution of lithologic features and composition

The compositions of individual zones of marble and of the resulting composite marble belt, shown in table 3, are deduced from study of

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Table 3. Approximate chemical and mineralogical composition of the marble between Sylvia Lake and Cedar Lake and its inferred composition prior to metamorphic changes.

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the Sylvia Lake-Cedar Lake marble belt around Sylvia Lake and just east of Fowler village. In this part of the marble belt the bedding is least obscured by metamorphic changes. Zone 2, the pyritic schist interlayer (table 2 and plate 1), is omitted from table 3. Pyritic schist must represent an argillaceous sandstone or siltstone, but the question whether the pyrite and carbon in it are of sedimentary or secondary origin has not been settled. This question is discussed on pages ~~50-60~~ 79-81.

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Table 3. Approximate chemical and mineralogical composition of the marble between Sylvia Lake and Cedar Lake and its inferred composition prior to metamorphic changes.

Stratigraphic Unit	Thickness in feet	Percent total thickness	Present very approximate mineral composition (by volume): K-Feldspar	Plagioclase	Diopside	Tremolite	Anthophyllite	Phlogopite	Serpentine	Talc	Quartz	Dolomite	Calcite	Pyrite	Anhydrite	Other
Rusty marble zone 15	150	8.0	20		18	2		1	tr	tr	5	tr	52	1		1
Footwall marble zone 14	220	11.7			12	3		tr	3	tr	28	tr	53	0.2		.8
Talc belt zone 13	230	12.2				49	7	tr	10	18	3		11	0.5	tr	1.5
Dolomite zone 12	210	11.2		0.5	5				7	tr	3	79	4	0.5	tr	1.0
Silicated marble zones 4 to 11 inc.	730	38.8	0.5		12	2		tr	3	tr	28	43	10	.1	.1	1.3
Dolomites zones 1 and 3	340	18.1		0.5	2				1.5	tr	1.5	89	5.0	.2		.3
Total	1880	100														
Weighted average			1.8	.1	8.4	8.1	.8	.1	3.8	2.2	15.4	41.3	16.8	.3	tr	.9

Stratigraphic Unit	Present approximate gross composition in weight precent: SiO_2	Al_2O_3	$FeO+Fe_2O_3$	MgO	CaO	K_2O	Na_2O	CO_2	H_2O	SO_3	Other	Inferred gross premetamorphic mineral composition marble, weight percent: Quartz	Dolomite	Calcite	Other
Rusty marble zone 15	36.3	4.4		6.0	28.2	2.1	1.2	20.4	.3	.6	.5	11.3	60.0	4.0	*24.7
Footwall marble zone 14	37.5	0.4		4.3	33.5	0.3	0.1	23.7	.2			18.6	76.0	2.0	3.4
Talc belt zone 13	51.1	0.4	.2	26.3	13.2			5.2	3.5	.1		15.0	84.0	1.0	
Dolomite zone 12	8.7	tr		21.1	28.4	tr		37.8	1.0	tr	3.0	3.7	90.0	1.0	5.3
Silicated marble zones 4 to 11 inc.	33.6		1.8	16.5	21.9	tr		25.03	0.42	.7	.05	21.6	73.0	2.0	3.4
Dolomites zones 1 and 3	3.2			21.6	29.3			45.6			.30	1.2	90.0	2.5	6.3
Total															
Weighted average	28.1			17.0	24.8			27.				15.0	78.0	2.0	5.0

* Chiefly clay mineral.

Figures are given in tenths of percent merely to facilitate calculations.

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Although the composite section represented in table 3 is thought to contain a minimum of material added during metamorphism, at least a third, or perhaps half, of the silica now present probably has been introduced, and probably one-fourth of the carbon dioxide was expelled in the formation of metamorphic silicates, principally diopside, tremolite, serpentine, and talc. The minimum changes in composition believed to have occurred in the marble can be found by comparing the present composition of the least altered marble with that believed to represent the same rock before it was metamorphosed. The Sylvia Lake-Cedar Lake marble belt appears to have been derived from moderately siliceous dolomite, containing about 17 percent MgO and 25 percent CaO, almost solely in the mineral dolomite, and 15 percent quartz (table 3). Iron, alkalis, and alumina probably made up less than 5 percent of the parent rock and possibly less than 3 percent. The only rocks in which either ferruginous or argillaceous material could originally have exceeded 3 percent are the pyritic schist, the median gneiss, and the so-called rusty marble (Zone 15, pl. 1, and tables 2 and 3). Most of the iron is in pyrite. Readily detectable hydrogen sulfide and members of the paraffin series occur in most zones, but it is uncertain whether these constituents are sedimentary or of secondary, metamorphic or magmatic origin.

Other noteworthy constituents of the Sylvia Lake-Cedar Lake marble belt are anhydrite and gypsum, which are associated with a little halite. The upper silicated marble (Zone 10) contains a thick and remarkably persistent bed of anhydrite. This anhydrite is rarely exposed at the surface, being found mainly in mine workings and drill cores (p. 104-105). Other layers of anhydrite may occur in less explored parts of the marble. The anhydrite appears to be of sedimentary origin, but it may possibly have been formed by remarkably selective replacement of a sedimentary bed. Some anhydrite, together with gypsum derived from it, also occurs in the commercial deposits of zinc, lead, and talc; this may either have been concentrated from nearby sedimentary interbeds or have been introduced from outside the marble during the formation of these mineral deposits.

The Sylvia Lake-Edwards marble belt was a little more siliceous before metamorphism than the Fullerville-Edwards marble belt. In both belts, silica appears to have been the only constituent abundantly disseminated in the dolomite and interlayered with it. The ratio of silica to dolomite appears to have varied both along and across the strike. There are few obvious, marked changes, however, in the character of individual beds along the strike or down dip ~~(see page —)~~; and if the entire sequence is considered, most of the variations along the strike tend to cancel out: an increase in the proportion of silica-carbonate in one zone is apt to be offset by a decrease in another zone or zones stratigraphically above or below. There are marked variations however, across the strike. Zones 4 to 11, for example, (pl. 1, tables 2 and 3) are estimated to have contained at least 20 percent silica, as quartz, before metamorphism, whereas Zones 1, 3, and 12 must have been almost pure dolomites. There is much greater uncertainity regarding the prem-metamorphic composition of Zone 15 (rusty marble), Zone 14 (siliceous calcitic marble), and Zone 13 (talc belt). These are all close to the migmatitic median gneiss, and were obviously much altered during metamorphism. Much silica and alkali have apparently been added to all three zones, especially to Zone 15 (rusty marble), while magnesia has been extracted, especially from Zone 14, and introduced into the now overlying, but probably older, talc belt. Zones 12, 13, and 15 are thought to have been somewhat less siliceous originally than the weighted average of Zones 4 to 11, inclusive (table 2).

In the zones of siliceous marble, much of the sedimentary silica reacted during metamorphism with associated dolomite to produce diopside, together with some tremolite and small amounts of other silicates. During these changes in mineral composition, silica was introduced abundantly over wide areas in the talc belts, in the units just northwest of the talc, and more sporadically elsewhere. This silica, like that already in the beds, contributed in part to the formation of metamorphic silicates, but much of it crystallized as quartz. Examples of carbonate-rich marble largely replaced by introduced silica are the quartz schist and quartz-mesh rocks. In addition to these silica-rich rock types of readily inferred replacement origin, many layers and nodular bodies of silica have very ragged boundaries, and tongues and rounded masses of quartz embay the enclosing marble. Many beds of siliceous marble grade into bodies of quartzose rock in which the layering is blurred or contorted (pls. 7C, 7D). These bodies are roughly lenticular and generally accordant with the stratification; in detail, however, they crosscut relict bedding and clearly replace pre-existing carbonate beds. Whether this silica was all introduced from a source outside the marble, or in part merely redistributed, is unknown. The calcitic footwall marble also contains innumerable small clusters and knots of quartz; most of them are alined in chain-like patterns, or as beads on curving strings (pls. 6B, 6D), but some occur almost at random (pl. 6A). Complete gradations in form and constitution appear to exist between quartz fragments formed largely by tectonic dispersal of beds, and quartz introduced into the marble and clustered around bed fragments and other small particles.

Alumina and alkalis were obviously introduced with some of the silica, especially adjacent to migmatitic interlayers such as the median gneiss, pyritic schist, and quartz-biotite gneiss. At these places and more sporadically throughout the marble, pegmatitic and granitic masses have formed, as well as disseminated grains of potash feldspar and phlogopite. Large bodies of granitic gneiss and perthitic pegmatites are abundant in the marble (a) north of Edwards, (b) north and northwest of Newton Hill in the vicinity of Talcville, (c) near the Hyatt mine, and (d) adjacent to the median gneiss throughout its extent. There are small pegmatites in and near the talc belts (a) in the Gouverneur Talc Company's mine north of Balmat Corners, (b) in the International Talc Company's 2 1/2 mine at Talcville, and (c) in both the Balmat and Edwards zinc mines of the St. Joseph Lead Company. In many places the rocks mapped as quartz schist or quartz-mesh rock contain as much as 25 percent of potash feldspar and in some places they also contain plagioclase. Most or all of the quartz in these rocks is believed to have replaced carbonates and marble.

The assumption that the early--or premetamorphic--carbonate was largely, if not solely, dolomite is mainly based on the mutual relations of carbonates, quartz, and the metamorphic silicates. Dolomite is still the principal carbonate in the marble, although calcite is locally more abundant, especially along and just southeast of the commercial talc zones in the central and southwest parts of the district. In most places where dolomite is associated with calcite, the dolomite is embayed or crosscut by calcite and silicates (pl. 5D). The calcite has formed irregular tongues extending along the dolomite layers (pl. 5B), or encloses clots and angular fragments of dolomite, quartz, and silicates (pls. 5A, 5C) or forms veinlets that cut them.

No outcrops are known, in which calcite marble is clearly older than dolomite, and there are hundreds of exposures throughout the district in which dolomite is the older carbonate. The relations of calcite and dolomite are more fully discussed later (see page 90). Some of the calcite has been left as a residue where magnesia originally in dolomite has been used in forming metamorphic silicates. The magnesium-bearing silicate that is commonly embedded in dolomite unaccompanied by calcite, is diopside or, rarely, forsterite. Crystals of younger silicates such as tremolite, anthophyllite, serpentine and talc, are closely associated with calcite, which commonly forms halos that separate them from obviously older enveloping dolomite (pls. 5A, 5C); and clots or layers of serpentine or talc commonly contain calcite, which separates them from associated dolomite. These relations are expectable unless appreciable quantities of magnesia are being added; for in all of these reactions except the formation of pure diopside from dolomite and silica, the ratio of MgO to CaO in the silicate is higher than it is in dolomite. This is shown in table 23, and by the equation below.

The reaction of dolomite with silica to form diopside may be expressed as follows:

$$\underset{\text{dolomite}}{CaMg(CO_3)_2} + \underset{\text{quartz}}{2\ SiO_2} \rightleftharpoons \underset{\text{diopside}}{CaMgSi_2O_6} + \underset{\text{carbon dioxide}}{2\ CO_2}$$

Since the ratio of CaO and MgO in diopside is approximately the same as in dolomite, neither CaO nor MgO forms as a byproduct of the reaction to re-combine with any released CO_2. A notable feature of the reaction is that if either CaO or MgO, or both, are being introduced as diopside forms, other minerals may be formed, unless the ratio of CaO and MgO is just that required for diopside. Silica introduced contemporaneously, on the other hand, forms no new phases other than quartz or diopside, but it decreases the loss in volume.

Tremolite can form through the reaction of wet sedimentary dolomite and quartz as follows:

$$5CaMg(CO_{\cancel{2}3})_2 + 8SiO_2 + H_2O \rightleftharpoons Ca_2Mg_5H_2(SiO_3)_8 + 3CaCO_3 + 7CO_2$$

dolomite silica water tremolite calcite carbon dioxide

At many places where tremolite occurs in the marble, calcite is about as abundant as it would be if pure dolomite had reacted with quartz to form the tremolite. The MgO in the tremolite was probably derived from the dolomite, with the excess CaO remaining in the form of calcite. In most instances, some silica appears to have been added. Much of the tremolite, for example, occurs in veinlike or other discordant masses of quartz, in pegmatite, or in highly siliceous layers such as the talc belts, which merge along the strike in both directions into less siliceous, carbonate-rich layers. In the talc belts, however, the volume of calcite is disproportionately small, and this suggests that much of it has been removed from the zone.

The serpentine that occurs in the marble replaces most other minerals except talc, but especially diopside, tremolite, and dolomite. As serpentine contains no CaO, its formation directly from siliceous dolomite, or from tremolite or diopside, releases large quantities of CO_2 and $CaCO_3$.

Possible volume change during metamorphism

If calcium- and magnesium-bearing silicates were formed by reaction of sedimentary constituents without addition of silica or other substances, decrease in volume would result. The amount of decrease would vary, of course, depending on whether the porosity changed, and on whether much excess carbon dioxide were removed. If only excess carbon dioxide is removed (and water in the case of diopside), the loss in volume when diopside and tremolite are formed from sedimentary quartz and dolomite, is roughly 50 percent. Similar volume decreases occur when talc or serpentine are formed from wet siliceous dolomite. Water and silica were almost certainly added, however, while the two last-named silicates were being formed and compensated, at least in part, for the shrinkage due to loss of carbon dioxide. Much smaller amounts of alkalis, metallic sulfides, halogens (especially fluorine), and other constituents have also been added. In many serpentinous and talcose parts of the talc belts, the absence of abundant calcite seems to indicate that as serpentine and the mineral talc formed in the talc belts, calcite also was removed from this zone, and probably from the marble as a whole.

The question arises, what were the overall volume changes in the marble? A partial answer is given by possible weighted average compositions of marbles posted in table 3. These field data indicate that neither MgO nor CaO has been added to or subtracted from the marble in appreciable quantity, except in the talc belts, and consequently that the ratio of Mgo to CaO is roughly the same now as it was before metamorphism.

Many authors have assumed that a great deal of magnesia was introduced into terranes very similar to that here discussed, but no support of this assumption has been found in the Balmat-Edwards district. There is no obvious external source of MgO or CaO, magmatic or otherwise. The belief that appreciable magnesia is derived from the invading granitic rocks does not seem justified either by observed geologic relations or by theory. There probably has been some loss of calcite, especially from the talc belts, but not in large amount. The amount of silica added to the weighted average composition of the marble (table 3) is estimated from detailed field studies to be 13.5 percent by weight. This is almost half the total now present. From these data the following calculations have been made (all data are from table 3):

Present mass of marble (Wp) = 1880 _/ ÷ 2.90 (Density)

Original mass of marble (Wo) = Wp - added silica + lost carbon dioxide

$$Wo = Wp - 0.135\,Wp + 0.383\,Wo - 0.27\,Wp$$
$$0.617Wo = 0.595\,Wp$$
$$Wo = 0.96\,Wp$$
$$Wo = 96\,\frac{(1880 \div 2.90)}{283}$$

Vo (volume original marble) = 0.983 of 1880 feet

_/ Thickness of marble in feet.

Little or no volume change may be involved if the previously stated assumptions are valid. Probably the most questionable of these assumptions is that as much as 40 or 50 percent of the silica now in the marble (apart from that in the mapped intercalated granitic masses) was introduced. J. S. Brown (oral communication) believes it likely that no more than 10 percent of the silica was introduced. If he is right, the total volume of marble has decreased by about 10 percent, largely through decarbonation and the loss of a little CaO during metamorphism.

Lithologic units in the Sylvia Lake-Cedar Lake marble belt.

Introduction

Excellent exposures on the surface and in five mines, together with many thousands of feet of drilling, have made possible an unusually complete understanding of the stratigraphy of the marble in the Sylvia Lake area. A detailed section of this marble has been worked out (Brown and Engel, 1956) and is reproduced in table 2. From the following discussion of the stratigraphy of this marble, and inspection of plate 1, it will become apparent that many of the same units probably extend northeastward from Sylvia Lake to the vicinity of Cedar Lake. It also seems probable that most other belts of marble, such as that extending from Fullerville to Edwards, are segments of the Sylvia Lake-Cedar Lake marble belt, much complicated and altered by faulting, folding and metasomatism.

Subdivisions of the marble are distinguished by brief descriptive names and zone numbers. The numbers are in order of age, the oldest zone being called Zone 1; but it is to be remembered that the beds are mostly overturned, so that a given bed is commonly "overlain" by an older bed. Because of the difficulty of correlating unfossiliferous rocks that are highly metamorphosed and greatly deformed, the zone that an exposure of marble represents is sometimes in doubt, and wherever an outcrop cannot be traced directly into the type section around Sylvia Lake, the zone number designating it on the map is followed by a query.

Dolomite, Zones 1 and 3

Distribution, form, and thickness

Zones 1 and 3, which consist largely of dolomite, can generally be distinguished with certainty. They are best exposed in the areas west and northwest of Sylvia Lake, where the pyritic schist (Zone 2) is underlain and overlain by pure dolomite marble (pl. 1). In this area Zone 1 is apparently about 125 feet thick, although northwest of Sylvia Lake (2.0 N., 15.0 W.) its thickness is presumably reduced by deformation to less than 50 feet. Although the contact of the dolomite of Zone 1 with the quartz-biotite-oligoclase gneiss is nowhere exposed, its location almost everywhere in the district can be estimated closely (pl. 1).

The zone of nearly pure dolomite on the eastern and southeastern side of the pyritic schist is mapped as Zone 3. Its thickness west of Sylvia Lake varies from 100 to about 350 feet. These wide variations in thickness, like those in Zone 1, are interpreted as being due to deformation of the marble in the large folds around Sylvia Lake. The dolomite exposed about 1,400 feet east of Fowler village (6.0 - 6.4 N., 6.4 - 7.0 W.) is correlated with Zone 3, for it clearly overlies the highly distinctive silicated marble of Zone 4 (although the latter is the younger). Blocks of cummingtonite-garnet-magnetite skarn, commonly found at and near the bottom of the pyritic schist, appear along the creek bank just northwest of the dolomite. If these blocks are part of an exposure, the dolomite of Zone 2 at this locality is less than 50 feet thick, for quartz-biotite gneiss is exposed about 200 feet north of the rock assigned to Zone 1, and almost along the line of strike from the blocks of skarn.

Northeast of the above-mentioned area, there are no exposures of rocks that certainly belong to either Zone 1 or Zone 3 nearer than 8.4 N., 1.4 W. This is just north of the West Branch of the Oswegatchie, about 4,400 feet southwest of its junction with the main Oswegatchie River. Here and for about 3,200 feet to the northeast, both pyritic schist and a dolomite layer immediately southeast of it (Zone 3?) are well exposed in and alongside the rapids of the West Branch of the Oswegatchie River. The exposures of pyritic schist and quartz-biotite-oligoclase gneiss at the rapids are so close together that Zone 1 must be less than 50 feet thick and may be locally absent. At 9.4 N., 0.85 W., however, where pyritic schist forms an island in the West Branch, a zone of dolomite at least 20 feet thick referred to Zone 1 is exposed during low water. Near 12.0 N., 1.5 E., in the flood-plain of the Oswegatchie River, drill holes have cored marble, probably of Zone 3, lying between the silicated marble of Zone 4 and the pyritic schist. Dolomite that is also thought to be in Zone 3 is exposed on a low ridge 800 to 1,200 feet north-northwest of the International No. 2 1/2 mine at Talcville (13.0 N., 3.6 E.) and also on the northwest side of Newton Hill from 14.8 N., 6.7 E., to 15.5 N., 8.0 E. At these localities dolomite (of Zone 3?) lies northwest of silicated marble tentatively correlated with Zones 4 to 11. No pyritic schist, and therefore no dolomite certainly belonging to Zone 1, is exposed here, and much of the marble is cut by granite, pegmatite, and quartz veins. Beyond Newton Hill (16.8 N., 8.8 E) and in the vicinity of Cedar Lake, the few exposures of marble immediately southeast of the granite and quartz-biotite gneiss are dolomite like that in Zones 1 and 3. Throughout the northeast half of the district, however, from the Hyatt mine to Cedar Lake, the northwestern part of the marble is so poorly exposed and so thoroughly metamorphosed that the form and distribution of these bodies of dolomite cannot be determined from surface exposures, and the correlation with Zones 1 and 3 is tenuous at best.

Lithologic features

In most exposures near Fowler and Sylvia Lake, Zones 1 and 3 consist of essentially pure dolomite. This is also true of the rocks in many exposures to the northeast that are tentatively correlated with these units. The occurrences in this zone of most other minerals and rocks, such as quartz and pegmatite and micaceous bodies, are irregular in pattern and of types readily inferred to be introduced. Moreover, there is no evidence that these dolomites grade along the strike into more siliceous marble, or any other rock reasonably ascribed to a sedimentary origin, or that any originally calcitic rock has been dolomitized. The nearly pure dolomite is therefore thought to be approximate, both chemically and mineralogically, the pre-metamorphic and possibly the sedimentary composition of Zones 1 and 3.

Weathered exposures of the dolomite rock are pale to medium gray, commonly faintly layered; fresh broken specimens are white to cream-colored. Scattered lenses and clots, most of partly serpentinized diopside, make up 1 to 5 percent of the rock. In some of these there are visible scattered grains of forsterite, plagioclase, pyrite, graphite, quartz, tremolite and phlogopite.

The approximate gross mineral and chemical composition of dolomite in Zones 1 and 3 is given in table 3, and the compositions of specimens typical of the least altered dolomite in Zones 1 and 3 are shown in tables 4, 5, and 6. Slight inconsistencies between the modes in table

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Table 4. Mineralogical composition in volume percent of some of the least altered specimens of Dolomite Zones 1 and 3.

Table 5. Accessory element concentrations in weight percent of some of the inferred least altered specimens of Dolomite Zones 1 and 3.

Table 6. Approximate chemical composition in weight percent of specimens of Dolomite Zones 1 and 3 as determined with the emission spectroscope.

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4 and the chemical analyses in table 6 are largely due to inaccuracies in spectrochemical analyses for major constituents and to the difficulty in obtaining good thin sections of the extremely friable dolomites. The percentages in table 6 may be in error by as much as 25 percent of the value given.

The extreme friability of the dolomites is apparent in plates 3A and 3C. These represent weathered outcrops southwest of Sylvia Lake (1.8 S., 13.4 W.) where dolomite sand 10 feet deep has been formed by disintegration of the dolomites in Zone 1.

Table 4. Mineralogical composition in volume percent of some of the least altered specimens of dolomite zones ~~one~~ 1 and ~~three~~ 3.

Mineral	Specimen No.												
	Gm 15	Gm 16	Gm 17	Gm 59	Gm 62	Gm 67	Gm 68	Gm 69	Gm 83	Gm 84	Gm 85	Gm 86	Gm 87
Dolomite	95.	94.3	99.5	97	99.1	99.5	96.3	98.5	99.5	93	99	97.3	99.8
Calcite		tr		tr			.3	.2	.3			.4	.2
Diopside	.5						.5	.2	tr			.2	
Tremolite				2.1									
Serpentine and talc	4.5	0.2	0.2	.5	.7	.3	.3	.1	.2	1.0		1.5	
Plagioclase feldspar		.3	0.3				2.0	1.4		1.0		.1	
K-feldspar		3.0					tr	tr			.2		
Quartz		2.0								1.5	.3	.7	
Phlogopite							.1	.1		1.0			tr
Apatite	tr		tr				tr						
Pyrite		.2	tr	.2	.1	.2	.7		tr	tr	tr		tr
Magnetite	tr			.2									
Sphene							tr			tr			
Graphite				.1	tr	tr	.3						
Iron oxide		tr					tr						
Chlorite and sericite							.2		tr	2.5	0.5	0.8	
Total	100	100	100	100	100	100	100	100	100	100	100	100	100

Table 4. Mineralogical composition in volume percent of some of ~~Descriptions of specimens in Tables 4, 5, 6.~~ the least altered specimens of dolomite zones 1 and 3 (cont.)

Gm 15	Massive dolomite, ~~in dolomite~~ zone 3, about 800 feet northeast of Kellogg's Corners (2.7 S., at 12.6 W.).
Gm 16	Massive dolomite, ~~in dolomite~~ zone 3, about 600 feet northeast of Kellogg's Corners (2.7 S., 12.8 W.).
Gm 17	Gray, highly friable dolomite, zone 1, just below the pyritic schist, ~~in the dolomite zone 1~~, 2800 feet north-northwest of Kellogg's Corners (1.45 S., at 14.0 W.).
Gm 59	Dolomite, zone 3, about 3400 feet west of the Ontario talc mine (5.5 W., 6.4 N).
Gm 62	Dolomite, zone 3, about 1600 feet north of Kellogg's Corners (1.8 S., 13.2 W.).
Gm 67	Dolomite, zone 1, about 1600 feet northwest of Kellogg's Corners (1.7 S., 14.0 W.).
Gm 68	Dolomite, zone 1, about 3500 feet north of Kellogg's Corners (1.7 S., 14.0 W.).
Gm 69	Dolomite, zone 1, about 2500 feet south of Mud Pond (2.5 N., 15.0 W.). ~~2.5 W.).~~
Gm 83	Buff and gray mottled dolomite, zone 3, about 1600 feet north of Kellogg's Corners.
Gm 84	Buff and gray dolomite, zone 3, about 1800 feet north of Kellogg's Corners.
Gm 85, 86, 87	Dolomite at and near top of ~~dolomite~~ zone 3, about 1200 feet south of Mud Pond (3.2 N., 13.9 W.).

Table 5. Accessory element concentrations in weight percent of some of the inferred least altered specimens of dolomite zones ~~one~~ 1 and ~~three~~ 3.

[A. H. Chodos, analyst]

Specimen No.[2]	Elements[1]									
	B	Ba	Cr	Cu	Mn	Sr	Ti	V	Zn	Pb
Gm 59 Sud.	0.0007	0.002	0.0002	0.0002	0.06	0.05	0.001	---	---	0.001
Gm 62 Sud	.007	.0005	---	.0002	.07	.01	.001	---	---	do
Gm 67 UD	.01	.002	---	.0002	.07	.01	.007	---	---	do
Gm 68 UD	.002	.0003	.0002	.0003	.09	.02	.002	---	0.05	do
Gm 69 UD	.003	.0002	.0009	.0001	.09	.02	.006	0.002	.05	do
Gm 83 SUD	.0006	.0001	---	.0001	.06	.01	.002	---	---	do
Gm 85 SUD	.0006	.0002	---	.0001	.04	.02	.002	---	---	do
Gm 86 SUD	.007	.0008	---	.0002	.08	.02	.002	---	---	do
Gm 87 SUD	.006	.0004	---	.001	.06	.04	.002	---	---	do

1/ Looked ~~Sought~~ for but not found: Ag, As, Au, Be, Bi, Cd, Co, Cs, Ga, Ge, In, La, Li, Mo, Nb, Ni, Pb, Pt, Rb, Sb, Sc, Sn, Ta, Th, Tl, U, W, Y, Yb, Zr.

~~A. H. Chodos, analyst.~~

2/ See notes following table 4.

Table 6. Approximate chemical composition in weight percent of specimens of dolomite zones ~~one~~ 1 and ~~three~~ 3 as determined with the emission spectroscope.

Specimen No. 1/	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O
Gm 68	1.6	.2	.2	22.	31	.2	.1
Gm 69	1.2	.2	.2	21.	30	.2	.1

[Cal-Research Corp., LaHabra, Calif., analysts.]

1/ See notes following table 4.

Grain sizes in the dolomite generally range from 0.5 to 5 mm, although in a few beds the dolomite grains are as little as 0.1 mm or as much as 10 mm in diameter. Some specimens are nearly equigranular, whereas others show mortar structure, with grains 3 to 6 mm in diameter in a matrix of grains 0.01 to 1.0 mm in diameter. The accessory minerals, quartz and feldspar, form scattered clusters of tiny grains (0.01 to 0.2 mm) enveloped by grains of dolomite. Both talc and serpentine were formed by alteration of diopside, forsterite, and tremolite, and they also replace dolomite to a slight extent. Other accessory minerals found in widely separated exposures include pyrite, apatite, graphite, phlogopite, magnetite, earthy and specular hematite, chlorite, and sericite (table 4).

The scarcity of Al_2O_3, Fe, and SiO_2 throughout the supposedly least altered dolomite of Zones 1 and 3 (tables 4 and 6) indicates that the original sediment contained very little quartz or argillaceous or ferruginous material. Calcite appears chiefly in association with rock types readily ascribed to a metamorphic origin, such as highly serpentinous clots, or in contact zones along bodies of quartz-mesh rock, granite and pegmatite.

Since the dolomite-rich beds may be very similar in mineralogical and chemical composition to the original sediment, the analyses of accessory elements in table 5 have considerable interest, for they may approximate initial sedimentary concentrations, and thus indicate how abundant certain elements were in the Grenville seas. It is noteworthy that amounts of any given element, such as strontium, manganese, titanium or barium, are quite uniform in specimens taken at widely separated places. For example, the spread between the arithemtic mean and mode of the analyses of strontium rarely exceeds the limits of error (25 percent for the spectroscopic analyses). The strontium and lead in the dolomite are particularly interesting because these elements have been thought to offer a means of dating the carbonate sediments (Wickman, 1947). The percentage of strontium is about the same as in many dolomites that are much younger, whereas less-metamorphosed parts of the same dolomite beds may contain appreciably more strontium. Possibly strontium has been driven out of the dolomite crystals during metamorphism, or trapped in other minerals such as plagioclase and anhydrite, which are irregularly distributed but more receptive to higher concentrations of strontium under the existing conditions of metamorphism.

Two alternative explanations for the low amounts of strontium invite consideration. It is now recognized (Odum, 1950; Epstein and Lowenstam, 1953) that many marine shell-secreting organisms may preferentially fix a good deal more strontium than is precipitated with inorganic dolomite or calcite. If organisms that secreted carbonate hard parts were scarce or absent in the Grenville seas, the ratio of strontium to magnesium and calcium carbonate deposited might be much lower than if such organisms were abundant. Still another possibility is that the Grenville seas were relatively impoverished in strontium. Odum (1950) believes that the quantity of strontium in the ocean has remained relatively uniform since the Cambrian, but even if this is correct, it does not necessarily follow that the quantity was the same during deposition of the marble, over a billion years ago.

The appreciable amounts of manganese (0.06 - 0.09 weight percent Mn) are of interest for, as most of the calcite was formed at the expense of the dolomite, three-fourths of the manganese was expelled from the newly formed calcite crystals. This impoverishment of manganese during calcification of the dolomite is fairly common throughout the marble. In the dolomites of Zones 1 and 3 there is not much calcite, but in calcitic siliceous marble such as Zone 14 where large amounts of dolomite were altered to calcite, much of the manganese has almost certainly gone into silicates, especially tremolite and anthophyllite. These processes are particularly obvious in Zones 14 and 15, and in the talc belts where the tremolite schists contain as much as 2 weight percent MnO.

Copper, chromium, and barium are scarce in the dolomite, as might be expected, because these elements rarely form detectable minerals in the marbles and do not substitute readily for calcium or magnesium in the dolomite crystal structure. C. Patterson, in the single determination made thus far of lead concentration in the Grenville marble, measured 2.14 ppm Pb, and he found the lead to be relatively high in the 207 isotope. A possible explanation for the composition of the lead is that exchange of substances such as lead and uranium have occurred between the dolomite and the graphitic pyritic ~~pyritic~~ schist (carbonaceous shale?) during metamorphism.

In contrast to the lead and other cations, however, the isotopic compositions of carbon and oxygen (the complex anion) in the dolomites deviate from those found in younger and modern sediments, possibly as a result of exchange with waters during diagenesis (Engel, Clayton, and Epstein, 1958). However, the data on oxygen in the dolomites seems to indicate the bulk of the metamorphism was fairly dry, with very little permeation or exchange of oxygen between magmatic or hydrothermal waters and marble as it recrystallized. This is not too surprising, for the amounts of oxygen introduced into most of the marble after diagenesis, by magmas or other metamorphic agencies, must have been small compared to that entrapped in the sediments during their deposition and lithification. Opportunities for isotopic exchange with a very different nonmarine carbon or oxygen would consequently be slight except in skarns enveloped by igneous rock, in ore deposits, or in highly leached surficial exposures. Some of the well exposed, highly weathered parts of the dolomite do show appreciable exchange with oxygen typical of rainwater.

Seemingly, the concentration and composition of strontium and lead in the marble were modified in important ways by metamorphism, whereas the oxygen was little modified.

The nearly pure dolomite which predominates in Zones 1 and 3 gives way, abruptly in most places but gradually in a few, to rocks containing much quartz, diopside, and locally much feldspar. Most of these changes appear to be due to addition of material during metamorphism. North of the International Talc Company's No. 2 1/2 mine at Talcville, for example, (12.8 N., 3.7 E.) and north of Newton Hill (14.75 to 14.9 N., 6.5 to 7.0 E.), pegmatites, quartz veins, and some bodies of granitic rock occur in the dolomite zones. Most of the veins of igneous-looking rocks cut across the bedding, but some layers of quartz-diopside rock, masses of bulbous milky quartz and pegmatites tend to follow the bedding. A few of the siliceous layers may actually be relict beds, especially those ~~of Hyatt~~ that extend for about 1,200 feet southwest of Hyatt (9.6 N, 0.25 W.). At those places layers of coarse-grained quartz and diopside rock make up at least a third of what seems to be a segment of Zone 3. Most of these siliceous layers are uneven in thickness and pinch out abruptly along the strike. They are commonly associated with feldspathic quartz-mesh rock and pegmatite, and at Hyatt with ore bodies containing pyrite, sphalerite, and galena. These relations favor the interpretation that most of the silica and alkalis were introduced or widely redistributed.

Pyritic schist (Zone 2)

Distribution, form, and thickness

The bodies of rock mapped on plate 1 as pyritic schist are widely scattered, and their relations to one another are largely uncertain. All of them, however, represent clastic rocks interbedded with marble, and it is possible that they all once formed parts of a single stratigraphic zone (Zone 2).

The two largest mapped areas of pyritic schist in the Sylvia Lake-Cedar Lake marble belt both lie near quartz-biotite gneiss, from which they are separated by the relatively pure dolomite of Zone 1.

The large, sickle-shaped area of pyritic schist which curves around Sylvia Lake on the north and west pinches out to the southwest in good exposures just northwest of Kellogg Corners (1.85 S., 13.25 W.). This pinching out is believed due to the complex deformation that accompanied granitic intrusion in this area, where many units are greatly thinned or even pulled apart. South and east of Kellogg Corners, along the inferred line of strike there are very few exposures of the schist, though dispersed fragments of it may be concealed under the alluvium. To the north there are several good exposures of the pyritic schist along and just north of the Kellogg Corners-Creek School road, at and north of 1.35 S., 14.0 W. In this area the distribution of the schist is much complicated by nearly horizontal close folds, along the apices of which one layer of the schist is apparently triplicated.

As the pyritic schist is followed northward toward the Sylvia Lake road (2.4 N., 14.75 W.) its width of outcrop increases, mainly because of a flattening dip. Its northwest margin also nearly impinges upon the quartz-biotite gneiss, probably as a result of rock flowage that occurred during the formation of the Sylvia Lake fold. In the area south of Mud Pond, the general form of the schist is obscured by the swamp, but between the pond and Fowler there are several exposures that extend almost entirely across the schist. The observed dips in these exposures are gentle, and although the width of outcrop is at least 500 feet, the true thickness of the schist is probably about 200 feet.

At Fowler, any pyritic schist that may be interbedded with the marble is obscured by remnants of the overlying Potsdam sandstone, but northeast of the village, there is a tiny outcrop of the skarn-like facies of the schist near 6.4 N., 6.5 W., and along the south bank of Turnpike Creek. Nearby, the ground is littered with slabs of grunerite-magnetite-garnet rock. These are almost identical with that in skarn-like layers exposed along the opposite margin of the pyritic schist, to the southwest. Rocks of this type are not found elsewhere in the district and these skarn layers appear at the stratigraphic position generally occupied by the pyritic schist. Dolomite is exposed on the low hill to the south, whereas the rock exposed across the creek to the northeast is quartz-biotite gneiss. If the skarn-like rock represents all of the pyritic schist in this area, the thickness of this unit must be far less than elsewhere; it cannot be over 50 feet. This would be consistent, however, with the abrupt thinning of all the marble zones in this area. If the pyritic schist west of Fowler is represented by this occurrence of skarn, Zone 2 is probably continuous or nearly so, under the alluvium and Potsdam sandstone of this area.

From there northwestward to the old Hyatt settlement on the West Branch of the Oswegatchie (9.4 N., 0.8 W.), there are neither surface exposures nor drill-hole data. At 9.4 N., 0.8 W., however, there are large exposures of pyritic schist on a picturesque island in the river and in the rapids nearby. Dolomite is exposed on the south bank of the river, and quartz-biotite gneiss at many places on the north bank. The gneiss is separated from the schist by another thin zone of dolomite, as it is west of Fowler, and this schist is accordingly mapped as an eastward continuation or a fragment of that exposed near Fowler. This interpretation is strengthened by other occurrences of pyritic schist at the same stratigraphic position in the marble along the old Emeryville-Edwards road, 12.35 N., 2.0 E., northwest of Talcville.

North and northeast of Talcville this part of the stratigraphic sequence is largely cut out by lobes of inequigranular granitic gneiss and no other extensions of the pyritic schist can be recognized with certainty. Several possible correlatives appear, however. In the apex of a fishhook-shaped cross fold north of Edwards (north of the northeast corner of pl. 1), there is a highly granitized gneissic and schistose interlayer just within the marble, south of its contact with the quartz-biotite gneiss. This contact is possibly to be correlated with a similar one in the southwest part of the district. If so, the granitized interlayer may be a part of the pyritic schist. It is not highly pyritic, but in other respects its composition is much like that of the pyritic schist. Pyritic schist like that around Sylvia Lake also occurs in a lens of marble in the quartz-biotite-oligoclase gneiss about half a mile northeast of Cedar Lake. Brown and Engel (1956) have suggested elsewhere that the marble and pyritic schist are infolded in the gneiss and belong to the Sylvia Lake-Cedar Lake marble belt.

Other scattered occurrences of pyritic schist are mapped, especially in the Fullerville-Edwards marble belt. The fragments which form a series of lenses along the bases of the low hills of gneiss just north of Highway 58 (10 to 13 N., 5 to 10 E.) are especially noteworthy. Cushing and Newland (1925) and Gilluly (1945) mapped this entire area of gneiss and schist as pyritic schist, but Brown (1936a, p. 242, and fig. 1, p. 237) concluded that this was an outlier of the garnet gneiss to the southeast. Most marginal lenses of schist in this area are lithologically very similar to the layers of pyritic schist discussed above, and unlike the garnet gneiss. On the other hand, most of the gneiss is similar to the quartz-biotite-oligoclase gneiss mapped to the northwest. The field relations suggest that this rock mass represents an anticlinal apex of quartz-biotite-oligoclase gneiss, whose margins are plastered with highly sheared remnants of pyritic schist. These relations are discussed in greater detail in the section entitled "Structural features of the Balmat-Edwards district."

Lithologic features

The pyritic schist contains an appreciable quantity of disseminated pyrite, as its name implies, and its dominant colors in most exposures are rusty brown, yellow, or brownish black. Locally it is colored a deep cherry red by admixed earthy hematite. Most of the pyritic schist is thinly although unevenly layered. A few parts of it are quite massive.

The layers vary widely in composition, and are commonly so fine-grained and so much discolored at the surface by the weathering of the pyrite that their mineralogy is obscure. In the better-exposed parts of the belt southwest of Fowler and northeast of Mud Pond, at least four general lithologic types can be recognized. These are (1) pyritic granulite rich in quartz and microcline, (2) pyritic biotite migmatite, (3) pyrite-bearing chlorite schist, and (4) garnet-grunerite-magnetite skarn. Each of these rocks appears to occur in irregular zones roughly parallel to the others and to the contacts of schist and marble. At least one crosses the contacts in the order named in the area 4 to 6 N., 10 to 13 W. To the west exposures are far too few to prove that they are persistent along the strike. In other places the above-stated order does not seem to hold; slabs of the skarn-like zone appear to weather out from central parts of the pyritic schist. In exposures on Turnpike Creek northeast of Fowler, skarn either makes up most of the pyritic schist zone, or lies near its base. The above-named rocks, however, or rocks intermediate to them, are typical components of most of the pyritic schist mapped in plate 1.

The massive feldspathic granulite member is exposed at intervals along the side of the island in the West Branch of the Oswegatchie at 9.4 N., 0.8 W. Somewhat similar rock also forms minor interlayers in the other members of this unit. The granulite commonly shows a well-defined layering, rudely parallel to its boundaries, that appears to be relict bedding. In many exposures, however, this layering is crossed by a faint to well-defined secondary foliation. In numerous outcrops west of Sylvia Lake this later foliation dips gently to the west and does not seem to be parallel to the axial planes of the more obvious folds in the schist.

The granulite is pale-gray to brown where it is least altered, and is dense and fine-grained. As shown in table 7, it consists chiefly of

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Table 7. Modal analyses of the several general types of pyritic schist. — p. 84 and 84a

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quartz and potash feldspar, but may contain as much as 15 percent of pyrite and 26 percent of muscovite. Where pyrite is fairly abundant, it commonly forms tiny lenses and veinlets along the relict bedding or follows secondary shear surfaces. The texture is in some places crystalloblastic; elsewhere it is that of a mortar gneiss or a mylonite.

Some of the feldspathic layers differ from the granulite chiefly in being more micaceous and more migmatitic. They contain numerous veinlets and lenticles of quartz and potash feldspar, chiefly along the prominent layering. The micaceous layers, or the argillaceous beds from which they were derived, appear to have been especially liable to lit-par-lit injection by granitic fluids. Garnet appears in them as does sillimanite, which is commonly replaced by pseudomorphs of sericite and quartz (table 7). Some of the sillimanite forms as disseminated needles or bundles of needles; most of it, however, occurs with secondary quartz and sericite, in augen and almond-shaped clusters, which tend to show both planar and linear parallelism. The lineation is accordant with the axes of the cross folds in the pyritic schist.

