THE GREEN MOUNTAIN ANTICLINORIUM IN THE VICINITY OF ROCHESTER AND EAST MIDDLEBURY, VT.

By

PHILIP HENRY OSBERG

VERMONT GEOLOGICAL SURVEY

CHARLES G. DOLL, *State Geologist*

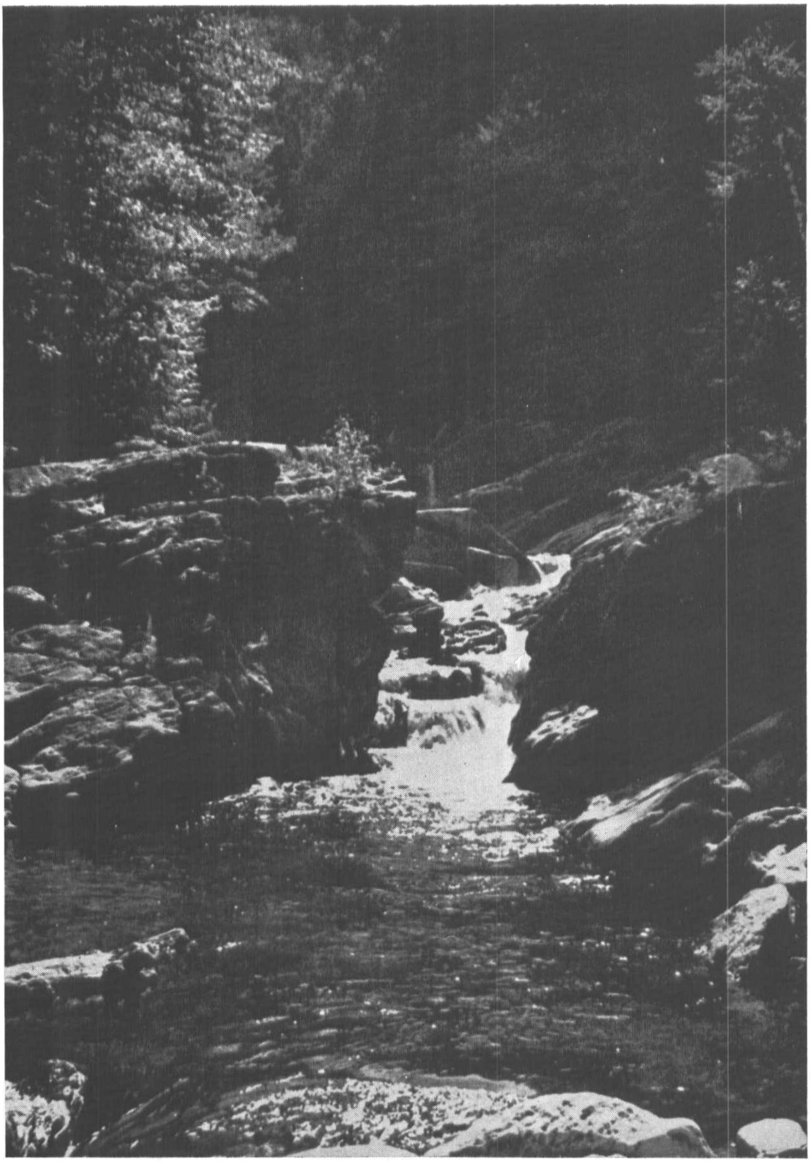
Published by

VERMONT DEVELOPMENT COMMISSION

MONTPELIER, VERMONT

SECOND PRINTING, 1964

BULLETIN NO. 5 1952



Mount Holly complex, South Branch of the Middlebury River, Ripton, Vermont

TABLE OF CONTENTS

Illustrations

THE GREEN MOUNTAIN ANTICLINORIUM IN THE VICINITY OF ROCHESTER AND EAST MIDDLEBURY, VT.

By

PHILIP HENRY OSBERG

ABSTRACT

The Green Mountain anticlinorium was studied in detail in the northern half of the Rochester quadrangle and in the East Middlebury quadrangle, Vermont. The rocks of this area belong to two Lower Paleozoic stratigraphic sequences, contrasting in lithology but more or less equivalent in time. Both sequences rest unconformably on the Pre-Cambrian complex.

The western sequence is composed dominantly of dolomite, quartzite, gray quartz-muscovite schist, albite-quartz-biotite-muscovite schist, and conglomerate. In the Rochester-East Middlebury area, this sequence is about 6,000 feet thick and is subdivided into eleven formations. Paleontological evidence suggests a Cambrian age for these rocks.

The eastern sequence consists of about 8,000 feet of albite-quartz-chlorite-muscovite schist, graphitic quartz-muscovite schist and albite-epidote-calcite-chlorite schist. These rocks are believed to be metamorphosed aluminous shales and interbedded graywackes and mafic volcanics. This eastern succession is divided into five formations. No fossils have been found but the relation of these rocks to the Pre-Cambrian unconformity and to a Middle Ordovician unconformity indicates that they are roughly equivalent in age to the rocks of the western sequence.

Most of the rocks of this region have been recrystallized in either the chlorite or garnet zones of metamorphism. They are subdivided into two compositional groups, the alkali-aluminous and the mafic-aluminous groups, and the mineral assemblages for each group are compared under conditions of increasing metamorphism. In the Rochester-East Middlebury area the index minerals for the alkali-aluminous rocks are chlorite and garnet, although in rocks that have a high K_2O to Al_2O_3 ratio, biotite is stable throughout the chlorite zone. The index minerals for

the mafic aluminous rocks are chlorite and actinolite. Actinolite and garnet form under approximately the same intensity of metamorphism.

The Green Mountain anticlinorium in this region consists of three anticlines. The westernmost is the Ripton anticline; the central anticline is the Bread Loaf; and the easternmost is the Burnt Hill anticline. The Ripton and the Bread Loaf anticlines have axis-culminations in the latitude of Ripton. Both of these anticlines are overturned toward the west. The Burnt Hill anticline has an axis-culmination that lies north of the Rochester East Middlebury area. This anticline is sharply overturned toward the west and may be, in part, recumbent. The east flank of the Green Mountain anticlinorium has a relatively gentle average dip to the east. Locally, however, the east flank is contorted by folds that are overturned toward the west and undergo abrupt reversals in plunge. The Middlebury synclinorium lies to the west of the Green Mountain anticlinorium and is overturned and thrust-faulted toward the west.

A variety of minor structures were observed, including drag folds, foliation, schistosity, cleavage banding, lineation, stretched cobbles and rotated porphyroblasts. The minor structures east of the Green Mountain anticlinorium are incongruous with the major structures; those west of the anticlinorium are congruous with the major structures.

The age of the folding of the Green Mountain anticlinorium is difficult to establish. Some evidence indicates that this region was folded during both the Taconic and the Acadian disturbances. No direct evidence of the roots of the Taconic thrust was found in this area.

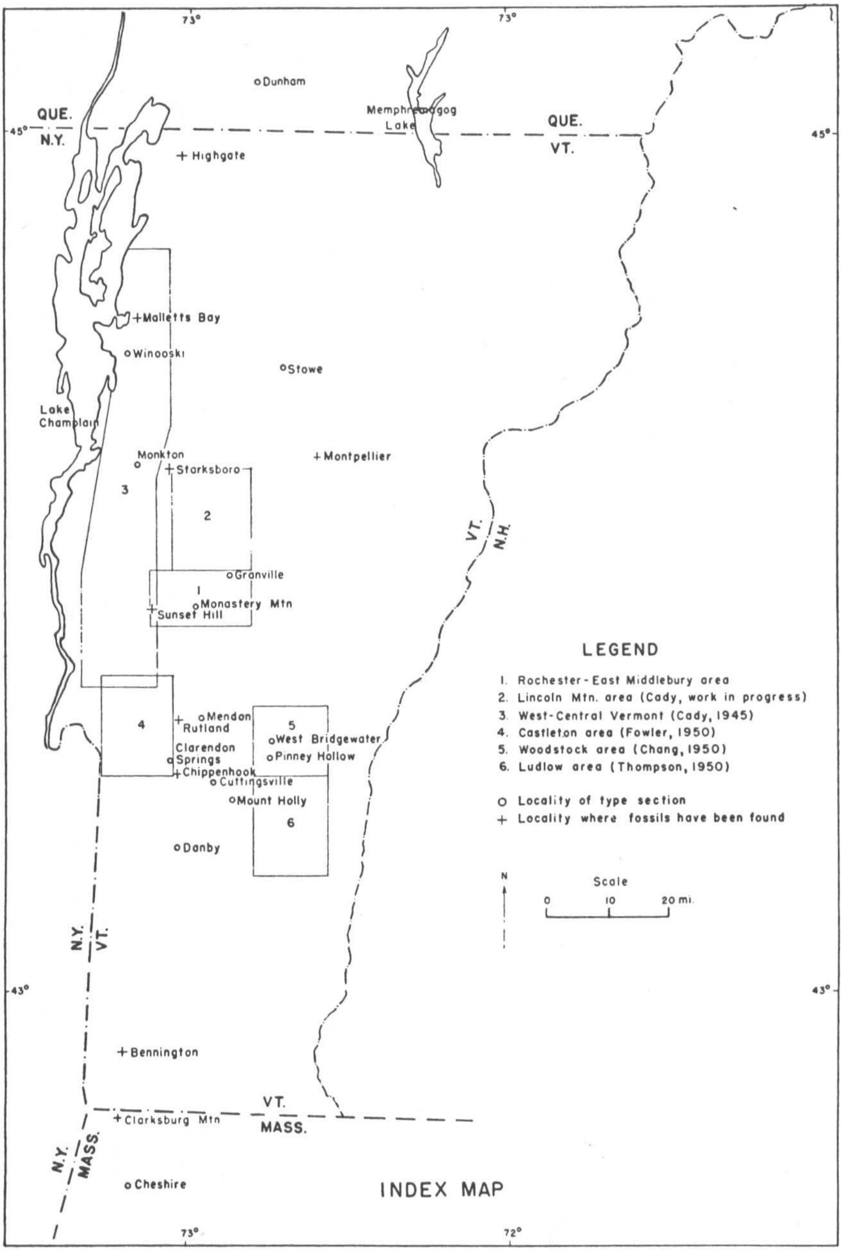
INTRODUCTION

Location

The purpose of this study is an analysis of the Green Mountain anticlinorium in the vicinity of Rochester and East Middlebury, Vermont (Figure 1). This area, which comprises the northern half of the 15 minute Rochester quadrangle and the entire 7½ minute East Middlebury quadrangle, covers approximately 167 square miles. It is bounded by parallels 43°52′30″ and 44°00′ north latitude and by meridians 72°45′ and 73°07′30″ west longitude.