Table 7. Modal analyses of the several general types of pyritic schist.

	Feldspathic granulite facies					Micaceous facies			Garnet-grunerite-magnetite Skarn						
	N5S	PS5	PS4	PS3	SPS	PS12	PS7	PS2	G5	PS8	PSS	PSA	PSB	G2	PSD
Quartz	37	32	28	40	49	39	51	35	60	13		4		2	9
K-feldspar	50	36	29	40	29	22	12	19						R	
Plagioclase	2	4	1	3	R	12			3						
Limonite and Hematite							6	2							
Carbonate			2			tr	2	1		1					
Serpentine		tr					5	15							
Tremolite					2		3	≠2		8					
Pyrite	7	15	7	13	9	7	2	5	1	12			2	R	
Pyrrhotite													1		
Apatite	1	R	R	3		1	1	2			2				
Biotite			4	1	5	5	7	11	5		4				
Muscovite sericite	2	13	26	R	3	7^{+}	10^{+}	4^{+}				3	1		2
Chlorite	1		2		1	7		3	2	6	5		4	1	
Augite									1		12	48	61		$^{+}31$
Grunerite											23	35	19	59	52
Garnet					2	R	1	1	2	14	19	3	7	36	2
Diopside	R	R	1			R				41					
Magnetite									26	5	28	5	4	2	3
Ilmenite		VR									7	2	1		1
Graphite								R	R		R			R	R
Zircon	R	VR													
Total	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100

≠ = a pale green hornblende R = rare
* = hypersthene VR = very rare
\+ = in part pseudomorphous after sillimanite

Location of specimens.

Specimens PS4, 5, SPS, PS7, G5, PS8, PSA, G2 and PSD from the belt of pyritic schist west and north of Sylvia Lake.

N5S and PS12 are from the pyritic schist along the south branch of the Oswegatchie River in the area (0.8 W., 9.4 N.).

PS3 and PS2 are from pyritic schist marginal to the Pleasant Valley gneiss, in the area (8.5 W., 11.5 N.).

PSS is from the small exposure of skarn (in pyritic schist) at (6.6 W., 6.4 N.).

(Table 6)

In parts of the micaceous feldspathic layers west and north of Sylvia Lake where sillimanite is abundant, there are almond-shaped clusters of sericite-quartz and sillimanite, which are of a pale cream or buff color, contrasting sharply with the dark rusty brown to gray groundmass. Rather similar spotted rocks appear at several places in the large mass of pyritic schist north of Highway 58 and southeast of Talcville (11.35 N., 6.5 E., and 11 and 12 N., 8 - 9 E.).

In most of the highly micaceous rocks, muscovite and biotite, or secondary chlorite replacing biotite, occur in flakes that range, in different specimens, from less than a millimeter to more than a centimeter in length. Some layers contain garnets up to 2 millimeters in diameter; some contain a little tremolite or hornblende (table 7). James Gilluly (unpublished data) reported hastingsite from one area west of Sylvia Lake. The plagioclase is almost invariably sericitized. Much of the quartz and feldspar form granoblastic layers between the leafy mica bands and sillimanitic scales.

The highly pyritic and chloritic parts of the schist commonly have a subconchoidal to blocky fracture and a rude fissility which is generally not parallel to the bedding but inclined to it at a small angle. Most of this rock when weathered is yellowish-black and contains much sericite, chlorite, and serpentine. When freshly broken it has a grayish-blue color due in part to fine-grained disseminated graphite. Accurate modal counts in sections of these rocks are almost impossible, for they are generally very fine-grained and much crushed as is shown by bent flakes of chlorite and strained or broken grains of quartz and feldspar. In general these rocks now contain much chlorite, sericite, and serpentine, and corroded relicts of quartz and feldspar. A few specimens contain as much as one percent graphite and a very little apatite and zircon. The pyrite content is as much as 14 percent in the specimens examined, but probably does not average more than 4 percent.

The layers of garnet-grunerite skarn appear to vary greatly in thickness, being possibly as much as 50 feet thick in some places, but are rudely conformable with the prominent layering in the pyritic schist and with the schist-marble contacts. They crop out most abundantly in the area 5.25 - 5.45 N., 12.0 - 12.7 W. They are massive to thin-bedded. In weathered outcrops they are dark brown to black; on freshly broken surfaces they are in part dark resinous brown, in part greenish and mottled with reddish-brown garnets. Grain size varies abruptly along and across layers. Splays or sheafs of grunerite can commonly be distinguished in even the fine-grained specimens and those of coarser pyroxene, grunerite, garnet and magnetite are easily recognizable, the pyroxene and grunerite being in crystals as much as half an inch long.

As may be seen from the seven thin sections examined (table 7), the above named minerals are the major constituents of the skarn rock and the accessories are ilmenite, muscovite (sericite), chlorite, biotite, and apatite.

The pyroxene is chiefly augite; its optical properties, determined by W. T. Schaller (Gilluly, James, unpublished data, 1934), are: α = 1.711, β = 1.718, and γ = 1.738 (all $\pm$ 0.002), extinction angle Z c 43°. In one specimen Schaller found a faintly pleochroic hypersthene, with α about 1.742, β = 1.747, and γ = 1.756 (Gilluly, James, unpublished data). The rock contains a little pale-green diopside (PS8, table 7). The amphibole is chiefly grunerite, but a very little tremolite is present locally. Grunerite is largely colorless to pale brown, and non-pleochroic; its extinction angle is 14° and ranges in different specimens between 1.701 and 1.713. The amphibole in the hypersthene-bearing rock has lower indices than the amphibole associated with the hedenbergite, and may be lower in iron. The biotite has γ index of about 1.667. Gilluly (unpublished data) noted that the chlorite which replaces the biotite has γ of about 1.662.

The garnet crystals are rounded and partly corroded. Some of the amphibole is clearly secondary after pyroxene, and the non-opaque minerals are all partly altered to sericite or chlorite. Some specimens contain calcite, and there is an abrupt but recognizable gradation between the more iron-rich grunerite-augite-garnet facies and diopsidic marble.

This garnet-grunerite rock was mentioned very briefly by Brown (1936a, p. 239), who concluded that it was formed by "contact"-metamorphic alteration of calcareous lenses in the pyritic schist. Several features, especially the mineralogy, the texture, the close association with contacts between schist and silicated marble, and the calcareous interlayers, mark the rock as a skarn. Grunerite-garnet-magnetite skarn is unknown elsewhere in the Grenville Lowlands, and the nearest known mass of igneous-looking granite is at least 1,000 feet northwest (pl. 1). But, as is commonly recognized, metasomatism of the kind shown in this rock is not always closely associated with intrusive bodies.

Other units of pyritic schist were evidently derived from sandy and shaly beds in the marble. In many outcrops of the schist the oldest layering is clearly bedding, now more or less blurred by discordant surfaces of cleavage.

The origin of the pyrite in these schists is far from clear. Most of it was either introduced into the schist, or widely distributed, for many lenses, leaves, and veinlets of pyrite follow secondary and tertiary cleavages, shears, and cracks. In places the pyrite is associated with as much as one percent graphite. Moreover, some of the marble zones contain appreciable amounts of H_2S, and of methane or closely related members of the paraffin series. The pyritic schist may conceivably represent a euxinic or "black shale" sedimentary type, deposited in stagnate sea water made rich in H_2S by large bacterial or algal populations. If so, the interlayering of zones rich in H_2S with layers of clean siliceous carbonate and anhydrite would require marked alteration, with time, of pH and redox potential at the interface of sedimentation. As was pointed out, however, on pages 33-34, certain of the pyritic concentrations in schists of the northwestern Adirondacks are peripheral to a well-defined zonal pattern of oxides and sulfides in the northwestern Adirondacks (Buddington, 1939, p. 179-180). The pattern is very similar to that found in certain metalliferous deposits, where there is ample evidence that the metals have been deposited from outward-spreading hypogene ore carriers or by outward diffusions at a late stage in the metamorphic history of the region.

A second point of considerable interest is that many different kinds of rocks, such as quartzite, limestone, dolomite, gneiss, granite, and gabbro, contain sporadic pyrite. In certain instances, and commonly throughout the pyritic schists in the district, the pyrite is accompanied by a distinctive kind of chlorite. Smyth first noted (1912, p. 167-169), and many other workers have agreed (Brown, 1936b, p. 336-342; Buddington, 1917, p. 12, 13), that the chlorite and the pyrite may be of the same age and of closely related origin. Much of the graphite also may be hydro-thermal or metamorphic.

The problem is complicated by the existence of a second chlorite clearly later than that formed with the pyrite (see also Smyth, 1912, p. 168). This second chlorite has long been thought (Brown, 1936b, p. 339-341, fig. 2) to have been formed by descending ground water. The earthy hematite that has accumulated locally in the marble near the pyritic schist and at contacts between marble and Potsdam sandstone obviously formed by the weathering of pyrite in the schist.

Silicated dolomite (Zones 4 to 11)

Distribution, form and thickness

The group of variously silicated layers of marble, with dolomitic interbeds and an anhydrite-gypsum unit, that are intermediate in age between the dolomite of Zone 3 and that of Zone 12, are called the silicated dolomite. In the Sylvia Lake area eight units have been distinguished within the silicated dolomite, Zones 4 to 11 inclusive. These encircle the lake in a series of folds. They have been differentiated and discussed in a paper by Brown and Engel (1956), and are briefly described in table 2. Three of them (5, 7, and 9) consist of relatively pure dolomite. Their aggregate thickness ranges from about 200 to 425 feet, and that of three silicated zones which give the unit its name ranges from about 200 to 1200 feet.

Zones 4 and 8 of the silicated dolomite are well exposed along the west and southwest sides of Sylvia Lake. Zones 4, 5, and 6 also appear in numerous excellent exposures north of the lake. The youngest zones, especially Zones 8, 9, and 10, form large, prominent outcroppings east and northeast of the lake as far as the large outlier of Potsdam sandstone. This sandstone covers nearly all of the silicated dolomite, except Zone 11 northwest of the Arnold talc mine, as far as the vicinity of 6 N., 7 W., east of Fowler. From the exposures east of Fowler, however, as far as 6.4 N., 3.1 W., it is clear that many beds of marble have become considerably thinner, or have even disappeared in the area of Potsdam cover; and the dolomite-rich zones are more poorly exposed and more difficult to distinguish from one another. None of these specific zones mapped in the Sylvia Lake area has been positively identified northeast of Fowler, although a correlation of Zones 5, 6, and 7 is attempted in the area 6.0 to 6.4 N., 5.3 to 6.7 W.

Both the width of outcrop and actual thickness of silicated dolomite decrease northeast of Fowler. The great thinning in width of outcrop of the silicated dolomite [in this area] is due to the steepening of dips and disappearance of large folds in the marble.

Thinning of the marble was doubtless further promoted by the bulbous mass of amphibolite southeast of Fowler. If this is a folded gabbro sill, as appears probable, it existed during most or all of the deformation of the marble, and the sill and the quartz-biotite gneiss and granite to the north may have formed a pair of huge pincers between which the marble was squeezed. It is likely that this pinching of the marble would be accompanied by the stretching and floating apart of some silicated beds in the carbonate and the pinching out of others.

Northeast of Fowler the silicated marble unit is about 600 feet thick, a decrease of roughly one half. Comparable thinnings also occur in other zones, such as the pyritic schist, in this area of transition between the flank and apex of the fold. Either the marble has flowed during the deformation, from the flank of the major fold in the area northeast of Fowler, into its apex around Sylvia Lake, or the flank area has been stretched laterally. In either process some component members of the silicated dolomite could well have been completely pulled apart or pinched out. The absence of certain distinctive beds northeast of Fowler that are present to the southwest is therefore ascribed chiefly to tectonic causes rather than to major variations in the original thickness of the beds.

Whether the silicated dolomite of the Sylvia Lake-Fowler area is continuous with that mapped on the wooded ridge to the east (8.5 N., 0.5 W.) is uncertain. This ridge does show a sequence of (1) dolomite, (2) silicated marble, (3) dolomite with a pyritic schist interlayer, and these divisions represent respectively dolomite Zone 11, silicated dolomite, and dolomite Zone 3. But no well-defined talc belt or siliceous calcitic marble is found on the ridge, and even the median gneiss is not positively known to continue to the ridge. Moreover, large parts of both the silicated dolomite and the dolomites above and below it are blurred by development of quartz-mesh rock and associated streaks and irregular masses of barren glassy to white quartz.

Farther to the northeast, the tracing of the silicated dolomite and some associated rocks becomes even more speculative. Exposures of these rocks are totally lacking in the vicinity of the Hyatt mine. Diamond-drill cores indicate, however, that at least three distinctive units--Zones 1 and 3, silicated dolomite, and Zone 12--found in the vicinity of Sylvia Lake are also present underground near Hyatt. These units are also identical in scattered outcrops north of the International No. 2½ mine, where the beds appear to have been squeezed into a very tight and irregularly silicified U-shaped fold. The tongues of granite which project into the marble here greatly complicate the lithologic pattern, and satisfactory understanding of the stratigraphy in this area will call for much additional work.

Between the No. $2\frac{1}{2}$ mine and the northeast edge of the area shown in plate 1 there are somewhat similar complications: tongues of granitic gneiss, pegmatite bodies, and quartz veins are common in the marble, which is much folded and silicated. These complicated outcrops are interspersed with large areas devoid of outcrops, such as the swamp between the International Talc Company Nos. $2\frac{1}{2}$ and 4 mines (13.0 to 13.5 N., 4.0 to 5.0 E.) and the tailings pond of the Edwards zinc mine (17.0 to 19.0 N., 9.0 to 10.5 E.).

In rather good exposures of the marble on Newton Hill (around 14.0 to 15.4 N., 6.5 to 8.0 E.) a great lens of silicated marble lies between dolomites that might reasonably be correlated with dolomite Zones 3 and 12. This lens appears to pinch out--or to be pulled apart--in either direction along the strike, but swells farther along the 5.0 E. grid line to a thickness of at least 500 feet. A similar sequence, much folded and thickened, appears to crop out northwest of the large fold between Edwards and Cedar Lake. The heavy forest cover there and the lack of a base map made it impossible to work out the stratigraphic details in the available time, but certain of the marble layers there are clearly duplicated or even triplicated in a series of large drag folds subsidiary to the great fold around Cedar Lake.

Lithologic features

The silicated dolomite comprises rocks of two major lithologic types. About 35 percent of it consists of nearly pure, medium to coarsely crystalline dolomite (grain size 0.1-10 mm) like that in Zones 1 and 3. This rock makes up Zones 5, 7, and 9 of the silicated dolomite. The other major rock type constitutes the silicated layers, lenses, and nodules; it consists mainly of quartz and partly serpentinized diopside interlayered with or enclosed in dolomite or calcite (Zones 6, 8, and 10 in the Sylvia Lake area). In weathered exposures of these rocks the silicates are etched out in strong relief above the carbonates. Examples of the silicate rocks, as they appear at widely separated localities, are illustrated in plates 3C, 3D, 4A, 4C, and 5C. A typical contact between silicate-bearing marble and

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Plate 3C. Folded contact of dolomite and silicated dolomite units southwest of Sylvia Lake. (in pocket)

Plate 3D. Contact of silicated dolomite Zone 4 and almost pure dolomite Zone 3 west of Sylvia Lake. (in pocket)

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unsilicated dolomite rock, exposed southwest of Sylvia Lake, is illustrated in plate 3D. Dolomitic rock is also shown in plate 3C. It is evident from these illustrations that the proportion of silicates to carbonates differs in different zones, and that the form and composition of the silicate masses varies widely. At some places, moreover, the ratios and kinds of both carbonates and silicates vary greatly along a given horizon. This is especially noticeable where silicate-bearing zones are traced through areas where the rocks have undergone different amounts of deformation or of metasomatic alteration.

Most of these major variations in form and lithology along a given horizon are probably of metamorphic origin. In many places, of course, sedimentary and diagenic processes must have been contributing factors, but their effects appear to be overshadowed by those of deformation and metasomatism. Where the rocks are least obviously deformed, most of the silicate masses are essentially uniform layers, symmetrical nodules and lenses. These conditions are illustrated in plates 4A, 4C, and 5C, pictures taken at widely separated localities.

Plate 4A. Silicated dolomite northeast of the Edwards Talc Belt and southwest of Cedar Lake. (in pocket)

Plate 4D. Diopsidic quartzite beds and laminated layers intercalated in moderately siliceous calcite of the calcitic Footwall Marble. (in pocket)

The silicates make up 25 to 50 percent of thicker zones, and perhaps 35 percent of the entire body of silicated dolomite. As the silicate-rich zones are traced along the strike into areas of complex folds, they assume more complex forms related in origin to associated folds or other structural features. The ratio of silicates to carbonates, also, may decrease or increase along a given horizon, going well above 1 to 1 or well below 1 to 4. Some typical examples of silicate masses in areas of pronounced deformation appear in plates 3C, 6C, 8A, 8C, and 8D. Many of the silicate lenses and layers have clearly been

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Plate 6C. Siliceous calcitic Footwall Marble containing a folded fragment of a silicated interbed. (in pocket)

Plate 8A. A dolomite-calcite-talc breccia in the marble. (in pocket)

Plate 8C. Diopsidic quartz schist which occurs in both dolomite and silicated dolomite zones in the marble. (in pocket)

Plate 8D. Closeup of the quartz schist mass in plate 8C. (in pocket)

pulled apart and variously dispersed by flow in the marble, as is illustrated by plates 4D, 6C; and in many areas either the silicate

Plate 4C. Massive diopsidic quartzite layers (beds) in Zone 6 of the silicated dolomite exposed along the north shore of Sylvia Lake (1.95 N., 12.1 W.). (in pocket)

or carbonate component of a layer is preferentially thickened at or near the apex of the fold, and thinned along one or both flanks. These relations are well exposed in the silicated dolomite southwest of Sylvia Lake (pls. 8A, 8C). The tendency of silicate layers to thin out along the short flanks of asymmetrical folds is especially apparent in plates 8C and 10B. In scattered examples around Sylvia Lake (pl. 3C) and also near Hyatt and Cedar Lake, the silicated layers are rolled into rodlike masses elongated parallel to the axes of cross folding. When viewed normal to the cross fold axes in areas of intense folding, the silicated masses exhibit intricately crenulated or bleblike structures (pl. 3C). When the silicate subzones are traced into areas of less intense folding, these structures are seen to grade into the more typical lenses and irregular layers.

The relation~~s~~ of mineral composition to textural and structural features varies widely. In the least altered silicate beds, a typical relationship is that of a core or inner zone of quartz, sheathed from the enveloping dolomite by diopside. The diopside embays the quartz and is clearly a product of reaction between quartz and dolomite (pls. 8C, 12A). Both very thin and very thick rims of diopside appear (pls. 8C and 8D). Elsewhere most or all of the quartz has been converted into diopside (pl. 4A), or less commonly into tremolite.

Wherever tremolite, steatite, and serpentine occur in the silicated marble, they are commonly separated from the surrounding dolomite by sheaths of calcite. This relation is illustrated in plate 5C. The nodules there shown are zoned, and consist, from the center

Plate 5C. ~~Near here.~~ in pocket

outward, of quartz, serpentinized diopside, and calcite, which is in contact with the enclosing dolomite. The paragenetic relations are quartz + dolomite → diopside → serpentinous diopside sheathed with calcite; the product at every stage was enveloped in dolomite. Forsterite appears in place of diopside at a few places where quartz is deficient. In rocks produced by reactions of these types, the ratio of calcite to magnesium-bearing silicate is generally about what one might predict from the ratio of MgO to CaO in the original dolomite. In most places, presumably, little MgO has been introduced and only CO_2 has been lost. Except for slight to moderate decarbonation, the rock has about the same bulk composition that it had before it was metamorphosed (neglecting CO_2 and H_2O) and the beds may have about the same volume and general form.

Plate 5C. Silicated marble showing the interrelations of serpentinous diopside lenses and beds to calcite and dolomite.

In some of the silicated dolomite, tongues and irregular masses of quartz-diopside rock diverge from the layering defined by nodules, lenses, and layers, and many of these divergent quartzose masses are not enclosed in continuous sheaths of magnesian silicates. Where these irregularities exist the ratio of total silica to magnesium and calcium rises abruptly over the rather stable minimum values characteristic of the rock in either direction along the strike or down dip. This more siliceous rock, which merges in places into quartz mesh rock or quartz schist, is believed to have been formed either by secondary addition or large-scale redistribution of silica (pls. 7C, 7D).

Large concentrations of silica, like those shown in plate 7C, are

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Plate 7C. ~~Near here.~~ in pocket

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of only local extent, but evidence of minor addition of silica appears in many exposures. In the silicated marble as a whole, about one-thir of the silica now present was introduced from widely separated sources either within or outside the marble. If so, this unit before metamorphism contained approximately 22 percent of silica by weight, the remainder being mostly dolomite (table 3). The present overall SiO_2 content, as estimated from field mapping and from laboratory analysis of drill cores and other composite samples, is roughly 34 percent. The calcite in the silicated dolomite--roughly 10 percent (table 3)--is interpreted as being largely produced by reactions between dolomite and quartz that formed silicates in which the ratio of magnesium to calcium was higher than it is in dolomite (see p. 90).

Plate 7C. Footwall marble at the area of abrupt transition from dolomitic to siliceous calcitic Footwall marble.

Rocks that show both abrupt and gradual increase in quartz content to values far above the inferred sedimentary average are well exposed southwest of Balmat and at Talcville. Some of the added quartz enlarges nodules, lenses, and layers. Quartz also forms irregular tongues and apophyses, many of them in contact with blocky masses of diopside, and some of them cut across and blur the layering. Partly to fully developed quartz-mesh rock is common in these areas. Clots of potassium feldspar and of diopside, or ragged sprays of tremolite, may be scattered through the quartz, but the quartz is not commonly sheathed with magnesium-silicates. The fact that these concentrations of quartz in the marble are accompanied by potassium feldspar, presumably contemporaneous with it, offers additional evidence of the metasomatic origin of the quartz in the more siliceous marbles.

Specific examples of silication in the silicated dolomite occur just southwest of Balmat Corners (1.0 S., ~~to~~ 9.7 W.), on the wooded hill southwest of Hyatt (8.25 to 10.0 N., 0.75 W. to 0.75 E.), west of the International Talc Company's No. 2½ mine (12.6 N., 3.5 E.), north and east of Sylvia Lake, and in the middle and upper parts of the silicated dolomite just south of Cedar Lake. Veins of pegmatite and quartz are common in the silicated dolomite on the north side of Newton Hill (14.75 N., 7.0 E.), north of which is a large area of granitic rock and pegmatite that lies athwart dolomite Zones 1 and 3.

Some typical chemical changes that accompany the formation of metamorphic silicates may be indicated with the aid of the four analyses of specimens Gm SL 1, Gm SL 2, Gm SC 1, and Gm 337 in tables 9 and 10. These are analyses of composite samples collected across

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Table 9. ~~Near here.~~ p. 116

Table 10. ~~Near here.~~ p. 117

Zones 4 and 6 in the silicated dolomite.

Table 9. Approximate chemical composition of parts of the silicated dolomite.

Specimen Number	Oxide (weight percent) SiO_2	Al_2O_3	$FeO+Fe_2O_3$	TiO_2	MgO	CaO	Na_2O	K_2O	SrO	BaO	MnO	SO_3	Ignition loss	CO_2	H_2O-	Total
1/ An-23A	2.2	.1	.5	.02	13.0	34.	.2	.1	.10	.015	.06	nd.	nd.	nd.		
1/ An-23B	46.0	10.0	4.6	.39	2.5	15.	1.1	2.3	.05	.055	.2	nd.	nd.	nd.		
1/ Gm 71	1.2	.3	.2	.02	22.	32.	.2	.1	See Table 10			nd.	nd.	nd.		
1/ Gm 72	.1	.1	.2	.02	22.	31.	.2	.1	See Table 10			nd.	nd.	nd.		
2/ Gm SC 1	63.90	.51	.25	nd.	14.01	12.86	.62	.9			.05	.04	6.42	4.95		99.56
3/ Gm SL 2	58.34	nd.	.90	nd.	13.93	19.04	.24	.16				nd.	nd.	nd.		
3/ Gm SL 1	44.32	nd.	.64	nd.	15.60	22.41	.24	.03				nd.	nd.	nd.		
4/ Gm 337	41.09	1.07		nd.	15.25	25.15	nd.	nd.				nd.	16.93	4	.23	99.89

1/ Analyses by emission spectrograph, courtesy Cal-Research Corp.

2/ Glen Edington, analyst.

3/ Orton Smalley, analyst.

4/ Celeste G. Engel, analyst.

Table 10. Concentrations of accessory elements in parts of the silicated dolomite.[1]

Specimen Number	Element[2] (weight percent)										
	B	Ba	Cr	Cu	Mn	Sr	Ti	V	Zr	Ni	Pb
Gm 58	.0008	.001	.0002	.0002	.05	.02	.0002	---	---	---	~.001
Gm 63	.004	.004	.0002	.0003	.07	.02	.008	---	---	---	~.001
Gm 71	.001	.0009	.0002	.0001	.05	.02	.002	---	---	---	~.001
Gm 72	.001	.01	.0001	.0001	.04	.02	.002	---	---	---	~.001
Gm 74	.0006	.003	.0003	.0002	.07	.05	.002	.0005	---	---	~.001
Gm SL 1	.005	.01	.0006	---	.06	.01	.004	---	---	.0003	~.001
Gm SL 2	.001	0.1	.0001	.0005	.02	.01	.001	---	.0004	---	.002

1/ Emission spectrograph analyses by A. H. Chodos.

2/ Sought for but not found: Ag, As, Au, Be, Bi, Cd, Co, Cs, Ga, Ge, In, La, Li, Mo, Nb, Ni, Pb, Pt, Rb, Sb, Sc, Sn, Ta, Th, Tl, U, W, Y, Yb.

Gm SL 1 is considered representative of the silicic Zone 4 northwest of Sylvia Lake, and Gm 337 of Zone 6 northeast of the lake. The two analyses are remarkably similar. Specimen Gm SL 1 is a channel sample across a uniformly layered part of Zone 4, taken about 1,200 feet southwest of the old Sylvia Lake Hotel (1.0 N., 14.1 W.). The principal silicates in it are quartz and partly serpentinized diopside; the diopside was presumably formed by reaction of sedimentary quartz and dolomite. The subsequent formation of serpentine was accompanied by the liberation of about 5 percent of calcite. The total silica content, 44.3 percent, is believed to be about the same as it was before the rock was metamorphosed. The ratio of magnesium to calcium is virtually the same as in dolomite, despite the extensive reconstitution that the rock has undergone. These facts, together with the field relations, suggest that the principal change in composition has been decarbonation, presumably accompanied by the addition of a little water and perhaps a very little silica.

Sample Gm 337 represents a rock that has not been studied in outcrops like those mentioned above, for it was taken from a diamond-drill hole. The analysis, however, and study of a thin section, show that this rock is not very different from Gm SL 1.

Sample Gm SL 2 is from the uppermost 30 feet of Zone 4, taken about 600 feet west of Sylvia Lake (0.1 N., 13.8 W.). In general, thi part of the zone is more silicic than other parts, but the field relations as well as the analytical data suggest that some silica may have been introduced. Slight swellings and apophyses appear in Zone l at this point, and the total silica is estimated to be some 10 percent higher than that present nearby along the strike in either direction. The ratio of magnesium to calcium, however, is nearly the same as in the mineral dolomite, and even if silica has been added, there is no evidence of preferential loss of either calcium or magnesium. Total K_2O is about twice as high as it is on the average in the silicated dolomites. This increase is interpreted as a secondary effect, incidental to the silicification of the rock, perhaps during the early stages of silicification.

A more advanced stage in replacement of dolomite by alkali silicates appears to be shown in the analysis of Gm SC 1, a channel sample cut entirely across the silicated dolomite about 1,300 feet northeast of Newton Hill (15.4 N., 8.0 E.). This locality is not far from an area in which a good deal of granitic and quartzose pegmatite has been introduced into the marble (pl. 1). In this area, bulbous and veinlike masses of coarsely crystalline, milky quartz are conspicuous. The silica content, 63.90 percent, is at least 15 percent higher than in most of the nearby evenly layered siliceous dolomite. Na_2O appears to be at least twice as high as in less siliceous parts of this unit, and K_2O almost twenty times as high. These alkalis occur chiefly in potassium feldspar, mica, and plagioclase, which are all uncommon in the inferred less altered siliceous dolomite. In sample Gm SC 1, moreover, the ratio of magnesium to calcium is much higher than that in either the less altered parts of this zone or in the mineral dolomite; these modifications appear to indicate marked changes in the composition of the rock, chiefly by decarbonation but also by silicification and introduction of magnesium or, more probably, by loss of calcium.

Although field relations appear to indicate that at least 30 percent by weight of the silica in the entire siliceous dolomite is of sedimentary origin, two other modes of origin of the silica in the silicates may be considered: (1) the silicate masses may represent either wholly or in part metamorphic concentration of silica originally disseminated more uniformly through the dolomite or (2) the silica is largely if not wholly introduced into the dolomite during metamorphism.

The redistribution of SiO_2 assumed under the first hypothesis seems inadequate as more than a moderate supplementary effect. The beds of silicated dolomite that appear to be least altered by metamorphism are in a characteristic sequence, which may be traced from place to place through the metamorphic overprint; it is not altogether obliterated by the uneven metamorphism caused by changes of temperature, pressure, stress, and composition of fluids.

The second hypothesis of introduction of most of the SiO_2 appears inadequate. If the siliceous bodies were largely or wholly of metamorphic origin, they should show a systematic relationship to other metamorphic phenomena, regional as well as local, instead of being blurred by them. If metamorphic agencies formed the lenses and layers of SiO_2, they must have done so before the silica reacted with the dolomite, for in these bodies as in the siliceous calcitic marble to be discussed later, the generally symmetrical relation of the diopside sheaths to the silica cores indicates that the cores were formed before this reaction took place. Moreover, any migration and coalescence of silica must have occurred before the metasedimentary rocks were deformed and injected by alkali silicate fluids, for many of the diopside-quartz masses are broken, folded, and dispersed by deformation, and the diopside is syngenetic with the granitic rocks and pegmatites. The advocate of large-scale metamorphic differentiation must assume that the mass movement of silica occurred before the emplacement of the granites and consequently before the carbonate sequence had been appreciably metamorphosed. It would seem that any large-scale metamorphic concentration of silica must have occurred before regional metamorphism, or soon after the metamorphism began. Nodules and lenses of silica might of course represent lenses of cherts or novaculite formed by replacement along the bedding shortly after the deposition of the rocks. This origin has been suggested for chert nodules and siliceous layers found in Paleozoic limestone such as that in the Boone and Burlington formations of the Mississippi Valley. It is hard to prove or disprove even when applied to unmetamorphosed Paleozoic rocks and hence is beyond development in the intensely metamorphosed rocks of the Balmat-Edwards district.

Although the hypothesis of metamorphic differentiation on a large scale seems inadequate to explain most of the siliceous masses, the process certainly operated on a small scale. There are, for example, streaks and clots containing diopside, tremolite, and serpentine in the silicated dolomite that not only lack the quartz core but contain cores or irregular intergrowths of carbonate, commonly calcite. The external form of these silicate masses is quite irregular, even where the marble has been but slightly deformed. Some of them are in layers that also contain cores of quartz encased in diopside, but others are not. Plate 5A, from a photograph taken on Newton Hill, shows a rock

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Plate 5A. ~~Near here~~. in pocket

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that contains only the irregular clots. The rock contains no relicts of a more uniform siliceous mass, and the clots are not uniformly distributed along the strike. Features of this kind appear to have formed, wholly or in part, during metamorphism by accretion of silica that had been widely disseminated through the dolomite.

Plate 5A. Highly serpentinous and diopsidic clots and irregular masses in partly calcitic, partly dolomitic marble.

One notable feature of the siliceous dolomite zones in the silicated dolomite is that all the quartz they contain seems to have been deposited in nodules, lenses, or layers that pinch and swell, rather than in uniform beds or laminae. The pinching and swelling shown in Plates 3C and 4A cannot have resulted from deformation of uniform layers and laminated interbeds of quartz; if highly regular layers or laminae were originally present in the silicated dolomite, relicts of them should still be there, as they are in the siliceous footwall marble (plates 4A and 4D). As the two zones consist of similar rocks, metamorphosed to about the same extent, the chances of survival of sedimentary features in both units should be about the same.

The relatively pure carbonate beds between and around the silicate masses in the silicated dolomite are similar in appearance and composition to those in Zones 1 and 3. These similarities can be appreciated by comparing the compositional data in tables 8, 9, and 10, for specimens

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Table 8. Near here.

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from dolomite zones in the silicated dolomite, with those for Zones 1 and 3, listed in tables 4, 5, and 6. The similarities are apparent not only in outcrops but in hand specimens and even in thin sections. The dolomite of Zone 7, in the silicated dolomite, forms a conspicuous exception, however, almost solely because of its appreciable H_2S content. When this dolomite is broken it emits the strong, characteristic odor of hydrogen sulfide. There are large exposures of this fetid dolomite or "stinkstone" (a) just southwest of Sylvia Lake, (b) along the north side of its outlet (1.5 N., 11.4 W.), and (c) in a series of exposures southeast of the Gleason zinc mine of the St. Joseph Lead Co. Zone 7 has been found in exposures and drill corings all around Sylvia Lake and almost surely is present northeast as far as the area of the Hyatt zinc mine. This zone may have greater continuity and strike length than the exposures described above, for stinkstones occur at about the same stratigraphic position in marble north of Newton Hill and in the vicinity of Cedar Lake. Stinkstone has also been encountered at different stratigraphic positions (Brown and Engel, 1956) in exploratory operations and in the mining of lead and zinc at both Edwards and Balmat, but that in Zone 7 is the thickest and most persistent known at present.

Table 8. Mineral composition in volume percent of some parts of silicated dolomite

Mineral	Specimen number								
	Gm 12	Gm 13	Gm 14	Gs 42	Gm 40	Gm 36	Gm 38	Na 4	Na 5
Quartz		.02		33.0	41	2.80		0.2	10.0
Diopside					59	96.0	98	94.0	86.0
Tremolite	.02								
K-feldspar	.01	0.5	1.0	50					
Plagioclase	.10	.12		Ab^{80} 17.0	tr		tr		
Phlogopite		tr						tr	tr
Serpentine and talc	1.25	2.0	2.7		tr	.9	2.0	5.8	4.00
Chlorite					tr				tr
Dolomite	98.5	97.0	96.0		tr				
Calcite	.2	1.6				.3	tr	tr	
Pyrite	0.1	tr	0.03	tr	tr			tr	
Apatite		0.2							
Magnetite									
Hematite and limonite		tr							
Graphite	tr	tr			tr				

Description of specimens for Tables 8, 9, 10.

NA4 and NA5.	Diopsidic layers at the locality illustrated in plate 4A, about 2000 feet south of Cedar Lake (13.2 N, at 11.6 E).
Gm 12	Massive ~~Fedit~~ fetid (H_2S-bearing) dolomite bed, silicated dolomite (zone 7) just northeast of Sylvia Lake (0.88 N, at 11.5 W).
Gm 13	Massive gray dolomite, silicated dolomite, southwest shore Sylvia Lake (1.25 S, at 12.25 W).
Gm 14	Massive dolomite interlayered with quartz-diopside layers east shore of Sylvia Lake.
Gm 36, 38 & 40	Massive, lenticular body of diopsidic and quartz-diopside marble partially enveloping granite (Gs 42) within the silicated dolomite, on the north side of Newton Hill (14.9 N, at 7.0 E).
Gm 58 & 72	Coarsely crystalline massive dolomite zone 5, about 2400 feet north of Sylvia Lake (12.8 W, 3.4 N).
Gm 63 & 71	Coarsely crystalline dolomite beds interlayered with quartz-diopside beds about 2800 feet north of Sylvia Lake (12.75 W, 3.6 N).
Gm 74	Massive dolomite beds exposed along Township road about 2400 feet northeast of Sylvia Lake (11.5 W at 3.65 N).
Gm SL 1	Channel sample of basal 200 feet of zone 4, taken in area about 800 feet northwest of Sylvia Lake (14.0-14.2 W, at 1.85 N).
Gm SL 2	Channel sample of uppermost 30 feet of zone 6, taken about 600 feet west of Sylvia Lake (.1 N, at 14.35 W).
Gm SC 1	Channel sample across silicated dolomite (approximately 200 feet thick) at locality about 1300 feet northeast of Newton Hill.
An 23A & 23B	Anhydrite interlayer in zone 10 northeast of Sylvia Lake (drillcore).
Gm 337	Chemical analysis of split diamond drill core across zone 6 northeast of Sylvia Lake. Vertical hole, ~~collaring at~~

Stinkstones appear to occur in the Grenville-like marbles in other widely separated areas. Brown (oral communication) has found them in marbles in the Gateneau River valley about 60 miles north of Ottawa; de la Rue (1948, p. 17, 21), in the Norminingue and Sicotte map area, Quebec; Buddington and Leonard, in the marbles southwest of Snyder Lake (Oswegatchie quadrangle); and the author, near Bancroft, Ontario, and just east of the Ottawa River near the type locality of the Grenville series in southwestern Quebec.

The H_2S, together with some associated methane and other members of the paraffin series, may be of either sedimentary or magmatic origin. They cannot have been introduced from the surface after lithification of the marble, for both H_2S and natural gas emerge from dense, unfractured marble in mine workings as much as 3,000 feet below the surface. If the methane or H_2S are of magmatic origin, they have very selectively infiltrated certain stratigraphic zones to the exclusion of others that seem equally permeable. The limited stratigraphic range of the stinkstone beds and their persistence along the strike suggest that the H_2S may be of sedimentary or diagenic origin.

Zone 10, which is in the upper siliceous marble and stratigraphically above the stinkstone, contains abundant anhydrite and gypsum. Although interlayers of anhydrite ($CaSO_4$) occur at several widely separated horizons in the marbles of the district, the above mentioned occurrence is the most persistent and widespread known. Brown (1932, 1938) was the first worker to record occurrences of calcium sulfate in the district, or, for that matter, in Grenville-like rocks.

The $CaSO_4$-bearing rocks are generally fissured to depths of 400 feet or more and also stained with hematite and locally replaced by supergene chlorite. The anhydrite is largely altered to gypsum and veined with gypsum (Brown, 1936b; Brown, and Engel, 1956). Most of the anhydrite rock has a medium-grained crystalline texture (0.4 - 4.0 mm) and is pale pink to pale lavender and massive. It includes a trace of halite. The layer of anhydrite rock lies parallel to the relict bedding in the silicated dolomite, and represents either a true bed or a very precise replacement of one. Only a single exposure appears at low water on the eastern shore of Sylvia Lake (1.7 N., 11.75 W.), but the anhydrite in the marble beneath the lake is as much as 200 feet thick. Here, however, it has been thickened along the apex of the major crossfold that bends around Sylvia Lake. N. J. Donald (oral communication~~, 19)~~ has noted that the thickened anhydrite zone along the apex of the fold appears to explain the location and surprising depth (as much as 150 feet) of Sylvia Lake. Rapid solution of the anhydrite and gypsum accompanied by slower solution in the associated marble undoubtedly produced a major karst, which has formed the lake basin. Since the axis of the Sylvia Lake crossfold and the longest axis of the bulged anhydrite mass plunge gently north, the above explanation of the lake's origin would lead to the prediction that the lake is deeper at its north end, as in fact it is. It is as much as 150 feet deep in its northern part, but nowhere more than 100 feet deep in its southern half, and it is shallowest along the south shore.

The source of the anhydrite and associated halite is of special interest, for if these substances were of sedimentary origin, the Grenville seas must have been abnormally saline (Engel and Engel, 1953a). The anhydrite zone in the silicated dolomite has a strike length of at least 2 miles, and possibly as much as 8 miles. Anhydrite also is present in both the Balmat and Edwards zinc mines, in other stratigraphic zones. It has been cored in the footwall marble of the talc in the Arnol and Wight talc mines, and at Hyatt. Outside the district, anhydrite in Grenville-like rocks has been noted in the northeastern Adirondacks (Zimmer, 1947), near Harrisville, N. Y. (J. S. Brown, oral communication), ~~19)~~, and in Quebec (Osborne, 1944, p. 17), commonly in close association with deposits of sulfide and iron oxide ores. This relation of anhydrite to ore, together with certain textural features, led Osborne to believe that the anhydrite as well as the gypsum are of secondary origin and were formed during or after the emplacement of base-metal sulfides. But the restriction of anhydrite to the vicinity of ores may be more apparent than real. Since anhydrite is dissolved away near the surface, it is rarely seen except in drill cores and underground workings made while prospecting and mining. Some metamorphic (hydrothermal?) redistribution, as well as thorough recrystallization of the sulfates has occurred, but this has happened to all constituents of the marble during the periods of metamorphism and ore deposition. At present it seems most likely that the anhydrite was derived from sedimentary beds of gypsum containing traces of halite and sylvite.

Footwall marble (Zone 14)

Distribution, form, and thickness

The marble Zone 14 that commonly overlies the rusty marble--or overlies median gneiss where the rusty marble is absent--forms the footwall of the principal talc beds and is commonly called the footwall marble. It must be remembered that the sequence is believed to be overturned, so that the footwall marble, although it dips under the talc rock, is presumably younger. The term "footwall marble" is useful in many mines, but complications arise in the usage of this term where there is no well-developed talc belt, as is the case southwest of Hyatt (11.0 N., 2.4 E.) and northeast of Newton Hill, or where folding has produced several rudely parallel talc belts, as there are northwest of Balmat Corners. The stratigraphic equivalent of the footwall marble in the Talcville area and northeast of it is somewhat uncertain, for there are few exposures in this area, either on the surface or in underground workings. A well-defined, though, less siliceous marble that lies between the talc rock and the rusty marble and median gneiss through this area is correlated provisionally with the footwall marble to the southwest.