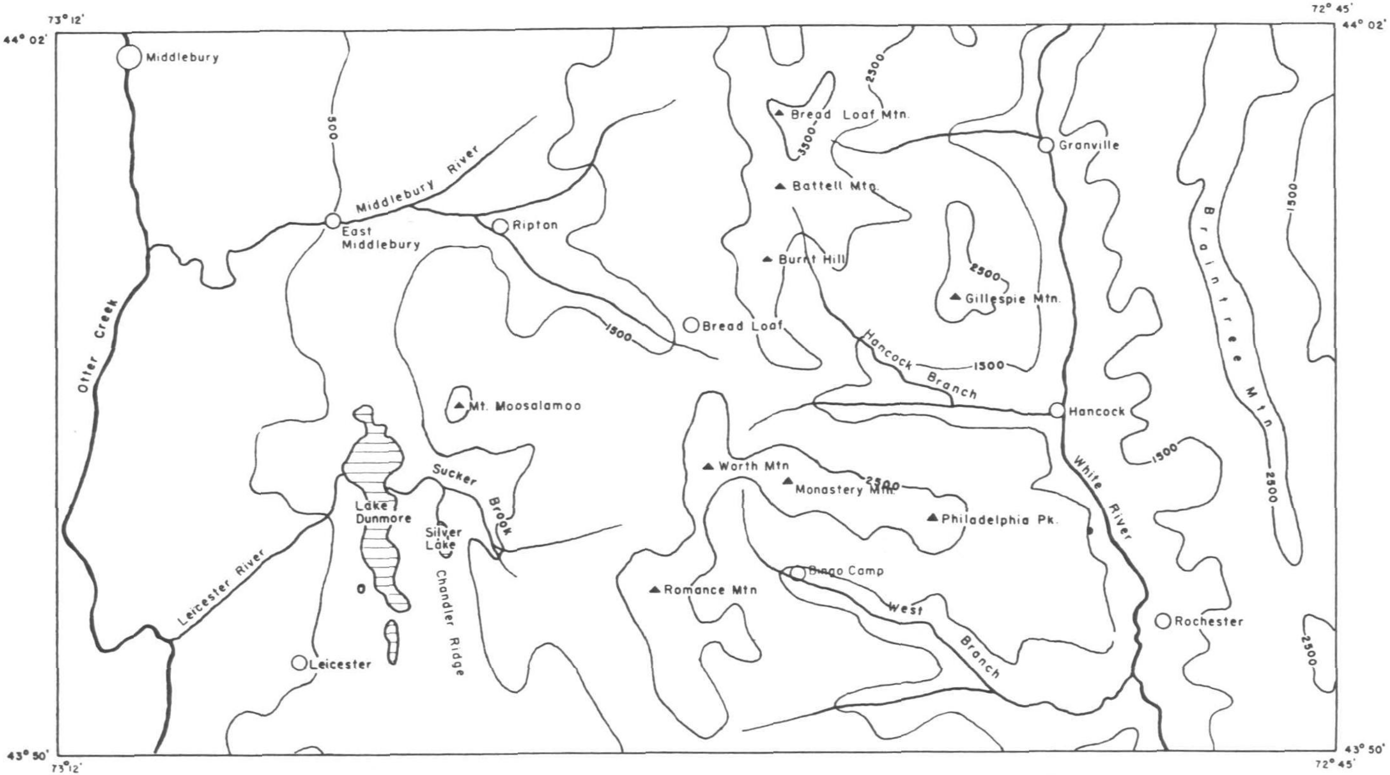
Topography and Drainage

A simplified map of the topography and drainage of the Rochester-



73°
73°
45°
45°
43°
43°
73°
72°
QUE.
N.Y.
QUE.
VT.
o Dunham
Memphremagog Lake
+ Highgate
+ Malletts Bay
o Winooski
o Stowe
Lake Champlain
Monkton
+ Starksboro
+ Montpellier
3
2
VT.
N.H.
o Granville
1
o Monastery Mtn
+ Sunset Hill
LEGEND
1. Rochester - East Middlebury area
2. Lincoln Mtn. area (Cady, work in progress)
3. West-Central Vermont (Cady, 1945)
4. Castleton area (Fowler, 1950)
5. Woodstock area (Chang, 1950)
6. Ludlow area (Thompson, 1950)
o Locality of type section
+ Locality where fossils have been found
4
+ Rutland
o Mendon
5
o West Bridgewater
o Pinney Hollow
o Clarendon Springs
+ Chippenhook
o Cuttingsville
o Mount Holly
6
o Danby
N
Scale
0
10
20 mi.
N.Y.
VT.
+ Bennington
VT.
MASS.
+ Clarksburg Mtn
N.Y.
MASS.
o Cheshire
INDEX MAP

Figure 1.



PHYSICAL FEATURES OF THE ROCHESTER-EAST MIDDLEBURY AREA

N

Scale

0 1 2 Mi.

Contour interval 1000'

Figure 2.

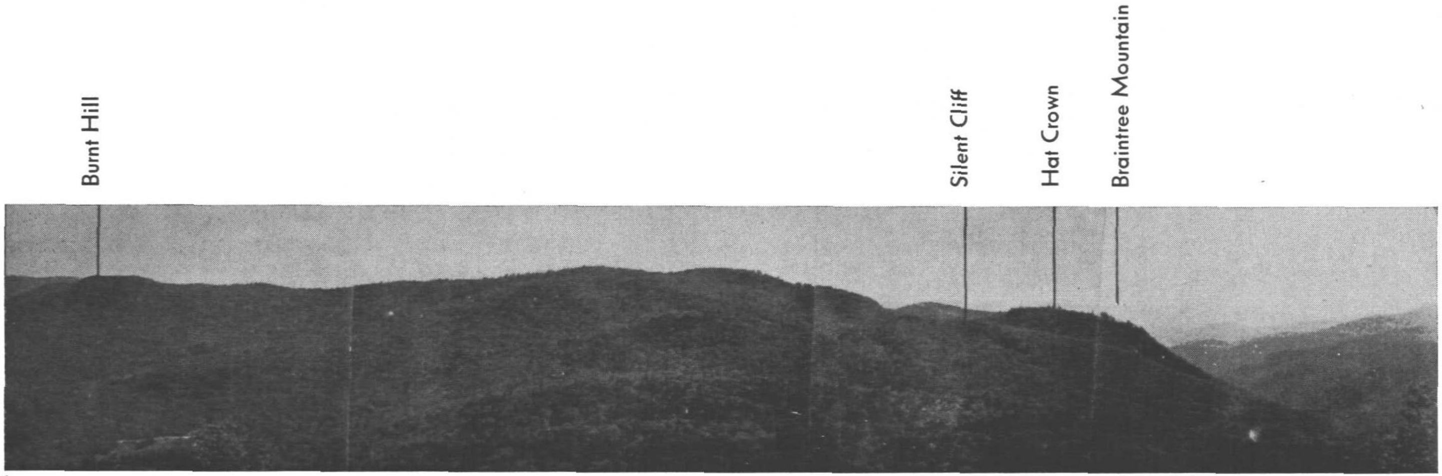


Figure 3. Panorama of the northern portion of the
Rochester quadrangle from Worth Mountain.

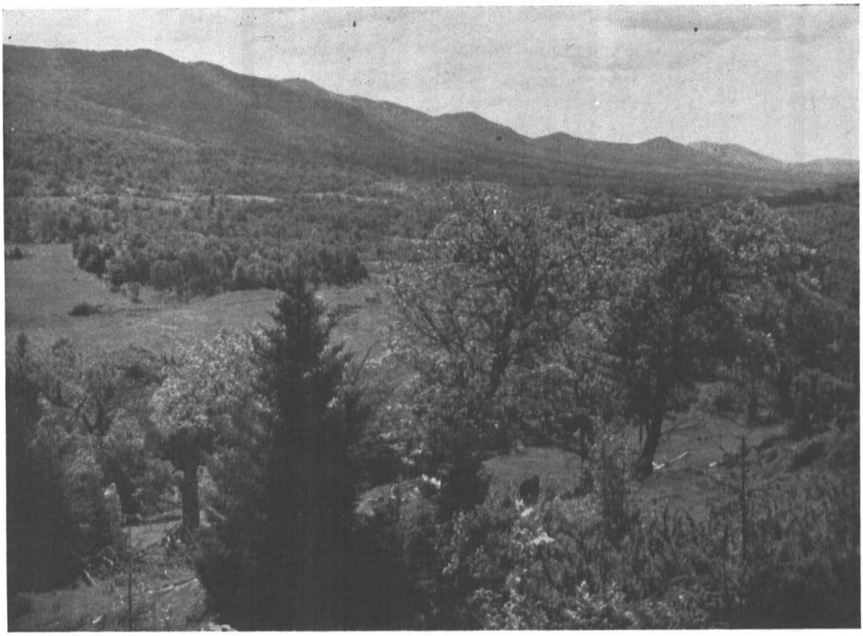


Figure 1. Braintree Mountain looking southeast from Old Sixty Hill.

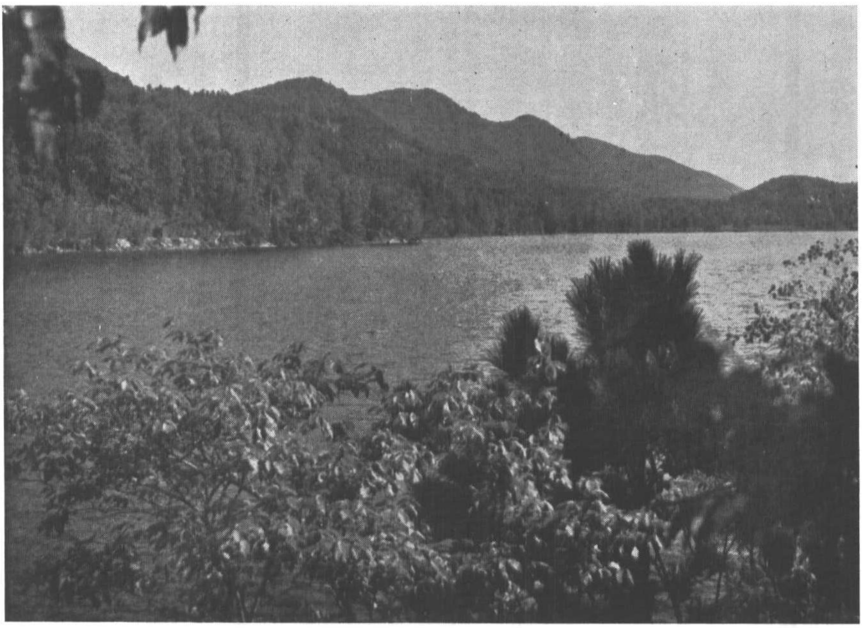


Figure 2. Chandler Ridge looking southeast from Hyolithes Point, Lake Dunmore.

PLATE 3

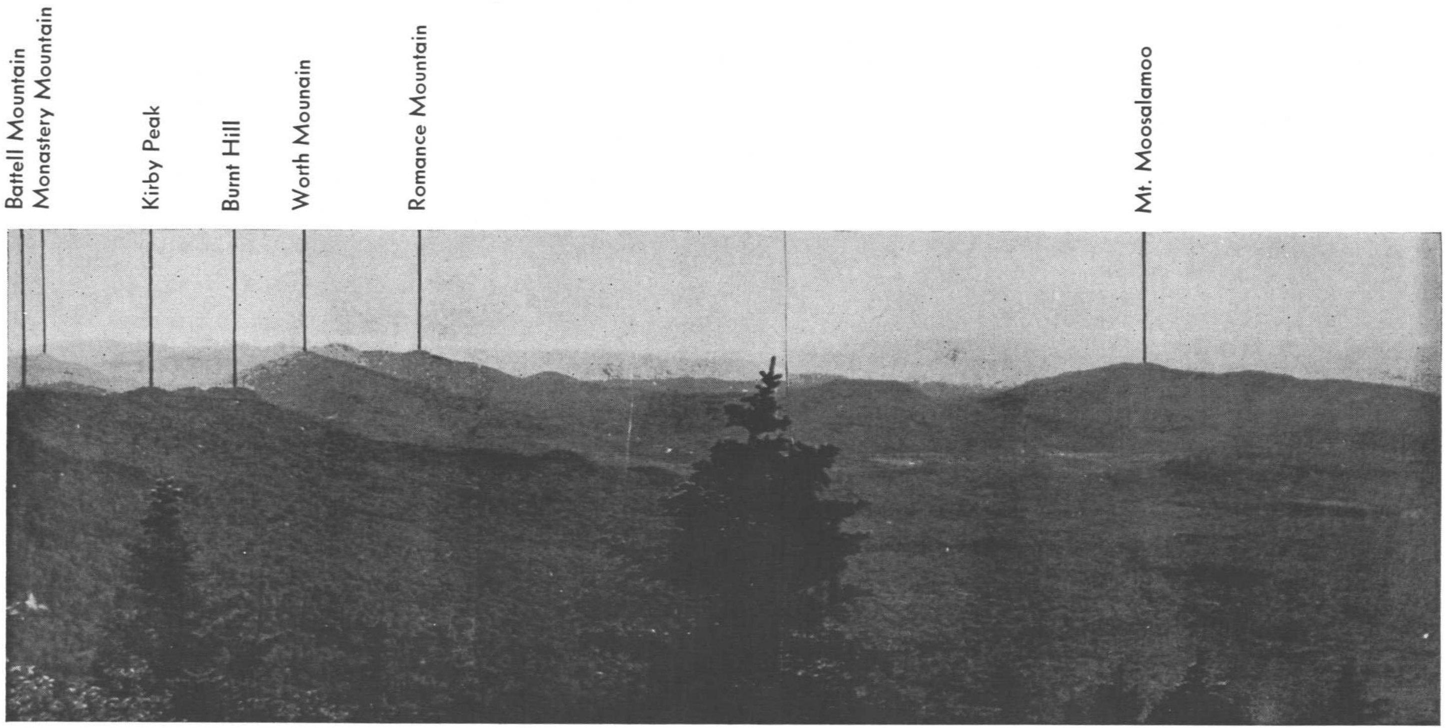


Figure 4. Panorama of the western portion of the Rochester–East Middlebury area from Bread Loaf Mountain.

East Middlebury area is shown in Figure 2. The eastern half of the area is drained by the White River and its tributaries. The White River rises on the southeast slopes of Bread Loaf Mountain. It flows eastward to Granville where it turns southward into the Hancock-Rochester Valley. At Hancock it is joined by the Hancock Branch, which flows eastward from the vicinity of Middlebury Gap. The western half of the area is drained by the Middlebury River, Sucker Brook, and the Leicester River. The Middlebury River drains the area north and west of Middlebury Gap and flows westward to Middlebury where it joins Otter Creek. A small area, north of Romance Mountain and west of Worth Mountain, is drained by Sucker Brook, which flows westward into Lake Dunmore. The Leicester River serves as the outlet for Lake Dunmore, joining Otter Creek at Leicester Junction.

The Rochester-East Middlebury area extends across the Green Mountains which, in this vicinity, consist of three north-trending ridges. Braintree Mountain (Plate 3, Figure 1) forms the eastermost ridge, where the summits, which range from 2600 feet to 2900 feet in altitude, rise 1600 feet above the town of Hancock. The central ridge (Figures 3 and 4) forms the backbone of the Green Mountains. It is an ancient upland consisting of the Battell Mountain-Worth Mountain ridge, Gillespie Mountain, Monastery Mountain, and Philadelphia Peak. It is about six miles wide, extending parallel to and west of the White River. This central ridge, culminating in Bread Loaf Mountain (3840 feet in altitude), is highly dissected by tributaries of the White River that have produced a relief of from 1700 feet to 2200 feet above the town of Hancock. Between the central ridge and the westernmost ridge of the Green Mountains lies an area of relatively low relief ranging from three to four miles in width. The western ridge (Figure 4, Plate 3, Figure 2), consisting of Mount Moosalamoo and Chandler Ridge, forms a series of low summits, 1700 feet to 2600 feet in altitude, which mark the western limit of the Green Mountains. The western slope of this ridge forms a steep escarpment that has a relief of 500 feet to 1000 feet above the Champlain lowland.

Purpose of Study

Considerable information has been compiled on the structure and stratigraphy of the Champlain Valley, and many data are being accumulated on the schists east of the Green Mountains. The present study is directed toward correlating the eastern and western lithologic sequences and determining the major and minor structural features of the Green Mountains. A further aim is to discuss in detail the petrography and metamorphism of this area.

Field Method

The writer spent two and one-half summers (1948 to 1950) mapping the Rochester-East Middlebury area. United States Geological Survey topographic quadrangle sheets were used as base maps. The Rochester quadrangle (1917 edition, reprinted in 1947) on a scale of 1/62,500 was enlarged to a scale of approximately two inches to the mile. The East Middlebury quadrangle (1946 edition) on a scale of 1/31,680 was used without enlargement. The locations of outcrops were plotted on the base maps by means of compass bearings and with the aid of an aneroid barometer, and where the slopes were gentle, by means of pace and compass methods. Field sketches were made at pertinent outcrops, and detailed pace and compass maps were made of some of the ultramafic bodies and of local areas of complex structure.

Acknowledgments

The writer is indebted to Professor Marland P. Billings under whose direction this work was pursued and who gave freely of his time for instruction and criticism. Thanks are also due Professor James B. Thompson and Dr. Wallace M. Cady for valuable field suggestions and for helping the writer gain an insight into the regional geologic picture. Members of adjacent field parties provided helpful discussions of common problems. Among these are P. H. Chang, J. L. Rosenfeld, P. Fowler, J. W. Skehan, and W. F. Brace.

The assistance of Bruce Nelson in 1949 and Bruce Bryant in 1950, as members of the field party, was invaluable, and their interest and cooperation greatly facilitated the work. F. A. Gonyer made the chemical analyses.