Southwest of the Balmat road, the structure is so complex that it is very hard to identify the footwall marble with any certainty. It seems likely that the marble separating the Fowler talc belt from the more southeasterly American talc belt consists in part, if not wholly, of the footwall marble, duplicated as shown in plate 1.

Northeastward from the Balmat road, the siliceous calcitic footwall marble is clearly defined by the underlying rusty marble and the overlying Fowler talc belt as far as the area of the Ontario talc mine, on the West Branch of the Oswegatchie (6.1 N., 3.5 W.), and throughout this area it is exposed in many places. It is cut by drill holes in the Woodcock and Ontario mines, some of which are shown on the subsurface maps (pls. 26, 30). Measurable sections completely across the footwall marble are exposed in the footwall crosscuts at the Wight mine (fifth level, 300 feet east of the shaft) and Arnold mine (eighth level, 235 feet east of the shaft).

Northeast of the underground drifts in the Ontario talc mine, in the West Branch Sand Hills area (pl. 1), there are no exposures of the footwall marble or of the adjoining units. This lack of exposures persists to the vicinity of the Hyatt zinc mine (11.6 N., 2.4 E.), where marble is exposed between median gneiss and the southwestern-most exposure of the Talcville talc belt. Between the Hyatt and International No. 2½ mines at Talcville (12.45 N., 3.8 E.), and again between the No. 2½ mine and the International No. 4 mine to the northeast (13.75 N., 5.25 E.), there are only a few scattered surface exposures of footwall marble and its form and distribution are largely inferred. Much information about its contact with the talc belt was obtained, however, from workings in the Hyatt, Wintergreen (11.8 N., 2.5 E.), and International No. 2½ and No. 4 mines, so that this contact was located with much more certainty than the contact with the rusty marble. The contact of footwall marble with rusty marble has some unusual features. Just east of the No. 2½ mine, at 12.65 N., 4.40 E., a great lens of quartz-mesh rock separates the footwall marble from the thin belt of rusty marble. Just southeast of the International No. 4 mine, at about 13.5 N., 5.75 E., an abrupt bulge in the footwall marble is well exposed. Immediately to the east, in the vicinity of Newton Hill (14.75 N., 6.4 E.) the footwall marble, the quartz-mesh rock, and the rusty marble are abruptly thinned and locally cut out, so that the talc belt lies in contact with the median gneiss. The map of Newton Hill (pl. 24) shows these relations on a larger scale than that of Plate 1.

Northeast of Newton Hill, scattered exposures of one or more thin talc-bearing zones appear 100 feet more or less above the base of the marble (as from 16.3 N., 8.75 E. northeast to 16.75 N., 9.1 E., and east of the Trout Lake road from 18.75 N., 10.75 E., to 19.0 N., 11.8 E. These occurrences of tremolitic rock are rudely alined along the strike, and they outline, very roughly, the upper limits of the footwall marble as far as the apex of the hook in the median gneiss (12.3 E., 19.3 N.).

Throughout the length of the footwall marble as defined above, there are numerous and in some places abrupt variations in its form and thickness. Between Sylvia Lake and the West Branch, there is a thick, bulging, swirled mass of footwall marble at and just east of the Balmat Road and a constriction south of the Van Amee shaft, near 5.6 N., 4.4 W. In between, however, much of the footwall marble is remarkably uniform in thickness and has a fairly straight outcrop except in the large drag fold just northeast of the Arnold mine shaft.

The footwall marble seems to be less deformed between the Wight and Arnold mine shafts than almost anywhere else in the district. Its boundaries in many places conform in considerable detail with the bedding, which is well preserved (pl. 4B). Even there, however,

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Plate 4B. ~~Near here.~~ in pocket

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many of the harder quartzose and silicated layers are broken and more or less dispersed by flow of the enveloping carbonate (pl. 4D). Any thicknesses recorded are probably only rude approximations to the original thickness. In the Wight mine crosscut, fifth level (pl. 28), the present thickness of the footwall marble is about 135 feet, but in the Arnold crosscut it is only about 90 feet (pl. 31). As this area lies on the northwest flank of the drag fold immediately to the northeast, the thinning is probably of tectonic origin. This view is borne out by the fact that fragments of the harder beds are pulled farther apart in the Arnold crosscut than at the Wight drift, where the layers seem to be partly duplicated close to the apex of the fold (pl. 1). Both northeast and southwest of the Arnold and Wight mines the footwall marble is even more deformed, so that measurements of its thickness would be still more unreliable.

Plate 4B. Diopsidic quartzite beds and laminae in the calcitic footwall marble, at the contact of this unit and the Fowler talc belt.

In the northeastern half of the district, variations in the thickness of the footwall marble are extreme: at Newton Hill there is no marble at all between the gneiss and the talc belt, but on both sides of the Hyatt mine shaft the marble forms complexly folded bulges. The bulges are due in part to changes in dip and to folds in this part of the zone. Footwall marble is perhaps most nearly uniform in thickness, and most nearly accordant in outline with the patterns of relict bedding, between Newton Hill and the apex of the Edwards hook in the median gneiss (19.3 N., 12.3 E.), especially in the area from 15.4 N., 8.25 E., to 17.0 N., 9.2 E.). There the units defined as footwall marble and Zone 12 combined, constitute in outline the approximate boundaries of a mappable stratigraphic zone about 190 feet thick. At 15.45 N. near 8.3 E. the base of the footwall marble is exposed in contact with median gneiss, and the northwest contact of Zone 12 is closely bracketed in position between uniform siliceous layers that are probably silicated dolomite.

Lithologic features

The footwall marble consists chiefly of two kinds of rock, namely (1) dolomite, in part diopsidic, and (2) quartzose calcitic marble (fig. 4). The calcitic marble grades locally into the rock designated

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Figure 4. ~~Near here.~~ in pocket

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quartz schist or quartz-mesh rock, which is discussed separately (p. 142).

Figure 4. Generalized map of parts of the Sylvia Lake-Cedar Lake marble belt, showing distribution of highly dolomitic and siliceous-calcitic parts of the footwall marble.

Between the Balmat Road and Talcville, the footwall marble is almost all quartzose and calcitic, but at the Balmat Road (1.5 to 1.75 N., 9.75 W.), and at Talcville, it grades into dolomite marble. To the southwest, very dolomitic marble separates the greatly thinned Fowler and American talc belts. Between the Balmat road and the sandy alluvium along the West Branch of the Oswegatchie River, the footwall marble is almost everywhere devoid of dolomite. A single exception is near the small talc prospect at 5.3 N., 5.25 W., where some dolomitic "islands" are associated with calcite and abundant silicates.

From the West Branch northeastward to the vicinity of the Hyatt zinc mine, the footwall marble consists mainly of highly siliceous quartz-calcite rock and includes little dolomite. Dolomite again becomes dominant, however--though again associated with calcite and silicates, chiefly altering to diopside, talc and serpentine--in the exposures along the railroad spur at Talcville (12.6 N., 4.0 to 4.3 E.), and it continues to predominate throughout the northeastern part of the district, though still accompanied locally by siliceous calcitic marble.

Within both of the major dolomitic parts of the footwall marble there are small bodies of calcitic rock which are believed to be products of dedolomitization of the general types described previously in the section on the silicated dolomite (p. 90). The calcite in these bodies either forms sheaths around grains of serpentine, talc, or tremolite, or completely fills the space between them. The calcite is that remaining when magnesium has been abstracted from dolomite to form silicates. Calcite of this origin constitutes as much as 20 percent of some of the dolomitic parts of the footwall marble. There seems to be an important genetic distinction between the calcite associated with abundant magnesian silicates and that associated with quartz in the quartz-calcite rock. In the latter kind of rock there is little or no dolomite, and quartz not only preponderates over silicates but is also more abundant than calcite; the percentage of quartz varies from 50 to 80 percent and averages perhaps 70 percent by volume. In the dolomitic parts of the footwall marble, on the other hand, talc, serpentine and diopside are much more abundant than quartz, which rarely makes up more than 5 percent of the rock by volume. The total silica content of the dolomitic rock rarely exceeds 18 or 20 percent by weight and averages perhaps 10 to 12 percent. Total K_2O plus Na_2O is less than 0.3 percent in the dolomitic rock but is about 0.8 percent in the quartz-calcite rock.

The composition of the quartz-calcite rock indicates it cannot be a product of local dedolomitization, in which magnesium-bearing silicates are formed by reaction between indigenous quartz and dolomite, leaving calcite as a byproduct. The total MgO content (about 5 percent) is far too low. The change of dolomitic marble to quartz-calcite rock requires the removal of a great deal of magnesium and the addition of a great deal of silica. Field relations strongly suggest that these large-scale alterations have actually occurred. Many lines of evidence suggest that the process may have been selective solution of $MgCO_3$ by carbonated alkali-silicate solutions of metamorphic or magmatic origin, which extracted and removed the magnesium and precipitated quartz.

The magnesia removed from the dolomite seems to have come to rest in the belts of talc rock. The footwall marble adjacent to most of the thick parts of the talc belts is siliceous and calcitic, but where the talc belts become thin or merge into silicated marble, as they do southwest of the Woodcock mine and in the Gouverneur Talc Co.'s mine, the footwall marble is dolomitic (fig. 4). One seemingly major exception appears in the vicinity of the International Talc Co.'s mines 3, 4, and 5, and at Newton Hill. At those places most of the carbonate in the footwall is dolomite, enclosing scattered grains and knots of silicates. The talc rock, however, is largely tremolitic, and contains a much lower ratio of MgO to CaO than that to the southwest, which is rich in serpentine and talc. The footwall marble in this area is clearly reduced in thickness and replaced in large part by bulbous masses of calcitic quartz-mesh rock, a process that must have freed a large amount of magnesia. These problems are dealt with in some detail in the sections on the origin of the talc rock (p. 216-225, 258-267).

As the dolomitic parts of the footwall marble are believed to be closer in chemical composition and lithology to the original sediments than the calcitic rocks are, they will be discussed first and the calcite rocks later.

Both in the areas northeast of Talcville and in the southwestern part of the district, southwest of Balmat road, large bodies of medium- to coarse-grained rock correlated with the footwall marble are commonly 70 to 90 percent dolomite, and so are the rare "islands" of dolomite in the footwall marble. Diopside, talc, and serpentine make up perhaps 10 to 12 percent of the remainder. In all the dolomitic rocks, quartz appears in the cores of knots and layers of diopside and serpentine, but it is only locally common, averaging less than 5 percent of the rock. Tremolite and forsterite are relatively rare, and no magnesite nor brucite appears here or in any of the marble.

Many of the layers, lenses, and nodules of serpentinous and diopsidic rock undoubtedly represent disrupted and reconstituted siliceous beds in the marble, but the origin of the others is obscure.

At the International No. 4 mine, at Newton Hill, and elsewhere, the silicates in the footwall marble are now largely serpentine and talc, which form disseminated splotches and occur in ellipsoidal and oblong nodules. Some of the nodules are enveloped in dolomite but have calcitic cores; others have a dolomitic core and sheaths of calcite. In these rocks, as in the silicated dolomite, the silicated clots seem to have formed by replacement or metamorphic differentiation rather than as original sedimentary features. Most of them lie along bedding or surfaces of flow, but these surfaces may have merely controlled the location of replacement or differentiation.

Rarely, quartz appears to have been introduced, though on a small scale. This quartz forms curved lenses and nodules several inches to several feet long, most of which extend along foliation surfaces in the dolomite. In many of these the quartz is coarsely crystalline and appears glassy or milky. Some masses are enveloped in dolomite and not associated with diopside or other metamorphic silicates, which indicates that the quartz in them was introduced at a stage of metamorphism in which reaction between dolomite and silica was no longer possible, but before hydrothermal activity had ceased. A little silica was probably added earlier in the footwall dolomite, as it was in other dolomitic zones, in the quartzose calcitic marble, and in the quartz schist.

In the dolomitic parts of the footwall marble, as in the dolomite of other units, little or no calcite appears where dolomite and quartz have reacted to produce diopside; but wherever the diopside is altered to serpentine or talc or those minerals replace carbonates, calcite is closely associated with them. The volume relations of calcite to silicates suggest that most or all the calcite that was formed as a byproduct of the reaction has remained in place or nearly so and that only a very little water was added and a little carbon dioxide lost. Where streaks and splotches of serpentine or talc are alined along foliation surfaces, the calcite envelopes these silicates and forms white or light-gray tongues in the darker-gray dolomite (pls. 5A, 5C).

Locally in the dolomitic parts of the footwall marble, the calcite rims around grains of serpentine or talc, or those around the sparse grains of tremolite, are disproportionately large. In these places more magnesia may have been extracted from the immediately surrounding rock than was fixed in the magnesian silicate, or $CaCO_3$ may have been added. Conversely, in scattered masses of dolomite that contain clots of serpentine, tremolite, or anthophyllite, neither cores nor fringes of calcite have been formed. There MgO has surely been introduced or CaO removed, for no magnesium-bearing mineral other than dolomite is known to occur in marbles that have undergone little or no metamorphism.

In all the examples of silication in dolomitic footwall marble described above, the oldest mineral is the dolomite and next oldest are diopside, forsterite and tremolite. These three minerals appear to be virtually contemporaneous, although a little tremolite appears to have formed as an alteration product of the diopside. Anthophyllite--a rare constituent--replaces both diopside and tremolite, and serpentine has formed at the expense of all pre-existing minerals. Much serpentine is in turn altered to talc. Calcite has formed in several generations as a byproduct of the reactions forming tremolite, anthophyllite, serpentine, and talc, and also presumably as a later vein filling (pl. 5D).

Another type of dedolomitization, involving only local redistribution of magnesia, appears in dolomitic marble that is considerably jointed and brecciated. The most noteworthy example is west of the Balmat zinc mine and just northwest of the American Talc Belt, (1.5 N., 10.2 to 10.5 W.), in what seems to be the southwesternmost dolomitic segment of the footwall marble (pl. 8A). The jointing and brecciation follow a stage of solid flow of marble, for parts of flowage folds and flow surfaces appear in the broken fragments. Some of the blocks were moved very little by the fracturing so that the foliation surfaces remain alined; in others the fragments have moved considerably, relative to one another. Pale gray to white calcite has formed along the edges of the blocks and in joints and cracks. It is readily distinguished from the dark-gray dolomite which forms the centers of the blocks. In the breccias the calcite is obviously encroaching inward from the fractures toward the cores and replacing the dolomite, a change that causes the surfaces of flow in the dolomite, as well as the joints and block outlines themselves, to blur or disappear. In many of the breccias where little of the carbonate seems to have been finely crushed before it was calcified, the pale calcite formed by replacement along fractures and prominent foliation surfaces has somewhat the appearance of a matrix (pl. 8A). In other examples a true matrix of finely granulated dolomite probably formed during the brecciation and was readily altered to calcite.

The cracks from which the calcification of dolomite began are now commonly occupied by veinlets and mammillary masses of buff to yellow-brown talc. An analysis of the talc is given in table 11 (T-1).

Table 11. Near here.

Seemingly most or all of the magnesia liberated in the alteration of dolomite to calcite is now fixed in this talc. The thickness of a talc veinlet is about proportional to the volume of calcite formed at the expense of nearby dolomite--a conclusion based on traverses made across the breccias with a measuring tape. There were probably only small changes of volume during the dedolomitization, and the principal compositional changes in the breccias, as in many other masses of slightly dedolomitized marble, probably were addition of small amounts of silica and water accompanied by modest decarbonation. Excellent examples of these calcite-talc veinlets along joints and fissures in the brecciated dolomite appear just west of the Woodcock mine in the dolomite of Zone 12, northwest of the Fowler talc belt.

Table 11. ~~Chemical and mineralogical~~ compositions of a composite sample across the calcitic, siliceous, footwall marble, of talc in dolomitic marble, and of the inferred pre-metamorphic or early metamorphic rock now altered to siliceous calcite.

	FMO	FMP	Gm 335	T-1
SiO_2	18.6	37.6	40.48	62.33
TiO_2				.01
Al_2O_3	n.d.*	0.4	.71	tr.
Fe_2O_3				.22
FeO				.12
MnO				.01
MgO	17.4	4.3	9.42	31.77
CaO	24.3	33.6	27.16	.13
Na_2O	n.d.	.13	.23	.05
K_2O	n.d.	.34	.57	.02
H_2O+	n.d.	0.2	.27	5.20
H_2O-	n.d.	n.d.	.06	.07
CO_2	38.5	19.75	20.06	tr.
Loss on ignition	n.d.	n.d.	1.09	n.d.
Cl	n.d.	n.d.	n.d.	.02
Other	1.2	3.68		
Total	100.00	100.00	100.05	99.95

Explanation of specimens:

FMO = Inferred approximate composition of footwall marble prior to metamorphism.

FMP = Present composition of calcitic footwall marble at the Arnold mine as calculated from modal analyses of its mineral constituents in thin section.

Gm 335 = Chemical composition (weight percent) of a channel sample taken across the calcitic footwall marble at the Wight mine. Celeste G. Engel, analyst.

T-1 = Pale buff, mamallary veinlets of talc cutting calcite matrix to dolomite breccia 200 yards west of the main shaft, Balmat zinc mine, Balmat, N. Y. Margaret Foster, analyst

*n.d. = no determination.

tr. = trace ($<.02$).

The formation of the breccia and joint systems in the dolomite separating the American and Fowler talc belts, and subsequent partial dedolomitization and formation of talc veinlets, must have occurred when metamorphism was complete or nearly so, for it was preceded by widespread flowage of the marble and formation of diopside and tremolite. The breccia may have developed during a stage of retrograde metamorphism in which there was widespread formation of serpentine and talc in the talc belts. During this process, however, as manifest in the talc rock and closely associated marble, some of the carbonate, serpentine, and talc was deformed by solid flow. Moreover, the talcose rock in the breccias does not look like that in the deposits of commercial talc, or like that which commonly replaces diopside in the marble, nor is it accompanied by serpentine, as is that in the talc deposits. These differences raise the possibility that the talc in the breccias was formed at a later time. In many places in the Grenville lowlands, brecciation of marble is related to much younger solutions that seem to have acted partly before and partly after the deposition of the Potsdam sandstone of Cambrian age. Most of this solution work seems to have been effected by ground waters at temperatures below 100° C.

Localities in the footwall marble at which dolomitic marble appears to merge into siliceous calcitic marble are critical to interpretations of the paragenetic and stratigraphic relations of these rocks (pl. 7C). The transition of calcitic to dolomitic marble near Balmat, just west of the Balmat road, at about 1.5 N., 9.8 W., is indicated by exposures of dolomite embayed and cut by tongues of quartz-calcite rock (pl. 7D). The calcite appears mostly along the relict bedding and flow surfaces in the dolomite, although cross-cutting them in places. The contacts of serpentinous and diopsidic dolomite with quartzose calcitic marble are exposed for some 50 to 100 feet beyond the first point at which tongues of calcite appear. Northeast of this locality at Talcville the footwall marble is virtually free from dolomite. A single exception, previously noted, is an "island" of dolomite containing talc and a little tremolite in the area southwest of the Van Amee talc mine (pl. 1). The dolomite clearly postdates calcite there, however, and also at several occurrences of calcite and quartz-mesh rock in the northeast part of the district, notably southeast of the International Talc Co. No. 4 mine (near 13.5 N., 5.9 E.) and northeast of Newton Hill (14.75 N., 7.25 E., and 15.4 N., 8.45 E.). The calcitic and dolomitic marbles merge at Talcville, although the precise relations are largely obscured by alluvial cover.

In the siliceous calcitic rock, the calcite is commonly medium to coarse grained, with grains ranging from 1/64 of an inch to 1/2 inch in diameter. Quartz usually makes up 15 to 30 percent of this rock, and locally as much as 50 percent. Diopside is less abundant than in the dolomitic segments, averaging perhaps 5 percent. Serpentine and talc are common but not dominant constituents, occurring largely as alteration products on the peripheries of diopside grains. The splotches and knots of serpentine and talc which are abundant in the dolomitic rock are more uncommon in the calcitic rock. Tremolite is more common in the calcitic than in the dolomitic marble but is everywhere subordinate to diopside. As in most parts of the marble, tremolite and diopside are almost never closely associated in large amounts, and one is commonly associated with quartz, almost or quite to the exclusion of the other.

Part of the quartz in the footwall calcitic marble forms sugary knots and ragged clusters in the calcite, and a part is mixed with diopside or tremolite in rather uniform layers and groups of laminae, which are disrupted in places. Typical examples of some of the larger clusters appear in plates 6A, 6B, and 6D. Many other clusters

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Plate 6A. In pocket ~~Near here.~~

Plate 6B. In pocket ~~Near here.~~

Plate 6D. In pocket ~~Near here~~

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are wholly quartz. Some are very small, consisting of relatively few minute grains; others are several inches in diameter. In many places the small quartz clusters tend to be alined along surfaces of foliation (pl. 6D); elsewhere they are distributed nearly at random in fairly massive marble (pl. 6A). Some clusters contain sparse grains or aggregates of diopside, tremolite, phlogopite, microcline, graphite, or pyrite; others consist almost wholly of quartz or of quartz with interstitial calcite. At several localities, notably just south of the International Talc Co. No. $2\frac{1}{2}$ and Freeman mines at Talcville, and just south of the Hyatt zinc mine, these quartz clusters increase in number and size until they merge to form quartz schist and quartz-mesh rock (pls. 8C, 8D). The textures viewed in thin sections indicate that the quartz has replaced diopside and carbonate. The contacts of quartz and carbonate are ragged and interpenetrating and whatever diopside appears has a random, highly irregular distribution. Many of the quartz clusters appear to have grown out from or formed at the expense of silicate grains and small, widely dispersed fragments of silicate layers.

Plate 6A. Siliceous calcite comprising highly deformed footwall marble north of Balmat, N. Y.

Plate 6B. Crumpled and alined quartz leaves and irregular lenses weathered into relief in a calcite matrix in the footwall marble.

Plate 6D. Highly quartzose clusters and knots alined in complex flowage folds in calcitic footwall marble.

Most of the more uniform quartzose layers (pls. 4B, 4D, and 9D) contain more or less diopside, tremolite, or feldspar. In some, diopside is as abundant as quartz, or more so, and in a few, there is more tremolite than quartz. Very uniform layers and laminae appear only rarely in the siliceous and calcitic footwall marble between the Balmat road and the Hyatt zinc mine, but they are abundant and beautifully exposed in the low ridges of footwall marble between the Wight and Arnold mines. Many of these laminae are in various states of dispersal and deformation (pls. 4D, 6C, 6D, and 9B). Where these silicate layers are least deformed, as in plate 4B, they can be seen to be relict beds in a well-defined sedimentary sequence.

Where fragments of quartzose layers are dispersed in the enveloping carbonate, at least a few of them have about the same form and composition as the typical small clusters of quartz, and locally the two forms appear to intergrade. In plate 6C, for example, most of the layers are more or less fractured, and the smaller fragments appear to be in the process of conversion to quartz clusters like those shown in plates 6A and 6D. Plate 11D illustrates another kind of transition; the smaller silicate masses in the center and right foreground are almost surely broken and dispersed laminae, yet a few are almost identical in form with the ragged quartz clusters of plates 6A and 6D. These are not clearly derived from fragments of layers, but such a derivation is suggested by their arrangement in groups and streams.

In general, however, complete gradation from uniformly laminated beds to clusters such as those in plate 11D is not common. An absence of transitional sizes and forms of bed fragments and quartz clusters is very apparent in the photographs. Moreover, in several areas of the footwall calcitic marble--for example near Talcville (12.1 N., 3.5 E.)--the aggregate volume of the clusters is far greater than that of both thin and thick siliceous layers in either direction along the strike. There also is little similarity in composition and no well-defined gradation between quartz-rich clusters such as those shown in plates 6A and 6B and bed fragments like those shown in plates 4D and 11D, which consist almost entirely of granoblastic to crystalloblastic quartz-diopside granulite. Tremolite locally takes the place of diopside, and a very little potassium feldspar may be present. Few of the layers contain less than 10 percent diopside, and the average diopside content is nearer 20 or 30 percent. In the more continuous beds and the larger fragments of beds, some evidence generally remains of layers and sheaths of diopside enveloping lenses and cores of quartz, although these relations are much blurred by subsequent redistribution and addition of silica.

There is little evidence that the widespread small quartz clusters (pls. 6A, 6B, and 6D) are fragments of quartz-diopside layers. If many of them are, the diopside in them is largely replaced by quartz and their original form and texture are obliterated by thorough recrystallization and irregular growths of quartz. The ratio of quartz to diopside in the clusters is rarely less than 25:1, and quartz is commonly the only siliceous mineral present.

In summary, the field and microscopic relations indicate the following paragenesis: beds of quartz reacted with interlayered dolomite until they were sheathed with diopside and thus cut off from further reaction. Subsequently, dolomite was dissolved, calcite precipitated, and a good deal of silica was redistributed and introduced, in part replacing diopside and carbonate. At the same time the marble was very slightly but irregularly feldspathized. All these reactions probably took place in the final stages of deformation of the marble. Small fragments of beds thus became dispersed in the marble, and along with other particles became centers for further additions of silica. Extreme silicification produced the quartz schist and quartz-mesh rock that is common alongside the footwall marble and within it (pl. 1).

The major chemical changes in the siliceous calcitic marble during its metamorphism included loss of most of its magnesia and of carbon dioxide. The substance added in largest amount was SiO_2, which was increased, on the average, by a factor of two; the composite weight percent of introduced $Al_2O_3 + Na_2O + K_2O$ is roughly 2 to 3 percent. The inferred changes are shown in table 11, where the compositions of the premetamorphic dolomite (FMO) and the existing dolomite (FMP, Gm 335) are contrasted. The nature of the inferred changes, and of those in adjoining parts of the marble, are discussed further in the sections on the origin of the talc rock belts (p. 216-225, 258-247).

Rusty marble (pyritic silicated marble, Zone 15)

Distribution, form, and thickness

Between the median gneiss and footwall marble in the area between the Balmat zinc mine and the Fullerville road (5.6 N., 13.6 W.) there is commonly a zone of pyritic, complexly silicated marble, here called the rusty marble (pl. 1). The rusty marble is not well exposed northeast of the Fullerville road, either in the West Branch sand hills or immediately east of them, or in the vicinity of Hyatt (11.0 to 11.4 N., 2.0 to 3.0 E.). Northeast of Talcville near 12.6 N., 4.5 E. there are intermittent exposures of a narrow zone of rock very similar to the rusty marble, in the correct stratigraphic position between the median gneiss and the footwall marble. This zone seems to be practically continuous throughout the northeast part of the district, at least as far as the southeastern flank of the hook-shaped fold of the median gneiss (about 1,000 feet north-northwest of Edwards), where it clearly inter-fingers with the northwesternmost layers of median gneiss.

The southwesternmost known exposure of the rusty marble is along the Balmat road just south of the Gouverneur Talc Co.'s shaft (0.8 to 0.9 N., 9.7 to 9.9 W.), where it appears to be duplicated in the apex of the large fan-shaped flowage fold outlined by the overlying marble and the Fowler talc belt. The rusty marble in this area is not underlain by any well-defined layers of median gneiss. Its sinuous form between the outcrops on the Balmat road and those southeast of the Wight mine shaft (0.9 N. to 2.85 N., 8.15 to 9.5 W.) is largely inferred from the form and structural features of the adjoining marble, especially that to the northwest. The apex of one fold in the rusty marble is exposed, however, in the woods immediately north of the Balmat zinc mine tailings pond (1.3 N., 9.4 W.). Another fold is inferred to exist about 700 feet east of this point, largely because of the form of the adjoining marble.

Northeast of the Wight mine, the rusty marble is exposed in a cross-cut in the Arnold mine and at the surface northeast of the Arnold mine shaft (4.0 N., 7.6 W.). Beyond this point the form and distribution of the rusty marble are well defined by closely spaced outcroppings of this and adjoining units almost to the Fullerville road (5.6 N., 4.0 W.).

The thinned extension of rusty marble shown northeast of this point and southeast of the Hyatt zinc mine, is purely hypothetical, for there are no exposures nor exploration data in the intervening area. None of the diamond-drill holes in the West Branch area extend through the calcitic-siliceous marble into the upper layers of median gneiss to clarify these relations. An outcrop of contorted rusty marble appears on Highway 58 (7 N., 0 E.) about in line with the Balmat fault. In this area, however, the exact geologic relations are obscure and could be interpreted in several very different ways. Several drill holes in the vicinity of the Hyatt mine were driven from marble into the median gneiss without coring any rock identified as the rusty marble, which therefore may have been pinched out or may never have been deposited in this area. Northeast of Talcville, on the other hand, the zone mapped as rusty marble is very like that in the southwest, though perceptibly thinned in its few exposures northeast of the Freeman mine (13.0 N., 4.9 E.). It again thickens abruptly, however, southeast of the International No. 4 mine (13.45 N., 5.75 to 6.00 E.).

Little if any rock of this type appears along the strike of the stretched and folded beds to the northeast until, near 8.75 E., 15.7 N., the rusty marble (or a rock very like it and apparently of the same stratigraphic horizon) reappears in a series of excellent exposures. These continue northeastward to 17.2 N., 9.4 E., where tailings from the Edwards zinc mine almost cover the bedrock. In this area a thin zone of gneiss first appears a little above the rusty marble. From 15.0 N., 8.9 E., to the vicinity of the Edwards zinc mine, a second zone of rock like the rusty marble is exposed on the southeast side of the median gneiss. The pattern is quite symmetrical (pl. 1), and seems to substantiate other evidence that the median gneiss may be duplicated in a tight isoclinal fold. This possibility is discussed in the section on structural features of the district (p. 190).

All the abrupt breaks and changes in form and thickness of the zones mapped as rusty marble at either side of the West Branch-Talcville area have probably resulted from deformation. Nearly all of the variations in thickness occur in folded or sheared parts of the zone, and conversely, those segments of rusty marble which are least folded are the most uniform in thickness.

The prominent bulges in the rusty marble just east of the Wight mine shaft (2.40 N., 8.75 W.), and near 4.25 N., 7.50 W. and 4.8 N., 6.9 W., are due to flowage and partial duplication--even triplication--of layers in tight sigmoidal folds. The folds near the Wight mine become increasingly prominent in the overlying marble and in the talc belt to the west, and those to the northeast involve adjoining units. Thickening and duplication of beds in rusty marble also are visible in the apex of the hook-shaped fold north of Edwards. Abrupt thinning of the rusty marble at 4.8 N., 6.8 W., and in the area 5.3 N., 5.1 W. to 5.6 W., and the less abrupt break around the large V-shaped fold at Newton Hill (14 N., 8 E.), and nearby, are interpreted as analogous to the squeezing and stretching of boudins and related features in individual layers (pls. 9A and 10A). All abrupt thinnings and bulges are in places where the rusty marble and adjoining units have flowed and have been so thoroughly sheared that their bedding and contacts with other zones have been blurred or obliterated.

The complicated forms of the rusty marble and associated units south of the Gouverneur Talc Co.'s shaft near 0.9 N., 9.9 W. and underground in the Balmat zinc mine are in areas where there has been profound flowage, and folding of calcite-rich marble. Although the exact pattern of disruption in the rusty marble in this area is unknown, the general relations are inferred by analogy with patterns on a smaller scale in outcrops such as those shown in plate 9C.

What is perhaps the least deformed part of the rusty marble in the area is exposed in a crosscut on the ninth level of the Arnold talc mine, 235 feet east of the shaft. In this crosscut, distinctive layers of feldspathic quartzite occur at and near the top and base of the rusty marble in almost the identical sequence exposed (1) in the outcrops near the railroad tracks 600 to 800 feet southeast of the Wight mine shaft, and (2) in exposures of the upper layers of the rusty marble in the footwall crosscut on the fifth level of the Wight talc mine, 300 feet east of the shaft. The quartzite beds, which commonly behave as the least mobile units in the rusty marble, are only moderately fractured in the Arnold crosscut, and the same is true of somewhat similar beds intercalated in the overlying footwall marble.

At this point the rusty marble measures 65 feet thick. It also measures approximately 65 feet about half a mile southwest of the Edwards zinc mine, at 16.1 N., 9.0 E., where it is little deformed.

Lithologic features

In weathered exposures the rusty marble is rusty-brown, rather friable, and faintly to distinctly layered. Most of its constituent grains range in diameter from 0.01 to 5 mm. It includes numerous interlayers, however, especially near the base and top, that are more coherent, finer-grained, and a fraction of an inch to several inches or, rarely, a foot in thickness. These are weathered into sharp relief above the more friable layers.

On weathered surfaces the rusty marble is mostly pale to deep green, buff, or pale to dark brown, but is locally gray. The thicker, more friable layers commonly contain 25 percent or more of calcite, which in places encloses brown phlogopite; stubby prisms of brown to honey-yellow tremolite; equant grains of cream-colored to green diopside; plagioclase; and a scattering of quartz and pyrite. The silicates are mostly interspersed with calcite grains; the inequant grains of tremolite and leaves of phlogopite are alined subparallel to the most prominent layering. In some exposures this type of rusty marble includes thin streaks and lenses in which tremolite, mica, or diopside predominate. Most of the finer-grained, more coherent layers contain abundant potassium feldspar, which is hard to distinguish megascopically from associated quartz.

In thin sections the rusty marble shows rather wide variations in texture and mineralogy, repeated in many widely separated outcrops. The modes of representative specimens from outcrops and mine exposures are given in table 12. Calcite is present in most specimens, commonly

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Table 12. Near here. (p. 169)

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forming between 25 and 55 percent of the rock, but locally as much as 80 percent. Where calcite is abundant, diopside tends to preponderate over tremolite, but tremolite preponderates where calcite is rare or absent. The abundance of potassium feldspar (microcline) is largely independent of the abundance of calcite. In a large number of thin sections there is a sieve-like replacement of large grains of diopside, or less commonly tremolite, by calcite or, in a few sections, by feldspar and quartz. Some of the large silicate grains are cracked, perhaps by rock movements during the replacement; and fragments of once-continuous grains are slightly bent and rotated out of crystallographic continuity. Much of the diopside is twinned and colorless. Most of the tremolite also is colorless but a little of it is green. A chemical analysis of the colorless tremolite is given in table 13 (TAH 5). The tremolite's composition reflects the relatively

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Table 13. Near here. (p. 170)

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high alkali content of the rusty marble as compared with other zones, such as the talc belts, in which the tremolites are poor in Na_2O and K_2O (tables 13 and 31). In most thin sections the calcite shows wide variations in grain size. Some of it is mixed with serpentine or talc in masses, which appear to replace diopside, tremolite, and sparse scapolite.

Table 12. Mineralogical composition in volume percent of layers of rusty marble in the Balmat-Edwards district.

	S79	S30A	S351	S702	CBSE	S13	S352	SARM	EF11	EFRM	AH-1	N18	N10	S18	S20	S35	S54	S59	N19	SBS	N30	HMW	Av.
Calcite	30	32		52	48		40	31	32	29	37	41	50	80	79	34	56	28	41	53	26	7	41
Diopside	32	25	6	19	22	5	15	30	14	32		21	20	3	5	14	14	22	25	5	25	74	18
Tremolite		5	53	green 14	green R	45	10		3		29	5		R			3		green 10		2		8
Quartz	12	2	8	5	4	15	8	11	12	4	4	R	2	1	2	14	2	5	1	40		2	7
K-feldspar	13	4	28	9	21	35	19	17	27	3	20	30	15	9	6	36	23	38	23	2	44	10	19
Plagioclase (Oligoclase-Andesine)	2	6			1		7	8	9	9	9		9	6	4			7≠					3
Scapolite	7					R		R	R	23													1
Phlogopite	1	23			2		1	2					4	.5	4		1	R				5	2
Sphene	1	R		1				R	2	R		R		R		R	.3	R	R	R	2	.2	
Tourmaline	2			R											R								
Zircon	R						R			R						R	R						
Pyrite		3	5	R	2			1	1	R	1	3		.3	R	2	.5	R	R	R	1	2	1
Apatite	R	R		R	R			R			R			R	R	R	R		R				
Totals	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100

R = rare (.05)

≠ = Albite

Table 13. Chemical composition of average samples of rusty marble, siliceous/calcitic Footwall Marble, adjacent talc belts and tremolite from the rusty marble.

Oxides	RMA*	Gm 338	Gm 335	T 4	Gm 356	TAH 5	T 15
SiO_2	36.3	47.10	40.58	59.80	59.40	55.78	57.86
TiO_2		.17		tr	n.d.	.82	.05
Al_2O_3	4.4	6.66	.71	.57	.74	1.79	.64
Fe_2O_3		.76		.05	.02	.51	.00
FeO		.86		.15	.12	.54	.14
MnO	.3	.07		.39	.20	.03	.06
MgO	6.0	10.09	9.52	27.25	30.09	23.10	24.73
CaO	28.2	19.38	27.16	6.70	4.94	8.45	13.27
Na_2O	1.2	1.85	.23	.10	<.1	4.92	.29
K_2O	2.1	3.40	.57	.15	<.1	1.86	.09
P_2O_5	.06	.01		n.d.	n.d.	.06	
F		.1		n.d.	n.d.	n.d.	.26
Loss on ignition		9.21	21.09	4.75	4.09		
H_2O+	.03		.27	.44	.47	2.32	2.38
SO_3				.07	.01		
CO_2	16.7		20.66	1.18	.31		
Total	95.29	99.66	100.13	100.02	99.90	100.18	99.77

Elements (parts per million/)

	RMA	Gm 338	Gm 335	T 4	Gm 356	TAH 5	T 15
B		.04	.0007	.02		.07	.02
Ba		.09	.04	.05		.0007	.0001
Zr		.001	.0006	.0001		.004	.0002
Cu		.002	.0003	.001		.0005	.0004
Mn		.05	.08	.3		.02	.3
Ni		.0004	.0002	.0003			.001
Pb		---	.02	.002			
Sr		0.1	.05	.02		.01	
Ti		0.1	.005	.01		.02	.04
Y		.005	.0007				

*More-feldspathic layers omitted.
/Spectroscopist Arthur Chodos.

RMA. Very approximate chemical composition of the rusty marble calculated from the mineralogical mode of an average sample collected southeast of the Arnold talc mine.

Gm 338 Channel sample across the rusty marble, collected in the shaft of the Gouverneur Talc Co. mine, Balmat, N.Y. Celeste Engel, analyst.

Gm 335 Channel sample of exposed calcitic, siliceous, footwall marble collected in the area southeast of the Wight Talc mine, Fowler, N.Y. Celeste Engel, analyst.

Gm 336 Channel sample across the Talcville Talc belt as explored in the International Talc Co. 2 1/2 mine, Talcville, N.Y. Glen Edgington, analyst.

Gm 356. Channel sample across the footwall zone of talc in the Wight mine, Fowler Talc belt, Fowler, N.Y. Glen Edgington, analyst.

TAH 5 Tremolite from the rusty marble, as exposed on the Balmat road (9.7 to 10.0 W at 0.75 to 1.0 N). Celeste Engel, analyst.

T 15 Tremolite from the Fowler talc belt, Arnold mine, Fowler, N.Y. Eileem K. Oslund, analyst.

Quartz constitutes, on the average, slightly less than 10 percent of the rusty marble, but makes up as much as 40 percent of a few layers and lenses. All the quartz is fine-grained and intimately associated with potassium feldspar and plagioclase. Most grains of potassium feldspar show faint to well-developed gridiron twinning, but perthitic feldspar is locally common. Most of the plagioclase is about An_{15}, although in some specimens the feldspar is more calcic than andesine.

The finer-grained coherent layers found especially near the base of the rusty marble approach and merge in mineralogical composition with the feldspathic layers of the median gneiss (pl. 11C). Typical layers contain 40 to 60 percent quartz, 30 to 40 percent microcline, and scattered biotite and scapolite. Others found very locally approach the pyritic schist (table 7) in appearance and composition. They contain abundant untwinned potassium feldspar, a fair amount of quartz, some tremolite, and more pyrite than is commonly found in the rusty marble.

The accessory minerals include sphene, tourmaline, zircon, pyrite, and apatite. Scapolite occurs locally, largely at the hook in the Edwards fold (18.65 N., 11.9 E.), where there seem ~~s~~ to be layers of rusty marble on both sides of the median gneiss. The tourmaline is of three varieties. One is a bluish-gray, another is drab-green, and the third is the brown magnesian tourmaline dravite, which is also found sporadically in the talc belts and in other zones of marble.

The rusty marble appears to have been derived from argillaceous sandy dolomite, intermediate between the clean siliceous dolomite ancestral to the footwall marble and the argillaceous sandstone from which the median gneiss was probably derived. Some light may be thrown on its origin and evolution by comparing available analyses of the rusty marble with analyses of marble from nearby older zones. In table 13, for example, a chemical analysis of rusty marble and the average composition as computed from its mineral constituents, are listed alongside an analysis of siliceous calcitic footwall marble and two analyses of talc rock. Each of the chemical analyses represents an average sample across the entire zone of rusty marble. Analyses Gm 338, Gm 335, and Gm 356 are, respectively, analyses of rusty marble, of siliceous calcitic footwall marble, and of the lowermost part of the Fowler talc belt, in the immediate vicinity of the Wight talc mine.

The relatively high Na, B, Sr, Y and Zr contents of the rusty marble strongly indicate additions to this zone from metamorphic or magmatic fluids. Argillaceous dolomites seldom if ever contain as much as 0.04 percent B, 0.1 percent Sr, or 0.001 percent Zr. The relatively high K_2O content is common in argillaceous dolomite, but 1.85 percent Na_2O and a $Na_2O:K_2O$ ratio of 0.6 are not.

Introduction of appreciable silica and alkalis also is suggested by the fact that the rusty marble contains pegmatites and many veinlets of feldspar, both of which increase in abundance toward the contact with the migmatitic median gneiss. It is also noteworthy that scapolite and tourmaline are most abundant and widespread in the rusty marble at the Edwards end of the district, which is nearest to the igneous rich, granitic massif of the central Adirondacks and contains the most highly metamorphosed rocks in the district (Engel and Engel, 1953a).

Both the silica content and the total alkali content of the rusty marble seem to have been doubled, perhaps tripled, during metamorphism, but this estimate would be very difficult to prove.

The origin of the abundant pyrite in the rusty marble is not clear. There is far more pyrite in the rusty marble than in the other marble zones but far less than in the pyritic schist. The distribution of the pyrite in the rusty marble is fairly uniform, but its abundance varies locally both across and along the bedding. In this respect the rusty marble resembles the pyritic schist. Very irregular patches of pyrite commonly occur in the quartz-biotite gneiss near its contacts with the pyritic schist and in the median gneiss near its contacts with the rusty marble. In fact, most of the gneiss and schist adjoining or interlayered with both the rusty marble and the pyritic schist contain pyritic patches, and this somewhat spotty distribution of the pyrite, as well as its paragenetic relations, suggests that if it is not secondary it has been redistributed and concentrated locally by secondary processes.

Although the zinc ore deposits represent local concentrations of pyrite that were definitely emplaced during metamorphism, both the pyritic schist and the rusty marble contain larger, more uniform and more persistent concentrations of pyrite than any other units in the district. Unlike the other marble units, both are derived from sediments that contained much clay and some detrital quartz--the schist, of course, much more than the rusty marble. Both are now injected and feldspathized, the schist much more so than the marble. Both contain skarnlike bodies of the so-called "contact metamorphic" silicates. It is also true, however, that except for the pyrite the premetamorphic composition of the pyritic schist appears to have resembled that of the median gneiss, which is pervasively feldspathized and injected lit-par-lit, but not markedly pyritic.

Quartz schist and quartz-mesh rock

Distribution, form, and thickness

Several large lenses and zones containing irregularly foliated, bulbous masses of quartz mapped as quartz schist and quartz-mesh rock, are interbedded with the marble of the Sylvia Lake - Cedar Lake belt. Only the largest and most distinctive of these quartz masses are shown on plate 1, but many smaller masses are scattered through the marbles of the district.

One of the largest masses of quartz schist lies just northeast of Talcville and is well exposed in the rounded ridges that extend northeastward from Talcville along the north side of the New York Central railroad track almost to Newton Hill (13.75 N., 6.45 E.). This bulbous zone of quartz schist lies partly within the rusty marble and partly between it and the dolomite of the footwall marble. Its western part swells, at least locally, to a thickness of 300 feet or more and almost wholly replaces the two adjacent zones of marble. It is much constricted at several places southeast of the International Talc Co.'s No. 4 mine, and west of Newton Hill it narrows abruptly to about 25 feet. On Newton Hill it forms thin discontinuous lenses, few of which appear to be more than 25 feet thick.

What seems to be a fairly persistent belt of quartz schist as much as 100 feet thick reappears in this general lithologic zone northeast of Newton Hill, between 15.6 N., 8.6 E., and 17.1 N., 9.3 E. How far this belt continues under the Edwards mine tailings pond is unknown, but there is very little quartz schist along the base of the footwall marble at the Trout Lake road. A few pod-like masses are exposed, however, east of the road, for example at 18.7 N., 11.75 E.

Southwest of Talcville, in the vicinity of the Hyatt zinc mine, lenses of quartz schist are exposed at intervals along and within the footwall marble, the largest and most distinctive being in the low knolls south of the railroad track and southwest of the tailings pond, near 11.0 N., 1.3 E. Several of these lenses are more than 200 feet thick and appear to extend across several zones of marble into the silicated dolomite. None seem to persist along the strike for more than a few hundred feet. Several pinch out abruptly. Others grade into calcitic marble that contains clusters of quartz. West of the Hyatt mine at 11.9 N., 1.3 E., quartz schist enclosing fragments of pyritic schist occurs near the top of dolomite Zone 1. Underground in the Hyatt zinc mine there are many additional exposures of the quartz schist. These masses seem to conform in a general way to the foliation and bedding of the enveloping marble, although they cross them locally and in detail.

Somewhat more irregular or complex masses of quartz are exposed just southwest of Balmat Corners, on the crest and south side of the low rounded hill near 1.0 S., 9.7 W. This locality is immediately north of the Balmat fault zone, in its hanging wall. The outlines of the quartz masses appear to be at least partly discordant with the bedding. Somewhat similar though less obviously discordant bodies of quartz schist lie at and near the contact of the silicated marble and dolomite Zone 3 southwest of Sylvia Lake (2.5 S., 12.4 W.). Many smaller bodies of quartz and quartz-diopside rock are exposed in and along the silicated marble north and east of Sylvia Lake, and also in dolomite assigned to Zone 3 and in silicated marble, northeast of the West Branch sand hills (8.5 to 10.0 N., 1.0 W. to 1.0 E.).

Lithologic features

Most of the quartz schist forms lenticular masses made up of somewhat irregularly crumpled layers (pl. 8C and 8D). Its dominant component is medium- to coarse-grained quartz (grain size 0.1-10.0 mm), in some places glassy, elsewhere milky and of sugary texture. Streaks and lenticules of calcite mixed with diopside are intercalated with the quartz in most exposures. Some of the quartz schist is feldspathic and at several places this feldspathic facies merges abruptly into coarse-grained quartz-microcline pegmatite. Scattered masses of uniformly layered and laminated quartz-diopside-feldspar granulite also occur within the quartz schist and on its margins. Quartz schist of this type locally grades into small masses of rock consisting of quartz mixed with diopside and carbonate, similar to the "quartz mesh" rock in the Canton quadrangle, described by Martin (1916, p. 23-26 and pl. 3). In these masses the quartz commonly forms irregular tongues that extend both across and along the foliation and consequently stand out on weathered surfaces in a meshlike pattern. The more deeply incised openings are made by the weathering out of carbonate, chiefly calcite, together with some diopside and tremolite. Quartz-mesh rock grades into quartz schist in places.

In most of the quartz schist northeast of Talcville, the component minerals are alined along somewhat irregular, highly crumpled layers and lenticules a small fraction of an inch to an inch thick. The axes of the crumples tend to be subparallel and to accord in plunge with axes of larger folds in the adjoining marble and gneiss. Calcite forms small disconnected blebs and pockets containing quartz and scattered diopside. Fifteen thin sections from specimens of this type of quartz schist collected between Talcville and the Trout Lake road contain 65 to 95 percent quartz, 2 to 15 percent calcite, 1 to 11 percent diopside, and random streaks of potassium and plagioclase feldspar, tremolite, and pyrite. Scattered grains of sphene, zircon, and apatite are present in some slides. These slides give only a rough idea of the normal proportions of the minerals because of the coarse and irregular texture of the material. They do show, however, that in spite of the crumpled appearance of the rocks, most of the quartz is unstrained and its outlines are gently rounded and only slightly embayed. The diopside forms both rounded grains and stubby blades; some of the grains are cracked and partly replaced by calcite or serpentine. The tremolite tends to occur in folia that enclose little or no diopside, but where these two minerals are associated, a few blades of tremolite seem to replace diopside.

In the ridges of quartz schist northeast of Talcville, lenses of calcite as much as 15 feet thick occur as intercalations, and equal or locally greater thicknesses of calcitic marble form partial fringes around masses of schist separating it from dolomite in Zone 14. Tongues of calcite from these fringes embay and project into the enveloping dolomite, from which the calcite is clearly derived. Most of this calcite is siliceous, and indeed there are all degrees of silicification of marble between quartz schist and the siliceous calcitic footwall marble (p. 111). Some of the best exposures of the transitional rocks are in the International No. $2\frac{1}{2}$ mine at Talcville and on the surface just south of it. The quartz clusters in the marble there grade into more elongate rodlike masses which in cross-sections resemble crinkled lenses and sheets (pls. 6B and 9B). The axes of the crinkles and the longest axes of the knots or rods lie parallel to the apices of crinkles in the nearby quartz schist. As the proportion of quartz increases, adjacent crinkles merge to form irregularly foliated quartz schist containing intercalated layers of calcite.

Similar relations may be observed in the vicinity of the Hyatt mine, although in that area very few relicts of dolomite remain in the marble, and much pyrite and a little sphalerite and galena occur in the quartz schist, where they appear to replace diopside, tremolite, carbonate and quartz.

Feldspathic quartz schist is exposed in the Hyatt mine and also in close association with a coarse-grained pegmatitic quartz-feldspar rock on the low ridge south of the International No. 4 mine near 13.45 N., 5.70 E. Thin sections of specimens from the latter exposure average roughly 60 percent microcline, 5 percent sodic oligoclase (An_{12}) and 35 percent quartz. In the coarse-grained parts of the exposure some of the quartz grains are as much as an inch across. There is no pronounced foliation. The quartz is partly glassy and partly milky. This pegmatitelike body has irregular margins with ragged tongues which interpenetrate and grade into the irregular folia of the surrounding quartz schist. Quartz schist and pegmatite in the silicated dolomite are associated in much the same way southwest of Sylvia Lake (2.35 S., 12.50 W.).

The quartz schist seems best explained as a product of replacement of marble by introduced silica. The quartz schist lenses appear abruptly in the marble, and the random patterns of some of these lenses, such as those at the Hyatt mine, are discordant with the bedding in the marble. The calcite lenticules which invariably occur in the quartz schist and the calcite fringes on its margins are derived from dolomite that does not contain siliceous beds comparable to the quartz schist. Most of the quartz schist is therefore interpreted as a result of extreme silicification of dolomitic marble that contained much less quartz, though some of it appears to have been formed by addition of silica to marble that was already highly quartzose.

The extremely crenulate patterns in much of the quartz schist must have resulted from flow of the marble before and during the replacement process. The stratigraphic position occupied by the quartz schist northeast of Talcville, at the base of the footwall marble, appears to be occupied normally by laminated siliceous beds. The very uniformly laminated quartz-diopside granulites found locally in the present quartz schist may have been derived from such beds.

The question whether much of the silica has been derived locally from siliceous beds in the marble seems especially pertinent with regard to quartz schist and quartz mesh rock exposed along the Balmat fault at Balmat, and in the silicated dolomite around and especially southeast of Sylvia Lake. There is good reason to believe that such large masses did not originate in this way; the amount of silica in them appears too great to have been derived from the surrounding marble. These large masses of quartz schist and quartz-mesh rock along major fault and shear zones and those associated with pegmatite are clearly not segments of stratigraphic zones. They seem instead to be related to channelways along which silica was carried in large volume. The Balmat fault, for example, locally cuts across the persistent strata of pure dolomite and silicated dolomite south of Sylvia Lake, and masses of quartz-mesh rock and pegmatite are distributed along the fault zone independently of the lithology of the adjacent wall rocks.

For these reasons Martin's conclusion (1916, p. 23) that the quartz mesh rock is of sedimentary origin seems invalid in the Balmat-Edwards district.

The great influx of silica believed to have been instrumental in forming both the quartz schist and the quartz-mesh rock seems to have occurred during several successive stages of metamorphism. The fact that diopside is associated with some of the quartz indicates that this silica was introduced at periods when diopside was forming or could persist as a relatively stable mineral. As in the calcitic part of the footwall marble, however, tremolite, calcite, and quartz replace some of the diopside, and some of the latest additions of silica appear to be contemporaneous with the formation of talc.

Local encroachment of quartz schist containing alkalis as well as silica into the base of the footwall marble is expectable, inasmuch as this schist is separated from the feldspathic median gneiss only by the relatively thin rusty marble. The feldspathic facies of the quartz schist were presumably formed by the same emanations, carrying silica and alkalis, that formed the migmatitic median gneiss, the pegmatites, and pegmatitic parts of the intervening rusty marble. These fluids also seem to have migrated to the southeast side of the median gneiss and have formed scattered bodies of feldspathic quartz schist along the contact with the marble of the Fullerville and Edwards belt.

Median gneiss (feldspathic granulite)

Distribution, form, and thickness

The layer of gneiss that separates the Sylvia Lake - Cedar Lake marble belt from the Fullerville - Edwards belt is called the median gneiss (Brown, 1936a, p. 238). The median gneiss is easily recognized and fairly well exposed; and because it almost everywhere lies within a few hundred feet of the commercial talc belts, it serves as an excellent reference horizon for geologic studies and explorations in the district. This rock has been distinguished in various forms on three earlier maps of the district (Cushing and Newland, 1925; Brown, 1936a; Gilluly, 1945). Brown gave the unit its name because of its apparent position as a distinctive interlayer separating the Sylvia Lake - Cedar Lake marble belt from the Fullerville - Edwards belt. Gilluly (1945), however, concluded that the gneiss in the southwestern part of the district is not at the same stratigraphic horizon as that to the northwest and mapped them respectively as "dark-colored" and "light-colored" gneiss. He shows an appreciable gap between the two in the area just northwest of the junction of the old Fullerville road and Highway 58 (6.5 to 7.5 N., 0.5 E. to 0.5 W.), but exploration drilling in this area has failed to show that any such gap exists.

The form of the gneiss in this area and farther northeastward may be much more complex than has heretofore been supposed. Possible complications in its form are indicated on plate 1, and also on figure 7, which is an attempt to diagram the inferred development of the structural features in the district. Those in the median gneiss are discussed on page 190. Briefly, however, what might seem to be a single thickness of median gneiss north of Fullerville near 7.0 N., 0.0 E., and thence northeastward to the northeast boundary of the district, may be gneiss duplicated in a very tight isoclinal fold. If this interpretation is correct, the marble of the Fullerville - Edwards belt is in reality a highly deformed extension of the marble of the Sylvia Lake-Cedar Lake belt, and the median gneiss a simple intercalation only in the area between Balmat Corners and the vicinity of the road to Fullerville (7.0 N., 0.0 E.).

Unfortunately, the only exposures of gneiss between Fullerville and a point about 1,200 feet southwest of the Ontario talc mine (5.45 N., 4.0 W.) are on and near Highway 58 (6.4 N., 2.2 W.). The form and distribution of the median gneiss shown on plate 1 are inferred from these exposures and from the logs of several drill holes to the north and east. Except in this area of very sparse exposures, the median gneiss is readily followed to the southwest as far as 2.3 N., 8.4 W., about 1,000 feet east-southeast of the Wight talc mine. Throughout this stretch it is bordered on the northwest by rusty marble and on the southeast by (a) thin, disconnected lenses of marble younger than the gneiss, (b) the large bulbous mass of amphibolite south of Fowler and (c) locally, east of the West Branch of the Oswegatchie (6.2 N., 2.2 W.), lenses of quartz-mesh rock. At 4.4 N., 6.85 W. the gneiss appears to be greatly thinned or even lacking. Elsewhere in this belt the width of outcrop of the gneiss appears to range from 30 feet to as much as 500 feet; this maximum is reached at 5.25 N., 4.4 W. In general, the zone thins to the southwest from the area south of the Van A~~l~~mee talc shaft (5.25 N., 5.0 W.) to the outcrop east of the Wight mine. Throughout this area the gneiss is almost surely a homoclinal unit at the top of the marble of the Sylvia Lake - Cedar Lake belt.

Southwest of the Wight mine and the Balmat mine (0.45 N., 10.25 W.) no surface exposures of the median gneiss are known. However, layers of gneiss that may represent this unit are cut by drill holes and underground mine workings in the area just east of the Balmat road and southwest of the most southwesterly surface exposure. If this rock is the median gneiss, it must form fragments embedded in siliceous calcitic marble, and is far removed from the rusty marble which normally adjoins the gneiss to the northwest. Such tectonic dispersal of fragments is typical of most stratigraphic units here. Presumably any major underground extensions of the median gneiss have been doubled back to the south and east and abut on or extend southeast of the large masses of amphibolite and granite northeast of Balmat Corners (figure 7).

Northeast of the sand hills along the West Branch of the Oswegatchie River, rock that is mapped as median gneiss appears in numerous exposures almost to the northeast boundary of the district north of Edwards (18.4 N., 12.4 E.). The width of the zone of gneiss defined by these exposures commonly ranges from 800 to almost 1,600 feet even though the dips are moderate to steep. Thus the gneiss appears to be greatly thickened. Just to the southwest, a marked anticlinal fold is well exposed just south of the intersection of the old Fullerville road with Highway 58 (7.0 N., 0.0 E.). It is therefore possible that the apparent thickening of the gneiss is due to a tight isoclinal fold of the type illustrated in figure 7 and in Brown and Engel, 1956, figure 1.

In addition to this widening of its outcrop, certain stratigraphic features of the gneiss to the northeast also suggest the presence of a major isoclinal fold. Northeast of Newton Hill, for example at 15.0 N., 8.9 E., and 15.7 N., 8.65 E., two thin streaks of pyritic silicated rusty marble are exposed. One is in the normal position of the rusty marble at or near the northwest side of the median gneiss. The other lies near the southeast side of the gneiss and may indicate duplication of the section in a refolded isocline. Another streak of marble at a different stratigraphic horizon appears in the gneiss east of the Edwards zinc mine and between two interlayers of pyritic silicated marble at 18.4 N., 11.5 E. This third marble is chiefly dolomite but contains moderate amounts of calcite and diopside and could be stratigraphically above the marble of the Sylvia Lake - Cedar Lake belt preserved along the axial plane of the fold in the gneiss.

If this segment of median gneiss has really been doubled up, the fishhook fold immediately north of Edwards is a folded fold, as indicated in figure 7. The extent of median gneiss beyond this area is very uncertain. The fact that an abrupt fold apparently affects the one rusty marble at 17.7 N., 12.0 E. suggests that the median gneiss also may be folded back northeast of that locality, and so does the presence of a small fragment of similar gneiss at 18.4 N., 12.4 E. Thin contorted and fractured slivers of gneiss resembling the median gneiss are enclosed in the marble of the Fullerville - Edwards belt, especially southwest of Edwards in the areas 13.5 to 14.3 N., 11.4 to 11.75 E.; 11.5 to 12.0 N., 9.4 to 9.8 E.; and south and southeast of Talcville, especially in the vicinity of 9 N., 4 to 5 E.; 9.5 N., 7.0 E.; and 11.75 to 12.75 N., 5.75 to 6.7 E.

The possible stratigraphic and structural relations of these widely separated slivers of gneiss are indicated in figure 7 and are discussed on p. 190 in the section on "Structural features of the Balmat - Edwards district."

Lithologic features

Most exposures of the median gneiss are distinctly layered, commonly gneissic although locally almost schistose rocks. They vary in color from pale buff, gray, and pink to rusty brown and even black. In general, the lighter colored rock is sugary-textured, equigranular, quartz-microcline gneiss or granulite (tables 14, 15). The intermediate

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Table 14. Near here. (p. 189)

Table 15. Near here. (p190)

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and darker colored parts of the gneiss are predominantly of two types, one biotitic and the other amphibolitic (table 15). The biotitic gneiss is chiefly composed of biotite, quartz, oligoclase, and microcline in various proportions. The amphibolitic gneiss is chiefly composed of hornblende, biotite, andesine, and quartz but contains some potassium feldspar (table 15); some also contains diopside, scapolite, and chlorite. The rusty-brown to yellowish-black exposures generally represent weathered pyritic facies of the gneiss.

Table 14. Modal Analyses of median gneiss

Biotitic layers

Specimen No.	S45-2	N88	S21-1	N11	S24-3	N62	N60	N16	N78	N68	N3	Av
Quartz	8	43	33	40	45	13	22	11	15	14	18	23
Plagioclase	*12	9	34	15	35		7	5	12	16	21	17
Microcline	68	35	*17	32	4	66	46	60	47	38	28	40
Biotite	9	10	11	12	15	20	24	24	26	32	33	19
Hornblende			4				1					
Pyrite	2	2		1	R	R	R	R	R		R	R
Sphene	R	R	1			R						R
Apatite	R	R	R	R	R	R	R	R	R		R	R
Tourmaline	R			R	R	1	R	R	R	R	R	R
Zircon	R	R	R	R	R	R	R	R		R	R	R
Garnet			R									
Scapolite												
Calcite												
Zoisite						R					R	
Diopside												
Muscovite	R	R	1.2	R	1	R	R	0.5	R	R	R	R
Total	100	100	100	100	100	100	100	100	100	100	100	99

Amphibolitic layers

Specimen No.	N3A	N7S	N8B	N17B	Av
Quartz	10	42	17	21	22.5
Plagioclase	2	3	16	32	13.2
Microcline		30	5	7	10.5
Biotite		R	17	28	11.2
Hornblende	22	13	38	9	20.5
Pyrite	R	1	R	1	.5
Sphene	1	R	1	1	.8
Apatite		R	1	1	.5
Tourmaline	R	R	R	R	
Zircon					
Garnet					
Scapolite	30	11			10.2
Calcite	4	R	2	R	1.5
Zoisite					
Diopside	31		3	R	8.5
Muscovite					
Total	100	100	100	100	99.9

Granulite layers

Specimen No.	N76B	N86	N76A	N76C	N49	N31	N23	N13	N67	N12	N14	N17	N5	Av
Quartz	40	48	48	50	50	50	50	53	54	55	57	63	64	52
Plagioclase	7	3	6	2	4	6	R	2	R	5	2	5	1	4
Microcline	50	44	40	44	46	45	46	40	40	37	37	+30	27	40
Biotite	2	4	4	1				3	2	2		R		1
Hornblende														
Pyrite	1	1	1	2	R	R	2	1	1	1	2	1	2	1
Sphene		R	R				R	R		R				
Apatite	R	R	R	R	R	R		R	3	R		R	4	
Tourmaline	R		R		R							R	R	
Zircon						R	R	R	R		R	R	R	
Garnet														
Scapolite														
Calcite														
Zoisite														
Diopside														
Muscovite	R	R	1	1	R	R	2	1	R	R	2	0.5	1	1
Total	100	100	100	100	100	100	100	100	100	100	100	99.5	100	99

* = Oligoclase

\+ = Untwinned

R = Rare

Table 15. Chemical analysis and mode of quartz-microcline granulite from the median gneiss. Specimen taken from exposure on Talcville Road just south of the bridge across the Oswegatchie River (12.1 N, at 4.02 E).

Chemical Analysis		Mode	
Oxide	MG1	Mineral	MG1
SiO_2	80.76	Quartz	50.3
TiO_2	.04	Microcline	38.5
Al_2O_3	11.11	Oligoclase-albite (Ab 90) (sericitic)	9.3
Fe_2O_3	.21	Biotite (chloritic)	1.4
FeO	.12	Magnetite	.2
MnO	n.d.	Pyrite	.3
MgO	.01	Total	100.0
CaO	.29		
Na_2O	2.18		
K_2O	5.04		
P_2O_5	n.d.		
H_2O+	.08		
H_2O-	.01		
Total	99.84		

Celeste G. Engel, analyst.

No very systematic stratigraphic distribution could be demonstrated for any of these kinds of gneiss. Several may be interlayered in a single large outcrop, and gradations between one kind and another are common. Each gradation can be well observed east, south, and southeast of the Arnold talc mine, and also in the fishhook fold north-northwest of Edwards. In general, the quartz-microcline granulite is predominant, as, for example, in the exposures southwest of the Ontario talc mine (5.0 to 5.5 N., 4.0 to 5.0 W.), southwest of the Hyatt mine (near 1 E., 9 N.), along the new road to Talcville (near 12 N., 4 E.), and west of Edwards (15.0 to 17.0 N., 9.0 to 10.5 E.). The biotitic rock is intermediate in abundance, and the amphibolitic rock is generally subordinate, forming only thin lenses or layers.

All these rocks have commonly been injected lit-par-lit or replaced by gray to pink seams and veinlets of pegmatite and granite. The permeation of the layers parental to the quartz-microcline rock is inferred to have been a very pervasive feldspathization (pl. 10C), although lenses, eyes, and braids of quartz and potassium feldspar extend along the dominant foliation (pls. 10B, 10D), and along obviously secondary or tertiary surfaces (fig. 10). Some of the quartz-microcline layers appear to have been especially ductile during this granitization (pl. 10C), whereas some of the amphibolite layers remained brittle and broke into boudins.

The more biotitic layers, especially those intercalated with amphibolite or quartz-microcline rock, were highly susceptible to lit-par-lit injection and shredding, whereas the more amphibolitic layers were generally migmatized unevenly, except where they were strongly sheared. In places the pegmatitic and more obviously granitic layers contain visible crystals of black tourmaline; for example, northeast of Talcville (at and near 13.3 N., 5.7 E.), in the fishhook fold north of Edwards, and east of the Wight mine (2.3 N., 8.4 W.). Fine-grained tourmaline is also an accessory constituent in many layers of typical granulite.

Pink garnet occurs, though very rarely, in the more migmatitic parts of the gneiss, especially northeast of Talcville and in the fishhook fold near Edwards. In this fold, but rarely elsewhere, sillimanite occurs in the highly migmatitic and schistose biotitic gneiss, and some granitic stringers in the granulite contain augen of quartz and sillimanite. Garnet, sillimanite and scapolite are extremely sparse or absent in exposures of the median gneiss southwest of 16.0 N. This appears to be due in part to the presence of layers of different composition in the gneiss near Edwards, but in addition the gneiss there has been more intensely sheared and recrystallized at higher temperatures than that to the southwest (Engel and Engel, 1958a). The two thickest marble interlayers in the gneiss at this place consist of the pyritic silicated rusty marble (p.133). There are smaller lenses of calcitic marble in the gneiss north of Edwards, south and southwest of the Hyatt mine, and possibly east of the Arnold mine; these lenses may well have had a greater extension along the strike prior to dynamic metamorphism.

At many contacts of median gneiss and marble, pale-green to cream-colored diopside and feldspar are the major constituents of the granulite. Not only the granulite but the biotitic and amphibolitic layers may contain augite as well as diopside, and they all contain more scapolite and sphene near the contacts than they do elsewhere. The marginal layers of granulite appear to have been highly siliceous and slightly calcareous sediments, now modified by the fluids that accomplished the migmatization.

Because the several major types of gneiss are intimately interlayered, no attempt was made to map them separately on plate 1. At several places, notably between the Fullerville road and the Wight mine, and especially between 5.0 N., 6.0 W. and 2.4 N., 8.4 W., most of the gneiss is biotitic or amphibolitic. In this area the gneiss and the gabbroic amphibolite which lies immediately to the southeast grade into each other. The gabbroic amphibolite appears to cut across the southernmost layers of gneiss. Clearly intrusive relations, however, are nowhere visible. The contact zone is granitized, highly sheared and foliated, and the foliation surfaces in the two rocks converge at very small angles. The rocks are partly granoblastic, partly mortar gneisses. A strong lineation, marked by parallel arrangement of hornblende prisms and elongate grains and clots of quartz, feldspar, and biotite, plunges northward parallel to the axes of most of the obvious folds in this area (pl. 1). Many of the highly biotitic, plagioclase-bearing lenses and layers in this area may be granitized amphibolite. In some places amphibolite layers even grade into coarsely porphyritic pink granite gneiss similar to Buddington's Hermon type of granitic gneiss (1939), which contains relict shreds of amphibolite.

Other amphibolite lenses and layers within and marginal to the granulite are exposed east of Edwards and northeast of Newton Hill and appear to have been derived from either tuffaceous or calcareous shaly sandstones. This conclusion is based upon the fact that these layers grade into diopsidic marble and contain diopside, sphene, and scapolite. None of these features is regarded as convincing proof of such an origin, for similar amphibolite occurs at the schistose and granitized margins of many bodies of mafic igneous rocks cutting marble in the Adirondacks. A great many of the amphibolitic lenses and layers in the granulite may be highly granitized and sheared gabbro sills or flows.

The origin of the feldspathic granulite in the median gneiss offers puzzling problems. Most of this rock is dense, sugary, uniformly layered, and has the appearance of quartzite, but the 13 specimens of granulite from which thin sections have been made contain 27 to 50 percent microcline, averaging 40 percent (table 15), and 40 to 64 percent quartz, averaging 52 percent. Accessory biotite, pyrite, apatite, tourmaline, sphene, albite, sericite, and zircon are common in the granulite. Much of the potassium feldspar is only faintly twinned microcline or untwinned and is not easily distinguished from the few grains of oligoclase. These rocks are wholly granoblastic except locally where they are affected by late or post-metamorphic brecciation and faulting. They grade in composition and texture into the more distinctly foliated biotitic facies on the one hand and into texturally similar but more calcic rocks containing little diopside on the other. The chemical analysis MG1 (table 14) is of granulite from the median gneiss at Talcville, where the rock has the appearance of a uniformly layered aplitic granite. The field relations of the granulites, however, and their high quartz content, seem incompatible with a magmatic origin.

The close association of feldspathic gneiss with biotitic and calcic gneiss and the local intergradation of these three types suggest that the layers containing abundant quartz and microcline were derived by feldspathization from either quartzose tuffs or quartz sandstones that contained small amounts of clay mineral. Both of these components, if distributed more or less uniformly through the sand, might have facilitated pervasive feldspathization, which was apparently retarded in the very pure quartzites.

An arkosic parent rock might be inferred except that (1) arkose is not likely to have been interlayered with very thick beds of pure limestone; (2) arkoses are commonly more calcic, hence would contain more plagioclase and a greater variety of accessory minerals than these gneisses do; and (3) a meta-arkose would be likely to show uneven layering (relict bedding) or even relict cross-bedding.

Quartz-biotite-oligoclase gneiss and migmatite

Distribution, form, and thickness

Quartz-biotite-oligoclase gneiss and migmatite (referred to in the following discussion as quartz-biotite gneiss) forms two very large masses, and may include several small isolated bodies. The largest mass, which adjoins the Sylvia Lake - Cedar Lake marble belt on the west and northwest, represents a very thick metasedimentary zone now much injected or replaced by gneissic granite and pegmatite (see also Engel and Engel, 1953b, 1958a). The other large body, which lies in the southeastern part of the district, is the northeastern part of a very large elliptical mass that adjoins the Fullerville - Edwards marble belt on the southwest (Brown, 1936a, fig. 1). Brown (1936a, p. 236-238), in his mapping of the district and adjoining areas, regarded these two bodies as separate stratigraphic units. He regarded the largest one, which he called the Upper Garnet gneiss, as the youngest stratigraphic unit in the district, and the other, which he called the Lower Garnet gneiss, as the oldest. More recent studies, however, indicate that these bodies are both parts of the oldest stratigraphic zone exposed in the district, and the more important lithologic features of this gneiss complex are consistent with this interpretation (Engel and Engel, 1953b, 1958a).

A fairly large exposure of gneiss, partly pyritic but otherwise very similar to quartz-biotite gneiss, occurs in the Fullerville - Edwards marble belt, at 11 to 12 N., 6 to 9 E., between the most southeasterly area of quartz-biotite gneiss mentioned above and extensive outcrops of the median gneiss. Although the rock in this exposure was mapped as pyritic schist by Cushing and Newland (1925) and by Gilluly (1945), most of it probably is part of the quartz-biotite gneiss surrounded by lenses of intricately folded pyritic schist. Its general structure is probably anticlinal, although it may be more complex (see p. 191).

Other exposures of gneiss assigned to the quartz-biotite gneiss occur north of Fullerville, in association with granite, marble, and amphibolite. The relations of these exposures to the larger masses of quartz-biotite gneiss, and to the structural features of the district, are shown in figure 7~~, and are discussed in detail on pages~~.

Excellent outcroppings of the quartz-biotite gneiss form low, linear ridges, separated by narrow strike valleys, throughout the area, from Kellogg Corners on the southwest to the vicinity of the Hyatt zinc mine, where the gneiss interfingers with inequigranular granitic gneiss. Emplacement of granite has obscured or obliterated the gneiss-marble contact zone from there almost to the Trout Lake - Edwards road (18.5 N., 10.0 E.). What is inferred to be the quartz-biotite gneiss reappears, however, in several exposures between this point and Cedar Lake, and still farther northeastward.

The quartz-biotite gneiss mapped on plate 1 is but a small part of a wide belt of migmatite that extends northwestward across the Grenville Lowlands for about 35 miles (Engel and Engel, 1953b, 1958a, 1960). Probably the part of the quartz-biotite gneiss shown on plate 1 constitutes the uppermost 800 to 1,000 feet of the unit whose total thickness, exclusive of later granitic veins and lenses, is about 2,800 feet.

The southeasterly mass of quartz-biotite gneiss (Brown's Lower Garnet gneiss) is well exposed throughout the wooded highland that extends southwestward from a point about $1\frac{1}{2}$ miles south of Edwards and continues well into the Lake Bonaparte quadrangle (Smythe and Buddington, 1926; see especially the quadrangle map). Outcrops of this gneiss terminate abruptly at the floodplain of the Oswegatchie River southeast of Edwards, so that its relations to the marble in this area are very uncertain (pl. 1).

Lithologic features

Throughout the district the quartz-biotite gneiss consists chiefly of migmatite, with numerous intercalations of pegmatite and inequigranular granitic gneiss. The typical migmatite consists of layers of quartz-biotite-oligoclase gneiss, a few inches to several feet thick, interlayered with veinlets and lenses of rock consisting of quartz, microcline and a little oligoclase. In many places, also, there is lit-par-lit intergradation in texture and composition between the quartz-biotite gneiss and granitic gneiss. An attempt is made in figure 5 to represent these relations graphically.

Thin interlayers of amphibolite occur through the gneiss, parallel to its dominant foliation and layering, and in several places the gneiss also contains thicker interlayers and lenses of dolomite marble. These features persist not only throughout the district but far beyond it. Their origin has been considered at length elsewhere (Engel and Engel, 1953b, 1958b, 1960).

The resulting gneiss complex is unlike other gneisses and schists in the metasedimentary series of the northwest Adirondacks and unlike most metasediments expectable in a sequence consisting mainly of thick layers of siliceous marble and quartzite. The distinctive compositional features of the quartz-biotite gneiss are summarized in tables 16 and 17.

Table 16. Near here. (p. 201)

Table 17. Near here. (p. 202

In table 16 Qb-LA is the average of 15 modal analyses of specimens of gneiss that are believed to represent its least-altered metasedimentary rocks. Qb-Av in table 17 is the average of three chemical analyses of similar material. Comparisons with modal analyses of other gneisses and schists in the district (table 16) show that the quartz-biotite gneiss represents a metasedimentary type that is unique in the district, as it is unique in the Grenville Lowlands as a whole. For example, the average modes of pyritic schist and median gneiss (table 16) appear to represent somewhat feldspathized argillaceous sandstones or sandy shales, such as specimen Sh-SS of table 17. The average chemical analysis of the quartz-biotite gneiss (Qb-Av, table 17), on the other hand, gives an Na/K ratio of 1.3--much higher than it would be in any mixture of quartz with clay minerals common in shales. This Na/K ratio more closely resembles that of graywacke-type sandstone (Gr-III, table 17).

Table 16. Average mineralogical composition of ~~five~~ four major segments of quartz-biotite gneiss and the approximate, average mineralogical composition of four ~~other~~ gneissic interlayers in the marble of the district.

	Qb-LA	Qb-NW	Qb-SE	Qb-SW	Mg-Gr	Mg-BF	PS-FF	PS-BF
Quartz	40.	38.	34.	33	52	23	39	41
~~Potash~~ Potassium feldspar	1	13.	15.	18.	40	40	29	19
Oligoclase	39.	32.	37.	35.5	4	17	4	5
Biotite	17.	15.	9.	11.	2	19	4	7
Muscovite	2.5				1	0.5	6	5
Pyrite		1.	0.25	0.5	1	R	9	5
Garnet	.3	0.5	3.	2.			0.5	
Sillimanite						R	0.5	1
Other accessories	.2					.5	8	17
Total	100	100	100	100	100	100	100	100

Qb-LA = Average of 15 modal analyses of chemically least-altered quartz-biotite gneiss northwest of the marble, Sylvia Lake-Cedar Lake.

Qb-NW = Average of 58 modal analyses of quartz-biotite gneiss as mapped northwest of Sylvia Lake, Hyatt and Talcville.

Qb-SE = Average 38 modal analyses of quartz-biotite gneiss as mapped south and southwest of Edwards.

Qb-SW = Average of 16 modes of quartz-biotite gneiss in the area southwest and southeast of Balmat.

Mg-Gr = Average of 13 modal analyses of least injected ~~layers of~~ granulose layers in the median gneiss.

Mg-BF = Average of 11 modal analyses of least injected ~~layers of~~ biotitic layers of the median gneiss.

PS-FF = Average of 8 modal analyses of least altered ~~layers of~~ feldspathic layers in the pyritic schist interlayer, marble Sylvia Lake-Cedar Lake.

PS-BF = Average of 6 modal analyses of least altered ~~layers of~~ biotitic layers in the pyritic schist, marble Sylvia Lake-Cedar Lake.

Table 17. Averages of chemical compositions of quartz-biotite gneiss, several graywacke sandstones, shales and composite sandy shales.

	Qb-Av	Gr-III	Sh-II	Sh-SS
SiO_2	70.74	69.69	58.10	65.50
Al_2O_3	12.87	13.53	15.40	12.00
Fe_2O_3	1.14	.74	4.02	3.10
FeO	3.73	3.10	2.45	1.70
MgO	1.95	2.00	2.44	2.00
CaO	1.98	1.95	3.11	4.5
Na_2O	3.50	4.21	1.30	1.0
K_2O	2.71	1.71	3.24	2.6
H_2O+	.36	2.08	5.00	2.9
H_2O-	.05	.26		
TiO_2	.39	.40	.65	.50
MnO	.05	.01		
P_2O_5	.10	--	.17	.1
CO_2			2.63	3.4
SO_3			.64	0.5
Total	99.57	100.01	99.15	99.8

Qb-Av = Average of three analyses of specimens of least altered quartz-biotite gneiss southwest and northwest marble Sylvia Lake-Cedar Lake. (See A. Engel and C. Engel, 1953b, table 3).

Gr-III = Average of 3 analyses of graywacke, Franciscan formation (after Taliaferro, 1943, p. 135).

Sh-II = Average shale (Clarke, 1924, p. 24).

Sh-SS - One part average sandstone plus two parts average shale (Pettijohn, 1949, p. 271).

Specimens representing the least altered quartz-biotite-oligoclase gneiss are equigranular gray rocks, faintly to distinctly foliated (see table 10 in Engel and Engel, 1953b). In outcrops north of Talcville, Hyatt, and Fowler the constituent biotite is greenish-brown; to the northeast and southeast it is predominantly reddish-brown. (For analyses, see table 18).

Table 18. Near here. p. 204

The reddish-brown biotite is accompanied by more garnet than the greenish-brown variety, by pervasive feldspathization of the gneiss, and locally by sillimanite. Garnet is most common in and alongside the pegmatites but tends to pervade all the gneiss north and south of Edwards and the gneiss immediately around the alaskitic granite southwest of Balmat. The garnet is dominantly almandite (table 18) with a fairly constant index of refraction near 1.80. Its composition, also, is fairly constant over large areas, although a manganese-rich variety occurs in the pegmatites of the Emeryville area (see Engel and Engel, 1953b, 1960).

The veinlets of quartz and potassium feldspar in the gneiss range in texture from very coarse to fine. These veinlets and also the thicker lenses of granite tend to follow the dominant foliation in the gneiss, but many cut across it. The dominant feldspars are microcline and microperthite and, as noted above, the pegmatites contain almandite garnet. A few sheets of pegmatite contain black tourmaline. Some of the granitic veins are very thin, and others are tens of feet thick. Some are quite uniform in thickness, but most of them pinch and swell abruptly. In areas where the foliation in the gneiss is contorted, many of the pegmatitic veinlets are equally crumpled, but some that are much crumpled appear to cut across the oldest foliation--probably bedding--in the gneiss.

Table 18. Analyses of garnets and biotites from the quartz-biotite gneiss north and west of Edwards, New York.

	B-237	B-216	Gar-4	Gar-8
SiO_2	36.16	35.37	37.96	38.16
TiO_2	3.18	4.33	.02	.07
Al_2O_3	17.40	17.50	22.13	22.09
Fe_2O_3	2.55	1.41	1.05	.88
FeO	17.71	19.25	31.14	31.13
MnO	.30	.04	1.21	.97
MgO	9.36	9.12	5.59	5.69
CaO	.07	.24	1.32	1.12
Na_2O	.12	.10		
K_2O	9.61	9.09		
H_2O+	3.30	3.27	--	--
H_2O-	.00	.01	--	--
P_2O_5	--			
F	.29	.26		
	100.05	99.99		
Less O for F	$\cong$ .12	$\cong$ .11		
Total	99.93	99.88	100.40	100.11

B-237 = Greenish-brown biotite from essentially uninjected and non-garnetiferous quartz-biotite gneiss 500 feet northwest of Emeryville. Lee Peck, analyst. Gamma index = 1.647

B-216 = Reddish-brown biotite from highly injected garnetiferous quartz-biotite gneiss at 2.4 N, 8.4 E. Lee Peck, analyst. Gamma index = 1.658.

Gar 4 = Garnet in migmatitic gneiss north of Hyatt, N. Y., Celeste G. Engel, analyst. Index = 1.796.

Gar 8 = Garnet at contacts of pegmatite and gneiss at Emeryville, N.Y. Celeste G. Engel, analyst. Index = 1.795

Gradation between dark-gray gneiss and gneissic granite is common and is frequently associated with well-defined veins of pegmatite and granite. The interrelations of the gneissic and granitic rocks are represented graphically in figure 5. Some of the best examples of garnetiferous migmatite appear in the southeasternmost body of gneiss, along the south boundary of plate 1, where, in fact, the gneiss is more thoroughly injected, feldspathized, and crushed by deformation than most of that along the northwest side of the district. Probably much or most of the granitic rock in the gneiss represents gneiss replaced by granitic fluids (Engel and Engel, 1953b, 1958a).