Professor Charles G. Doll, State Geologist of Vermont, arranged for financial assistance by the Vermont Geological Survey for the writer in 1949 and 1950. Acknowledgment is also made to the Shaler Memorial Fund of Harvard University for its generous grants, which were used to defray the expenses of the assistants, and to the Department of Mineralogy and Petrography for furnishing the chemical analyses and thin sections.

Previous Work

The first comprehensive study of this area appeared in a report in 1861 made by Edward Hitchcock under the auspices of the Geological Survey of Vermont. He recognized the anticlinorial structure of the Green Mountains and successfully delineated the broader lithologic units. His report describes gneiss on the central ridge of the Green Mountains and conglomerate at Ripton and Goshen. The rocks east of the Green Mountains were referred to as talcose schist and clay slate.

C. L. Whittle (1894), as a member of the Archean Division of the United States Geological Survey, confirmed the anticlinorial structure of the Green Mountains east of Rutland. He attempted to subdivide the rocks exposed in the core of the anticlinorium into Archean and Algonkian.

T. N. Dale (1918) studied the area between Ripton and Bread Loaf. He described the Ripton conglomerate and inferred that it lay at the base of the Cambrian. His map showing its distribution is fairly accurate. In 1915 Dale briefly described the marble at Rochester and Hancock.

The Rochester-East Middlebury area was briefly examined by W. G. Foye (1918). He corroborated the evidence for an anticlinorial structure of the Green Mountains in this area. He also described the various rocks found within the Rochester quadrangle.

C. H. Richardson (1918 through 1926), in a series of papers, traced an unconformity east of the Green Mountains from Canada to Massachusetts. Although he considered this unconformity to be the contact between the Cambrian and Ordovician, it is now thought to be mid-Trenton in age.

E. L. Perry (1928) studied the stratigraphy of Bridgewater and Plymouth townships. A lithologic sequence assigned to the Pre-Cambrian, Cambrian, and Ordovician was established.

A regional study of the Green Mountains was made by Arthur Keith (1932). He suggested that the schists and slates of the Taconic Mountains were allochthonous, and that the roots of this allochthon might be found on the east side of the Green Mountains. He also suggested that the fossiliferous Lower Cambrian quartzite on the west flank of the Green Mountains rested with regional unconformity on the underlying schists and marbles. His conclusion was based on the assumption that several thousand feet of schists and marbles lay between the Lower Cambrian quartzite and the gneisses of the core of the Green

Mountains in the vicinity of Ripton. On the other hand, only a hundred feet of schists lay between a similar quartzite and the gneisses of the core of the anticlinorium on Clarksburg Mountain, Massachusetts. Hence he considered that the Lower Cambrian quartzite truncated more than a thousand feet of rock between these two localities.

H. E. Hawkes (1940) examined the northern half of the Rutland quadrangle and the southern half of the Rochester quadrangle. His chief contribution was in tracing the formations described by Perry in the Bridgewater-Plymouth area northward to Rochester.

W. M. Cady (1945) described the geology of the Champlain Valley between Brandon and Milton. He established the stratigraphic succession and the synclinorial structure of this region. His work was carried eastward to the west slope of the Green Mountains.

LITHOLOGY AND STRATIGRAPHY

General Statement

In the vicinity of the Rochester-East Middlebury area, the Pre-Cambrian rocks of the Green Mountain anticlinorium are flanked on the west by a dolomite-quartzite sequence and on the east by a schist sequence. These two lithologic sequences are equivalent in age and represent two contrasting environments of deposition. The western succession is a shelf sequence composed of sediments that have undergone prolonged winnowing and consequent removal of unstable minerals. These sediments have been derived predominantly from the west and are typical of a miogeosyncline (Kay, 1947). The eastern succession was derived from an eastern source that contributed great quantities of sediment and is typical of a eugeosynclinal assemblage (Kay, 1947). Some evidence of interfingering of the two sequences exists.

Most of the rocks of the area are metamorphosed and are of sedimentary and volcanic origin. The rocks west of the Pre-Cambrian complex are dolomitic marbles, calcitic marbles, quartzites, graphitic quartz-muscovite schists, and conglomerates. This western succession is 5,000 to 6,000 feet thick and is subdivided into eleven formations. No fossils were found by the writer, but previous workers have reported Cambrian and Ordovician fossils (Walcott, 1888; Wolff, 1891; Cady, 1945).

The metasediments east of the Pre-Cambrian are characterized by graphitic quartz-muscovite schists, quartz-chlorite-muscovite schists (with or without albite or garnet porphyroblasts), albite-quartz-biotite-muscovite schists, conglomerates, thin quartzite beds, and small lenses of marble. Epidote-albite-carbonate-chlorite schists probably represent metavolcanics. This sequence, in the Rochester-East Middlebury area, is about 8,000 feet thick and is subdivided lithologically into seven major units. Although no fossils have been found, a Cambrian age may be postulated.

The metasediments of the eastern and western sequences rest unconformably on a basement of more highly recrystallized and metamorphosed rocks.Gneisses, schists, marbles, and quartzites predominate in this older succession which, because of its stratigraphic relation to the overlying rocks, is considered to be Pre-Cambrian.

Metamorphosed ultramafic bodies occur within the Cambrian rocks of the eastern sequence. These bodies, for the most part, have been recrystallized to serpentine, talc and carbonate.

Three varieties of dikes intrude the Rochester-East Middlebury area. Meta-kersantite dikes in the Pre-Cambrian gneisses have been involved in the folding. Albite porphyry dikes intrude the Cambrian succession on the east limb of the Green Mountain anticlinorium. These are slightly cleaved but not crumpled. Mafic dikes, younger than the folding, intrude all the rocks of the area.

Rock Nomenclature

Many of the terms used in the nomenclature of metamorphic rocks are loosely defined and are therefore ambiguous. The classification employed in this paper is purely descriptive, being based on texture and mineral content.

Texture as used here refers to grain size, relative shapes of grains, and relative orientation of adjacent grains. If some of the individual minerals are conspicuously larger than the surrounding minerals, the large minerals are termed "porphyroblasts," and the smaller minerals are termed "groundmass." If the individual mineral particles approach a common size, they are termed "grains." Minerals are also grouped according to their relative shapes—equidimensional minerals and nonequidimensional minerals. The nonequidimensional minerals are either "platy" or "needle-shaped."

The following textural names are used in this paper:

Schist— a recrystallized rock consisting of minerals that can be resolved megascopically. At least 30 per cent of the non-equidimensional minerals have a common orientation.

Gneiss— a recrystallized rock consisting of minerals that are large enough to be recognized megascopically. Between 30 and 10 per cent of the nonequidimensional minerals have a common orientation.

Granulite— a recrystallized rock in which the minerals are recognized megascopically. Less than 10 per cent of the nonequidimensional minerals have a common orientation.

Marble— a granulite consisting dominantly of calcite or dolomite.

Quartzite— a granulite consisting dominantly of quartz.

Amphibolite—a gneiss or a schist consisting essentially of amphibole.

The full rock name consists of a textural name preceded by a mineral name that consists of the names of the essential minerals of the rock linked together by hyphens. Thus, if a rock has the texture of a schist and if it is composed of quartz, albite, chlorite, and muscovite, the full rock name is albite-quartz-chlorite-muscovite schist.

Stratigraphic Nomenclature

Stratigraphic nomenclature is loosely defined. The delineation of rock bodies used in this paper is briefly outlined below. The "Committee on Stratigraphic Nomenclature" (Ashley, 1933, pp. 429–431) defines formation as the fundamental unit in the local classification of rocks. Formations are distinguished as far as possible on the basis of lithologic unity and continuity of areal extent. They should be discriminated so that they may be readily traced in the field and represented on geologic maps. Furthermore, formations must have a considerable geographic extent in order to be useful tools with which to express the geologic development and structure of an area. As defined by the "Committee on Stratigraphic Nomenclature" (Moore, 1947; pp. 513–528) the formation has no time connotation; that is, the formation boundaries do not necessarily coincide with hypothetical planes representing equivalent time.

Members (Ashley, 1933, pp. 420, 438) are defined as subdivisions of a formation and have a considerable geographic extent.

In this paper, the definitions of formations and members are based on

1) approximate unity of chemical composition, 2) practicability of mapping, and 3) geographic extent.

Rock units are delineated by changes in bulk chemical composition as reflected in the mineral assemblage. This is necessarily true in metamorphic rocks, because original textures are commonly obliterated, and a variety of metamorphic textures may be impressed subsequently on the rock. Furthermore, purely petrographic features are not sound criteria for their subdivision. A rock unit which was originally mineralogically homogeneous may, after metamorphism, consist of two or more different mineral assemblages in different localities, because of differences in metamorphic conditions even though the bulk chemical composition remains the same. Consequently, stratigraphic subdivision of recrystallized rocks must be based on approximate chemical composition as determined from a megascopic examination of the mineral assemblage. It follows that an intimate knowledge of the chemical composition of the constituent minerals is necessary.

Such rock units may be either formations or members. They are differentiated on the basis of practicability of mapping and the geographic distribution. Rock units that are at least 150 feet thick can be mapped (on a scale of 1/62,500) under all circumstances, even if the bedding is vertical. Rock units that are less than 150 feet thick can not be mapped if the bedding is vertical, because under such circumstances the rock unit would appear on the map with a width of a pencil line or less. On the other hand, if the bedding is inclined from the vertical, the breadth of outcrop may exceed 150 feet, and in such cases the rock unit is shown to scale on the map.

Finally, formations should have great geographic extent, perhaps being traceable for a distance of at least 50 miles. Members have a much more limited distribution.