Figure 5. Generalized curves showing the interrelations between the mineralogical, chemical, and textural features of granitic gneiss and quartz biotite gneiss. (in pocket)

Most of the amphibolite interlayers in the gneiss are only a few inches thick, although several are as much as 200 feet thick; the thicker ones are shown on plate 1. Their contacts with the gneiss that is believed to be least deformed are commonly sharp. Amphibolite makes up about 8 percent of the gneiss complex by volume. Most of it is a hornblende-andesine rock which contains pyroxene locally; biotite, chlorite, quartz, ilmenite, sphene, and potassium feldspar are common accessories. The core of the thicker, least injected and deformed amphibolites averages about 65 percent hornblende, 20 percent andesine, 10 percent quartz and 2 percent ilmenite. Many thinner or more deformed and injected amphibolites contain as much as 15 percent biotite as well as potassium feldspar, both of which clearly formed during the final stages of injection, deformation and granitization of the gneiss. The least-altered amphibolites are similar in bulk chemical composition to quartz-bearing diabase or basalt, but they are not ophitic in texture and have no other features that prove them to be of igneous origin.

One of the larger marble lenses in the quartz-biotite gneiss lies north of Trout Lake, just outside of the limits of plate 1. Very thin interlayers, too small to be shown on plate 1, occur along the line of strike southwest of 9.0 N., 8.0 E. Brown (1936a, p. 237, fig. 1) interpreted these to be squeezed relicts of the same marble beds that now form the large embayment in the gneiss directly northeast (10.0 to 12.0 N., 10.5 to 13.8 E.). This marble may be either an interlayer in the gneiss, greatly thickened along the apex of a fold (see fig. 7), or a part of the Fullerville - Edwards marble belt, exposed in the highly deformed and eroded crest of a major southwestward-plunging fold. The marble in these lenses is mainly dolomite, much of which contains quartz or diopside, or both. Locally, however, especially in the embayment described above they contain calcitic marble, but the calcite in them seems to have formed at the expense of dolomite by preferential extraction of magnesia during the metamorphism. The process is believed to be similar to that discussed in some detail in conjunction with the origin of the calcitic marble of Zone 14 and the talc deposits. The interlayers of marble in the gneiss presumably represent rare, intermittent lapses in the deposition of the clastic sediments now represented by the quartz-biotite gneiss.

The origin of the quartz-biotite gneiss presents many puzzling problems. As is shown on plate 1, the contacts of gneiss and marble are accordant with relict bedding in the marble where it is least disturbed by deformation. Where the gneiss is least modified by granitic injection, replacement, and deformation, its uniformity of composition is matched by a uniformity of textural and physical features. Presumably the sediment now altered to gneiss was a clastic of medium to fine grain size; that is, a silt, clay, sand, or tuff, or some intermixture of these.

Regional studies indicate that the uniformity of the gneiss is widespread and that no major unconformity separates the clastic sediment from the adjacent carbonate-rich sediments. This relationship, without evidence of marked unconformities and without coarse clastics, suggests that the sediments from which the gneiss evolved were derived from maturely weathered rocks. Curiously enough, however, the bulk composition of the least-altered gneiss is that of graywacke, dacitic tuff, or sodic shale, none of which are commonly associated with thick, laterally continuous beds of siliceous carbonate. Conceivably the excess of Na_2O over K_2O in the gneiss was added to the parent sediment at the interface of sedimentation in seas that were abnormally saline. This suggestion is prompted largely by the occurrence of what seem to be beds of evaporite in the enclosing marble. Other alternatives, perhaps less easily reconciled with the geologic features, are that (1) the parent sediment consisted of dacite volcanic ash or (2) Na_2O was added to the rock, or K_2O removed from it, after sedimentation as a diagenic or metamorphic phenomenon.

Irrespective of the origin of the quartz-biotite gneiss, the thin sheet-like interlayers of amphibolite present still another problem. Their composition and mode of occurrence are consistent with derivation from either basalt sills or flows, beds of basic tuff, or pure or impure limestone, but no gradation between amphibolite and any of these rocks has been observed. The fact that many of the amphibolite layers, although very thin (1/4 to ½ inch) are laterally extensive and can be seen in many good exposures to be essentially accordant with the oldest foliation would seem to indicate that some if not all of the amphibolite is of sedimentary origin. But if the parent sediment was a basic tuff, one would expect it to be associated with basic dikes and lavas. If it was a calcareous rock, some carbonate relicts should remain, and none have been found.

Granitic rocks

Introduction

Granitic rocks, largely inequigranular, biotitic, and gneissoid, are abundant in the district where, at least locally, they inject or replace all of the metasedimentary rocks, together with associated amphibolites. The granitic rocks with two exceptions appear to be all of similar origin and history and may be correlative with the Hermon gneissic granite complex of Buddington (1939, p. 142-161; 1948, p. 32-33, 36-39; Engel and Engel, 1953b). One exception is a rock that occupies only a small area within the district, namely, the alaskite granite which forms what Buddington has described as the California phacolith (Buddington, 1929, p. 61-65). The northeastern tip of this phacolith projects into the southwest corner of the district, south of Kellogg Corners. A second, somewhat dubious exception is the irregular prong of granite exposed in garnetiferous quartz-biotite-oligoclase gneiss in the southeastern part of the district (7 to 9 N., 7 to 8 E.). This granite exhibits certain features common to both the Hermon gneissic granites and the alaskite. The age relations between the two rocks are unknown but the alaskite is believed to be slightly younger than most of the Hermon complex.

Gneissic granite complex

The granite and granitic gneiss of the Hermon type as defined by Buddington (1939) commonly forms rounded hills--roches moutonnées--which are readily distinguished from the more angular ridges of the distinctly layered metasedimentary gneiss and schist. In most exposures the granitic gneisses are either deep-pink or gray and contain pink augen of microcline. These rocks vary widely in physical and chemical features, but they may be divided into several general types, each grading at least locally into the others, all intimately mingled and seemingly of almost the same age. Their most marked variations of texture and composition are presumably due to their composite origin and differing degrees of deformation. Most of them have been deformed in the solid or nearly solid state, and much of the granite appears to be of metasomatic origin. In some areas the granitic rocks appear to grade into all of their siliceous host rocks. This is especially true of the contacts between the granite and the quartz-biotite gneiss and other paragneisses and schists, but it is also true of a few contacts of granitic gneiss with amphibolite. Almost all contacts of the granite with marble, on the other hand, are sharp.

The gradations in both chemical and mineral composition between the quartz-biotite gneiss and associated granitic rock are shown graphically in figure 5 and have been discussed in separate publications (Engel and Engel, 1953a, 1958a). Similar graphs might be constructed to illustrate gradation of this granitic gneiss into amphibolite, median gneiss, and pyritic schist, but these graphs would have lines with more abrupt slopes indicating less subtle transitions in composition. In general, the granitic rocks contain accessory minerals characteristic of the associated pre-granitic rock. Where this is amphibolite, the granite near the amphibolite commonly contains amphibole, and either more plagioclase or more calcic plagioclase, and less quartz, than where the wall rock is median gneiss. In contrast, samples taken well within all the larger masses of granite indicate a relatively uniform composition, irrespective of the associated older rock (table 19). For example, the averages of the granite

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Table 19. p. 213 ~~Near here.~~

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around Clear Lake (along the 8E. coordinate, south of 7N.) and the granite near Talcville and Cedar Lake along the northeast boundary of the district are very similar to those of the granites between Fullerville and Balmat; yet the granite masses last mentioned invade almost every kind of pre-existing rock, whereas the other two groups of granite are largely confined to quartz-biotite gneiss. All these averages are fairly close to the averages of the modes (table ~~19~~ 20) of the Hermon type granitic gneiss emplaced in quartz-biotite gneiss along the northwest side of the district, between

p. 213a

Kellogg Corners and Cedar Lake. This average, 32 percent quartz, 28 percent microcline, 33 percent oligoclase, and 5 percent biotite, is very different from that for the essentially non-biotitic, low-plagioclase, microperthite granite of the so-called Alexandria type (Buddington 1939, p. 145), which occurs in the California phacolith southwest of Kellogg Corners.

Table 20. Chemical analyses and modes of inequigranular granitic gneiss (Hermon type granite of Buddington, 1939) in quartz-biotite-oligoclase gneiss and in marble. 23-213a

Specimen No.	Qb-11	PG-18	EG-28
SiO_2	64.18	72.35	69.40
Al_2O_3	16.19	12.94	15.63
Fe_2O_3	1.87	1.13	0.60
FeO	3.98	1.52	0.62
MgO	2.17	0.72	0.14
CaO	2.69	1.33	0.18
Na_2O	4.64	4.22	5.03
K_2O	3.48	4.90	7.73
H_2O+	0.24	0.18	0.18
H_2O-	0.06	0.05	0.04
TiO_2	0.34	0.10	0.04
MnO	0.06	0.03	tr.
Total	99.90	99.47	99.59

Modes

Mineral	Qb-11	PG-18	EG-28
Quartz	29	35.5	29
Plagioclase	44 (AB^{75})	41.5 (AB^{82})	37 (Ab^{86})
Biotite	15	4	2
Potassium ~~Potash-~~feldspar	11	18	31
Magnetite, apatite, zircon, pyrite, tourmaline	1±	1±	1±

Qb-11 Porphyroblastic facies of granitic gneiss from the point 5.0 W., 1.289 N. Ledoux and Cox, analysts.

PG-18 Porphyroblastic facies of granitic gneiss from the point 3.2 E., 7.6 N. Ledoux and Cox, analysts.

EG-28 Alaskitic facies of granitic gneiss in quartz-biotite gneiss at 14.3 N., 1.8 W. Ledoux and Cox, analysts.

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Table 19. Modal analyses of granites and granitic gneisses in the district and an average of 75 modal analyses of granite in the quartz-biotite-oligoclase gneiss.

	Granite in Clear Lake area					Granite in Talcville-Cedar Lake Area											Granite in Lost Pond area				Granite in Balmat-Fuller~~ton~~ville areas													Granite enclosed in marble - N 39	Av. of 75 modes of granite in G
	Gs53	Gs60	Gs61	Gs47	Av.	Gs34	Gs38	Gs40	Gs56	Gs51	Gs78	Gs52	Gs57	Gs58	Gs 6	Av.	Gs73	Gs75A	Gs76	Av.	Gs12A	Gs14	Gs23	Gs33	Gs24	Gs22	Gs27	Gs17	Gs19	Gs32	Gs29	Gs 31	Av.		
Quartz	44	29	33	30	34	33.5	23	39.5	18	32	37	48	30	27	20	30.7	29	34	38	34	32	35	31	30	40.6	26	32	40	36	38	30	18	32.4	40	32
Microcline	18	22	41	55	34	15	14	17	18	20	28	33	41	48.7	60.8	29.3	11	24	28	21	16	19	20	21.5	22	25	26	27	29	29	44	55	27.8	29	28
Oligoclase	31≠	43.5≠	25≠	15	28.6	30	43	40	50	41	32	19	20	23	*19	31.7	44	29	30	31	42.8	39	45	41	32	44	34.5	32	31	33	23	23	35	30	33
Biotite	2.5	5.1			1.6	21	18.6	3.1	12	2	1		6	1		6.4	15	10	3	9.1	9	6	2	7	5	5	7		3	R	2	2	4	.5	5
Chlorite	.5	.4	R	R	.2	.5	.4	.4	2	.5	R		2			.6	R	2	1		R	R	2	R	R	R	R	1	1	R	1	R	.2		
Hornblende							1			3				VR										R	R		R		R		R	2			
Magnetite	R					R		R		1			.6	R		.2	1	R	R					R										R	
Apatite		R				R	R	R	R	R	R		R					R			R	.6	VR	R			R								
Sphene		R				R	R		R	R															R							R		0.3	
Zircon						R		R	VR	R			R	VR				R			R		VR		R		R								
Garnet	3.5		1		1.1																						VR								2
Pyrite	R			R		R	R		R		R							R	R															R	
Muscovite	.5	R	R	R	.1	2	R			.5	2		.4	.3	VR	.3																			
Allanite						R																													
Clino-zoisite						R									R																				
Serpentine	R												R					R					R												
Total	100	100	100	100		100	100	100	100	100	100	100	100	100	99.8+	99.2	100	99.+	100		100	99.6+	100	99.5+	99.6+	100	99.5+	100	100	100	100	100		99.8	100

≠ = sodic andesine
* = calcic albite
R = between o.1 and o.4 percent
VR = less than 0.1 percent
= North side of Newton Hill (6.9 E, 14.9 N).

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(213 as follows)

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The marked variations in the proportion of each major constituent mineral, especially of biotite, in smaller masses of the Hermon type granitic gneisses, are unknown in the Alexandria type granite except very locally in contact zones. The numerous accessories in the Hermon type granitic gneisses, which include garnet, sillimanite, calcite, and hornblende, suggest either a metasomatic or hybrid origin. There are many areas of pyritic granite, in which the percentage of pyrite, by volume, ranges from 0.1 to 0.4. The garnet (mostly almandite) and tourmaline (schorl) are largely confined to pegmatite fractions intruded into the quartz-biotite gneiss, where the garnet is a reaction product (Engel and Engel, 1953b, p. 1072-1073).

In figure 5 an attempt has been made to relate some of the dominant textural features of the Hermon type granitic gneiss to its composition and to that of the enclosing quartz-biotite-oligoclase gneiss. The inequigranular textures of the granitic gneiss masses reflect the appearance in the adjoining gneiss of large pink eye-shaped to rudely rectangular crystals of microcline or plagioclase. These crystals range from 1/8 to 2 inches in length, and average about $\frac{1}{2}$ inch. They are partly or wholly enclosed in small lenses, lenticules, and aggregates of quartz and biotite, all set in a matrix of microcline, oligoclase, quartz, and biotite.

Most of the augen forms of the large microclinesappear to have resulted from deformation. The augen are enveloped in a mortar of feldspar and quartz and have tails of feldspar, largely recrystallized. Subhedral to almost euhedral crystals locally replace foliated gneiss that abuts against their ends. Grains of quartz, biotite, and other minerals are enclosed in many of the large feldspar crystals, which appear to have been formed by coalescence of smaller grains. The large tailed augen of feldspar tend to be roughly alined, and together with subparallel flakes of biotite and polycrystalline leaves of quartz, they define a faint foliation.

The granitic gneiss commonly has a stronger foliation at and within a hundred feet of its contacts with marble. Much granitic gneiss adjacent to marble is mortar gneiss verging on mylonite. These marginal zones of granitic gneiss may also possess a distinct lineation (pl. 1), formed by alinement of the feldspar augen and of elongate quartz leaves and biotite flakes. The lineation is further emphasized by intersections of foliation surfaces.

The crushing and mechanical deformation of the granites are commonly accompanied by partial disappearance of biotite and other mafic minerals. In texture and composition these deformed granitic gneisses which evolved from the Hermon type closely resemble sheared marginal facies of the equigranular alaskitic granite found in the California phacolith.

In some places the dominant gneissic augen-bearing granite may be either cut by or grade into both coarse-grained pegmatite and medium-grained, nearly equigranular pink alaskitic granite containing little or no biotite. The last-named rock does not seem to be a product of shearing and recrystallization of a porphyritic granite, and it may indeed be a product of direct solidification of magma. Masses of this alaskitic rock form a part of the large, gently curved sheet of granite in the quartz-biotite gneiss just north and west of Fowler village, and several others lie within the quartz-biotite gneiss just north of the area shown on plate 1.

In thin sections of the granitic gneiss the oligoclase is much more altered--especially to sericite--than the microcline and has more irregular outlines. Most of the microcline, both coarse and fine-grained, shows distinct gridiron twinning. Microperthite and albite commonly form thin sheaths around oligoclase crystals and veinlets within them. In the microperthite (plotted as potassium feldspar in fig. 5), the albite forms irregular spindles, beads, and splotches. Analyses of two of these perthites taken from pegmatite cutting marble and gneiss are given in table 21.

Table 21. Near here. p. 217

Most of the biotite is brownish-red, but some is green. Both kinds are sometimes found in a single hand specimen.

Table 21. Wet chemical and spectroscopic analyses of perthites from pegmatities in marble and in quartz-biotite-oligoclase gneiss.

Wet chemical				Spectroscopic analysis		
Oxide	FP-4	FP-1	N-11-24	Element	FP-4	FP-1
SiO_2	66.54	65.22	61.60	Li	.0001	.0004
TiO_2	.01	.01	.01	Rb	.09	.04
Al_2O_3	18.85	19.05	19.74	Cs	--	.001
Fe_2O_3	.10	.04	.04	Sr	.03	.3
FeO	--	--		B	--	.02
Mgo	tr	.05	.52	Pb	--	.003
CaO	.30	0.21	.05	Mn	--	.1
Na_2O	2.64	4.87	2.88	Ga	--	.0003
K_2O	13.48	9.14	10.30			
BaO	.05	1.07	3.95			
H_2O+	.02	.16	.19			
Loss on ignition	.05		.04			
H_2O-		.07				
Total	99.94	99.89	99.91			

FP-4 Perthite from pegmatite (up to 5 feet wide) accordantly interlayered with quartz-biotite-oligoclase gneiss.
Celeste G. Engel, wet chemical analyst; A. A. Chodos, spectroscopist.
Indices = 1.520, 1.524, 1.528; -2V = 84°.

FP-1 Perthite from pegmatite (up to 3 feet wide) intruded discordantly across layering in marble and talc, 6th level, American Talc mine.
Celeste G. Engel, wet chemical analyst; A.A. Chodos, spectroscopist.
Indices = 1.523, 1.527, 1.631; - 2V = 83°.

N-11-24 Perthite from partly serpentinous pegmatite in talc belt, 5th level, International 2 1/2 Talc mine, Talcville.
Katherine Oslund, wet chemical analyst.

Buddington (1948, p. 39) has suggested that the Hermon granitic gneisses of the Grenville Lowlands were formed by volatile-rich magma and alkali-silicate fluids from the magma, which shredded, intruded, and replaced rocks of the Grenville series. This general statement appears completely applicable to the granite in this district but it is often unclear whether a particular part of the granite is metasomatic or igneous.

Alexandria type granite

The pink fine-grained equigranular alaskitic granite which forms the northeastern tip of the California phacolith, southwest of Kellogg Corners, is described in the following pages as alaskitic granite (Alexandria type of Buddington, 1939). In most exposures this granite is distinctly foliated, although locally it is nearly massive. In eight thin sections from specimens of the least-altered alaskitic granite collected in the southwest corner of the district the percentage of quartz ranges from 28 to 45, that of microperthite from 35 to 44, and that of oligo-andesine from 10 to 31. Most of this granite contains accessory ilmenite and zircon. As Smythe and Buddington have noted (1926, p. 47), plagioclase appears to become more abundant toward some of the borders of the granite. Southwest of Kellogg Corners (near 4.0 S., 13.7 W.), the northeasterly tip of the phacolith is cut by the Balmat fault and is extensively crackled. There, the granite, like other feldspathic rocks along that fault, is much albitized and locally contains as much as 4 percent chlorite. In thin sections this crackled granite shows a cataclastic texture. Elsewhere the granite is granoblastic.

Smythe and Buddington (1926, p. 47) and subsequently Buddington (1929, p. 61-65) have emphasized that much of the phacolith has a well-defined aureole. In going outward from the granite southwest of Kellogg Corners one crosses a greenish granite with schlieren of pyroxenic gneiss, and then a garnet-sillimanite gneiss with one or more interlayers of diopsidic and enstatitic gneiss. The aureole varies in width up to a maximum of about 150 feet. The contacts of its members with each other and with granite and enveloping metasediments are well-defined and readily mapped. Parts of the aureole and the granite are cut by pegmatites that are only mildly deformed but contain tourmaline and some garnet.

Within the granite are thin seams and layers of biotitic amphibolite that follow the dominant foliation. These range up to several feet thick, are quite persistent along the strike, and in places form swarms in which the amphibolites are separated by several inches or several feet of biotitic granite.

Buddington (1929) has concluded that the granite is phacolithic and of igneous origin. The limited observations possible during this study neither strongly substantiate nor disprove this interpretation. If the granite is igneous, the amphibolite interlayers must represent successively injected mafic hoods as Buddington suggests (1929). The extreme uniformity of the amphibolite sequences seems most difficult to explain in this way, but alternative explanations are hardly justified without additional field data. Smythe and Buddington (1926, p. 47) originally believed that much of the granite in the core of the phacolith was undeformed, but detailed study shows that most if not all of the granite has been slightly to moderately deformed and recrystallized in the solid state.

In discussing the phacoliths of alaskitic granite, Buddington (1929, p. 74-75, fig. 33) included the granite at Clear Lake, a prong of which extends into the district from the south along the 8 E. coordinate. In both composition (table 19) and texture, however, this granite is intermediate between the Alexandria and Hermon types of Buddington (1939). There is no contact aureole along its contact with the enveloping quartz-biotite gneiss, and much of that contact is gradational. Neither of these features is typical of other phacolithic bodies of alaskite in the Grenville Lowlands (Buddington, 1929).

Structural geology

Regional structural features

The northwest Adirondack Mountains are divisible into two major structural (and lithologic) elements, the Adirondack massif and the Grenville Lowlands (pl. 2). The massif consists largely (80-85 percent

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Plate 2. ~~Near here.~~ In pocket

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by volume) of anorthosite, gabbros, syenites and granites, formed in that order (Buddington 1939; 1948; 1952). All of these igneous-looking rocks except the granites seem to have been emplaced as semi-conformable sheets and thick lenses into a sequence of little-deformed Grenville-like metasedimentary rocks (Buddington, 1952). The granites formed during and between epochs of profound deformation. The metasedimentary rocks now appear as thin screens, skarns, roof pendants and wedges occupying less than 20 percent of the surface area of the massif. They are metamorphosed to upper amphibolite and lower granulite facies rocks. Many are considerably altered in composition, largely by introduction of substances from granitic magma and associated fluids.

Plate 2. Generalized geologic map of the northwestern Adirondacks, showing orientation and distribution of linear elements.

Buddington (1952) has inferred that the deformation of the rocks in the Adirondacks massif occurred mainly in two periods, separated by a period of slight or at least lesser deformation. He believes that the earlier of the two major deformations occurred slightly before and during the initial emplacement of granites. A second, severe deformation is indicated during the culmination of granitic intrusions. His principal reasons for so believing are that (1) most of the incoming granites appear to have cut across the bedding and across close folds in the syenitic sheets, (2) some parts of these granites, especially within the massif, contain but slightly deformed granitoid textures and fabrics, (3) other parts of the above mentioned granites and other granites are much deformed, having been converted to lineated mortar gneisses and mylonites.

Two major periods of deformation also are indicated in the Lowlands region (Brown and Engel, 1956). The Adirondack massif dips at moderate angles (25°-65°) northwest under the Grenville Lowlands, where metasedimentary rocks predominate (Engel and Engel, 1953b, p. 1052-1058, and 1958a). The contact between massif and lowlands actually is a zone of intense shearing and displacement (Buddington, 1939, p. 251-281; 1952; Engel and Engel, 1958a, p. 1375). Minimum stratigraphic discontinuities in the metasedimentary sequence along the boundary between the lowlands and massif probably exceed one or perhaps two miles. Data from many parts of the lowlands indicate that most of the early rock movements (post-syenite-syn-early granite) seem to have involved rolling and thrusting of the metasedimentary rocks of the lowlands southeastward and possibly over a synclinorial structure of syenites and granites along the perimeter of the massif. Major structural elements formed at this time were folds overturned toward and against the massif, with subhorizontal axes that trend northwest.

The dominant structural element induced in the metasedimentary rocks of the lowlands seems to have been a bulbous anticlinorium (fig. 3). Prior to further deformation, most or all of the constituent folds in the anticlinorium plunged gently northeast and southwest. Their axial planes, in northwest to southeast vertical sections, formed an asymmetrical fan-like pattern (Buddington, 1939, p. 237-243; Engel, 1949, p. 768; Engel and Engel, 1953b, p. 1052-1059; 1958a, p. 1372-1376).

Later rock movements along the outer parts of the massif and in the lowlands seem to have been dominantly strike slip. Many major structural features indicate that rocks of the lowlands were dragged northeast relative to those of the massif and great refolds having a common direction sense (asymmetry) were superimposed upon the already folded complex in both lowlands and massif.

Gently plunging northeast-trending folds in syenite and gabbro sheets, granites and metasediments were refolded along northwest- and southeast-plunging axes. Several belts of refolding are especially well defined. Two of particular interest lie on opposite sides of the Grenville Lowlands. The belt on the northwest side includes the Rossie, Hammond and Black Lake areas in which the folded folds plunge steeply (65°-90°) downward along the nearly vertical contact of metasediments and granite of the Alexandria "batholith" (Gananoque-Brockville-Mallorytown areas; see Buddington 1939, fig. 25 and pl. 2; Engel, 1948, fig. 1).

The second belt, on the southeast side of the Grenville Lowlands, varies in width from 5 to 15 miles or more and includes both metasediments of the lowlands and syenites, gabbros, granites and metasediments in the northwestern margin of the massif (Brown and Engel, 1956, p. 1620-1622; Buddington, 1939). The refolds take the form of sigmoidal and hook-shaped structures superimposed on northeast-trending, closely appressed folds with subhorizontal axes. Many crossfolds have a wave length or amplitude of 1 to 5 miles (Brown and Engel, 1956). In general, the axes of refolding in this belt plunge north to northwest at angles of 20° to 60°, subparallel to the outer surface of the massif. Consequently, the axes of refolding commonly cross the axes of pre-formed accordant folds at high angles (pl. 2). The form of each of these refolds is consistent with the previously noted movement of the lowlands rocks northeastward relative to the Adirondack massif. Because axes of the folded folds along the southeastern side of the lowlands plunge northwest athwart the northeast regional lithologic trends imparted by the first deformation, the refolds are referred to as cross folds.

The two zones of cross folding in the lowlands are separated by a medial zone in which crossfolding is only locally or incipiently developed. Most folds in the medial zone retain their northeast trend and gentle northeast to southwest plunges, characteristic of the features of the initial deformation (pl. 2). Northeast-trending folds of this type are referred to as accordant folds.

The Balmat-Edwards district lies largely within the southeastern belt of cross folds. As indicated on plate 2, however, the most northwesterly parts of the district coincide with the zone of transition between cross folds and northeast-trending folds of the central lowlands. Spectacular examples of refolding of accordant folds into cross folds are common throughout the district.

Structural features of the Balmat-Edwards district

The inferred structural evolution of the Balmat-Edwards district is diagrammed in figure 7 and has been discussed in detail by Brown and

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Figure 7. ~~Near here~~. ~~In pocket~~

Engel (1956). Plan 3 of figure 7 shows in a schematic way the geologic features of the area included on plate 1. Plans 2 and 1 are, respectively, reconstructions of preceeding stages in the development of the structural features. Briefly, Plan 1 indicates that the original major structure in the district was a large syncline extending from Sylvia Lake to Edwards. This syncline was tightly mashed together and then refolded as shown in Plan 2. Two of the largest and best defined cross-folds resulting from the refolding are (1) the large complex fold in the marble around Sylvia Lake and (2) the great hook-shaped fold in the gneisses and marble at Edwards.

Figure 7. Diagrammatic sketches illustrating the inferred development of structural features in the Balmat - Edwards district.

Sylvia Lake fold complex

At first glance the forms of the folded complex around Sylvia Lake do not seem to be the product of the same rock motions that induced the hook-shaped fold at Edwards. But if the trace of the axis of the initial accordant syncline is followed in detail, it is clear that the rocks northwest of Sylvia Lake have been rolled northeast relative to the southeast side (see fig. 11 and Brown and Engel, 1956, pl. 2). From a point on the trace of the axis of the Sylvia Lake syncline east of the Balmat road to the culmination of the syncline at the northeast end of Sylvia Lake, the opposite flanks of the parent fold are refolded and fused together. The deposits of zinc ore in the Balmat mine, and also in the Balmat talc belt as a whole, lie about parallel to the axis of the initial accordant syncline and very close to it. The cross folds induced in this area involve both flanks of the parent syncline for distances of hundreds or even thousands of feet as measured normal to its axis. One of the largest of these cross folds (refolds) is apparent in the marble and the Fowler talc belt at the Woodcock mine, and there are many others immediately southeast and to the west (see fig. 11 and pl. 1). A striking feature of all of these second generation folds is their asymmetric form; their axial planes have a south to southwestern trace, and their southeast flanks are short and commonly disrupted or attenuated compared to their northwest flanks (pl. 9). They plunge in general to the north and north-northeast at angles varying from several to 70° (pl. 1). Inasmuch as certain of these folds in the Sylvia Lake area involve both flanks of the initial syncline and have a common direction sense, they clearly are superimposed upon the initial syncline. The form of each of these cross folds is therefore entirely consistent with the interpretation that the rocks on the northwest side of the lake have moved northeast relative to the rocks on the southeast side.

Edwards crossfold

The hook-shaped cross-fold at Edwards has many subtle features (Brown and Engel, 1956, p. 1612-1614) but its general form is readily visualized (pl. 1 and fig. 10). On the short southeast flank of this fold the metasedimentary beds are either stretched, torn apart or obliterated, whereas on the long northwest flank they are continuous and well defined (pl. 1). The structure is an enormous drag-fold, presumably resulting from the northeastward rolling or shear of the northwest side of the structure relative to the southeast side. The axes of rotation are the axes of cross-folding, indicated by numerous minor folds and omnipresent lineations which plunge northwest at about 40°. As the southeast flank of this fold coincides with a major stratigraphic and structural discontinuity in the Grenville series, the rock movements may have included large displacements.

Folded forms southwest of Edwards

Many of the structural features in the vicinity of the Edwards fold, especially south of it, are due to movements of this type. One feature of considerable interest is the large mass of quartz-biotite gneiss at point 8 of figure 7, hereafter called the South Edwards gneiss. A second structural feature complicated by cross folding is the island of contorted, partly pyritic schist at 7 in figure 7, hereafter called the Pleasant Valley gneiss. Both these gneiss bodies appear to have been accordant anticlines of quartz-biotite gneiss, now variously folded.

The northern tip of the South Edwards gneiss appears to be part of an anticline whose axis plunges about 45° to the northwest. The axial plane of this fold strikes northeast and dips 45° northwest. This part of the fold appears to be a typical cross element, plunging almost directly athwart the general northeast trend of the immediately surrounding gneiss and marble. As this fold is traced to the southwest into the Lake Bonaparte quadrangle (Smythe and Buddington, 1926) its axis diverges in its direction of plunge toward the west (pl. 2). In its southwestern apical portion the fold has the same apparent anticlinal form, but its axis plunges southwest at about 45°. This major fold, then, when traced southwestward, appears to change from a northeasterly-striking cross-fold into a southwesterly-plunging nearly accordant fold.

The view that the broad structural form of the South Edwards gneiss is anticlinal or domical was first advanced by Cushing and Newland over thirty years ago (1925). The southwest half of the gneiss certainly has the form of a very simple broadly rounded accordant fold, plunging about 50° west-southwest. The northeastern end of the fold is, however, far less simple. Cushing and Newland (1925, p. 58-62) believed that the northeast end of the gneissic dome was a "closely pinched anticline which pitches north," and that the two large northeast-trending tongues of gneiss, shown in the southeast corner of plate 1 and at 6, figure 7 were two anticlines separated by a synclinal mass of marble (see especially Cushing and Newland, 1925, figs. 9 and 11). They thought the southeast tongue formed a main anticline and the northwest tongue formed a subsidiary anticline. This view is in general consistent with alternative 1 (fig. 9), but so are at least two other quite different interpretations. If these prongs of gneiss do represent anticlines, these are accordant structures now largely obliterated and sheared out by cross-folding movements.

FIG. 9 Diagrammatic sketches of the form of the elliptical body of garnetiferous quartz-biotite gneiss south of Edwards, New York (South Edwards gneiss), showing three possilbe interpretations of its sturctural ~~fore~~m .

(add ek) 25m.

A second hypothesis, suggested by Brown (1936, fig. 1, and p. 236), is that these prongs of gneiss represent the opposite flanks of the northeastern apex of the dome in the gneiss. This interpretation is based on certain field relations that seem to indicate that the marble at 6, figure 7, was the squeezed remnant of a carbonate interbed in the gneiss, now forced into the northeastern apex of an anticlinal nose in the gneiss. This is alternative 2 of figure 9. Brown assumed that the prongs were connected at their northeast ends and thus enveloped the large intervening tongue of marble. This is in many respects the simplest explanation of the known facts because the southwestern part of the gneiss (at and southwest of 9, fig. 7) definitely has the form of an anticlinal nose plunging 45° southwest, and it embraces a mass of marble somewhat similar to the one between the prongs of gneiss to the northeast.

But not all the features of the gneiss northwest of 9, figure 7, can be reconciled with this explanation. In the first place, the junction of the two prongs of gneiss is covered with alluvium so that its exact position is unknown. Moreover, the layers of gneiss just to the southwest apparently form a northwest-plunging anticline whose convex side faces southwest. As shown in figure 9, alternative 2, the same general pattern of a cross-fold convex to the southwest seems to persist throughout the area between 6 and 9, figure 7. The attitude of the layering suggests that the prongs of gneiss may be the greatly thinned flanks of a single crossfold. The change in direction and amount of the plunge of this fold between its core and its apex might be due to a progressive decrease toward the southwest in the dip of the layers at and near the axial plane of the fold.

If the axial plane of a plunging fold of this type was projected upward to the northeast and arched downward again into the quartz-biotite gneiss northeast of Edwards, it could account for the structural features there. In each of these hypotheses, the marble at point 6, figure 7, plays a critical role. In explanation 1 (fig. 9), this marble is assumed to be part of the Sylvia Lake-Cedar Lake belt. In explanation 2 (fig. 9) it is assumed to represent an interlayer near the top of the gneiss and to consist of impure marble older than the gneiss. This lower marble would presumably belong to the central or Gouverneur marble belt.~~(p. ——)~~.

One reason for assuming, in alternative 2, that there is a marble interlayer in this gneiss, is that there seemed to be such an interlayering east of Trout Lake (at 3, fig. 7). Further study suggests that the marble at the latter place encloses slivers of what seem to be median gneiss, pyritic schist, and other rocks that are associated with marble of the Sylvia Lake - Cedar Lake belt. The distribution of these rocks suggests that the marble at 3, figure 7, is in a shallow accordant syncline, separated from the marble belt of Sylvia Lake by an accordant anticlinal fold in the gneiss(at 2, fig. 7), and that both these folds were contemporaneous with the isoclinal syncline in the median gneiss and were refolded with it to form the Edwards fishhook fold.

Some of the known and readily inferred features of the Pleasant Valley gneiss (at 7, fig. 7) are shown in figure 8. Its central part consists of quartz-biotite-oligoclase gneiss, which seems to form the apex of a small initially accordant anticline. It should be emphasized that the precise forms of the initial accordant fold or folds are blurred and even obliterated in many places by superimposed cross-folds. In fact, the cross-folds recorded on plate 1 and figure 8 are the principal structural features in this area. The much-disrupted fragments of pyritic schist peripheral to the Pleasant Valley gneiss are correlated with the pyritic schist described in the Balmat section as Unit 2. Pyritization is by no means confined to the peripheral slivers of pyritic schist but pervades much of the gneiss itself. The pyritization of much of the gneiss in this area led Gilluly (1945) to correlate all of this gneiss with the pyritic schist around Sylvia Lake, but this correlation seems inconsistent with the mineralogical and chemical composition of the gneiss (Engel and Engel, 1953b).

The syncline in the marble between the Pleasant Valley and South Edwards gneisses (at 5, fig. 7), also has been refolded, like the accordant folds in the gneiss to the north and south. The marble encloses contorted segments of what seem to be the median gneiss, rusty marble and siliceous dolomite of *zones* 4 and 6.

Fowler amphibolite

The large mass of amphibolite southeast of Fowler is interpreted as a thick sill emplaced in the marble stratigraphically above the median gneiss prior to any pronounced deformation of the sedimentary rocks. It is now mashed tightly together as a synclinal knot (fig. 7). The ubiquitous lineations on the flanks of the amphibolite plunge northward parallel to associated cross-folds (pl. 1) and were induced by the cross-folding motions.

Figure 8. Inferred form of the peripherally pyritic, quartz-biotite-feldspar gneiss in the Fullerville-Edwards marble. This interpretation is based largely upon the geologic features shown in plate 1 together with petrographic studies of the rock layers

California phacolith

The California phacolith at the southwest end of the district is still another accordant anticline, whose form is complicated by cross folding. The northern nose of the phacolith, as defined by the gneisses that envelop it, plunges north-northeast (pls. 1 and 2 fig. 6). This plunge is about as steep as that of the associated crossfolds but its direction is intermediate between that of the cross-folds and accordant folds. As the axis of the California phacolith is traced southwest, it flattens in its central part, becomes accordant with the regional strike and plunges about 20° SW. at its southwest end. There the phacolith is a typical accordant fold.

Shear and fault zones

The combination of solid flow and tearing apart of layers which characterized the more ductile periods of deformation in the district has been described in conjunction with the discussion of stratigraphic units. Many of the thin zones and some major units are either dispersed or locally obliterated by shearing motions; the bulk of the major rock motions and thus most dominant planes of secondary shear appear to have followed the sedimentary fabric. Because sedimentary bedding remained the dominant guide to the shearing and penetrative movements in metasedimentary rock units, the most prominent foliation and lithologic layering induced by metamorphism is, in most areas, a quasi-bedding foliation (Engel, 1948). Faults in the sense of discontinuous ruptures displacing mappable units are common, but most appear to be small and to have formed in the final stages of the deformation described above. Insofar as these were observed and of mappable size they appear on plate 1.

Balmat fault

One fault that may involve large displacements of units was first noted by Smyth and Buddington (1926, p. 96), and subsequently described by Brown (1936a) as the Balmat fault, plate 1. Brown (p. 243-244) says:

> "Field evidence of faulting is shown by the displacement of the garnet gneiss and by pronounced discordance in structure between the gneiss and limestone.....(southwest of Balmat). Obvious brecciation and granulation are not conspicuous, presumably because the fault was formed at a depth where flowage was as important as fracturing. However, in its clearest exposures, for a distance of several miles, the fault zone~~s~~ is 200 or 300 feet wide with either minute shattering and jointing, or strong flow banding and microgranulation of the rocks involved. There is sporadic reticulation by quartz veinlets. The displacement of the fault is probably horizontal rather than vertical, with a maximum of about 3 miles. The fault is clearly later than the porphyritic granite, but nothing in the field evidence was found to prove that it is later than the non-porphyritic granite to which it seems closely related in origin and effects."

These comments of Brown do not completely reflect his present beliefs nor those of the author. As Brown points out, there is ample field evidence of a northeasterly-trending zone of crackled and brecciated rock, a couple of hundred feet wide, just south of Kellogg Corners and Balmat Corners. From the exposures of crackled rock along its margins, the fault zone appears to be a belt of discontinuous rupture and brecciation, which partly coincides with the dominant foliation and flow surfaces in both the metasediments and the phacolith granite.

These relations were noted by Gilluly (1935) who said that the

> ".....(Balmat fault) must be younger than the igneous injections, for it cuts off and brecciates both the California phacolith (Alexandria type granite) and the porphyritic (Hermon type) granite west of Fullerville."

Gilluly also recognized that the discontinuous crackling and brecciation were induced under conditions of much lower pressure and temperature than those that prevailed when the phacolith was emplaced and the other major structural features were formed, for the folds were formed in large part by flow of solid rocks while minerals of the amphibolite facies were forming, whereas the brecciated rocks along the Balmat fault were being partly replaced as they broke, by albite and chlorite. As may be gathered from inspection of plate 1 and fig. 7, the entire south side of the Sylvia Lake - Edwards isocline(s) was thoroughly mashed and partly sheared away from the rock masses to the south during the more ductile stages of metamorphism and intrusion. It was this deformation and penetrative flow that caused the thinning and at least partial obliteration of the marble south of Sylvia Lake and near Balmat and the plications and flowage phenomena in the quartz-biotite-oligoclase gneiss near Kellogg Corners and Balmat. Intense flowage is also apparent to the northeast, especially along the southern margins of the amphibolite and in the thickened and contorted median gneiss. All of the solid flow and penetrative shearing is clearly older than the crackling and retrograde metamorphism along the Balmat fault, which, however, tends to follow the surfaces of flow. Some of the displacement attributed by Brown (1936, p. 244) to the Balmat fault per se may have resulted from earlier flow shearing. Brown (Brown and Engel, 1956) would now accept this suggestion, at least as a working hypothesis.

The relations are roughly comparable to those that might result if a southwesterly-trending breccia zone at least superficially like the Balmat fault zone formed along, say, the southeastern barb or short flank of the fishhook fold in median gneiss and marble north of Edwards. Even though no actual fault displacements resulted during the stage of discontinuous deformation, a very large displacement might mistakenly be ascribed to the period of brecciation--for the metasedimentary units along the southeast flank of the Edwards fold are in part thinned and dispersed, in part obliterated, by distributive displacements due to solid flow.

Unfortunately the part of the Balmat fault (as defined by discontinuous deformation) that was mapped in conjunction with the present study does not exhibit obvious displacements nor any well-defined subsidiary ruptures or drag folds that would indicate relative displacement. The zone of crackling and shear could, however, mark either a thrust fault or a normal fault that dips about 65° NW. and has a displacement of perhaps 500 or more feet in a direction near the dip.

Talc deposits

Distribution, form, and thickness

The commercial talc occurs in sinuous belts as shown on plate 1. Probably all of these "belts" between Sylvia Lake and a point south of Cedar Lake are segments of the same stratigraphic zone, now much folded and in places pulled apart by regional deformation. For purposes of description, however, each well defined segment of the talc has been given a name. The major segment of the talc-bearing zone south of Fowler, which extends northeastward from a point near the southeast shore of Sylvia Lake and contains the Woodcock, Wight, Arnold, Johnson, Van Amee, and Ontario mines, will be referred to as the Fowler talc belt. The shorter segment of the talc zone south of Fowler, which is exposed south of the Woodcock mine and extends from the vicinity of the old American talc mine shaft southwestward to the extreme southeast shore of Sylvia Lake, is here called the American talc belt. The major segment of talc exposed at and near Talcville, which is traced by workings of the Wintergreen Hill mine, the International Talc Co.'s mines Nos. 2, $2\frac{1}{2}$, 3, 4, and 5, and the Newton Hill workings, will be referred to as the Talcville belt, and the short belt exposed just east of the Trout Lake road and north of the hook-shaped fold in the median gneiss as the Edwards belt.

The fifth body of commercial talc, just northwest of Balmat Corners, has no well-defined outcrops but has been outlined by drilling and by the subsurface workings of the Balmat zinc mine and those of the Gouverneur Talc Co. This body of talc is elongate parallel to the American talc belt, and southeast of it. Since it is best exposed in the Balmat mine, where the talc is closely associated with zinc ore (pl. 31), it will be referred to as the Balmat talc belt.

Fowler talc belt

The data on the extent, form, and thickness of the Fowler talc belt have been utilized in preparing plates 20, 26, 27, 28, 29, 30, and 31, and figures 11 and 12. Plate 1 indicates the known surficial extent of the belt, and shows its outcrops and its inferred form between outcrops. The data for all but the northeast end of the Fowler belt are summarized in figure 11. For a three-dimensional conception of major parts of the Fowler belt, figure 11 and plate 1 should be compared with plate 20. Details of the mined part of this belt are shown in plates 26 to 31, which depict the geologic features of the Woodcock, Wight, Arnold and Ontario talc mines.

As may be seen in plate 1, the southwest "end" of the Fowler belt is just east of Sylvia Lake, and its northeast "end" under the West Branch sand hills. It is not definitely known whether the talc pinches out or is pulled apart at these "ends." At the Sylvia Lake end, data from drill holes suggest that noncommercial extensions of the Fowler belt curve abruptly to the east and connect with the American talc belt. This belt in turn is probably refolded at the southwest to form the Balmat talc belt.

The relations at the northeastern end of the Fowler belt (pl. 1) are inferred in part from the known extent of the talc in the Ontario mine and in part from findings in two drill holes put down by the U. S. Bureau of Mines (Engel, 1948). In this area the Fowler belt has been traced about 1,000 feet east of the West Branch by drifts driven from the Ontario shaft (pls. 29, 30). The total width of the belt rarely exceeds 100 feet and most of the talc rock is low grade. What is inferred to be the Fowler talc belt appears to thin almost to extinction in the area of the West Branch sand hills (drill hole W.H.2), where it consists of about 15 feet of tremolitic marble. Presumably this is about as far northeast as the Fowler belt is recognizable as a talc-bearing unit, for no tremolite nor talc is present in exposures of this zone about 3,200 feet farther northeast (8.3 N., 0.4 W.).

Assuming that its opposite ends are at Sylvia Lake and at the drill hole W.H.1, the Fowler belt has a strike length of almost 20,000 feet. If the contorted segments described as the American and Balmat talc belts are added, the length of the resultant talc-bearing zone is increased to about 6 miles. At the Sylvia Lake end of the Fowler belt its known extent down-dip is at least 3,000 feet, as calculated from drill data of the St. Joseph Lead Co. (J. S. Brown, and N. H. Donald, personal communication). Down-dip extensions of the talc farther northeast are probably at least this deep or deeper.

The general form of the Fowler belt is that of an undulating sheet, which in general strikes northeast and dips northwest at moderate angles, averaging perhaps 45°. The form of the talc is complicated by a series of subparallel drag folds or corrugations that generally plunge 15° to 50° north to northeast (see pl. 20). The plunge of successive folds in the belt increases from southwest to northeast. Near Sylvia Lake folds in the talc plunge 10° to 20° N. (pls. 1 and 20); at the Woodcock and Wight mines (pls. 26 and 28), and at the Ontario mine (pl. 30) the folds plunge about 30° to 40° N. and locally plunge as much as 80°N. These folds therefore converge downward. They also show a common asymmetry: their east flanks are short and diverge at almost right angles from the general strike of the talc belt, whereas their west flanks are long and tend to diverge very little from the general strike. A marked exception to this rule is the complex fold at the west end of the Wight mine (pls. 1, 20, 28).

Several of the most pronounced local thickenings of the Fowler belt appear in the apical parts of these folds (see pls. 2, 26, 27, and 30). Other local bulges occur in slightly sinuous segments of the belt; these alternate with narrower segments or "waists." The bulges in the apical parts of folds tend to preserve their form at successive levels down the axes of the folds, so that the longest axis of a thickening mass lies parallel to the plunge of the fold rather than to the dip of the layers.

In the apical part of the fold at the northeast end of the Woodcock mine and at the southwest end of the Wight mine, the talc belt is from 350 to more than 500 feet thick (pl. 27); away from this fold, in the southwest end of the Woodcock mine and the northeast end of the Wight mine, it is only from 150 to 250 feet thick. Other significant bulges in the Fowler belt appear at the apices of the less marked folds in the Ontario talc mine (pl. 30). There the axis of the more westerly of the two largest folds passes near the shaft. In the apical part of this fold and in another fold apex 300 feet to the northeast, the footwall zone of commercial talc is about twice as thick as in most other parts of the mine. It should be noted, however, that there is no good correlation of apical bulges with commercial talc. Some of the bulges in apices of folds are noncommercial talc, whereas thinner segments along the long flanks of folds are minable.

Several of the apical bulges in the Fowler belt shown on plate 1 are inferred rather than observed, being in areas where there are only scattered exposures. One of these is southwest of the Woodcock mine, near 0.75 N., 10.75 W., and another is northeast of the Arnold mine, near 4.8 N., 7.1 W.

Distinct from the bulges at apices of folds are the many swellings of the Fowler talc belt, and especially of zones within the belt, that occur in nearly straight or only slightly sinuous parts of the belt. Many of these bulges are not greatly elongated in any direction, and where perceptible elongations do appear they form no systematic pattern. Bulges of this type are common in the Arnold mine, in the northeast end of the Wight mine, and between the several incipient folds in the Ontario mine. At these places some layers within the talc belts pinch and swell along the strike and down dip somewhat randomly. The swells may increase the width of a zone by as much as 50 percent, and their two larger dimensions may be as much as 10 times the width of the bulge. Near areas of folding the bulges tend to become more elongate, with their longest dimensions subparallel to the axes of folding. In many places there are bulges and thinnings of intermediate types which have the form of boudins, or sausagelike swellings. These commonly occur along the flanks of folds. They are generally parallel to the axes of folds and lie in or near slight warps in the rock layers. Typical boudins occur throughout the central and southwestern parts of the Woodcock mine and on the flanks of the folds in the Ontario mine. At these places most of the bulges and associated pinches in talc zones can be traced from one sublevel to the next, or even from level to level, along lines about parallel to the axes of folding (see especially pls. 26 and 30).

Locally the elongate bulges or boudins may plunge at about right angles to the axes of the cross-folds. Two notable examples of these bulges are major structural elements in the footwall zone of commercial talc rock at the Arnold mine (pl. 31). The very large bulge on the 8th level, some 200 to 400 feet west of the shaft, appears farther and farther west on successive levels below level 8 but is said to be very near the shaft in the now inaccessible higher levels of the mine. Similarly, the marked constriction of the footwall zone that passes near the shaft on the 15th level of the mine is encountered farther and farther east on successively higher levels; on the 8th level it lies about 1,000 feet northeast of the shaft. The direction of plunge and the form of these bulges and thinnings are best seen in the plan of the mine workings. They rake about 20° to the west and plunge about 50° SW. Their longest axes in this area are in the plane of the talc zone but about at right angles to the axis of cross-folding. Features of this same type also appear in the siliceous interlayers in the marble nearby (pls. 10A, 13A). Excellent examples are exposed near 1.5 N., 9.85 W. just west of the Balmat road, where talcose boudins in the siliceous marble lie parallel to one another on the flanks of folds, with the longest dimension of each parallel to the flank and about perpendicular to it. These boudins appear to have formed by stretching of the layers along lines parallel to the fold axes, during the folding. The large bulge and the nearby constriction in the footwall zone of the talc at the Arnold mine are much larger but appear to be otherwise similar.

American talc belt

The form, distribution, and thickness of the American talc belt are shown (pls. 1 and 20, fig. 11). Near the southeast shore of Sylvia Lake this belt is abruptly refolded southeastward to form the so-called Balmat talc belt. The eastern end of the commercial part of the American talc belt is near the Balmat road at 0.75 N., 10.0 W. Workings of the Gouverneur Talc Co.'s mine follow the American belt this far northeast but no talc is cored in drill holes several hundred feet farther northeast, and no similar talc rock is found in the outcrops east of the road. It is probable that this talc belt is folded abruptly west and pulled apart, and was originally connected to the Fowler talc belt as shown on plate 1.

Assuming that opposite "ends" of the American talc belt are at Sylvia Lake and the Balmat road, it is about 2,800 feet long. Its width varies considerably, ~~and~~ reaching a maximum of at least 100 feet in the vicinity of the Gouverneur Talc Co.'s mine. This belt, like the overlying Fowler talc belt in this area, extends down dip about 3,000 feet, although below 1,500 feet much of the talc is non-commercial.

The American talc belt lies less than 1,000 feet southeast of the Fowler talc belt (pls. 1 and 20, figs. 4 and 11). At the surface, well-defined segments of these two talc belts are separated by a zone of partly calcitic, partly dolomitic marble some 500 to 800 feet thick. At depth, however, the two belts tend to converge, as shown in figure 11 and by structure contours along the footwalls of the two belts (pl. 20). Drill-core data indicate, however, that these two belts do not join at depth, but remain separated by a thin zone of marble.

Talcville talc belt

The Talcville belt seems to be at the same general stratigraphic position as the Fowler belt. The southwest end of the commercial talc in the Talcville belt is at the Hyatt zinc mine. The most westerly exposure of the talc is at 11.65 N., 2.25 E. Southwest of this point, extensive drilling and mine workings have failed to encounter more than scattered wisps of tremolitic talc, and it seems unlikely that the talc belt is a continuous or well-defined unit there.

The northeasterly termination of the Talcville belt is on Newton Hill (pl. 1), where the layers of tremolitic talc rock of the Talcville belt merge along the strike into layers of highly diopsidic and siliceous dolomitic marble. The gradation between these rocks occurs along a strike distance of about 200 feet (pl. 24) and is fairly well exposed.

The intermediate parts of the Talcville belt, as exposed and inferred on the surface and in mine workings are shown on plates 1, 20, 21, 22, and 23. Its strike length from the Hyatt mine to Newton Hill is approximately 10,000 feet; its extent down dip to the northwest is not known. The deepest mine workings and drill holes at the International Nos. $2\frac{1}{2}$ and 4 mines have cut talc rock at depths of 800 to 1,000 feet. The thickness of the Talcville belt varies greatly and reaches about 300 feet in some places.

The Talcville belt is less distinctly zoned than the Fowler belt and contains fewer and less extensive interlayers of marble (compare figs. 12 and 13). Some folds in the Talcville belt also diverge considerably in form from the asymmetric drag folds that characterize the Fowler belt. In the Wintergreen and Hyatt mines, for example, the folds have very irregular forms. Most of the folds in the International Nos. $2\frac{1}{2}$ and 4 mines, however, are more similar to those prominent elsewhere in the district. They tend to have short east or northeast flanks and long northwest flanks that more nearly conform with the regional trends of all rock layers.

In the Talcville belt, as in the Fowler belt, bulges in the talc rock tend to be associated with the apices of folds. This tendency is illustrated at the Wintergreen Hill and International Nos. $2\frac{1}{2}$ and 4 mines (pls. 22 and 23), and in the folded rocks of the Talcville belt west of the shaft on Newton Hill (pl. 24). At major folds the belt is thickened at least 50 percent. These relations are especially apparent at Newton Hill and in the vicinity of the U. S. shaft of the International No. $2\frac{1}{2}$ mine. At Newton Hill (pl. 24) the Talcville belt is perhaps 75 feet thick except in the area about 300 feet west of the shaft. There the layers of talc rock are crumpled into an asymmetric fold, which dies out abruptly across the strike in the enveloping marble. At the apex of the fold the talc belt swells to a thickness of about 125 feet. This bulge appears to plunge at a moderate angle northwestward, about parallel to the associated folds in other rock layers. The greatly thickened node in the central part of the International No. $2\frac{1}{2}$ mine, opposite the U. S. shaft (pl. 22 and fig. 13) is a tight sigmoidal fold in the talc belt. The short flank or central part of the sigmoid fold is largely obliterated, and the apices of the folds are squeezed together in a large knot (fig. 13). At this place the hanging wall is separated from the footwall by about 250 feet of the talc rock, measured normal to the dip, whereas nearby in both directions along the strike, the talc belt is less than 150 feet thick.

Because the Talcville belt contains more closely spaced major folds it has fewer of the randomly oriented and less elongated bulges than does the Fowler belt. With the possible exception of the most westerly large fold in the International No. $2\frac{1}{2}$ mine (pl. 22 and fig. 13), the folds in this belt plunge and rake either slightly northwest of the dip or very nearly down the dip. They tend to lie subparallel to one another or to converge slightly downward. If the plunges of these folds are compared with those of the folds in the Fowler belt, it is found that the two groups of folds converge toward a point several miles down dip (pls. 1 and 20).

Stratigraphic relations

Most parts of the talc belts are essentially conformable with associated relict beds and stratigraphic zones and there seems little doubt that the talc has replaced pre-existing bedding. There are of course, some marked and spectacular complications of the sedimentary relations of beds in and along the talc belts, as in other lithologic units. These complications include abrupt thinning and thickening of the belts, evidence of tearing apart and shearing off of layers and zones, profoundly incongruent or dissimilar folds which rapidly swell or die out if traced across the layering, and so on. Almost without exception these features appear to be metamorphic in origin and superimposed upon regular and persistent bedding.

Perhaps the most obvious disturbance of the original stratigraphic relations may be seen at Newton Hill (pl. 24). Both west and northeast of this hill the Talcville talc belt is separated from the feldspathic median gneiss by two major stratigraphic units, the pyritic silicated marble and the siliceous calcitic marble (pl. 1). Quartz schist also is found at these places but is believed to represent, in part at least, a silicified phase of the siliceous calcitic marble rather than a distinct stratigraphic unit. Both the pyritic silicated marble and the footwall marble can be traced for long distances on either side of Newton Hill, but they thin toward the hill, and near its top they are completely absent. This pinching out is clearly due to deformation, for it is associated with a steeply plunging fold in all the units, and the limestones are only locally present on the crest of this fold. As a result of this deformation, the lowermost layers of the Talcville belt on Newton Hill lie directly in contact with the top of the median gneiss. This relationship persists to depths of at least 500 feet, as indicated by drill holes through the talc.

Analogous relations on a smaller scale are apparent at most exposures of contacts between the talc belts and marble. Typical examples appear in figures 12 and 13, which are sketches of parts of the contacts at the International No. $2\frac{1}{2}$ mine, Talcville, and at the Woodcock mine near Balmat. These contacts are of much the same type as the one at Newton Hill but involve units less than 10 feet thick rather than major stratigraphic zones.

The shearing out of layers and differential slippage between them has not only caused some stratigraphic discordances, but has resulted in the reconstitution of parts of some beds as "talc," while other parts of these same beds, in either direction along the strike or down dip, were less deformed and altered and still consist of marble.

At or near the contacts of talc belts and marble, distinctive, competent beds commonly have not been obliterated by shearing and therefore can be used as datum planes in deciphering structure. Some examples are the quartzitic beds enclosed in the footwall marble of the Fowler, American, and Balmat talc belts (pls. 4B, 4D), and the highly graphitic beds at the Woodcock mine. The relations of the distinctive quartzitic interbeds in the footwall marble immediately below the Fowler and American talc belts indicate a remarkable degree of conformity of those contacts along most of their extent. In the Fowler talc belt, for example, the footwall contact of talc and marble is well exposed in workings of the Wight and Arnold mines for a distance of some 4,000 feet along the strike, and in some places for almost 1,000 feet down dip. Throughout this area, the contact of the talc rock and marble does not deviate more than 40 feet from one stratigraphic horizon. In the Arnold mine a single zone of diopsidic quartzite layers--locally fragmented and somewhat pulled apart but readily identifiable--lies within 20 feet of the contact for a lateral distance of 750 feet. A similar, close approach to conformity along the footwall of the Fowler talc belt is found in the Ontario talc mine where the talc is underlain by distinctive layers of diopsidic quartzite.

At all of these places the basal layers of the Fowler talc belt, commonly containing much tremolite, lie in contact with thin beds of diopsidic quartzite and marble (pl. 4B). The contact is abrupt, with little interlayering of talcy layers and marble. It is commonly slickensided because shearing has generally taken place along the contact or in a narrow zone that includes it. Along most of its extent the contact is gently undulated, but in some places, for example at the southwest end of the Wight mine, it is very sharply folded. In the Wight mine the layers of diopsidic quartzite in the underlying marble are contorted, shattered, and complexly dispersed in the enveloping carbonate (pl. 6C). On the surface nearby, at the southwest apex of the complex fold in the talc rock at 1.75 N., 9.7 W., and across the Balmat road to the southwest, rock movements have been so complex as to obscure the interrelations of talc rock and the marble.

The contact of the Fowler talc belt with Zone 12 (hanging wall dolomite) is less sharp than that on the footwall, but the zone of transition or interlayering of talc rock and marble is rarely more than 10 to 15 feet thick. The hanging-wall dolomite persists throughout the extent of the Fowler talc belt; that is, from Sylvia Lake to a point east of the West Branch of the Oswegatchie River, a distance of about 6 miles.

The Talcville talc belt and adjacent marbles are more complex in structure and perhaps less conformable with each other than the talc belts and marble to the southwest. They also are less well exposed than the Fowler talc belt, and their precise interrelations are less well known except in the International Nos. 2½ and 4 talc mines and at Newton Hill.

The relations of the Talcville belt to the disrupted footwall marble at Newton Hill were described on page~~s~~ 309 ~~151-152~~. The marble on its hanging-wall side is not fractured and its contact with the talc is persistently conformable with relict bedding not only at Newton Hill but as far west as the International No. 4 talc mine. There, however, the hanging-wall marble is much squeezed and is cut by numerous bodies of pegmatite, thin sheets of granite, and quartz veins. Stratigraphic zones cannot there be traced with certainity from the scattered surface and underground exposures of altered marble, but in most outcrops where bedding can be identified the talc layers are roughly conformable with the bedding. This also is true for the footwall side of the talc belt at the Nos. 3, 4, and 5 shafts of the International Talc Co., where the footwall, elsewhere usually calcitic and siliceous, is mostly dolomite with scattered splotches, lenses, and beds of serpentinized diopside.

Southwest of the No. 3 shaft and northeast of the immediate vicinity of the No. $2\frac{1}{2}$ mine there are no exposures of the talc zone or of the marble in either the hanging wall or the footwall, but in the one hole drilled into the talc from the granite on the northwest the hanging wall is dolomite like that to the northeast.

At the No. $2\frac{1}{2}$ mine, surface exposures are abundant and many contacts are exposed in the mine workings. In an attempt to determine the relations of the talc belt to original bedding, careful search was made for distinctive marker beds, many of which appear from place to place in the mine; and two of them, a graphitic layer in the marble and a micaceous amphibolite in the talc, were chosen as being especially distinctive (fig. 13). The graphitic layer is almost certainly a bed. The amphibolite is probably a reconstituted mafic tuff or other sediment. Where least altered, it is a diopside-hornblende-oligoclase-quartz schist. However, most of the amphibolite in the No. $2\frac{1}{2}$ mine is strongly sheared and has been converted to biotite schist. Both units appear useful in determining to what extent the contacts between talc and wall rock, and the secondarily formed lithologic zones and layers in the talc belt, conform to the bedding of the wall rock.

The general agreement in the trends of the primary layering and laying of metamorphic origin is striking, especially in view of the complex folding and slippage in both the talc belt and the marble. In the west-central part of the No. $2\frac{1}{2}$ mine, however, there is marked discordance on contacts of talc and marble with both the graphitic bed and the amphibolite. In the vicinity of F, in figure 13, for example, the graphitic bed is almost obliterated, but its probable position in this area (E-F) is indicated. The relations imply a marked local discordance between the attitude of the talc-marble contact and bedding near F, and the attitude of the footwall at and east of C. At F, a long sliver of what is normally the hanging-wall marble appears to have been folded into the talc belt and converted to talc rock. Farther east, at B, the sliver has undergone similar but less advanced alteration. These relations in the No. $2\frac{1}{2}$ mine indicate a greater discordance between the original bedding and layering of metamorphic origin than has been observed at any other place in the district.

North of the Edwards zinc mine, at and east of the Trout Lake road near 18.7 N., 11.1 E., talc rock is exposed in a belt that is perhaps too small to be of commercial value, but whose relations to the marble are of some geological interest. The contacts and foliation of this talc belt conform as closely with associated relict bedding as those of the Fowler talc belt. The marble in both the hanging wall and the footwall is predominantly dolomitic; less than 15 percent of it consists of interbedded siliceous or silicated calcite marble. Some quartzitic and diopsidic beds form a prominent zone about 50 feet above the hanging wall of the talc. As shown on plate 1, these and the beds of siliceous marble in the hanging wall are essentially parallel to the upper contact of talc and dolomite. The footwall is likewise parallel to the few siliceous interbeds in the underlying marble, which are in turn parallel to the contact of this marble with the feldspathic median gneiss.

The alined positions of the Fowler, Talcville, and Edwards talc belts, at approximately the same distances above the median gneiss and below the silicated marble zones, constitute strong evidence that these talc belts are all at roughly the same stratigraphic horizon. There seems to be little doubt that the median gneiss and silicated marble are continuous stratigraphic units which clearly bracket the position of the three talc belts.

Both the footwall and the hanging-wall marbles of these talc belts vary somewhat in composition along the strike, and, as shown later, there appear to have been lithologic variations along the strike in the beds now replaced by talc. Probably, however, neither the top nor the base of the zone along which this talc has formed varies in stratigraphic position by more than 200 feet of original beds. In view of the degree and extent of deformation and alteration in the marble, this is impressive evidence that the formation of talc rock along a strike length of almost 10 miles was stratigraphically controlled.

Lithologic features of marble replaced by talc

The nature of the parent marble now replaced by the talc belts is speculative for many reasons. If the physical conditions required for talc formation persisted for a protracted period, most or all of the marble susceptible to alteration to talc rock may have been so converted. If this is true, the remaining masses of marble within or alongside the talc belts escaped alteration because of critical differences in composition, and the existing intergradations between the talc rock and marble, either along or across the strike, reflect a gradation in some critical compositional property of the parent rock. On the other hand, it may be argued that the talc-forming processes were interrupted at least locally by critical changes in temperature, PCO_2, or other aspects of the physical environment, and in these areas there are relicts of marble compositionally much like that converted into talc. If this supposition is correct there still exists the problem of recognizing which are the relicts of marble most like that converted to talc.

The composition of most of the marble zones in the district is remarkably simple and apparently did not vary abruptly or irregularly along the strike before metamorphism. The same is probably true, therefore, of the rocks parent to the talc belts.

It is unfortunately also true that the chemical and physical features of the parent rock recognizable today are not precisely those of the rock during talc formation. Physical properties which are strongly dependent upon temperature or pressure certainly have changed; and compositional changes have been imparted by changes in amounts and kinds of pore fluids, and so on. Thus any subtle variations in porosity or permeability, for example, that might have been critical in providing access to metasomatizing fluids or otherwise localizing talc formation could have been altered or obliterated during or after metamorphism.

Parent rocks of talc deposits

What seem to be relict beds within the talc at various stratigraphic positions suggest that talc has replaced marble of widely different composition. Most of these relicts are carbonate-rich. Relicts of highly quartzose beds are not found either interlayered or widely and abundantly distributed throughout the talc. At least locally, however, especially in the footwall zone of the Fowler talc belt, the rock now altered to talc seems to have consisted of interbedded quartzite and dolomite (pls. 13C and 13D). That near the hanging wall, on the other hand, not only in this belt but in considerable parts of the Talcville and Edwards belts, appears to have been massive dolomite that contained only a little disseminated and bedded quartz.

In the American talc belt, also, dolomite appears to have preponderated over siliceous marble in the footwall zone, although quartzitic interlayers were apparently more abundant than in much of the Talcville belt. In the Balmat talc belt, however, what seem to be relict beds contain from 15 to 30 percent of silica in quartzitic laminae, layers and possibly nodules.

The marble altered to talc apparently contained almost no argillaceous interbeds, nor is there any reason to believe that much of it contained particles of silica uniformly disseminated through the carbonate. Instead, the silica appears to have been concentrated in layers and lenses alternating with thicker layers of carbonate, all of which appear to have originally been dolomite before the rocks were metamorphosed.

Summing up, at least 75 or 80 percent of the marble now replaced by the talc belts seems to have been originally almost pure dolomite, the remainder being chiefly silica in quartzitic or cherty interlayers.

This conclusion is based upon several kinds of relations observed during field mapping of the talc beds. In some places part of a bed in the wall of a talc belt was altered to talc, whereas other parts of the same bed remained unaltered. Where this has occurred the lithology of some of the marble converted to talc rock is readily inferred.

What is perhaps the largest clearly recognizable example of this relation is in the west-central part of the International No. $2\frac{1}{2}$ mine, where the hanging wall consists mainly of dolomite, Zone 12, with less than 10 percent of quartzitic beds. Near F on figure 13, an irregular lens of this slightly quartzitic dolomite, at least 25 feet and possibly as much as 50 feet thick, which lies immediately below the graphitic bed employed as a datum in mapping, has been converted into a mixture of talc, tremolite, and anthophyllite. Less extensive alteration of dolomite to talcose rock appears in other parts of the mine, especially in the vicinity of B, figure 13. As there are no other abrupt changes in the composition of the hanging-wall marble except in the small area where it is altered to this talcose and diopsidic rock, it is unlikely that the dolomitic marble now dominant in the hanging wall contained irregular lenses of some other rock that was replaced by talc. More likely, lenticular segments of the hanging-wall dolomite were replaced by talc because of changes in physical environment, which may in turn have facilitated metamorphic changes in composition.

Analogous examples of the conversion of dolomite to talc, though on a smaller scale, may be observed along the hanging wall in every mine in the Fowler and Edwards talc belts and along the footwall of the American talc belt. Parts of highly dolomitic beds are also converted to talc along the footwall side of the Talcville belt (1) at the International Nos. 3, 4, and 5 shafts northeast of Talcville, (2) at Newton Hill, and (3) north of Edwards. At these places there are no relicts of siliceous marble in the walls of the talc belts, nor any fragments of such rock within them, so that all the rock converted to talc may originally have been highly dolomitic. North of Edwards, the marble in both the footwall and the hanging wall is dolomitic. Moreover, to the northeast, virtually all the beds on the strike of this talc consist of dolomite with less than 10 percent of silica.

Layers of carbonate-rich rock alternating with layers and lenses of talc rock are also especially common in the Fowler talc belt. They appear in every mine on this belt, and many of them are 10 to 25 feet thick. They are now calcitic, but they are inferred to have been dolomitic before talc rock was formed ~~(see p. 51)~~. In a great many of these lenses, talc and carbonate grade into each other both along and across the strike. Some of the clearest examples of this gradation appear in the Woodcock mine (see fig. 12). Large parts of the two intermediate talc zones shown in figure 12 appear to have replaced carbonate, for they enclose and merge into irregular remnants of carbonate rock. There, as in the places noted earlier, it seems clear that the talc has replaced a nearly pure dolomite.

The quartzitic marble which underlies the Fowler and Balmat talc belts and also some parts of most other talc belts, has been converted to talc rock in many places. This alteration has been observed along the footwall of the talc belt in the Ontario, Arnold, Wight, Woodcock, and Gouverneur Talc Co. mines. Both gradual and abrupt transitions of tremolitic rock to quartzitic interlayers are fairly common in these mines. Relations of this type are well shown just east of the shaft in the Wight mine and also along the margins of the Balmat talc belt in the Gouverneur Talc Co.'s mine. Relict lenses of quartzitic marble within the talc belts are visible just above the footwall in the Wight and Arnold mines, and are especially conspicuous at the eastern end of the American talc belt in the Gouverneur Talc Co. mine (see pl. 13C). Quartzitic beds have been partially replaced by tremolite, anthophyllite, serpentine and talc both along and across the strike but in almost every exposure that shows this replacement the associated interlayers of carbonate and of diopsidic rock have been more completely replaced than the quartz.

Other clues to the nature of the marble that was replaced by talc rock would seem to exist at what are regarded as the ends of the talc belts, especially at the places where commercial talc rock fingers out in marble along the strike. The Fowler, Talcville, American, and Edwards talc belts finger out at both ends into rocks that range from almost pure dolomite to marble rich in quartz or in tremolite and diopside. The northeast ends of the Edwards and Fowler talc belts tongue out in dolomite that contains only a little quartz and diopside. At the west end of the Fowler talc belt, on the other hand, data from drill cores indicate that layers of quartz and diopside are abundant along strike from the talc. The east end of the Talcville talc belt grades into highly siliceous and diopsidic marble, which in turn fingers out into dolomite. Farther northeast along the strike, however, quartzitic and diopsidic layers reappear along the talc-bearing horizon, and still farther northeast, near the Trout Lake road, there are some lenses and thin layers of talc rock.

It has been noted that perhaps half of the quartz in the marble is introduced. Consequently, the composition of the lateral extensions of the talc belts also may have been modified by metamorphism as or after the talc was formed, and the nearby talc may have replaced nearly pure dolomite. There seems to be no means at hand to resolve this question, just as there is no positive evidence that in some places at least the ends and sides of bodies of talc rock do not coincide with abrupt changes in the composition of the parent rock. But some exposures appear to give almost unequivocal evidence that some talc has replaced nearly pure dolomite, whereas other talc has formed at the expense of siliceous dolomite of sedimentary origin.

This conclusion raises an important question. If the beds in which talc was forming varied in composition from place to place, and were in general like the composition of some other parts of the marble not altered to talc, what was the nature of the stratigraphic control? The answer is not clear. Subtle variations in texture, of course, may have been major influences in talc formation, but there seems to be no way to clarify whether an original textural control existed. Another possibility is that marked physical control was exerted by a major zone of shearing more persistent and pervasive than others in the marble. This suggestion carries with it the question of why a different kind of shear zone evolved in this specific part of the marble. I see no very concrete or definite answers to these questions, but they are better considered in the light of mineral and chemical changes accompanying formation of the talc, a discussion of which follows.

Mineral and chemical changes accompanying formation of talc rock

The average composition of the sedimentary rock now replaced by the talc is thought to have been that of dolomite containing about 15 to 20 percent silica. If that is true, the ratio of MgO to CaO in the marble now altered to talc was about 7 to 10 (table 24), and the ratio of SiO_2 to MgO reached a maximum of perhaps 3 to 1 and averaged about 1 to 1.

Table 24. Near here. (p.266

Table ... chemical, mineralogical and density data employed to calculate changes in constituent oxides in the average talc belt during metamorphism. A constant thickness of 230 feet is assumed for the belt.

	Tremolite	Talc	Serpentine	Anthophyllite	Quartz	Calcite	Dolomite	Total	Weighted average density	Total original weight*	Total percent* weight
Mineralogical composition, average talc belt, approx. volume percent	50	18	10	7	3	12	tr	100			
Density, constituent minerals (Dp)	3.1	2.75	2.51	3.00	2.65	2.72	--		2.92		671
Inferred mineral composition of marble altered to talc					15		85	100			
Density constituent minerals (Do)					2.65		2.85		2.82	649	
Approx. chemical composition	SiO_2	MgO	CaO	CO_2	H_2O						
Average talc belt, percent oxides	51.3	26.4	13.4	5.3	3.6			100			
Inferred chemical composition of marble altered to talc	15.	18.5	25.8	40.7				100			
Inferred original weights, oxides	97	120	168	265				650		649	
Present weights oxides	343.5	177.1	90	35.5	24.2			670.3			
Percent additional(+) or less(-) in weight at constant volume	+246.5	+57.1	-78	-229.5	+24.2						

*Calculations based upon an average thickness of talc belts of 230 feet which is assumed to remain constant during metamorphism; thus 230 x Do (2.82) = 649; 230 x Dp (2.92) = 671.
thus 230 feet (thickness) x density original zone (Do) = 649
and 230 f[illegible]thickness) x density present talc belts (Dp) = 671

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If the talc belts formed by reaction and replacement of dolomite and siliceous dolomite the process must have involved changes in composition and probably of volume. The present ratio of SiO_2:MgO:CaO in the talc belts is about 4:2:1 (table 24). No possible mixture of sedimentary silica and dolomite could give this ratio. Even greater changes would be involved if some of the sedimentary carbonate in the marble was calcite. As was previously pointed out ~~(p.——)~~, no evidence has been found that the marbles contained magnesite either before or during its alteration to talc.

One way to point up the kind and amount of change involved in the formation of the talc is to compare the present weighted average chemical and mineral composition and specific gravity of the rock in the talc belt with those of the siliceous dolomite from which it is assumed to have been derived (table 24). The listed figures for the talc belts include all of the rock, both commercial and non-commercial, mapped as talc belts on plate 1. The data are based on microscopic and chemical analyses of specimens from the talc belts, taken from surface exposures, mine openings, and drill cores. The mineralogical and chemical compositions of the talc rocks and also those inferred for the parental marble, are listed in table 3, along with calculated compositions of other units in the marble. For additional comparisons, figure 14 shows the percentages of SiO_2, MgO, and CaO in the talc beds, and also in the marble zones of commercial talc. The figures show that the evolution of talc belts from marble, and of commercial talc from the average rock of the talc belts, involves successive increases in the ratio of SiO_2 + MgO to CaO (or of SiO_2 + MgO:CaO + CO_2).

Assuming that the values for talc rock and for inferred parent rock are about right (table 24), the transformation of marble to talc rock involved an increase in density from about 2.82 (for siliceous dolomite) to 2.92 (weighted average for the talc belts) (table 24). If the transformation took place without appreciable change in volume, there were major additions of silica, magnesia and water. Assuming that the average thickness of the talc belts is 230 feet, SiO_2 increased about three-fold and MgO almost 50 percent. Just how much water persisted in the marble during and after the high-temperature stages of metamorphism is hard to estimate.

That the volume of rock remained constant while the talc rock was being formed is doubtful, but unless the talc belts increased in volume, about 90 percent of the carbon dioxide and nearly half the CaO were expelled during their evolution. A decrease in volume is to be expected, but unless this decrease was on the order of 75 percent the requisite MgO and SiO_2 cannot have been derived from the parental marble. The alteration of dolomite to the siliceous calcitic marble of the footwall appears to offer a nearby source of magnesia, silica, and water for the talc-forming process. If the previous account of the changes in composition of the rusty marble and siliceous footwall marble is about correct, enough MgO could have been leached from these units to satisfy the requirements of the talc belts with only slight decrease (10-20 percent) in their volume during metamorphism.

Chemical composition of the deposits

The chemical compositions of channel samples from the talc deposits are given in tables 25 and 26 and figure 14. The analyses in table 25

Table 25. Near here. p.270

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represent the average composition of each of five of the largest and most uniform and persistent bodies of commercial talc in the district; those in table 26 represent smaller zones and sub-zones that have been

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Table 26. Near here. p.271

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mined, or that could be mined, as independent units. The samples analyzed include some of the most valuable talc rock that has been mined in the district and some that can scarcely be mined at a profit. Inasmuch as value is reflected in large part by composition, it follows that very little rock is mined and sold whose composition does not fall somewhere between the limits defined by the analyses in table 26.

Table 25. Chemical analyses of channel samples across each of 5 major belts and zones of commercial talc in the Balmat-Edwards district.

	T1	T2	T3	T4	T5	Average of 5
SiO_2	57.46	57.58	59.40	59.40	66.13	60.07
TiO_2	.04	n.d.	n.d.	n.d.	n.d.	n.d.
Al_2O_3	1.14	.57	.74	.57	1.05	0.81
Fe_2O_3	* .23	.03	.02	.05	.13	.09
FeO	.05	.09	.12	.15	.22	.12
MnO	.51	.31	.20	.39	.16	.31
MgO	29.18	28.65	30.09	27.25	25.71	28.19
CaO	6.50	6.86	4.94	6.80	2.26	5.49
H_2O-	.34	.54	.47	.44	.25	0.41
H_2O+	3.98	5.39✓	4.09✓	4.75✓	3.86	3.86
CO_2	.29	1.28	.31	1.18	.56	0.72
P_2O_5	n.d.	n.d.	n.d.	.06	.07	n.d.
SO_3	.14	.07	.01	.07	.01	0.06
BaO	.04	n.d.	n.d.	n.d.	n.d.	n.d.
Cl	n.d.	n.d.	n.d.			n.d.
F	n.d.	n.d.	n.d.			n.d.
Total	99.91	100.09	100.08	100.11	100.08	

*Probably in large part contamination from steel grinding equipment.

✓Total loss on ignition (includes CO_2).

Table 26. Compositions of commercial talc, Balmat-Edwards district. Each represents an analysis of a channel sample across a specific, minable zone of commercial talc within the talc belts.

	CT 1	CT 2	CT 3	CT 4	CT 5	CT 6	CT 7	CT 8	CT 9	CT 10	CT 11	CT 12	CT 13	CT 14
SiO_2	56.83	57.05	62.03	56.33	60.74	64.50	59.63	63.17	62.94	66.73	56.40	56.10	57.2	57.98
Al_2O_3	2.40	1.30	1.54	1.01	.82	1.31								
Fe_2O_3	.17	2.01	.46	.12	.21	.07	1.16	.76	1.28	.66	1.60	1.4	1.2	.54
FeO														
MnO						.41	.69	.22	1.13	.28	.86	.35	0.4	.22
MgO	28.17	29.08	24.57	30.45	32.63	27.60	27.62	21.43	25.56	23.95	28.92	30.15	30.50	30.90
CaO	7.44	7.05	7.41	7.53	1.67	4.01	7.06	8.09	4.36	3.84	6.40	6.40	4.50	4.58
H_2O-	.59	.55	.39	.77	.42	.49	.34	.40	.40	.37	.51	.59	.46	.46
H_2O+	4.41	2.73	2.48	3.82	3.15	1.93	2.81	2.73	2.51	3.69	4.50	4.59	4.84	4.87
CO_2	0.46	0.27	1.31	0.23	0.21	0.20	.42	3.36	1.62	.72	0.70	.30	.80	.24
SO_3						tr			tr		.06		.09	tr
Total	100.47	100.04	100.19	100.26	99.85	100.52	99.73	100.16	99.80	100.24	99.95	99.88	99.99	99.79

	CT 15	CT 16	CT 17	CT 18	CT 19	CT 20	CT 21	CT 22	CT 23	CT 24	CT 25	CT 26	CT 27	CT 28	Average	
SiO_2	60.4	64.38	62.92	64.01	64.58	67.00	68.32	54.74	57.80	58.80	63.99	58.37	64.84	70.60	61.38	SiO_2
Al_2O_3										.85						Al_2O_3
Fe_2O_3	0.6	1.02	1.12	1.58	0.90	1.4	0.70	.46	.48	.32	.94	1.02	0.50	0.64		Fe_2O_3
FeO																FeO
MnO	.44	0.28	.25	0.60	0.15	0.8	.49	.21	0.42		0.83	.96	2.48	2.21		MnO
MgO	30.60	23.82	25.14	23.60	27.23	24.40	22.80	29.80	28.70	28.30	23.60	24.20	26.27	22.40		MgO
CaO	2.80	5.54	5.86	4.60	2.30	2.10	3.58	7.08	6.80	6.10	6.50	11.18	2.78	0.96		CaO
H_2O-	.39	.39	.36	.36	.41	.31	.33	.78	.51	.21	.36	.41	.30	.27		H_2O^-
H_2O+	4.20	4.50	3.51	3.65	3.77	2.70	3.24	5.00	4.70	4.40	2.66	3.31	2.79	2.65		H_2O^+
CO_2	.80	0.33	0.73	1.60	0.54	1.30	.39	2.04	0.56	.95	0.90	0.62	0.05	0.05		CO_2
SO_3		tr	tr	.06	tr	.07	tr	tr	.05		tr		0.08	0.08		SO_3
Total	100.23	100.26	99.89	100.06	99.88	100.08	99.85	100.11	100.02	99.93	99.78	100.07	100.09	99.86		Total

CT 1 - CT 6 inclusive, Ledoux & Company, analyst.
CT 7 - CT 16 inclusive, Orton Smalley, analyst.
CT 17 - CT 21 inclusive, Celeste Engel, analyst.
CT 22 - CT 28 inclusive, Orton Smalley, analyst.

The chemical composition of the talc reflects its mineral constituents. In table 25, for example, the percentage of silica ranges from 57.4 to 66.2, of MgO from 25.7 to 30.1, and of CaO from 2.3 to 6.9. If we compare these percentages with those in the constituent minerals (table 23) we see that the large bodies of talc

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Table 23. Near here. p. 273

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rock contain about the same percentages of these oxides as tremolite, and in fact much of the talc rock does consist predominantly of tremolite. Silica percentage could be raised to about 63 percent by increasing the proportion of the mineral talc, or of quartz, or of both. Some small bodies of ore rich in the mineral talc (see table 26) contain 1 to 7 percent more silica than is found in pure talc (63 percent by weight); most of these bodies contain free quartz. Low silica percentage indicates the presence of serpentine, which contains about 44 weight percent SiO_2. All the larger ore bodies and most of the small ones contain more than 55 percent SiO_2, which indicates that little of the talc rock contains a large proportion of serpentine. This same fact is reflected in the range of MgO from 25.7 to 30.1 percent in large talc bodies and from 21.4 to 32.6 percent in small ones (tables 25, 26). Serpentine-rich talcs would contain up to 43 percent MgO.