The above discussion leads to the following stratigraphic usage:

Formation is a rock unit that is at least 150 feet thick and is amenable to mapping over a geographic area of at least 50 miles.

Member is a rock unit that forms a distinctive part of a formation. In general it is 1) less than 150 feet thick and may or may not have a geographic extent of 50 miles, or 2) over 150 feet thick but is not amenable to mapping over an extent of 50 miles.

Names of formations and members have been employed according to the rules set forth by the "Committee on Stratigraphic Nomenclature" (Ashley, 1933, pp. 433–434). The names are binomial, consisting of a

TABLE 1

Western Sequence

Age	*Formation*	*Lithology*	*Thickness*
	Clarendon Springs dolomite	Massive, gray dolomite.	200′ ±
	Danby formation	Interbedded gray quartzite and dolomite.	400–600′
	Winooski dolomite	Pink dolomite.	800′ ±
	Monkton quartzite	Red quartzite separated by beds of pink dolomite.	400–500′
Cambrian	Dunham dolomite	Buff to white dolomite.	1000′ ±
	Cheshire quartzite	Massive white quartzite.	1000′ ±
	Mendon formation	Conglomerate, albite-quartz-biotite-muscovite schist, albite-quartz-chlorite-muscovite schist. Forestdale member in middle: buff to white dolomite. Moosalamoo member at top: graphitic quartz-muscovite schist.	800–1800′
		UNCONFORMITY	
Pre-Cambrian	Mount Holly complex	Gneiss, schist, quartzite.	?

geographic name followed by a descriptive lithologic name, but if no single lithologic term is appropriate, the term formation is used.

The rocks of the Rochester-East Middlebury area are divided into an eastern sequence and a western sequence. The formations established within each sequence are shown in Tables 1 and 2.

Pre-Cambrian Rocks

MOUNT HOLLY COMPLEX

Name: Whittle (1894) used the term Mount Holly series for a heterogeneous group of rocks exposed in the core of the Green Mountain anticlinorium in the townships of Mount Holly and Shrewsbury (Figure 1). It is proposed that the name Mount Holly series be emended to Mount Holly complex. Present stratigraphic usage defines series as the rocks deposited during an epoch. The rocks exposed in the core of the

TABLE 2

Eastern Sequence

Age	*Formation*	*Lithology*	*Thickness*
	Stowe formation	Pale-green quartz-muscovite-chlorite schist. Brackett member: dark-green albite-epidote-carbonate-chlorite schist.	2000′ ±
	Ottauquechee formation	Graphitic quartz-muscovite schist and quartzite, dark-green albite-epidote-carbonate-chlorite schist.	1800–2500′
Cambrian	Pinney Hollow formation	Pale-green quartz-muscovite-chlorite schist and albite-quartz-muscovite-chlorite schist. Hancock member: dark-green albite-epidote-carbonate-chlorite schist.	1000–1500′
	Granville formation	Graphitic quartz-muscovite schist.	400′ ±
	Monastery formation	Pale-green albite-quartz-muscovite-chlorite schist. Tyson member: conglomerate and albite-quartz-biotite-muscovite schist. Battell member: graphitic quartz-muscovite schist.	1700–2100′
		UNCONFORMITY	
Pre-Cambrian	Mount Holly complex	Gneiss, schist, quartzite.	?

Green Mountain anticlinorium are contorted and recrystallized and contain no known fossils. Therefore these rocks are not amenable to subdivision into time-rock units representing such a short span of time. The term complex, on the other hand, does not have any time connotation. The "Committee on Stratigraphic Nomenclature" (Ashley, 1933, p. 445) prescribes the following usage for the term complex: "Where a large mass is composed of diverse rocks of any class, or classes, and is characterized by highly complicated structure, the term complex may be used."

In the type locality, this complex contains fine-grained biotite gneiss, augen gneiss, garnet-quartz-muscovite schist, rusty muscovite schist, amphibolite, quartzite, and saccharoidal marble. The Mount Holly complex, as used here, includes units lithologically similar to those of the original Mount Holly series

Distribution: In the northern part of the Rochester-East Middlebury area the Mount Holly complex occupies the area between the western escarpment of the Green Mountains and the ridge formed by Battell Mountain and Burnt Hill (Plate 1). To the south the area of Pre-Cambrian narrows into three belts that trend roughly north-south. The easternmost belt extends southward from Middlebury Gap through Monastery Gap and to Bingo Camp where it extends beyond the limits of the area. The central belt is exposed on Romance Mountain and in Sucker Brook. The westernmost belt, between Mount Moosalamoo and Sugar Hill, probably ends in the vicinity of Goshen, although its contacts are largely inferred because of poor exposure.

Description: Granulites and gneisses of various kinds underlie much of the area mapped as Mount Holly complex. Typical modes are given in Table 3. Of these, quartz-feldspar, biotite granulite, finely banded quartz-feldspar-biotite-muscovite gneiss, and porphyroblastic microcline-quartz-biotite-muscovite gneiss are most abundant.

Good exposures of quartz-feldspar-biotite granulite crop out in the Middlebury River about one mile west of Ripton. The granulite is gray to green on a weathered surface but is buff on a fresh surface. Feldspar makes up the largest part of the rock, about one-third of which is oligoclase with zonally arranged inclusions of epidote and sericite, and the other two-thirds of which is microcline. Tourmaline-bearing pegmatites are commonly intrusive into these granulites, and in some localities, the exposures are more than half pegmatite

Gray, finely banded quartz-feldspar-biotite-muscovite gneiss is common along the west slopes of Burnt Hill and Kirby Peak. The quartz and feldspar are in rounded to augen-shaped grains distributed in layers 1 to 2 millimeters thick. These layers are separated by thin septa of muscovite, biotite and chlorite. In these, muscovite always predominates, and biotite and chlorite may be in any proportion. Locally these rocks grade into quartz-feldspar-muscovite gneisses.

Outcrops of porphyroblastic microcline-quartz-biotite gneiss are exposed at Ripton. These rocks are coarse grained, with microcline porphyroblasts ranging from 0.8 mm. to 3.5 mm. in longest dimension. The porphyroblasts are set in a matrix of quartz, altered oligoclase, and unoriented biotite and muscovite. Some of these gneisses take on an augen texture by a clustering of the microcline into pod-shaped areas.

Schists in the Mount Holly complex are dominantly quartz-biotite-muscovite schist and porphyroblastic albite-quartz-biotite-chlorite-muscovite schist. Modes of these schists are given in Table 4.

Exposures of quartz-biotite-muscovite schist crop out about one mile east of Romance Mountain. These schists are gray and have a well developed schistosity. Quartz is segregated into layers one millimeter thick separated by biotite and muscovite. Calcite grains are associated with the quartz layers.

Albite-quartz-biotite-chlorite-muscovite schist is exposed in West Branch near Bingo Camp. The color is green to gray, and they contain porphyroblasts of albite about one millimeter in size. Where garnets are present, the chlorite has recrystallized to biotite. In some localities, however, both the garnet and biotite are partially or entirely altered to chlorite. Most of the garnets, which range in size from 8 mm. to 1 cm., are broken and crushed.

Bright-green actinolite-chlorite schists are interbedded throughout the Mount Holly complex (Plate 4, Figure 1). The actinolite schists have varying amounts of chlorite, epidote and zoisite. Oligoclase is usually the dominant feldspar.

Locally amphibolites are found within the Pre-Cambrian. These are composed essentially of needle-shaped hornblende and small amounts of oligoclase. Epidote is a minor constituent.

Thin beds of quartzite and marble are found within the Mount Holly complex. The quartzites are glassy and contain varying amounts of sericite. The marbles are relatively pure calcitic marbles, although a few tremolitic types are present.

Large pegmatites are confined to the Pre-Cambrian rocks (Plate 4, Figure 2). They are commonly course grained and are composed primarily of microcline, sodic plagioclase, quartz, muscovite, and biotite. Tourmaline is invariably associated with them.

Age: The Mount Holly complex is assigned to the Pre-Cambrian, because this highly metamorphosed and contorted complex lies unconformably beneath a less metamorphosed sequence of rocks containing Lower Cambrian fossils.

Correlation: The Mount Holly complex is similar to the rocks that occupy the core of the Green Mountains to the south. Ancient gneisses

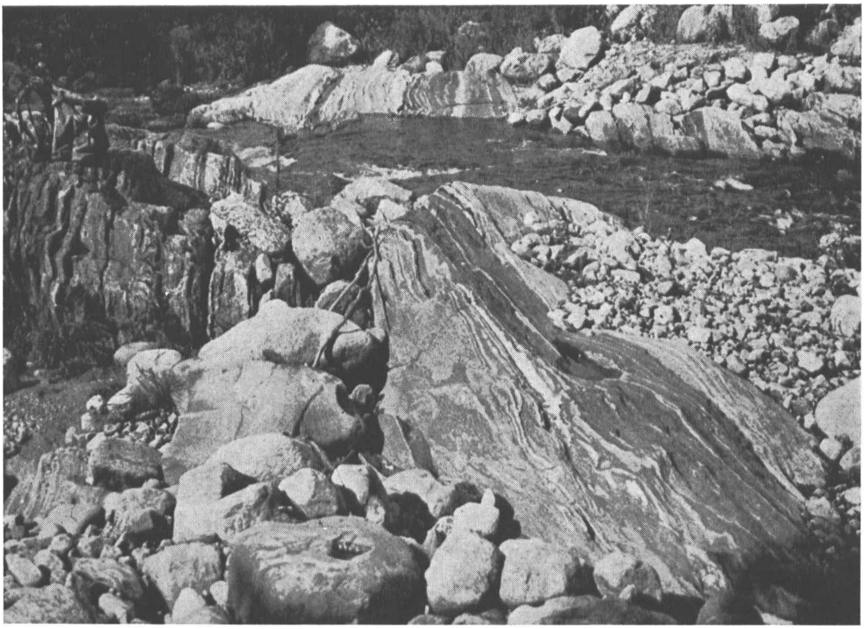


Figure 1. Banded amphibolite, Mount Holly complex, 0.7 mile S8°E from Ripton.