Table 23. Approximate chemical compositions of the common minerals in the talc belts and in closely associated marble.

	Formula	Percent by weight of major component in mineral formula					Approximate specific gravity	Grams of each oxide in 1 cc of mineral (assuming density as indicated)				
Mineral		CaO	MgO	SiO_2	H_2O	CO_2		CaO	MgO	SiO_2	H_2O	CO_2
Diopside	$CaMgSi_2O_6$	25.9	18.5	55.6			3.23	0.84	0.60	1.79		
Tremolite	$Ca_2Mg_5H_2(SiO_3)_9$	12.5	27.5	58.	2.0		3.00	0.37	0.83	1.74	.06	
Anthophyllite	$5Mg_7H_2(SiO_3)_8$	3.0	35.	60.	2.0		3.00	.09	1.05	1.80	.06	
Serpentine	$H_4Mg_3Si_2O_9$		43.0	44.1	12.9		2.51		1.08	1.11	0.32	
Talc	$H_2Mg_3Si_4O_{12}$		31.7	63.5	4.8		2.75		0.87	1.75	0.13	
Dolomite	$CaMg(CO_3)_2$	30.4	21.7			47.9	2.85	0.87	0.62			1.36
Calcite	$CaCO_3$	56.0				44.0	2.75	1.54				1.21
Quartz	SiO_2			100.0			2.65			2.65		

The CO_2 content of the talc rocks, which ranges from about 0.2 to as much as 3 percent or very locally more, is an index of their calcium carbonate content, for they contain very little dolomite or other carbonate than calcite. When enough CaO to combine with this CO_2 (as $CaCO_3$) is subtracted from the total CaO, an appreciable amount (up to 7 percent CaO) commonly remains, in the omnipresent tremolite.

Samples CT 27 and CT 28 (table 26) are of interest in that they contain much amphibole, yet their total CaO percentages, 2.78 and 0.96, respectively, are so low as to indicate an abundance of anthophyllite rather than tremolite. The anthophyllite in the talc rock does not contain more than about 2 percent CaO. The anthophyllite content of the average talc rock is about 10 percent.

The values Al_2O_3 + total iron are very low, ranging from about 0.5 to 2.5 percent. In most instances where the total percentage of iron (shown as FeO + Fe_2O_3) exceeds 0.3, the rock contains either iron oxides, such as hematite (Fe_2O_3), or, more commonly, pyrite (FeS_2). These minerals are not widespread and mostly occur near deposits of zinc or lead sulfides, where pyrite is abundant.

The MnO content of the talc rock is generally between 0.15 and 1.0 percent but is 2 percent or more in a few places. In the larger bodies of commercial talc rock (table 25) the maximum percentage of MnO is about 0.5 percent and the average about 0.3. Nearly all of this manganese occurs in anthophyllite and tremolite, though a very small amount is in manganese oxides which are abundant in many of the other rocks. Their scarcity in the talc rock is fortunate, since the discoloration they cause greatly reduces the commercial value of talc.

The SO_3 in the talc rock is largely contained in anhydrite and gypsum. The amounts of these sulfates increase appreciably at depth and in several parts of the Fowler, American and Balmat talc belts anhydrite and gypsum are common enough to jeopardize the use of the rock as commercial talc.

Mineralogy of the deposits

Introduction

It has been said facetiously that talc is anything that can be ground into powder and sold. The origin of this comment might be found in the complex mineralogy of the talc deposits of the district.

Distinctive features of the minerals that are common in the talc belts are noted below, and brief comments are made regarding the uncommon minerals in the talc which are of critical importance because they impart either favorable or unfavorable characteristics to commercial talc, or because their presence is especially indicative of geologic processes important in the origin of the talc. All the minerals that have been found in the talc belts are listed in table 22, with data on their abundance, their occurrence, and certain of

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Table 22. Near here. (p. 276)

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their physical and chemical features.

Table 22. Annotated list of minerals found in the talc belts. ~~The minerals are listed in alphabetical order with notations on abundance and physical appearance.~~

Silicates	Abundance	Geologic environment
Actinolite	rare	A few nests found along margins of pegmatite and amphibolite (metadiabase?) in International 2½ mine.
Andesine		In amphibolites in talc in all talc belts.
Anthophyllite	common and ubiquitous	See discussion in text.
Antigorite	rare	See ~~discussion in text under~~ serpentine.
Apatite	locally common accessory mineral	Green and blue prismatic crystals found in micaceous schist and pegmatite in Talcville, Fowler and Balmat talc belts.
Biotite	locally common major constituent	Thin sheets of biotite schist in all talc belts. Some merge into amphibolite and represent alteration products thereof.
Chlorite	Locally common major constituent	Widely associated with supergene alteration of sulfides in talc (See Brown, 1936B) and locally as retrograde alteration of biotite schist.
Chrysocolla	very rare accessory mineral	Stains on talc.
Chrysotile	abundant and ubiquitous	See ~~discussion in text under~~ serpentine.
Datolite	very rare accessory mineral	Scattered, honey yellow crystals found in talc at Woodcock mine, Fowler talc belt and in pegmatite at International 2 1/2 mine.
Diopside	locally common	See discussion in text.
Hornblende	locally common major constituent	In amphibolites in all talc belts.
Microcline	locally common major constituent	In scattered pegmatites in all talc belts.
Mountain leather (serpentine)	locally common accessory	Chrysotile serpentine along scattered fault surfaces and joints in all talc belts.
Perthite	locally common major constituent	Major feldspar in pegmatites in talc. See also table 18 and plate 14D.
Phlogopite	locally common accessory to major constituent	In nests in and along some pegmatites and at other widely separated spots in all talc belts.

Silicates (cont.)	Abundance	Geologic environment
Prehnite	very local scattered crystals	Fine-grained replacement of calcite in brecciated impure talc, 10th level, Arnold mine, 262 feet west of shaft.
Quartz	abundant and ubiquitous	See discussion in section on mineralogy.
Scapolite	locally common accessory mineral	Associated with several pegmatites and as isolated nests.
Serpentine	abundant and ubiquitous	See discussion in section on mineralogy.
Sphene	locally common accessory mineral	In pegmatites, especially at International 2 1/2 mine, Talcville.
Talc	abundant and ubiquitous	See discussion in section on mineralogy.
Tourmaline	locally common accessory mineral	See discussion in section on mineralogy and paragenesis.
Tremolite	abundant and ubiquitous	See discussion in sections on mineralogy and paragenesis.
Zircon	locally rate accessory mineral	In pegmatites in talc.
Zoisite	locally rare accessory mineral	In Amphibolites in talc.

Carbonates		
Aragonite	locally common	Fibrous vein fillings along fault surfaces.
Azurite	locally very rare	Stains on talc.
Calcite	abundant and ubiquitous	See discussions in text in sections on "Lithologic features of marble replaced by talc."
Dolomite	locally common	See discussions in text in sections on "relations of talc belts to associated stratigraphic units."
Malachite	very rare	Thin, rare stains and crusts on talc.

Chlorides		
Halite	very rare but widespread	A few crystals up to 1/2 inch across and traces of halite, usually $<$ .1 percent by volume, occur sporadically in and along layers of talc.
Sylvite	Trace, locally	Minute traces of sylvite occur with the rare, scattered crystals of halite.

Elements	Abundance	Geologic environment
Native copper	one locality known	One radiating group of crystals found at contact of serpentinized amphibolite and talc on the 5th level, 2 1/2 mine, Talcville.
Graphite	widespread, traces to accessory	As microcrystalline to visible flakes disseminated in scattered layers in all talc belts.
Fluorides		
Fluorite	very local, trace to accessory	Associated with a pegmatite in talc at Talcville and in isolated occurrences at the Woodcock and Balmat mines.
Oxides		
Goethite	local traces to rare	Crusts of fibroid crystals at the Woodcock, Ontario and American Mines, Fowler and American talc belts.
Hematite	locally common to abundant	Stains, coatings and pervasive replacement of talc, especially in western part of the Fowler belt and in the American and Balmat belts.
Limonite	local traces to rare	As thin stains and crysts on talc.
Magnetite	very locally rare to accessory	In amphibolites in talc and associated with talc and oxidized sulfides in Balmat mine (see Brown, 1936B).
Pyrolusite	local accessory to common	In micaceous amphibolite in talc at Talcville; as radiating crystals in nests in Fowler talc belt.
Wad	widespread trace to accessory	Traces and sooty stains widely distributed, especially in Fowler and Talcville belts; associated with oxidized parts talc belts.
Phosphates		
Apatite	locally common accessory mineral	In and along pegmatites and locally in nests with phlogopite at International 2 1/2 and 4 mines.
Sulfates		
Anhydrite	locally common trace to major constituent	As pale pink to gray layers, lenses and grains associated with parts of the Balmat, American and Fowler talc belts.

Sulfates (cont.)	Abundance	Geologic environment
Barite	very local trace to accessory	Found sporadically as disseminated grains and microscopic veins in all talc belts (see also Brown, 1936B).
Celestite	very local trace to accessory	Scattered blue-gray crystals in Balmat, American, Woodcock, and Arnold mines in Balmat, American and Fowler belts.
Gypsum	locally common trace to major constituent	As white commonly sugary to powdery veins and clots formed by oxidation, hydration of anhydrite.

Sulfides		
Chalcopyrite	locally common trace to accessory	As minute inclusions in sphalerite and associated with galena in the Balmat and Fowler talc belts (see Brown, 1936).
Galena	Local trace to accessory	As small grains and nests associated with pyrite and sphalerite in Balmat, Wight and International No. 4 mines.
Molybdenite	very rare	One large (2 inches) nest in serpentinous tremolite talc, fifth level, International 2 1/2 mine, Talcville.
Pyrite	widespread trace to major constituent	As single grains and small clusters of crystals in scattered layers of all talc belts.
Pyrrhotite	locally common trace to accessory mineral	Associated with pyrite and chalcopyrite in Balmat talc belt and at International No. 4 mine, Talcville.
Sphalerite	very locally common-trace to accessory	Closely associated with talc in Balmat mine and locally present in scattered grains and tiny veins in Woodcock, Wight, International No. 2 1/2 and No. 4 Mines.

Anthophyllite

Distribution

Anthophyllite is widely though unevenly distributed throughout the talc rock. It is found in every talc mine and in almost any well-exposed section across the talc belts. Many zones of commercial talc contain less than 3 percent anthophyllite, but rock from a few selectively mined lenses and zones contains 15 to 25 percent or even more. In the period 1945 to 1950, anthophyllite constituted about 10 percent of the total talc mined. The percentage of anthophyllite mined in the future will decrease, however, unless great efforts are made to mine it selectively, for it decreases in abundance downward to the lowermost levels of most of the larger ore bodies.

Anthophyllite is abundant at Newton Hill, where the Talcville talc belt contains two zones up to 40 feet wide and 100 feet long (as cored in holes 1 to 8 drilled on the Thompson property by Roy Brown of Gouverneur in 1952) which average at least 40 percent anthophyllite. In the Woodcock mine, roughly equivalent concentrations of more fibrous anthophyllite have been mined in the hanging-wall zone ("back vein") between the second and fifth levels. There, in several elongate stopes, the talc rock contains from 13 to 58 percent anthophyllite, averaging roughly 30 percent throughout a zone 6 feet wide and 200 feet long. The anthophyllite content of this zone decreases with depth, however, and on the sixth level of the mine it averages only 15 percent. Innumerable other, smaller lenses, layers, and knots of anthophyllite talc occur at the above-cited localities and in other mines throughout the district. Near the hanging-wall in the International No. 2½ mine a zone of "fibrous talc" several feet wide contained as much as 80 percent anthophyllite locally, and averaged about 20 percent anthophyllite as mined from a stope on the fifth level (some 80 feet west of the No. 2½ mine shaft). In the footwall zone of the Arnold talc mine, above the tenth level, there are many layers and lenses of talc rock several feet wide and 15 to 50 feet long that contain from 20 to 75 percent anthophyllite. In the Ontario mine, similar lenses of anthophyllite talc rock occur near the footwall at the small fold 1,200 feet east of the shaft (measured on the fifth level), and in the hanging-wall zone or "back vein" between the third and the sixth levels. At the International No. 4 mine, rock that contains about 10 to 20 percent of anthophyllite occurs in most parts of the so-called "dollar ore" mined from the trough of the large open fold 600 feet east of the No. 4 shaft.

Fibrous anthophyllite also occurs in places along the hanging-wall side of the Balmat talc body, especially on the 500-foot level of the Balmat zinc mine. There it constitutes the uppermost layers of talc rock, which average about 20 percent anthophyllite.

In many other places in the talc belts, especially along well-defined shear zones, there are small, irregularly distributed knots and thin sheets composed almost entirely of anthophyllite enveloped in almost pure talc. Most of these bodies contain only a few tens or hundreds of pounds of anthophyllite.

The decrease in the abundance of anthophyllite with depth is of course most apparent when data are compared from upper and lowermost levels in the deepest mines. The rate of decrease is far from uniform, but commercial and non-commercial talc rock together contain at least 20 percent less anthophyllite at depths of 400 to 600 feet vertically below the surface than on the highest levels of the same mines. The total amount of commercial talc rock, however, remains about the same on the deeper levels, but both anthophyllite and talc become increasingly displaced by tremolite.

Appearance

Some of the anthophyllite in the deposits is highly fibrous and readily confused with fibrous talc, while some is bladed or columnar and not distinguishable by the naked eye from much of the tremolite (see, for example, pls. 17A and 19A and figs. 16, 17, 18, 19). Its distribution has been determined in part by microscopic examination but chiefly by study of 100 X-ray powder diffraction photographs, which afford the best means of identifying the anthophyllite in the talc deposits. Use of the petrographic microscope for that purpose is limited because optical properties of the highly fibrous anthophyllite are so similar to those of the fibrous talc which replaces it and is pseudomorphous after it.

These two minerals in turn are intimately associated with tremolite that has been partly replaced by anthophyllite or by anthophyllite and talc. In many places alteration of tremolite to anthophyllite and talc has had little effect on the general appearance of the rock in hand specimens. If this altered material is powdered, however, it may show a markedly fibrous texture, and often fluffs or "teases" (pl. 19A). As may be seen under the microscope, the alteration of tremolite begins commonly with the growth of fibers of anthophyllite, or of both anthophyllite and talc along cleavages of the tremolite. As the alteration proceeds, the tremolite blades evolve into highly striated, fibrous bundles of anthophyllite and talc that have the outward form of the replaced crystals of tremolite. Various stages in the alteration of tremolite to fibrous anthophyllite are shown in plate 17A. Neither the

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Plate 17A. ~~Near here.~~ In pocket

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color nor the texture of the tremolite rock are appreciably changed by this alteration, although some anthophyllite rock is appreciably softer and less brittle than unaltered tremolite schist (fig. 17). The color of some of the fibrous anthophyllite in this district is the deep buff to clove-brown commonly described in mineralogical texts, but much of it, like the tremolite and talc, is off-white or of various shades of light and dark gray, grayish-green, pale pink, or tan. Much of it also is stained brownish, red, or black by iron and manganese oxides.

Plate 17A. Photomicrograph of highly fibrous talc showing paragenetic relations.

Although most of the anthophyllite is fibrous, some of it is coarsely bladed. There is a large mass of this type of anthophyllite on Newton Hill. At least half of the anthophyllite in this mass is white to pale pink, coarsely bladed, and indistinguishable in hand specimens from bladed tremolite. Like much of the tremolite schist it is triboluminescent, as may be seen by striking with a pick. There is no convincing reason, however, for supposing it to have been formed by replacement of tremolite. Microscopically this anthophyllite is easily distinguished from either tremolite or talc because the individual grains are large enough to permit determinations of 2V and indices of refraction, and to observe the parallel extinction ~~(see p.)~~.

Chemical composition and optical properties

The anthophyllite in the talc deposits belongs to a variety about as high in SiO_2, MgO, and MnO as any yet found. It is also about the lowest in FeO of naturally occurring anthophyllites and has the lowest indices of refraction. These relations are apparent from inspection of figure 20 and tables 27 and 28. Table 27 lists optical and some other

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Table 27. Near here. p. 285

Table 28. ~~Near here.~~ p. 286

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physical properties of four specimens of anthophyllite whose composition is shown in table 28. The data on composition are based upon two analyses previously published (Kunitz, 1930; Allen and Clements, 1906), coupled with analyses for iron and manganese on two other specimens made in conjunction with this study. All of the analyses of anthophyllite appear to have been made on impure material. Principal impurities are tremolite and the mineral talc, which are intimately interleaved with the anthophyllite. Probably, however, the composition and optical constants of all four specimens are very nearly the same. The indices of refraction have been plotted against weight percent FeO + Fe_2O_3 + TiO_2 + MnO in figure 20, where corresponding data are also given for anthophyllites from other localities as compiled by Rabbitt.

Table 27. Physical properties of anthophyllite from the talc deposits.

Anthophyllite specimen number	Color	Indices of refraction α	β	γ	Sign and 2V	Orientation	Distinctive qualities and crystal habit	Remarks
1	colorless	1.596	1.615	1.621	(--) 62°	0°	coarse blades splits to fine fibers	X-ray diffraction pattern coincides with that of "standard" Anthophyllite (Rabbitt, 1948).
2	very pale pink	1.598	1.619	1.6222	(--) 60°	0°	pseudo crystals 3x5 cm which split into fine fibers	Ditto
3		1.598	--	1.623		0°		
4				1.62				

1. Newton Hill

2. International 2 1/2 mine, 5th level.

3. As described by Kunitz from Edwards, N. Y. (1930).

4. As described by Allen and Clement from Edwards, N. Y. (1906).

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Table 28. Partial chemical analyses of anthophyllites from the talc deposits. Numbers of specimens are the same as those in table 27.

	Anthophyllite			
	1	2	3	4
SiO_2	n.d.	n.d.	58.82	59.29
Al_2O_3	"	"	.66	.59
Fe_2O_3	.33	.29	.52	.29
FeO	.28	.18		.06
MgO	n.d.	n.d.	30.81	30.98
CaO	"	"	3.45	1.26
Na_2O	"	"		.37
K_2O	"	"		.19
H_2O+	"	"	2.64	3.80
H_2O-	"	"		
CO_2	"	"		
TiO_2	"	"		.03
Cl	"	"		
F	"	"	.28	.20
MnO	2.87	2.48	2.53	2.77
Total			99.71	99.83

1. Newton Hill, Margaret D. Foster, analyst.
2. International 2 1/2 mine, 5th level, Celeste G. Engel, analyst.
3. As described by Kunitz from Edwards, N.Y. (1930)
4. As described by Allen and Clement~~s~~ from Edwards, N.Y. (190~~8~~6)

The relatively high manganese content of the anthophyllites from this district is especially noteworthy. Rabbitt (1948) has noted that manganese is not, in general, an important constituent ~~constituent~~ of the anthophyllite series, but some of the anthophyllites of the talc belts, though almost colorless, contain from 2.5 to almost 3 percent MnO, and where weathered they give rise to a good deal of manganese stain in the talc.

Diopside

Distribution

Diopside is uncommon in the commercial talc, and less than 1 percent appears in most commercial grinds or blends of talc. It is common, however, in the carbonate-rich layers intercalated in some parts of the commercial talc, and is abundant in the marble that encloses the talc belts. Near the footwall contact of the Fowler talc belt, for example, the lowermost layers of commercial talc rock consist in many places of slabby tremolite schist, and these layers are in sharp contact with marble containing diopsidic lenses and layers like those figured in plates 4B and 13A. Most of the diopside is clearly a product of interaction between pre-existing quartz and dolomite. At several widely scattered localities tremolite layers appear to replace sheared diopside rock. Little or no diopside is found, however, in thin sections of this tremolite schist, although most of it, like most of the adjacent diopside rock, contains considerable quartz.

Very large, irregular bodies of diopside occur locally at and just above the contact of the Fowler talc belt with the overlying dolomite of Zone 12. In the Wight mine, for example, a large lenticular body of rock that is 70 to 80 percent diopside adjoins the talc and encloses the replacement veins of pyrite, sphalerite, and galena which were found between the third and fifth levels (pl. 28).

Layered and laminated quartz-diopside-carbonate marble forms the footwall of the Fowler talc belt and both walls of the Balmat talc belt southwest of Talcville. Northeast of Talcville the footwall marble is not well exposed but apparently all of it contains diopside. At the International Nos. 3, 4, and 5 shafts, for example, disseminated grains and rounded clots of diopside, now partly serpentinized, are disseminated throughout the marble that underlies the talc belt. Most of the hanging-wall marble here and southwestward to the vicinity of the No. $2\frac{1}{2}$ mine is highly diopsidic and contains many lenses and more irregular bodies of quartz-diopside rock, some of which are as much as 10 feet thick and 50 feet long.

The several carbonate-rich layers that separate the talc zones in the Fowler talc belt all contain from several to 10 percent diopside, some of them as much as 10 percent, usually in grains and clots distributed along layers of calcitic marble.

Appearance

Almost all of the diopside is white, cream-colored, or very pale green. Much of it occurs in sub-rounded to slightly elongate grains from 0.4 mm to 3.0 mm in diameter. Microcrystalline diopside, however, is commonly intergrown with quartz or included in it, and is mixed with quartz in layers and clots. Coarsely crystalline bright-yellowish-green diopside forms a few veinlets or irregular knots adjacent to the talc (pl. 17D). One such veinlet in the Wight mine cuts across quartz-diopside layers and associated tremolite.

Most of the diopside is more or less altered to serpentine and some of it to tremolite, anthophyllite, and talc. With increasing serpentinization the color of the diopside commonly changes either to a much deeper lustrous green, or to shades of buff, yellow, and deep brown.

Chemical composition and optical properties

Optical studies of many diopside grains in and along the talc belts indicate little variation from the following: nX 1.665, nY 1.672, nZ 1.692, 2V 60°, maximum extinction 38°. These data, compared with those of Hess (1949) on Adirondack skarn pyroxene, indicate at least roughly the chemical composition of the diopside (fig. 21). The relatively low indices of refraction suggest that the diopside in and

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Figure 21. ~~Near here.~~ In pocket

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along the talc belts is exceptionally low in ferrous and ferric iron and alumina.

Figure 21. Optical properties of diopsides from the Balmat - Edwards district (crosses) compared with those of other diopsides as recorded by Hess (1949). (In pocket)

Quartz

Distribution

Quartz commonly constitutes from 1 to 5 percent, or rarely as much as 20 percent of the talc, but several large masses of commercial talc rock contain hardly any quartz. Most of the highly tremolitic talc rock contains one to several percent quartz, and all of the larger zones rich in foliated talc contain 2 to 10 percent or more quartz.

In the Talcville talc belt, quartz tends to be most abundant along the hanging wall. In the Fowler talc belt, however, quartzitic lenses and knots are as common near the footwall as in higher layers. As the miners try to avoid taking quartz in appreciable amounts, many commercial talc bodies are limited on one or both sides by highly quartzitic lenses and layers, such as those near the hanging wall and footwall of the Fowler talc belt. Many thin, slabby layers overlying the commercial talc in this belt contain from 25 to 95 percent quartz, together with manganese-bearing tremolite and manganese oxides. Much of the tremolite schist near the hanging wall of the Talcville belt northeast of Talcville is too siliceous to be good ore.

Most of the diopside rock that adjoins the talc belts, especially near Talcville, is highly siliceous, containing from 10 to 25 percent quartz. At Newton Hill, where the rocks of the Talcville belt interfinger with layers of diopsidic marble, both rocks contain from 10 to 30 percent quartz.

Appearance

Most of the quartz occurs in medium-sized grains, 0.1 mm to 1.0 mm in diameter; a little coarse-grained quartz is very common. In some places the grains are disseminated through the silicates; elsewhere they form thin lenses and layers in less quartzose rocks. Tiny knots and clusters of quartz are interspersed in places with lenses of foliated talc (pls. 13C and 13D). Where the talc belts are much deformed, as is the Fowler belt at and just east of the Woodcock talc mine, many quartz layers are pulled apart or knotted and contorted into shapes like those in highly deformed marble. Plate 13D shows some of these quartz knots in highly crumpled foliated talc rock.

In the Woodcock mine, and very locally in most of the other mines, coarse-grained irregularly shaped bodies of quartz as much as 5 feet thick are rudely accordant with the layering but pinch and swell abruptly and send out short flamelike tongues into the surrounding less siliceous rock. Their constituent grains are a quarter of an inch to half an inch in diameter, and most of the masses are milky white. One of the largest of these masses in the Woodcock mine contains scattered clusters of deep-reddish-brown tourmaline crystals (pl. 12A).

A small amount of quartz occurs in pegmatites which are very sparsely scattered through the talc belts. The pegmatites have the form of ellipsoidal knots and lenses, some of which lie parallel to the layers of talc rock, while others cut across the layers (pls. 12D, 14C, and 22). The quartz content of these pegmatites varies from 10 to 50 percent. Perthitic feldspar (see table 21) is associated with the quartz in pegmatites on the fifth level of the International No. $2\frac{1}{2}$ mine at Talcville (pl. 14D), on the tenth level of the International No. 4 mine northeast of Talcville, and on the sixth level in the American talc belt, near the workings of the Gouverneur Talc Co.'s shaft. Other minerals associated with quartz in these pegmatitic bodies include sphene, apatite, biotite, datolite, phlogopite, anhydrite, halite, pyrite, and molybdenite.

Serpentine

Distribution

Serpentine, like all the other important constituents of the talc belts, is widespread but irregularly distributed. It constitutes about 10 percent of the talc rock. Spots, streaks, and irregular splotches of buff to deep-brown or locally green serpentine are abundant throughout the lower two-thirds of the Talcville talc belt, and they occur at irregular intervals in the upper third of this belt. The serpentine forms as much as 15 to 18 percent of the talc rock in this belt, where most of it has replaced tremolite (pls. 13B, 16A, and 18B). Serpentine has likewise replaced highly tremolitic talc rocks in much of the footwall talc zone in the central parts of the Fowler talc belt, and in the lower half of the explored parts of the Balmat talc belt.

Appearance

Most of the serpentine is massive and microcrystalline, and in the form of either rounded specks and larger splotches or elongate streaks and lenses which have replaced diopside, tremolite, and carbonates. Much of it forms either the cores of knots, clusters, and layers of other silicates, especially diopside and tremolite, or sheaths around such clusters ~~(pls. 14C and 14D)~~. Slip-fiber, and more rarely cross-fiber, serpentine occurs along a few shear surfaces and joints.

The microcrystalline serpentine in the talc belts is mostly resin-colored to amber-brown, or various shades of green, especially a dark watery green. A white to cream-colored serpentine also is abundant, especially in the footwall zone of the commercial talc in the Arnold talc mine. A very little of the serpentine is blue, pink, gray, reddish-brown, or black. Most of the serpentine in the carbonate-rich layers in and adjacent to the talc belts is green.

Chemical composition, optical properties, and thermal analysis

Chemical analyses of three specimens of serpentine from widely separated parts of the talc belts are given in table 29. All three of

Table 29. ~~Near here.~~ p. 296

these specimens have been identified on the basis of thermal analysis[1] and

[1] The differential thermal analyses of these serpentines were made by George Faust of the Geological Survey.

X-ray diffraction studies as chrysotile. Standards used for comparison include synthetic chrysotile (Bowen and Tuttle, 1949), natural chrysotile from the Grand Canyon, and antigorite from Antigorro, Italy. Specimen S-1 is a typical example of the amber to resinous-brown chrysotile that replaces tremolite in the central and lower parts of the talc belt in the International No. 2½ mine at Talcville. Under the microscope this material is seen to be micro- to cryptocrystalline and to have an aggregate index of refraction of about 1.547. George T. Faust notes (written communication) that this specimen gave off water at low temperatures, and he suggests that it may consist of a member of the series of chrysotile-deweylite. It encloses 1.68 percent by weight of calcite, 1.5 percent hydrated iron oxide, and a very little halite.

Specimen S-2, table 29, is dark-green, very lustrous and translucent on thin edges. It forms lenses and layers a fraction of an inch to several inches thick, which replace tremolite, diopside, quartz, and carbonates, in and adjacent to the American talc belt at the shaft on the fifth level of the Gouverneur Talc Co.'s mine. It consists of chrysotile, which, like that from the International No. 2½ mine, is micro- to cryptocrystalline, but its aggregate index of refraction is approximately 1.554. It gives essentially the same X-ray diffraction pattern and thermal curves as S-1.

Table 29. Chemical composition of serpentines from the Balmat-Edwards district.

	S-1	S-2	S-3
SiO_2	42.03	43.04	44.60
Al_2O_3	.35	.00	
Fe_2O_3	.26	.87	.96
FeO	.06	.16	
MgO	39.95	40.64	39.79
CaO	1.53	.85	.96
Na_2O	n.d.	.07	n.d.
K_2O	n.d.	.01	n.d.
H_2O+	12.23	13.76	12.63
H_2O-	1.04	.41	
CO_2	1.13	.74	.83
TiO_2	.16	.02	n.d.
Cl	.04	.05	n.d.
F	.55	.00	n.d.
Ca_2O_3	.01		
MnO	.49	.15	.12
BaO	.00	n.d.	n.d.
P_2O_5	n.d.	.00	n.d.
	99.83		
Less	.24		
Total	99.59	100.03	99.89

Trace Elements	S-1*	S-2*
Cu	.0009	.0005
Zn	.01	.03
Mn	.08	.1
Ti	.002	.004
B	.03	.06

*Looked for but not found: Be, Ag, Au, Pt, Mo, W, Ge, Sn, Pb, As, Do, Bi, Ti, Cd, Ni, Ga, In, Cr, V, Sc, Y, Yb, La, Zr, Nb, Ta, U, Sr, Ba, Co. A. H. Chodos, spectroscopist.

S-1 Serpentine, variety chrysotile, from the 5th level, International No. 2 1/2 mine, Talcville. X-ray and differential thermal analysis suggests this is a member of the series chrysotile-deweylite. Eileen K. Oslund, analyst. Cream color; negative (-); 2V moderate, mean index 1.55, B - .004.

S-2 Serpentine, variety chrysotile, from the 5th level, Gouverneur Talc Company, mine at shaft, Balmat, N.Y. Margaret D. Foster, analyst. Deep green color; negative (-); 2V moderate; mean index 1554.

S-3 Serpentine, variety chrysotile, from diamond drill core into talc belt just north of Hyatt mine shaft, Hyatt, N.Y. Orton Smalley, analyst.

Specimen S-3, which has replaced knots of diopside in the carbonate-rich talc rock at the Hyatt zinc mine, gives an X-ray pattern and thermal curve similar to those of S-1 and S-2, but it consists of cream-colored porcelanous chrysotile.

Many other samples of the serpentine in the talc rock were X-rayed, studied optically, and analyzed thermally. All these samples, except two from the International No. 2½ and Wight mines, appear to consist of chrysotile. The two exceptions have X-ray patterns and thermal curves like those of the antigorite from Antigorro, Italy.

Talc

Distribution

The mineral talc, like serpentine and anthophyllite, is widely though very irregularly distributed in the talc deposits. In the period 1945-1950 only about 20 or 25 percent of the rock mined and sold as commercial "talc" was the mineral talc. Certain zones and irregular bodies are selectively mined to obtain a relatively high or low proportion of talc, as compared with tremolite, anthophyllite, and serpentine, and thus to attain desired physical or chemical properties.

Ores containing as much as 75 percent talc have been sold under certain designations of grade and type. Little if any of the ore mined in the district contains more than 75 percent talc. This percentage is approached in parts of the footwall zones of the Fowler talc belt within 500 feet of the surface, measured down the dip of the belt (pl. 18C). Some parts of the intermediate and footwall zones at and near the west end of the Fowler talc belt also contain as much as 75 percent talc, and several intermediate and hanging-wall zones in the remainder of the Fowler talc belt average as much as 50 percent. Minable lenses and zones in the upper one-fourth of the Talcville talc belt at Talcville also contain as much as 50 to 65 percent talc. The footwall side of the Talcville talc belt, northeast of Talcville, probably averages almost 75 percent talc. Zones and lenses even richer in talc are common enough, but most of them are no more than 2 or 3 feet thick and 25 or 50 feet long. There are several of these small bodies near the hanging wall of the American talc belt above a depth of 300 feet (measured down-dip), and also in the upper half of the footwall talc zone in the Fowler talc belt. In some of these, the percentage of talc is between 80 and 90 percent by volume. Even more nearly pure talc forms thick films of gouge, and foliated or fibrous sheets, lenses, and knots along shear zones in all the talc belts; some of these are as much as 98 percent talc (pl. 16C).

Several irregular veinlets of coarsely crystalline talc occur within the talc belts. One of these veins in the Arnold mine varies in width up to 3 feet and cuts diagonally across the layering of highly brecciated marble that is partly replaced by talc. The talc in the vein forms colorless, nearly transparent leaves an inch or more across.

Talc, like anthophyllite, decreases irregularly in abundance at depths greater than a few hundred feet in the talc belts.

Appearance

The talc in the deposits appears in three habits. Much is in foliated aggregates, more or less crumpled and sheared (pl. 18C); some is in minute fibers or blade-like bundles of fibers; some forms micro- to cryptocrystalline masses, translucent on thin edges. Talc of the last-mentioned habit occurs on highly slickensided surfaces and shear zones and forms lenticules and ships of gouge, often resembling chips of water-soluble wood glue. As noted on page 233 , much of the fibrous talc is pseudomorphous after anthophyllite, and the blade-like bundles of anthophyllite and talc are pseudomorphous after tremolite (pl. 19A).

Talc of any of these habits may be either white or very pale green, pink, blue, tan, or gray. Various intermixtures of the above-mentioned three varieties can be found in a single hand specimen. The foliated talc and the fine-grained lustrous talc are almost invariably associated with fibers of both anthophyllite and tremolite.

Chemical composition and optical properties

A fairly complete chemical analysis of a specimen of foliated talc has been made by Margaret Foster (table 30). Optical examination

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Table 30. ~~Near here.~~ p. 301

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suggests that the specimen contains a little tremolite, but the analysis shows that the amount was very small (less than 2 percent), for the composition of the sample corresponds closely with the stochiometric formula for talc. The nZ index of this material is 1.588.

In addition to the above sample, 32 samples of foliated, massive, and fibrous talc have been X-rayed during the present investigation, as have over fifty other mixtures of talc with anthophyllite, serpentine, or tremolite. About a dozen differential thermal analyses of talc rock that contained more or less of the mineral talc have been made by George T. Faust. Talc intimately mixed with serpentine, anthophyllite, and tremolite, seems to be as readily identified by differential thermal analysis as by X-ray methods.

Table 30. Chemical and spectroscopic analysis of talc from Balmat, N.Y.

	T-1		T-1
SiO_2	62.33	Cu	.002
Al_2O_3	.00	Zn	.2
Fe_2O_3	.22	Mn	.01
TiO_2	.01	Ti	--
CaO	.00	B	.008
MgO	31.77		
Na_2O	.05		
K_2O	.02		
MnO	.01		
P_2O_5	.00		
CO_2	.00		
Cl	.02		
F	.00		
H_2O-	.07		
H_2O+	5.20		
Total	99.82		

Looked for but not found: Be, Ag, Au, Pt, Mo, W, Ge, Sn, Pb, As, Sb, Bi, Tl, Cd, Ni, Ga, In, Cr, V, Sc, Y, Yb, La, Zr, Nb, Ta, U, Sr, Ba.

Quantative spectroscopic analysis for specified minor elements in talc from Balmat, N.Y.

Material analysed is pale buff, massive talc in irregular veinlets and mammillary masses. The index of refraction n_Z of this talc is about 1.588± .003. The locality is 200 yards west of the main shaft, Balmat zinc mine, Balmat, N.Y. Wet chemical analysis for major constituents by Margaret Foster; spectroscopic analysis for minor constituents by E. L. Hufschmidt.

Tremolite

Distribution

Tremolite constitutes about half of the commercial talc rock sold each year, and is common in most parts of each talc belt. Some of the most uniformly tremolitic rock is found in (1) large parts of the Balmat talc belt, (2) the central parts of the footwall zone of the Fowler talc belt at depths of 500 feet or more down dip, and (3) the northeast half of the hanging-wall side of the Talcville talc belt. At all these places the talc rock is virtually a tremolite schist, containing up to 10 percent quartz and less than 20 percent, in all, of serpentine, talc, anthophyllite, and other silicates. At least locally, however, much or even most of the tremolite has been replaced by other minerals, and in some parts of these talc belts, especially the ends of the Fowler talc belt, tremolite makes up less than 40 percent of the talc belt as a whole, and less than 20 percent of some of the commercial zones. Approximately average percentages of tremolite--that is, some 40 to 60 percent--are common in large segments of the commercial talc rock of the Talcville belt, at and northeast of Talcville in the American talc belt, and in the central parts of the middle and hanging-wall zones of the Fowler talc belt.

The proportion of tremolite to other talc-forming minerals increases irregularly downward in at least four of the commercial talc zones. In the major producing zone--the footwall zone--of the Fowler talc belt, the proportion of tremolite increases markedly with depth, and there appears to be a less marked downward increase in tremolite content throughout the Talcville and American talc belts. For the district as a whole, the percentage of tremolite in the talc rock mined from the district increases perceptibly as far down the dip as mining had gone in 1945. For all we know, however, this tendency may cease, or even be reversed, at greater depths; too little information is available on large parts of the talc belts below, say, 700 feet to warrant a definite conclusion. No well-defined, persistent decrease in tremolite content with depth has been observed in any of the mines.

The percent of tremolite has not been found to vary systematically either across or along the beds in any talc belt. In certain parts of some talc belts, however, within 300 feet of the surface, well-defined zonal variations in tremolite content may sometimes be observed within a single mine, and the same kind of variation may persist throughout a large segment of a talc belt. In the Balmat talc belt, for example, the median and footwall zones contain a higher percentage of tremolite than the uppermost fourth of the belt. In the part of the Fowler talc belt explored by the Arnold mine, also, the commercial talc in the footwall zones is more tremolitic than that in the intermediate and hanging-wall zones. Farther west, at the Woodcock mine, the hanging-wall zone of the Fowler talc belt contains about as much tremolite as the intermediate and footwall zones, and this is also true at the northeast end of the Fowler belt.

To points 500 feet vertically below the surface, the commercial talc mined in the Balmat, American, and Talcville belts contains more tremolite per foot of depth than that mined in the Fowler talc belt.

The variations in the proportion of tremolite to other minerals are in many places too slight or too complex to be followed in extracting ore. Many highly tremolitic zones of commercial talc include layers, lenses, or irregular bodies of much less tremolitic rock, mostly in shear zones or near faults; other intimate intermixtures of tremolite and anthophyllite, serpentine and talc are on the apices or the shorter flanks of folds; still others have no obvious relation to structural features.

Appearance

The tremolite in the talc belts commonly forms elongate prisms, flattened blades and needles, or fine fibers, stubby laths, or almost equant grains (see pls. 14A, 14B, 17A, 17B and 18D). The longest dimension of the prisms and blades is usually between 1/32 and 1/4 inch, and the more equant grains are commonly from 1/4 to 1/16 inch in diameter. Individual grains larger and smaller than these make up less than 20 percent of the talc rock. The largest crystal of tremolite that I have seen was in the International No. 2½ mine at Talcville: it measured 37 inches parallel to the c axis, and 8 3/4 inches across. Tremolite too fine grained to be recognized with the naked eye but visible under the microscope at magnifications of 100 to 450, occurs at many places in the talc belts but rarely composes more than 15 percent of the talc rock by volume.

Elongate crystals of bladed to needlelike tremolite are commonly intergrown or interlocked in flattened, reticulated, or matted layers. In many places, the longest axes of the tremolite crystals are oriented at random in the planes of foliation (pl. 14A). In other places they are matted together (fig. 22) with their longest axes roughly parallel to one another, to the dominant layering, and to the axes of associated folds. Linear parallelism of any of these kinds shows that the tremolite grew under marked shearing stress. Rarely, the elongate prisms of tremolite lie parallel to the layering and about perpendicular to the fold axes. Tremolite blades also may be alined at right angles to the dominant rock layering, or at some intermediate angle, as is shown in plates 15A and 15B. In some massive, unfoliated rocks, prisms

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Plate 15A. ~~Near here.~~ In pocket

Plate 15B. ~~Near here.~~ In pocket

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or blades of tremolite form sheafs, sprays, or stellate clusters. Much more rarely, the tremolite occurs in nearly equant grains, as a constituent of sugary-textured, granular rocks that superficially resemble coarse-grained quartzites. Most or all of the various types of tremolite rock noted above can be found, either interlensed or interlayered, in most mines and many large exposures along the talc belts.

Plate 15A. Tremolit~~e~~ic marble sectioned normal to the foliation. (in po

Plate 15B. An advanced stage in the replacement of marble by blades of tremolite growing at right angles to the prominent layering in the marble. (In pocket)

The tremolite in this district, where it is not stained gray, black, or brown by oxides of manganese and iron, is partly colorless or white, partly gray or tan, and partly tinted in pastel shades of pink, lavender, yellow, green, and blue. The pink to lavender tremolite has long been known as hexagonite (Goldsmith, 1893; Koenig, 1876; Dana, 1893, p. 385-389).

Much of the tremolite in the district is triboluminescent; that is, where struck with a hammer it glows for a second or more.

Chemical composition and optical properties

The chemical composition and optical properties of representative samples of tremolite are recorded in tables 31 and 32. Table 31 gives

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Table 31. ~~Near here.~~ p. 309

Table 32. ~~Near here.~~ p. 310

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new analyses, by A. E. J. Engel, of two samples of colorless tremolite from large commercial talc deposits, and 13 partial analyses by Margaret Foster and J. O. Clark of tremolite from the various talc belts. Table 32 gives the optical properties and colors of the tremolite in many of these samples, of the tremolite in several unanalysed samples, and of a "tremolite standard" from Baltimore County, Maryland (U. S. National Museum No. 16, 554).

Table 31. Chemical analyses of tremolites from the Balmat-Edwards district and from Baltimore County, Maryland.

	1+	2+	3+	4+	5+	8+	9a+	9b*	10+	11+	15*	16≠	17≠	18≠	19≠
SiO_2								57.83			57.76			58.67	58.10
Al_2O_3								.57			.64				
Fe_2O_3	.64	.35	.37	.31	1.19	.96	.18	.20	.27	2.22	.00				
FeO							.11	.11	.17	.60	.14				
MgO							24.88	24.25			24.73			24.50	24.63
CaO							13.53	11.42			13.27			11.18	11.10
Na_2O								.17			.28				
K_2O								.14			.09				
H_2O+								2.09			2.38				
H_2O-								.06			.13				
CO_2															.12
TiO_2								.03			.05				
Cl								.01			.02				
F								.41			.26				
MnO	.13	1.80	.56	1.34	.28	.36	.035	2.81	.008	.028	.06	.085	0.82	.96	.72

100.10

+Margaret D. Foster, analyst; *A.J. Engel, analyst; ≠J.O. Clark, analyst.

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Specimens 1 to 10 inclusive are the same as the specimens 1 to 10 inclusive whose physical properties are listed in table 32.

Specimens 1 to 5 inclusive and specimen 8 are from the hanging wall zone of commercial talc in the Woodcock talc mine, Balmat, N.Y. Specimens 6 and 7 are from impure talc just above the footwall zone in the Wight Talc mine. Specimen 9b is from level 14, Arnold talc mine, and 9a is from the 7th level of the International No. 4 mine, Talcville. Specimen 10 is a U.S. Geol. Survey "standard" from Baltimore County, Maryland. Specimen 11 is from Natural Bridge, N.Y., courtesy W. Schaller.

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level of the International No. 4 mine, Talcville.

Table 32. Physical properties of tremolites from the talc deposits of the Balmat-Edwards district.

Tremolite specimen number	Color*	Indices of refraction			Sign and 2 V	Orientation Z ∧ C	Form	Remarks
1	colorless	1.599	1.614	1.625	(--), 81°	17°	Thin laths, Transparent	Tremolite schist with scattered brown tourmaline†
2	bright pinkish lilac	1.601	1.613	1.624	(--), 65°	15°	Laths which split to fine fibers	Slightly altered; 2 V variable palest hexagonite
3	pale lavender	1.599	1.614	1.627	(--), 86°	14°	Columnar, coarse	Typical hexagonite
4	Lavender	1.600	1.615	1.627	(--), 85°	14°	Columnar, coarse	Typical hexagonite
5	pale vinaceous	1.603	1.618	1.631	(--), 83°	15°	Thin laths	
6	pinkish lilac	1.601	1.616	1.628	(--), 85°	17°	Thin laths	
7	pale persian lilac	1.603	1.618	1.630	(--), 83°	16°	Thin laths	
8	brownish vinaceous	1.604	1.617	1.629	(--), 83°	13°	Thin laths	Darkest colored hexagonite
9a	colorless	1.5999	1.615	1.624	(--), 80°	19°	Columnar	
9b	colorless	1.603	1.618	1.628	(--), 80°	13°	Columnar	
10	colorless	1.599	1.615	1.626	(--), 82°	16°	Bladed	"tremolite standard"

*Color determined from Ridgeways color standard. †Tourmaline associated with this tremolite is listed as WoTo2 in table 33.
All optical constants checked by Jewell Glass, U. S. Geological Survey.
Partial data on chemical composition of these specimens are given in table 31.
Specimens 1 to 5 inclusive and specimen 8 are from the hanging wall zone of commercial talc, Woodcock Talc mine.
Specimens 6 and 7 are from impure talc just above the footwall zone in the Wight Talc mine. Specimen 9b is from the 14th level Arnold mine, and 9a from the 11th level of the International No. 4 mine, Talcville. Specimen 10 is a U.S.G.S. "standard" from Baltimore County, Maryland. Specimen 9b is from the 7th level of the International No. 4 mine, Talcville.

The specimens of tremolite numbered 1 to 8 in both tables are arranged in order of color, the first being colorless and the others more and more deeply tinted with lavender. Koenig (1876) and many other workers have assumed that the lavender color of the so-called hexagonites was imparted by manganese, which presumably replaced magnesium in the crystal lattice, but comparison of the two tables indicates that there is no simple relation between the intensity of the lavender color and the percentage of manganese; in fact the tremolite containing the most MnO is colorless. W. D. Schaller (written communication, ~~19~~) has suggested that the lavender color may be due primarily to small amounts of the strongly chromatic oxide Mn_2O_3, but may be greatly deepened by Fe_2O_3, and this suggestion is borne out by the fact that the most deeply colored samples listed in table 31 (numbers 5 to 8) have a brownish tint and are highest in Fe_2O_3. Possibly the percentage of MnO recorded in the older analyses of tremolite (Dana, 1893, p. 393, anal. 12 and 13) are too high. The four newly analyzed hexagonites in table 31 contain, in order of increasing intensity of color, 0.56, 1.34, 0.28, and 0.36 percent MnO, whereas the hexagonites cited by Dana are said to contain 1.37 and 2.39 percent.

The indices of refraction of tremolite appear to be much more greatly increased by small amounts of Fe_2O_3 than by similar amounts of FeO, MnO, or FeO + MnO. The three analyzed samples containing the most Fe_2O_3 (No. 8, with 0.96 percent; No. 5, with 1.19 percent; and No. 11, with 2.22 percent) also have the highest maximum indices of gamma prime, namely 1.629, 1.631, and 1.644, respectively. There is no well-defined correlation of indices of refraction with percentage of manganese; samples No. 2 and No. 9b are high in MnO (1.80 and 2.81 percent) but do not have correspondingly high refractive indices. W. T. Schaller (written communication~~, 19~~) has suggested that these two samples are higher in Na_2O than the others, and that the sodium proxying for calcium may lower the indices.

Tourmaline

Distribution

Magnesium-rich tourmaline is a minor constituent of the talc rocks in the Fowler talc belt and locally in other talc belts. In the Ontario, Arnold, Wight, and Woodcock mines it forms small augen and grains disseminated through the talc. The tourmaline is most abundant in layers and lenses of off-color, highly impure talc that adjoin or alternate with layers of commercial talc. The tourmaline rarely forms more than a few percent of any layer of talc, but at some places in the Woodcock mine it constitutes as much as 10 percent of some layers of quartzitic, anthophyllitic, and tremolitic schists.

Appearance

The tourmaline is generally deep wine-red, reddish-brown, or yellowish-brown in color. It is partly in microscopic grains, but these grade in size into zoned crystals several inches in diameter; the estimated average diameter is 1/16 inch. Most tourmaline crystals are nearly equant. The very large crystals (plate 12A), have

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Plate 12A. ~~Near here.~~ In pocket

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been found only in several quartzose veins on the fifth level of the Woodcock mine. More common are tiny rounded to augen-shaped grains that lack any semblance of prismatic form.

Plate 12A. A cut and polished surface across a quartz-tourmaline vein which cuts the Fowler talc belt in the Woodcock mine. (In pocket

Chemical composition and optical properties

The indices of refraction of much of the tourmaline in this district are higher than those of any magnesium-rich tourmaline hitherto described. In twenty-three specimens the values of nE range from 1.645 to 1.691. In the large zoned crystals illustrated in plate 12A, marked though unsystematic variations were found in the indices of different zones. Partial chemical analyses of one of the thicker zones in a large crystal and of a fine-grained aggregate of tourmaline are given in table 33, together with optical data. The association of manganese 33, p. 315a oxides with much of this tourmaline and its abnormally high indices of refraction, suggested that these might be somewhat manganoan tourmalines, but the analyses showed only 0.32 and .01 percent of MnO, and the total iron content (expressed as 8.56 and 7.30 percent Fe_2O_3) is not unusually high for tourmaline. It is possible that a relatively high titanium content (1.68 percent in one crystal) may be characteristic of these tourmalines and account for their high indices of refraction.

(p. 315 a follows)

Table 33. Optical properties and partial chemical analysis of tourmalines from the Woodcock mine, near Balmat, N.Y.

Property	Specimen	
	WoTo 1	WoTo 2
SiO_2	36.44	
TiO_2	1.68	
Al_2O_3	25.88	
Fe_2O_3	7.30	8.56
FeO		
MgO	11.77	
CaO	1.61	
MnO	.01	.32
Loss on Ignition	3.55	
Color	Deep wine red	Reddish brown
Indices of refraction	1.679 1.689	No 1.676 Ne 1.687
Sign and 2V	(—) 0-70	(—) 0-50
Crystal form	Large equant crystals	rounded aggregates
Remarks	specimen shown in plate 12A	indices vary on different aggregates

Optical constants checked by Jewel Glass.

WoTo 1, Celeste Engel, analyst. WoTo 2, Margaret D. Foster, analyst.

Paragenesis

Tremolite is the earliest-formed of the major minerals in the talc deposits. This is clearly true at all places in the talc belts except Newton Hill, where tremolite and coarsely bladed anthophyllite are intimately intergrown, suggesting that the two minerals are contemporaneous. There, as elsewhere in the district, some of the tremolite also is replaced by pseudomorphous fibrous anthophyllite, but some of the coarsely bladed anthophyllite also is replaced by fibrous anthophyllite. It seems probable, then, that the tremolite and the coarse anthophyllite are contemporaneous and the fibrous anthophyllite younger than either. Presumably anthophyllite forms at temperatures or pressures, or both, below those obtaining when tremolite forms, possibly in a system that contains less water than the minimum necessary to convert the material to serpentine.

Everywhere else in the talc belts that these two minerals occur together, tremolite is in various stages of replacement by fibrous anthophyllite. This relation is clearly shown in plate 17A.

Both tremolite and anthophyllite are replaced by serpentine and by talc. Where talc and serpentine are associated the serpentine is usually in various stages of alteration to talc. Rarely the relations of the two suggest that they were formed at about the same time, but in no observed instance does serpentine clearly replace talc. Talc therefore seems to be the last magnesian-bearing silicate to form in the talc belt. Plates 17A, 17B, 17C, and 18A, 18B, and 18D are photomicrographs that illustrate these relations as seen in various kinds of talc rock.

Plate 17B. Near here.

Plate 17C. Near here.

Plate 18A. Near here. in pocket

Plate 18B. Near here.

Plate 18D. Near here.

The successive paragenetic steps in the evolution of the talc deposits are: tremolite→anthophyllite→serpentine→talc. The alteration is commonly incomplete and irregularly developed. Some tremolite, for example, is almost completely unaltered, and some is slightly altered either to anthophyllite or directly to serpentine. Some is replaced directly by talc, neither anthophyllite nor serpentine being present. Much anthophyllite also alters directly to talc. These relations are expectable, seeing that these minerals were all formed during the deformation of the talc belts, under conditions that did not permit the establishment of equilibrium for any phase except locally.

Plate 17B. Tremolitic talc partly replaced by microcrystalline grains of the mineral talc.

Plate 17C. Slightly serpentinous and talcose commercial talc.

Plate 18A. Diopside (D) veined by serpentine(s), the whole partly replaced by microcrystalline talc (T).

Plate 18B. Commercial talc composed of tremolite (TR) altered by serpentine (s).

Plate 18D. Highly tremolitic commercial talc.

Anthophyllite, serpentine, and talc have replaced some carbonate, quartz, and diopside. At least nine-tenths of the anthophyllite, however, and three-fourths of the talc and serpentine, have replaced tremolite. Serpentine and talc replace a far wider variety of minerals than anthophyllite, including quartz, feldspar, tourmalines, and micas (see pls. 14C and 14D).

Quartz appears to have formed and recrystallized at all stages in the evolution of the talc deposits under the entire span of physical conditions occurring in the deposits, from the initial tremolite stage through the final talc stage.

Relations to associated rocks and ores

The minerals that make up the talc deposits have been derived from dolomitic marble and associated quartzitic interlayers through a sequence of changes that began with the formation of tremolite and ended with the formation of talc. These changes must have required appreciable time and a wide range of physicochemical conditions. The evolution of the minerals in the talc deposits is closely paralleled by related changes in the associated Grenville-like rocks.

In the marble that encloses the talc deposits, a whole series of mineral and chemical transformations, described on pages 45 to 55 must have occurred just before and during the formation of the talc deposits. Part of the diopside, for example, appears to have been formed before the tremolite and part of it contemporaneously with tremolite. At a few localities in the marble that encloses the talc belts, a small amount of tremolite coexists with diopside, and in that environment neither mineral clearly replaces the other. At the contacts of marble with talc rock, however, and at widely separated places within the talc deposits, tremolite can be seen to replace diopside, though in a few other places the two minerals are intergrown and apparently of the same age.

The contact zones of the pegmatites in the marble commonly contain a lot of diopside, and both field relations and textural features observed in thin sections indicate that most diopside was contemporaneous with the potassium feldspar. The pegmatites in marble cross the boundaries between clastic metasediments and marble and are continuous with the feldspathic veinlets and pegmatites in the migmatitic rocks. Migmatization, granitization, and the formation of diopside were virtually contemporaneous, and they all preceded the formation of most of the anthophyllite and all of the serpentine and talc in the talc belts.

The constituents of the gneisses and schists that are of the same age and metamorphic facies as the serpentine and talc are chlorite, sericite, and albite, all of which were formed at a time when the rocks of the Grenville series were being silicified and cut by quartz veins, before the emplacement of igneous and igneous-looking rocks.

The age relations of zinc and lead ores to talc rock may be studied in several localities where these deposits are intimately associated. They are perhaps clearest near Balmat, where large pyrite-sphalerite-galena deposits form a partial sheath around a body of commercial talc in the Balmat talc zone. There and at other localities, such as the Wight talc mine and the workings at Hyatt, where base-metal sulfides and talc rock are juxtaposed, the sulfides replace the tremolite and scattered grains of anthophyllite but appear to be of the same age as some of the anthophyllite and much of the serpentine. Some of the serpentine, chlorite, and talc at these localities is clearly younger than the sulfides.

Age relations of talc-forming minerals to deformation of the Grenville series

Deformation of the talc belts accompanied the evolution of their component minerals. The field relations, and textural and structural features of the talc deposits indicate that the intensity and pervasiveness of the deformation decreased with time, though not at a uniform rate, and ceased at about the same time that talc formation ceased, or soon thereafter. In the early stages of deformation, during the growth of diopside, tremolite, and some of the anthophyllite, the rocks were being deformed by profound solid flow ~~(see p. —)~~. Most lineated and foliated masses of tremolite were produced during this ductile stage (pl. 15C, figs. 17, 18, and 22). Most of the foliation induced by shear is parallel or subparallel to the bedding, but some

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Plate 15C. ~~Near here.~~ In pocket

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shear surfaces cut abruptly across relict bedding and lie roughly parallel to axial planes of associated large folds. In much of the tremolite-rich schists of the talc belts, the crystals and blades, needles, and fibers of tremolite are largely parallel to the nearby axes of folding. This lineation is also commonly parallel to the dominant foliation; in fact, it defines the foliation. In a minority of instances, however, the linear and associated planar features tend to cut across and blur an older foliation (pl. 14B).

Plate 15C. Anthophyllite-tremolite schist split along the dominant foliation.

Most of the lineated tremolite schists are medium to coarsely crystalline, granoblastic mats. Some are only thin layers which have formed along shear zones, but a majority have appreciable thickness as well as great length measured along the plunge of major folds. The lineated anthophyllite, on the other hand, is more commonly in relatively thin skins that have formed along shear surfaces where the preexisting tremolite has been ruptured without complete recrystallization (pl. 15C, figs. 18, 19). The anthophyllite fabrics therefore appear to be products of much less pervasive deformation, in which considerable rupture accompanied the intra-crystalline flow. In the anthophyllite schists, some lineations accord with those of the tremolite and fold axes, but most of the lineations, especially those in which there is obvious granulation and rupture, lie along shear surfaces nearly normal to associated axes of folds (pl. 15C, figs. 18, 19). The anthophyllite fibers thus lie in the direction of maximum shear and slickensiding (the A structural coordinate) rather than parallel to axes of rotation (the B structural coordinate). Some anthophyllite blades seem to be at least partly rotated into this position, whereas others which have replaced quartz and tremolite have grown in this orientation. These relations all seem to indicate that the deformation during the growth of the tremolite involved more pervasive solid flow (ductility) and recrystallization than the stages at which the anthophyllite formed.

Increasingly brittle response to deformation is characteristic of the periods in which serpentine and talc formed. These two minerals occur largely along surfaces and zones of shearing, which include many well-defined faults. In some areas of marked folding these minerals are also localized along some particular part of the fold, such as its axis or its shorter flank. Elsewhere the maximum formation of talc cannot be related to any obvious structural feature but stratification, which it tends to follow. Both talc and serpentine are also granulated, sheared and extensively crumpled in places. The crumples are in reality second or third order folds whose axes parallel those of large parent folds, and most of the associated shear surfaces are systematically related to other major structural features that began to be formed during or before the development of tremolite (pls. 11B, 12B, 16B, and 16C). The deformation of the serpentine and talc is a final stage in the orogeny in which the earlier minerals grew. This relationship is further indicated by the formation of some fibrous serpentine which forms slip-fiber asbestos along the shear surfaces. Many of the fibers are parallel to the shear surface, with their longest axes normal to associated fold axes. The resulting lineation commonly obliterates the anthophyllite lineations but is similarly oriented.

Small masses of crushed and crumpled talc rock have recrystallized into coarse-grained rock with talc folia oriented at random, but most of it retains a strong foliation and lineation induced by crumpling and is more or less twisted or broken. These relations are the principal basis for the conclusions that the talc was formed during the final, mild stages of regional deformation, at lower temperatures and confining pressure.

Comparison of paragenetic sequence in the talc deposits with that in the magnetite deposits of the igneous massif

There are some interesting parallels between the paragenetic sequence in the talc deposits and in certain of the St. Lawrence County magnetite deposits that have replaced marble on the outer border of the Adirondack massif (Leonard and Buddington, in press). These magnetite deposits include complex assemblages of Ca-, Mg-, Fe-, and Mn-bearing silicates, together with hematite and iron sulfides. Leonard and Buddington have shown that the magnetite deposits, like the talc deposits in the Grenville Lowlands, were formed during and after the emplacement of the granitic rocks, pegmatites, and migmatites in the massif. They note that iron-rich pyroxene, amphibole, "andraditic" garnet, and mica skarns, generally in the order named but with some overlap, appear to have replaced diopsidic marble. These skarns are in turn replaced by magnetite, by sulfides, and finally by dominantly hydrous minerals.

The diopsidization of marble, followed by development of amphibolitic and micaceous skarns, constitutes a retrograde sequence analogous to the paragenetic sequence diopside$\rightarrow$tremolite$\rightarrow$anthophyllite in the talc deposits, and the iron oxides and sulfide in the skarns appear to be of about the same relative if not absolute age as the sphalerite, galena, and pyrite associated with the talc deposits. The concluding hydrous-mineral stage in the skarns may have as its analogue in the talc deposits the appearance of serpentine and talc. A striking difference between the deposits is the abundance of the iron- and aluminum-rich pyroxenes, amphiboles, and mica in the skarns as contrasted with the iron-poor, magnesian varieties of these minerals in the talc deposits.

Leonard and Buddington (in press) conclude that the magnetite deposits of St. Lawrence County are undeformed; yet from its paragenetic relations the magnetite would seem to have been emplaced at the same time or before the talc in the Grenville Lowlands. Possibly the final rock movements indicated by deformed talc did not extend eastward into the magnetite deposits.

Description of mines

Talcville talc belt

In the Talcville talc belt two mines, the International No. $2\frac{1}{2}$ or Freeman mine at Talcville and the International No. 4 mine northeast of Talcville, both owned by the International Talc Co., were actively operated in 1945.

International No. $2\frac{1}{2}$ or Freeman mine

The No. $2\frac{1}{2}$ mine is on the north bank of the Oswegatchie River, at the northwest edge of Talcville, Edwards Township, St. Lawrence County, N. Y. (pls. 21 and 22). The property has been a major producer of excellent quality talc since 1877, when mining began in the Talcville area. Upper levels in the mine were turned from the No. 2 or U. S. shaft, which is now badly caved. The lowest levels used in 1945 extended from the No. $2\frac{1}{2}$ shaft, an unevenly inclined shaft from which eight levels had been turned. The lowermost level, the eighth, about 600 feet vertically below the surface, was then incomplete and all material mined was being hoisted from the seventh level. The levels above the seventh are largely caved and are cut by raises and stopes. Much of the fifth level was accessible in 1945.

The known and inferred form and structural features of the talc belt at the No. $2\frac{1}{2}$ mine are shown in plates 21 and 22 and figure 13. These maps show that the segment of the Talcville belt explored in this mine varies considerably in thickness and form, largely because of the opposed anticlinal and synclinal folds in the vicinity of the U. S. shaft. These folds plunge from 20° to 40° NW. The brecciated rock in the syncline along the footwall of the talc belt is well exposed in drifts and raises between the fifth and seventh levels. The complex anticlinal fold above it is cut by a single drift on the seventh level. Because of the juxtaposition of these folds the talc belt is about 325 feet thick at one place on the seventh level, but it extends from its thickest part to the northeast and southwest in flat V's and for some 450 feet in either direction is relatively unfolded and uniformly layered. Shears developed in the course of the folding diverge from the crests of both folds; they generally follow the layering and form sheeted zones on the flanks of the folds. Along many shears there is a talcose or serpentinous gouge, which is usually a fraction of an inch thick, but it widens in places to several inches or a few feet and rarely to as much as 15 feet. Where these shears are numerous in the more tremolitic layers east and west of the large syncline, talc and serpentine are abundant enough to reduce the specific gravity of the rock and increase its oil absorption factor. Relative movements along single shears are commonly measured in inches or a few feet, but on several of those that cut the micaceous amphibolite about 200 feet west of the large folds, movements of as much as 12 feet are recorded by the displaced amphibolite. The movements indicated by most of the shears are inclined about 10° to 18° from the horizontal measured in the place of shear.

The abrupt thinning of the talc belt at the No. $2\frac{1}{2}$ shaft and 500 to 600 feet west of the major folds occurs in folded and complexly sheared rock. Where the talc belt is thinned the talcose and serpentinized tremolite which form the bulk of the rock mined is cut out entirely, and much of the remainder is discolored and calcareous. The total thickness of the talc rock at either end of the mine workings as developed in 1945 was less than 50 feet.

The amphibolite sheets near the No. $2\frac{1}{2}$ shaft are generally parallel to the schistosity but locally cut across it. The relations of the most persistent sheet to talc rocks and to the enclosing marble have been discussed on page 313. Much of the highly fibrous talc in this area is intimately interlayered with the amphibolite and therefore contaminated beyond use. Pegmatite nodules and lenses are interlayered with the talc near the No. $2\frac{1}{2}$ shaft, especially east of it.

International No. 4 mine

The No. 4 mine is about half a mile northeast of the No. $2\frac{1}{2}$ mine and Talcville (pls. 1 and 23). It is on a part of the Anthony tract, which was first mined in 1877. Its upper levels connect with drifts turned from the now abandoned No. 3 shaft to the west and the No. 5 shaft to the east. All three shafts follow the somewhat variable dip of the talc layers. Eleven levels had been turned from the No. 4 shaft in 1945, at vertical intervals of 50 to 60 feet. The lowest is 600 feet vertically below the collar of the shaft, and is driven in both directions along the talc belt under the old No. 3 and No. 5 mines, which are badly caved.

Most of the talc rock at the No. 4 mine is relatively free from the minerals that commonly constitute impurities in the talc. Quartz, however, is fairly common in the tremolite schist along the upper side of the belt, though usually rare elsewhere. Some hexagonite, phlogopite, and chlorite are interlayered along the hanging-wall side of the belt, and locally near the footwall. Scattered pegmatite nodules and bunches of serpentine, diopside, sphene, apatite, and garnet, occur locally--for example, on the upper side of the belt 300 to 500 feet east of the shaft. Pyrite grains and bunches are scattered along the footwall of the belt, and about 250 feet west of the shaft there is a good deal of pyrite in the basal tremolitic layers.

The major folds in the belt, such as the large open S-fold followed by the tenth and eleventh levels east of the shaft and the abrupt flexure several hundred feet west of the shaft, have a moderate, fairly uniform plunge to the northwest. Subsidiary folds throughout the belt have a somewhat similar orientation.

The abrupt northward curve of the footwall marble about 550 feet east of the shaft on the tenth and eleventh levels pinches the talc belt appreciably, and mining there was limited in 1945 to the west side of the belt. What appears to be the same curve is exposed in the sump hole at and just northeast of the No. 5 shaft (see lowest level of the No. 4 mine, pl. 23). Here a drift penetrates what appears to be the same footwall marble, passing into talc rock beyond. The folded talc belt appears to be cut and offset here by a fault along which rocks on the west have moved relatively northeastward and upward. Northeast of this apparent fault the talc belt has been followed for several hundred feet by a drift from the old No. 5 shaft, but in this interval it is much narrowed, and much more crumpled and sheared than the rock mined to the west. It consists of tremolitic and more talcose rocks irregularly interlayered but contains abundant impurities, chiefly carbonates, quartz, mica, and carbonaceous or graphitic matter.

Small displacements of the talc belt and adjoining marble apparently have occurred along the north-south shears at the abrupt northward curvature of the footwall marble, several hundred feet west of the No. 4 shaft. Throughout the mine, the shears and folds reflect the action of the NE-SW couple discussed on pages 183-191 in the sections on structural features of the district and region. Pronounced differential shearing and intricate drag and flowage folding are widespread in the talc belt. In the explored parts of the belt, these features are more conspicuous east of the No. 5 shaft. Some layers of schistose tremolite are broken or sheared loose from adjoining layers or from the marble, and folded independently of them. The complexly swirled layers exposed on the surface just north and northeast of the No. 5 shaft and on the tenth and eleventh levels, 300 to 500 feet northeast of the No. 4 shaft, are largely the result of this differential shearing and folding. Many of the more complex structures are hardly discernible owing to the absence of distinct color layering in the talc rock. The rocks west of the No. 4 shaft, however, and especially the serpentinous tremolite exposed beyond the abrupt fold several hundred feet west of the shaft are more uniformly layered than those along the strike to the northeast, and are relatively free from complex folding.

Shear surfaces and joints are the most prominent planar structures throughout the mine and are everywhere followed by drifts, raises, and stopes. In areas of least complexity, as north and west of the shaft, the prominent shears either follow the layering in the belt or diverge from it only slightly. In the intricately folded talc rock east of the shaft the larger shears cross many crumpled layers, although in some places they are parallel to the margins of the talc belt (pl. 23). The more prominent shear zones, such as the one striking diagonally eastward across the talc belt east of the No. 5 shaft on the tenth and eleventh levels, and that rudely paralleling the talc layers about 50 feet northwest of the shaft, are locally as much as 12 feet wide. They enclose gouge, irregular chips, and lenticular blocks cut by fracture or slip cleavage. Much of the sheared rock is altered to a pale-green to gray glassy-appearing serpentine, or to talc. Most of these shear surfaces and zones apparently involve displacements of only a few feet, and they appear to merge with fractures of little or no displacement. West of the shaft, diagonal and longitudinal fractures are numerous, well-defined and apparently undisplaced. East of the shaft longitudinal joints are conspicuous almost everywhere, but clear-cut distinction between joints and faults is very difficult.

Wintergreen Hill mine

The Wintergreen Hill talc mine, abandoned and flooded in 1945-47, is about 2,300 feet west of the No. $2\frac{1}{2}$ mine, on the north bank of the Oswegatchie River. It was operated in the period 1911-1921 by a group called the Uniform Fibrous Talc Company.

Two shafts inclined to the north, approximately down the dip of the talc layers, have been sunk at Wintergreen Hill. The original shaft, said to be 350 feet deep, is immediately north of the river. Reportedly because of caving in this shaft, the operators sank a second shaft parallel to the first, but some 200 feet farther north, on about a 42° incline, to a vertical depth of 153 feet. Four short levels, totalling about 600 feet in length, are turned from this shaft to the southwest, toward the river. Plate 25 shows part of these workings. There appears to have been little development eastward from the shafts, which are separated from the northwesterly drifts in the No. $2\frac{1}{2}$ mine by at least 800 feet of unmined and unprospected ground. The Wintergreen Hill mine is said to have been abandoned because of financial and managment difficulties rather than because of the absence of commercial grades of talc. Cushing and Newland (1925, p. 113) report that between 30 and 40 tons of commercial talc rock was being mined each day prior to June 1921 when the mine was closed. Some rock exposed at the surface on Wintergreen Hill consists of tremolite, serpentine, and talc, and is very similar to that mined along the lower side of the belt at the No. $2\frac{1}{2}$ mine. Other exposed layers are chiefly composed of relatively unaltered tremolite, with some interlayered quartz and hexagonite.

Newton Hill and East Anthony properties

The Newton Hill and East Anthony talc properties are on the south slope of Newton Hill, which rises abruptly about 160 feet above the Oswegatchie River, about 1 mile (airline) northeast of Talcville, in Edwards Township (pl. 24).

Much prospecting and a little mining are said to have been carried on at Newton Hill between 1880 and 1890. The most easterly and deepest shaft was driven about 350 feet down the dip of the talc belt during this period, according to John Wallace of Gouverneur (oral communication). Apparently some talc was mined from pockets along the shaft, but no levels were turned. In 1942 Mr. Wallace, who had leased the property from its owner, W. R. Thompson, had three diamond drill holes driven near the shaft.

The Newton Hill mine is apparently near the northeastern limits of the Talcville talc belt. The proportions of tremolite, talc, and serpentine vary considerably both across and along the strike at Newton Hill. Along the footwall side of the belt, talc, or less commonly serpentine pseudomorphous after tremolite, generally predominates over tremolite. On the upper or hanging-wall side of the belt, the rock consists chiefly of tremolite, less than a fourth of which is altered to talc and serpentine, but contains as much as 10 percent each of quartz and carbonates. On the surface the proportion of quartz and carbonates increases appreciably east of the shaft, and some 300 feet east of the shaft diopside becomes conspicuous. In this area and farther east the belt narrows somewhat, and the layers of tremolite and fibrous talc become subordinate in quantity to interlayered quartz, diopside, and carbonates. Hexagonite, phlogopite, and scattered pyrite occur as accessory constituents, partly within the belt but mainly along its margins.

The talc belt at Newton Hill dips 34° to 70° northward, its average dip being about 45°. It attains its maximum thickness, about 85 to 90 feet, some 300 feet west of the shaft, about on the property line between the Anthony and Newton tracts. This bulge appears to result largely from differential shearing and folding along the belt, which is here curved. At the bulge the talc is complexly folded, and sinuous curves and small crumples occur throughout the belt. Most of the folds at the bulge and west of the shaft pitch 5° to 25° west of north, at angles of 23° to 58°, and the bulge itself probably has a similar orientation. East of the shaft there is more divergence in the orientation of folds, but most of them plunge north or slightly west of north.

Fowler talc belt

The Fowler talc belt was being actively mined in 1945 at the Woodcock, Arnold, and Ontario mines of the Loomis Talc Corp., and at the Wight mine of the International Talc Co.

Woodcock mine

The Woodcock talc mine is located about 1,500 yards north of Balmat Corners and 1 3/4 miles southwest of the village of Fowler, St. Lawrence County, N. Y. (pls. 1, 26, 27). Talc prospecting was begun on the Woodcock property in about 1883, and some mining was attempted shortly thereafter (Parker, 1890). The present shaft, a straight one inclined at 55°, was begun in 1934 by the Loomis Talc Corp., and is near the site of the old operations. Four levels, numbered 1, 3, 4, and 5, were turned by 1945. A sixth level, some 300 feet vertically below the surface, was being driven in 1955. Talc was being taken from stopes between the third and fifth levels, and trammed from the fourth and fifth levels.

The talc belt explored at the Woodcock mine ranges from about 125 to 225 feet in thickness (measured normal to the dip). It includes three explored and variously productive talc zones, one near the base and two in the upper half of the belt, which are commonly separated by equal or greater thicknesses of impure talc and marble.

The least-folded parts of the talc belt strike northeast and dip steeply to moderately northwest. The entire belt, as well as specific zones within it, show varying degrees of crumpling and shear, generally increasing in intensity northeastward. At the northeast end of the mine the layers are contorted into close accordion folds whose crests and troughs, especially in the more talcose layers, are appreciably thickened by migration of material from the highly compressed flanks. The folds in this area and southwest throughout the mine plunge approximately north at angles between 15° and 60°, averaging perhaps 35°. The orientation of the folds is sufficiently uniform, however, to warrant the belief that they continue to depths as yet unmined without much change in form and position.

Cross joints are especially conspicuous in the more fibrous talc layers; they make angles not exceeding 25° with the fold axes and lineation. Locally these fractures appear to pass into one or both sets of diagonal fractures.

Arnold mine

The Arnold mine, operated by the Loomis Talc Corp., is about three-quarters of a mile south-southeast of the village of Fowler and Highway 58, on a relatively unfolded segment of the Fowler talc belt (pl. 31). In the mine, which is one of the oldest and largest producers in the district, 15 levels, aggregating some 2,700 feet in length, are turned from a shaft which follows the somewhat variable dip of the minable talc. The levels are from 45 to 65 feet apart, and number 15, the lowest existing in 1945, is about 170 feet below sea level. Many levels, raises, and stopes are badly caved because of the blocky nature of the jointed, sheared, and layered talc rock. Still partly or completely open in June 1945 were levels 4, 5, 6, 7, 8, 10, 12, and 13 east of the shaft, and 4, 10, 14, and 15 west of the shaft. The workings explore the minable talc zone for about three-fifths of a mile along the strike and for more than 1,000 feet down dip. Almost two-thirds of the strike distance is west of the shaft, from which the bulk of the talc has been mined in the past, and where all operations were concentrated in June 1945. Rock was being taken from stopes and raises between the eighth and fifteenth levels and was trammed and hoisted from levels 10, 14, and 15.

The talc belt at the Arnold mine ranges in thickness from about 75 to 200 feet. Most of the talc mined to 1945 was taken from the lower side of this belt, in a zone exceptionally free from impurities. Overlying zones, where cored on the eighth level and crosscut on the fourth, were found to be variously contaminated with carbonates, hexagonite, mica, quartz, and oxides of iron and manganese, and were not considered worth mining.

Major shears or slips are common in the upper margin of the mined talc zones, and some are found near the lower margin. Locally within the zones, subparallel partings and incipient to well-defined shears slice the rock into sheets. Sheeting is especially prominent (1) on the tenth level from 200 to 500 feet east of the shaft and from 400 to 1,100 feet west of the shaft, (2) on the fourth level 50 to 150 feet west of the shaft, and (3) on the fourteenth level from the shaft west about 250 feet.

There are all gradations between partings and the major slips which parallel the color bands and textural layers in the talc or diverge from them at acute angles. The walls of the shears are flat to sinuous or variously bulged. Locally the walls are slickensided; the slickensides are generally parallel to the dip of the shear or diverge from it by less than 15°. Displacements along curved shears can rarely be measured, but some movements appear to have brought outward or inward bulges on the hanging-wall shear into contact with reverse bulges in the footwall. Pinches in the mined zone, which are interpreted as the result of movements along curved shears, occur at the northeast headings on levels 7 and 15, and also about 700 feet east of the shaft on level 8.

Major folds, which constitute one of the dominant structural features elsewhere along the talc belt, are absent in the Arnold mine, and the mined layers have a relatively constant strike to the north-northeast throughout the mine. Fifty feet east of the shaft and at the west heading on the tenth level, and 200 feet west of the shaft on the fifteenth level, there are minor north-south crumples that plunge 15° to 50° N., like the folds in the talc belt east and west of the Arnold mine. The largest syncline or roll on the fifteenth level, which is flat to gently plunging and lies 150 feet west of the shaft, seems to pass eastward into a moderately plunging fold parallel to the crumples.

Fracture sets involving longitudinal, cross, and diagonal joints are common in the Arnold mine (pl. 31), where they cut the talc rock into blocky masses and constitute~~s~~ a real mining problem.

The dips of the explored zones vary considerably, ranging from almost vertical at the east heading on the sixth level to a low of about 10° at the west end of level 10. These extremes, however, are merely local, most of the dips being moderate and as uniform as they are in any comparable length of talc belt else~~some~~where in the district.

Wight mine

The Wight mine, operated by the International Talc Co., is about 1 1/8 miles south of the village of Fowler, on the Balmat - West Branch talc belt (pls. 27 and 28). In 1945 the present Wight shaft was sunk about 265 feet, at an incline of from 30° to 45°, down the dip of the basal talc layers. Six levels have been turned from the shaft, at vertical intervals of 25 to 75 feet. The major development and production is along the basal layers of the talc belt, but on levels 3 and 5, two crosscuts each about 350 feet long have been driven completely across the talc belt. From these crosscuts other talcose zones have been followed for short distances along the strike.

The talc belt at the Wight mine is about 200 feet thick (measured normal to the dip). Five zones within the belt are pure enough to warrant consideration as commercial talc. The basal zone, immediately above the marble, is by far the purest and most uniform. This zone is presumably continuous with the productive zone at the Arnold mine to the east. A moderate proportion of the tremolite in this zone has been altered to talc; layers of unaltered or slightly altered tremolite occur at the base of the zone, but much less commonly than in the Arnold or Ontario mines.

Shear surfaces invariably define the upper margin of the basal talc zone and occur locally within it and along its footwall. Foliate talc occurs sporadically in lenses and layers 1 to 3 feet wide, mostly along the upper shears and immediately below them. A sheared, micaceous, and serpentinized amphibolite occurs in the basal zone, about 100 feet west of the shaft on the sixth level and somewhat farther west on higher levels.

The explored and locally mined layers above the basal zone, with the exception of the uppermost, are highly sheared and relatively talcose, usually containing less than 25 percent residual tremolite. The uppermost zone contains numerous highly tremolitic layers.

The explored part of the talc belt in the Wight mine as in the Arnold mine to the northeast, is only slightly sinuous or curved. Shears and local layers within the belt are likewise nearly flat or gently curved surfaces and sheets. Crumples and tiny folds are uncommon and widely scattered. They pitch moderately to the northeast, in conformity with the major folds in the talc belt at the Woodcock mine to the west. The dips of the layers and nearly parallel shears in the talc belt near and east of the shaft are also moderate, ranging from 25° to 50° NW. About 250 feet west of the shaft the exposed basal layers steepen abruptly and pass through the vertical, and at the west heading on the fifth level, about 425 feet west of the shaft, the footwall layer dips about 65° SW. This reversal in dip is probably the first of the structural complications related to the complex folding and shearing of the talc belt at and east of the Woodcock mine (pl. 27).

Ontario mine

The Ontario mine (also called the No. 4, Potter, or Little York mine) of the Loomis Talc Corp., is on State Route 58, just west of the West Branch of the Oswegatchie River, and 1 3/4 miles east of the village of Fowler, in Edwards Township (pls. 29 and 30). The mine was operated until 1917 by the Ontario Talc Co., who turned the four upper levels. Thereafter it was closed until the spring of 1940, when the Loomis Talc Corp. acquired it and resumed mining. A fifth and sixth level have been developed, and in June 1945 rock was being mined between the third and sixth levels, with tramming and hoisting from the fourth, fifth, and sixth levels.

The talc belt at the Ontario mine varies from 30 to about 150 feet in width. The bulk of the commercial talc is mined from the lowermost zone of the belt, whose thickness ranges up to 25 feet, but a narrow zone of relatively fibrous talc and tremolite near the top of the talc belt has been mined above the fourth level.

The lowermost productive zone is followed down dip by the Ontario shaft (pl. 30) and has been drifted on to the northeast and southwest for more than 1,100 feet. It consists chiefly of highly sheared, crumpled, fibrous to foliate talc, mixed with a smaller amount of serpentine and residual tremolite from which the other minerals were derived. The quantity of tremolite varies considerably but averages perhaps 20 or 25 percent. Layers of schistose to nearly massive tremolite commonly underlie, and sometimes overlie, the more talcose layers, separating them from the marble below or from discolored and less talcose rocks above.

The thinner zone of talc rock mined near the hanging wall of the talc belt has been followed west of the shaft for about 300 feet on the second level and for shorter distances on the third and fourth levels. In this zone layers of a tough, fibrous rock consisting of talc, anthophyllite, and tremolite are more abundant than in the lower mined zone; they are intercalated with layers of more foliated talc.

The talc belt at the Ontario mine strikes roughly east-northeast and dips moderately northward. The most conspicuous structural features within it are shear surfaces subparallel to the layering. The "hour-glass" form of the mined talc zone, which results from relative movements of curved marginal shears, is apparent in plate 30. Tiny crumples and drag folds, developed concurrently with the shears, are present everywhere in the talc belt. The strongest folding along the belt may be seen at the southwest end of the level and in the eastern parts of levels 5 and 6. In these places both large and subsidiary folds are subparallel, and like most of the crinkles throughout the mine they plunge about 30° to 40° NE.

Outlook of future mining

There are sufficient reserves of talc in the Gouverneur district to last, under resourceful mining methods, for several generations at the present rate of production. In 1945 mines had penetrated the talc belts to depths ranging from several hundred to more than 800 feet and the deeper drill holes along the talc belts had cored as much talc beneath the workings as had been mined.

In addition to the talc exposed, mined, or inferred to exist beneath the mines, there are extensions or continuations of the belts between the operating mines. These are masked on the surface by Quaternary sands, drift, and river alluvium. More precise knowledge of their form, extent, and quality must necessarily be derived from exploratory drilling and mining operations. It is clear, however, that much commercial talc occurs at shallow depths between existing mines.