Figure 2. Quartz-feldspar granulite intruded by pegmatites (encircled in ink), Mount Holly complex, South Branch of the Middlebury River, Ripton.

PLATE 4

TABLE 3

Modes of the Mount Holly Complex

	1	*2*	*3*
Porphyroblasts			
Microcline	..	..	48
Groundmass or grains			
Quartz	23	39	14
Microcline	53	..	..
Oligoclase	22	8	3
Muscovite	1	13	10
Biotite	1	13	23
Epidote	..	25	tr
Clinozoisite	..	1	2
Apatite	..	..	tr
Zircon	tr	..	tr
Opaques	..	1	tr
Grain size in mm.			
Porphyroblasts			0.8–3.5cm.
Groundmass or grains	0.5–1.5	0.1–1.2	0.1–1.2
Texture	Gr*	G**	G

1. Quartz-feldspar-biotite granulite (one mile W. of Worth Mt.)
2. Banded quartz-feldspar-epidote-biotite-muscovite gneiss (0.25 mile S. of Middlebury Gap)
3. Porphyroblastic microcline-quartz-biotite-muscovite gneiss (one mile N. of Bread Loaf)

*—granulose

**—gneissose

and schists are exposed in the Berkshires of Massachusetts, in the Housatonic Highlands of Connecticut and New York, and in the Hudson Highlands of New York.

Cambrian Rocks of the Western Sequence

MENDON FORMATION

Name: The rocks immediately overlying the Mount Holly complex at Mendon (Figure 1) were named the Mendon series by Whittle (1894). Under present stratigraphic usage the application of the term series to these rocks should be abandoned, because there is no evidence that

TABLE 4

Modes of the Mount Holly Complex

	1	*2*	*3*	*4*
Porphyroblasts				
Oligoclase	..	44	..	..
Garnet	..	10	..	..
Groundmass or grains				
Quartz	22	4	..	..
Oligoclase	..	10	23	24
Muscovite	57	tr	..	..
Biotite	7	18	1	..
Chlorite	..	tr	tr	
Epidote	tr	..	17	4
Actinolite	..	..	57	..
Hornblende	..	..	..	66
Apatite	..	tr	..	1
Sphene	..	..	2	..
Calcite	13	14	..	..
Opaques	1	tr	..	5
Grain size in mm.				
Porphyroblasts		2.0		
Groundmass or grains	0.1–0.6	0.3–1.2	0.1–0.5	0.3–3.0
Texture	S*	G**	S	G

1. Quartz-biotite-muscovite schist (0.5 mile S45°E from Romance Mt.)
2. Garnet-oligoclase-biotite gneiss (0.5 mi. N52°W from Bingo Camp).
3. Actinolite-epidote-plagioclase schist (one mile N46°W from Middlebury Gap)
4. Amphibolite (1.2 miles S75°W from Middlebury Gap)

*—Schistose
**—Gneissose

they were deposited during the time interval which is appropriately spoken of as an epoch. Indeed, the Mendon series is delineated on the basis of lithology, and consequently the term series should be replaced by formation. Accordingly, it is proposed that the Mendon series be emended to Mendon formation.

Whittle (1894) described the Mendon "series" as including from the base upward a "conglomerate-schist" member, a "pebbly, crystalline limestone" member, and a "mica schist" member. The upper two members are characterized by lack of persistence. The "pebbly, crystalline limestone" will be referred to as the Forestdale member (Keith, 1932) of the Mendon formation, and the "mica schist" member will be referred to as the Moosalamoo member (Keith, 1932) of the Mendon formation.

Distribution: The Mendon formation crops out along the upper part of the western escarpment of the Green Mountains and underlies the summit and eastern slopes of Mount Moosalamoo (Plate 1). To the south the width of outcrop expands and includes the area between Chandler Ridge and Romance Mountain. A synclinal area of Mendon underlies the Hogback-Sugar Hill ridge as far north as the South Branch of the Middlebury River. Two small areas of Mendon occur about 1.4 miles N40°E from Ripton.

The Forestdale member, about 600 feet above the base of the Mendon formation, forms a thin belt in the northern part of the area. It is absent between the Middlebury River and the northern ridge of Mount Moosalamoo but is present as a narrow belt south of Mount Moosalamoo. A small culmination exposes it over a large area in the vicinity of Sucker Brook. The Forestdale member also crops out on the west slope of Hogback Mountain, but this belt pinches out to the north in the vicinity of Sugar Hill.

The Moosalamoo member is a lenticular body of schist in the upper part of the formation exposed between the Middlebury River and Silver Lake. It is absent both north and south of this area. Moosalamoo is also found on Sunset Hill, and it caps the Hogback-Sugar Hill ridge.

Description: Conglomerate, quartz-muscovite schist, porphyroblastic albite-quartz-biotite-muscovite schist, and quartz-chlorite-muscovite schist predominate in the Mendon formation. Modes of these lithologic types are given in Table 5.

Lenticular bodies of conglomerate are most abundant near the base of the formation (Plate 5, Figure 2). The coarse constituents range in size from pebbles 2 to 3 millimeters in diameter to blocks six feet in longest dimension. The pebbles and cobbles tend to be well rounded and have an average sphericity of 0.76. Locally the cobbles are deformed into flat ellipsoids that lie parallel to the foliation, and in some cases the cobbles are indented by adjacent cobbles. The cobbles are composed chiefly of quartzite with lesser amounts of vein quartz, pegmatite, gneiss, and slate. The matrix is commonly quartz-chlorite-

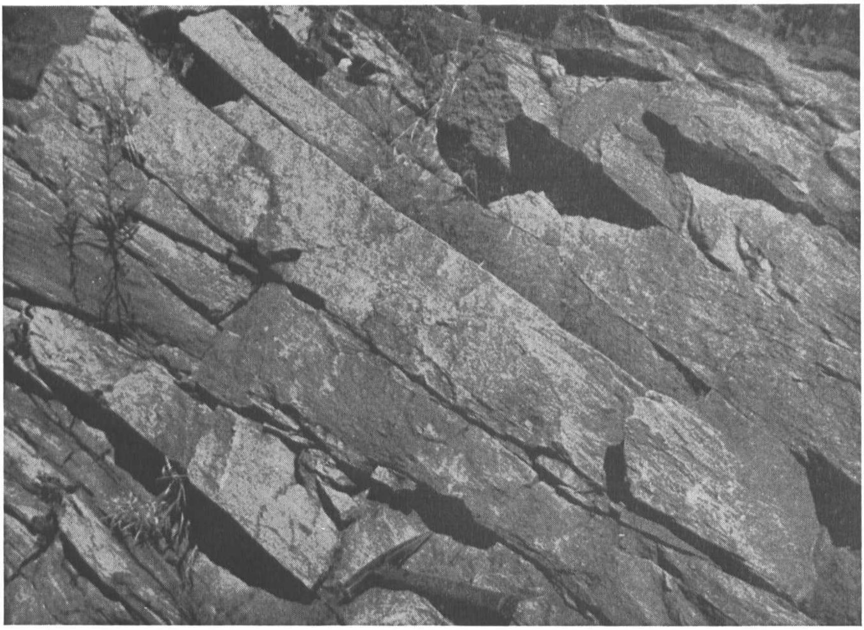


Figure 1. Arenaceous-quartz-muscovite schist, Mendon formation, elevation 1320 feet in Sucker Brook.

Figure 2. Schistose conglomerate, Mendon formation, 0.7 mile east of Mount Moosalamoo.

PLATE 5

muscovite schist or quartz-muscovite schist. Magnetite is locally abundant.

Arenaceous quartz-muscovite schist is associated with conglomerate (Plate 5, Figure 1). It is fine-grained with sporadic porphyroblasts of albite ranging from 0.2 to 0.4 millimeter in diameter. Ilmenite and magnetite are abundant and cause the gray color. Locally small amounts of graphite are present.

Porphyroblastic albite-quartz-biotite-muscovite schist has a wide distribution within the Mendon formation (Plate 6, Figures 1 and 2). Albite porphyroblasts reach two millimeters in size and locally small, blue quartz grains form indistinct layers. These quartz grains are thought to be relicts of sedimentary origin.

Pale-green quartz-chlorite-muscovite schist has a limited occurrence within the Mendon formation. It is interbedded with quartz-pebble conglomerate in the Middlebury River about one mile west of Ripton. The rock contains abundant quartz grains arranged in layers which are separated by muscovite and smaller amounts of chlorite. Magnetite is abundant.

The Forestdale member is a buff to rusty-weathering, white to gray dolomitic marble (Plate 7, Figure 2). A mode is given in Table 6. Characteristically, small blue quartz grains are disseminated throughout the rock, and in places lenses of gray quartz-muscovite schist are interbedded. Pyrite is commonly a secondary mineral. The dolomite is fine grained and shows little evidence of stratification.

Gray to black quartz-muscovite schist is the dominant rock in the Moosalamoo member (Plate 7, Figure 1), although lenses of pale-green quartz-chlorite-muscovite schist are sparingly present. A typical mode is given in Table 6. Locally, laminae of ankeritic quartzite ¼ to ½ inch thick indicate the bedding. Near the top of the member massive white quartzite beds are as much as five feet thick. The Moosalamoo grades into the overlying Cheshire quartzite. By an increase in the amount of quartz, the gray quartz-muscovite schist becomes a gray muscovitic quartzite that grades into a pure white quartzite. The complete transition takes place within a stratigraphic interval of 50 to 100 feet Small lenses of pebbles are located near the top of this member. The contact at the top of the Moosalamoo is placed at the imaginary surface where the quartzite beds and the graphitic quartz-muscovite schist beds each make up 50 per cent of the strata.

Thickness: West of Ripton the breadth of outcrop of the Mendon

Figure 1. Boundary of large detrital quartz grain and small rounded grain of microcline. Mendon formation. Large biotite crystals penetrate the quartz grain. Nicols crossed. X 50.

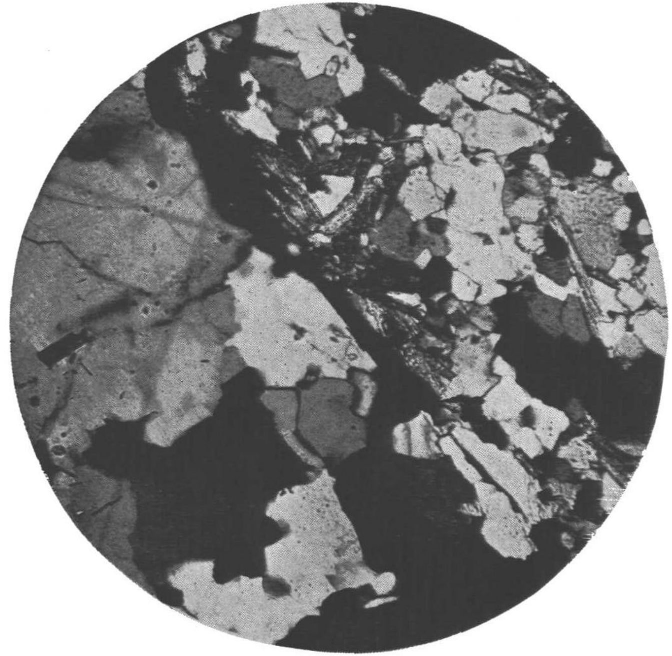


Figure 2. Boundary of large detrital quartz grain. Note sutured boundaries within quartz grain. Mendon formation. Nicols crossed X 50.

PLATE 6

Figure 1. Graphitic quartz-muscovite schist. Note plications and cleavage. Moosalamoo member of the Mendon formation. Nicols are not crossed. X 50.

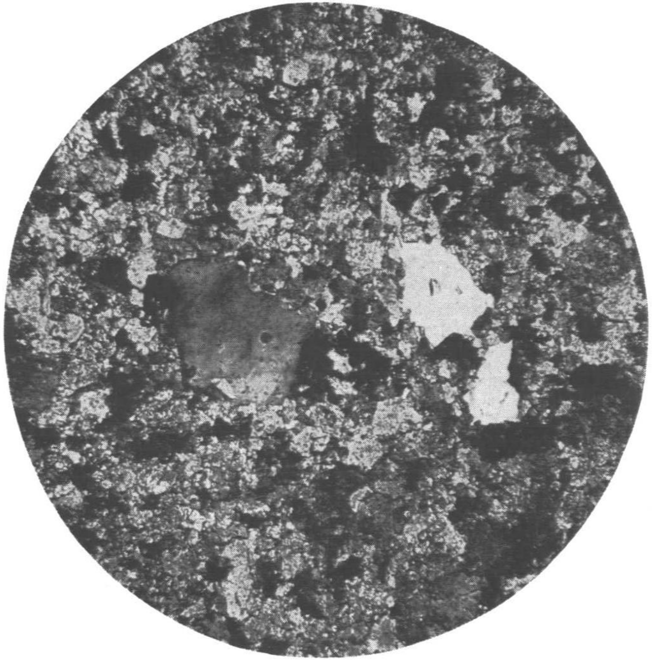


Figure 2. Dolomitic marble. Note quartz grain. Forestdale member of the Mendon formation. Nicols are crossed. X 50.

PLATE 7

TABLE 5

Modes of the Mendon Formation

	1	*2*	*3*	*4*
Porphyroblasts				
Albite	..	..	13	..
Groundmass or grains				
Quartz	60	51	46	46
Albite	16	5	5	..
Muscovite	18	40	24	42
Biotite	4	1	11	..
Chlorite	tr	2	..	10
Apatite	tr	tr	tr	
Epidote	..	..	..	tr
Zircon	tr	tr	tr	tr
Ankerite	..	..	tr	..
Opaques	2	1	1	2
Grain size in mm.				
Porphyroblasts			1.0–2.0	
Groundmass or grains	0.2–0.3	0.1–0.4	0.02–0.5	0.1–0.5
Palimpsest grains	2.0–7.0	1.0–2.0	0.8–1.5	
Texture	S*	S	S	S

1. Quartz-pebble conglomerate (0.7 mile W. of Romance Mountain)
2. Quartz-muscovite schist (0.7 mile W. of Romance Mountain)
3. Porphyroblastic albite-quartz-biotite-muscovite schist (1.7 miles W. of Hogback Mountain)
4. Pale-green quartz-chlorite-muscovite schist (1.7 miles E. of East Middlebury)

*—schistose

formation is 720 feet. The bedding here dips about 80°NE, and therefore the true thickness is about 705 feet. This thickness agrees with an estimated thickness of 640 feet 1.5 miles N30°E from East Middlebury. On the west slope of Mount Moosalamoo, its thickness is probably as great as 1800 feet.

The thickness of the Forestdale member ranges from about 115 feet to a feather edge. A section measured in the Middlebury River indicates a breadth of outcrop of 117 feet, but because the bedding dips 80 to 85° NE, the true thickness is about 114 feet. This member appears to be absent in places.

TABLE 6

Modes of the Mendon Formation

	5	*6*
Porphyroblasts		
Pyrite	tr	1
Groundmass or grains		
Quartz	4	29
Albite	1	3
Muscovite	..	63
Chlorite	..	tr
Dolomite	91	3
Calcite	4	..
Epidote	..	tr
Tourmaline	..	tr
Graphite	..	1
Grain size in mm.		
Porphyroblasts	2.0–3.0	3.0 –4.0
Groundmass	0.1–0.4	0.02–0.3
Texture	Gr*	S**

5. White to gray dolomitic marble (Forestdale member) (North shore of Silver Lake)
6. Gray quartz-muscovite schist (Moosalamoo member) (Elevation 980′, Sucker Brook)

*—granulose
**—schistose

The Moosalamoo member, like the Forestdale, is not persistent along the strike. Its thickness ranges from 900 feet to a feather edge, the maximum thickness occurring on Mount Moosalamoo. An exact thickness cannot be obtained because of minor duplication by folding.

Age: No fossils have been found within the Mendon formation in the Rochester-East Middlebury area. However, Walcott (1888) found fragments of *Olenellus* in a quartzite bed 100 feet above an unconformity with the Pre-Cambrian on Clarksburg Mountain in Massachusetts (Figure 1). This quartzite was correlated with the Cheshire (Whittle, 1894; Keith, 1932), and it was not until Prindle and Knopf (1932) resurveyed the locality that its true stratigraphic position was determined. They state:

"Above these quartzites (containing *Olenellus*) are micaceous phyllites, in places carbonaceous. . . . The phyllites on Clarksburg Mountain are overlain by pure heavy-bedded quartzite, which forms the lowest slope of the mountain. . . ."

The "micaceous phyllites" underlying the "massive-bedded quartzites" in some measure duplicate the stratigraphic relations between the Moosalamoo member of the Mendon formation and the Cheshire quartzite in the Rochester-East Middlebury area. If this interpretation is correct, beds of quartzite containing *Olenellus* underlie the equivalent of the Moosalamoo, and therefore must represent quartzites interbedded with the equivalent of the lower part of the Mendon formation. Consequently, it is suggested that the rocks lying above the Pre-Cambrian and below the Cheshire quartzite are of Lower Cambrian age.

Relations with the Mount Holly complex: The Mendon formation rests with apparent unconformity on the Mount Holly complex. However, at many localities this unconformity is difficult to delineate.

Regional mapping indicates the presence of an unconformity between the Mount Holly complex and the Mendon formation. The Mendon rests on several different lithologic units; in some places it rests on quartz-muscovite schist; in others it rests on gneisses or granulites.

The unconformable relations are exposed in a small saddle at the east foot of Mount Moosalamoo, 0.7 mile from the summit. At this locality a conglomerate containing flattened boulders of quartzite and gneiss rests on a porphyroblastic microcline-quartz-biotite gneiss. Although the actual contact is not exposed, the unconformable relations are suggested by the difference in metamorphic grade between the conglomerate and the underlying gneiss and by the fact that the conglomerate contains boulders derived from the porphyroblastic microcline-quartz-biotite gneiss.

This unconformity also is exposed immediately below the dam in Sucker Brook at an elevation of 1320 feet. Conglomerate and arenaceous quartz-muscovite schist rest on more highly metamorphosed quartz-microcline-biotite granulite and biotite schist. The conglomerate lies in tight synclinal troughs, and in one place the older rocks appear to be discordant with the younger rocks. This discordance, however, can be explained by either an unconformity or a small thrust fault. The exact position of the unconformity is difficult to determine in spite of the good exposures in the brook.

The difficulty in delineating the unconformity is in part due to structure and in part due to metamorphism. Both of these factors tend to obliterate the original erosional surface. The development of the Green Mountain anticlinorium has rotated the beds beneath this erosional surface into parallelism with the younger rocks, and metamorphism has, in some localities, produced similar mineral assemblages above and below the unconformity.

Cheshire Quartzite

Name: The type locality of the Cheshire quartzite is in Cheshire, Massachusetts (Emerson, 1892) (Figure 1). In that locality it is a pure-white, massive quartzite.

Distribution: Exposures of the Cheshire quartzite are confined to a belt that, for the most part, conforms to the lower slopes of the western escarpment of the Green Mountains (Plate 1). Cheshire crops out on Sunset Hill and in an isolated anticline just east of Mud Pond.

Description: The Cheshire is typically a massive white quartzite. Modes are given in Table 7. Some beds are as much as 10 feet thick, but more commonly the beds range from 1 to 3 feet in thickness. The quartzite in the thicker beds is massive and pure-white, whereas the thinner beds have small grains of ankerite scattered through them. Rarely, small quartz grains can be distinguished. Thin seams of graphitic quartz-muscovite schist are in places poorly rippled. The base of the Cheshire grades downward into the Moosalamoo member of the Mendon formation by an increase in the amount of feldspar, muscovite, and graphite. The top of the quartzite is transitional to the Dunham dolomite through a zone of interbeds. Cady (1945) and others have noted that the stratigraphically higher part of the Cheshire is composed of massive beds, whereas the lower part is characterized by thinner, more ankeritic beds. This distinction is not so apparent in the Rochester-East Middlebury area.

Thickness: Keith (1932) estimates a thickness of at least 1,000 feet for the Cheshire quartzite at Bristol. However, Cady (1945) indicates that it is about 850 feet thick at Bristol, whereas both Fowler (1949) and Thompson (1950) conclude that the Cheshire is about 400 feet thick near Rutland. These differences are primarily due to an actual thinning of the quartzite toward the east.

TABLE 7

Modes of the Cheshire Quartzite

	1	*2*
Grains		
Quartz	99	60
Microcline	..	3
Albite	..	1
Muscovite	tr	35
Biotite	..	tr
Tourmaline	..	tr
Opaques	1	1
Grain size in mm.		
Grains	0.2–0.4	0.1–0.3
Palimpsest grains		1.0–3.0
Texture	Gr*	S**

1. White quartzite (0.5 mile E. of East Middlebury)
2. White quartz-muscovite schist containing a few relict sand grains (0.1 mile E. of the Falls of Lana)

*—granulose
**—schistose

The thickness of the Cheshire is difficult to measure accurately in the Rochester-East Middlebury area. An estimate made on Sunset Hill indicates a thickness of 800 to 900 feet. The minimum thickness possible on the west slope of Mount Moosalamoo is 900 to 1100 feet.

Age: Fossils are scarce in the Cheshire and the writer found no organic remains. However, Walcott (1888) found fragments of *Hyolithes*, *Nothosoe*, and *Olenellus* near Bennington, Vermont, and specimens of *Hyolithes* were found on Sunset Hill in company with Seely (1910) (Figure 1). Previously, Hitchcock (1861) had reported *Lingula* from the Cheshire at Starksboro (Figure 1). These fossils fix its age as Lower Cambrian.

Relations with the Mendon formation: Foye (1919) and Keith (1934) described an unconformity just above the Falls of Lana in Sucker Brook. They postulated that the Moosalamoo member of the Mendon formation was cut out by an erosional unconformity and that the

Cheshire rested directly on the Forestdale. These relations were interpreted as indicating a major stratigraphic break.

The writer has found, however, that this unconformity is of minor extent, and careful mapping shows no angular relations (Figure 5.) It is true that here the Moosalamoo is absent and that the Cheshire rests directly on the Forestdale with slight erosional unconformity, but the same relationship occurs north of the Middlebury River. If the Moosalamoo were cut out by erosion, it seems reasonable that somewhere along the unconformity the thin Forestdale dolomite should also be cut out. However, at every point where the Moosalamoo is absent, the Forestdale is present. Thus the unconformity noted at the Falls of Lana is thought to be a local phenomenon, and the absence of the Moosalamoo is mainly due to the lack of deposition.

Dunham Dolomite

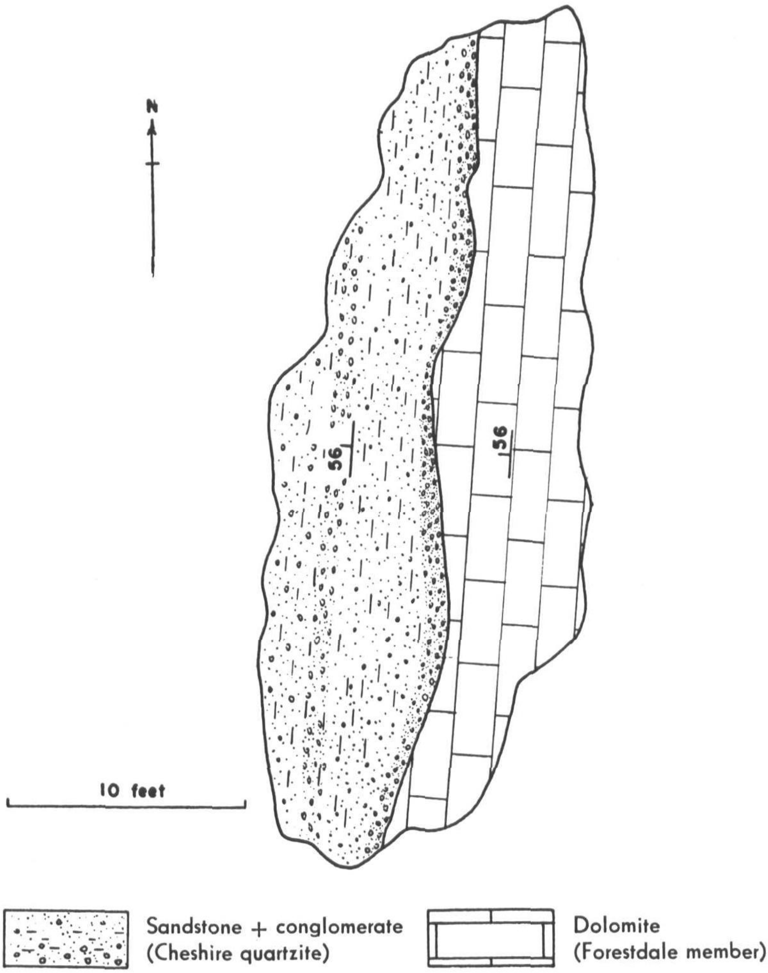
Name: The Dunham dolomite was named from exposures in Dunham Brook on Oak Hill, Quebec (Clark, 1934) (Figure 1). The dolomite "is almost everywhere a dark gray rock on the fresh surface, sometimes almost black. The surface zone of brown discoloration is seldom more than a millimeter thick. Rarely is it cream or buff coloured, and more rarely does it weather white."

Distribution: The distribution of the formations above the Cheshire is modified from Cady (1945) (Plate 1). A poorly exposed belt of Dunham runs the entire length of the East Middlebury quadrangle. A syncline underlying Lake Dunmore also exposes Dunham dolomite.

Description: As the writer studied only the lower beds of the Dunham, the lithologic description is taken from Fowler (1949).

"The Dunham . . . is a gray- and buff-weathering, compact, siliceous, gray dolomite containing irregularly distributed quartz grains. . . . The massive lower two-thirds of the Dunham is so irregularly jointed that true bedding planes are hard to distinguish. The upper third, which is similar in appearance to the Winooski dolomite, is lighter gray, cream-colored, and generally less sandy. . . . The upper Dunham just below the Monkton quartzite has a faint pink color and contains thin pink dolomitic quartzites."

Thickness: In the East Middlebury quadrangle, the Dunham is about 1,000 feet thick.



DISCONFORMITY EXPOSED IN SUCKER BROOK
Elevation 790′

Figure 5.

Age: Fossils identified as *Olenellus* (Resser and Howell, 1938) and *Ptychoparella* (Schuchert, 1937) were found near Highgate, Vermont (Figure 1). Wolff (1891) reported *Kurtorgina* and *Salterella* from the marble near Rutland, and Dale (1892) found *Hyolithes* near Chippenhook. These fossils suggest a Lower Cambrian age for the Dunham.

Monkton Quartzite

Name: Keith (1923) named the Monkton from exposures in the township of Monkton, Vermont (Figure 1). He describes the formation as consisting of reddish-brown, brick-red, and purple quartzite in layers from a few inches to three feet thick. A few intercalations of reddish or purplish shale and, toward the top, a few layers of gray to pink dolomite form a transition into the overlying Winooski. Cross-bedding and ripple-marks are common.

Distribution: The Monkton forms a narrow belt that extends from the northern to the southern border of the area in the west part of the East Middlebury sheet (Plate 1). The map pattern is modified from Cady (1945).

Description: The Monkton contains fine- to coarse-grained, pink to cream colored quartzite interbedded with various thicknesses of gray dolomite (Fowler, 1949). Cady (1945) describes the Monkton south of New Haven, Vermont as follows:

". . . . it becomes a schistose rock in which the red color is apparent only in certain zones 'protected' from deformation; . . . Nevertheless, the dolomites interbedded with the Monkton, although extremely squeezed, can be recognized by the pink or cream color on the fresh surface and the thin black to red siliceous partings which separate individual dolomite beds."

Thickness: Cady (1945) estimated the thickness of the Monkton at Middlebury as 400 to 500 feet.

Age: Lower Cambrian fossils including *Bonnia, Antagmus, Kurtorgina, Nisusia, Paterina, Acreteta, Helcionella, Hyolithes,* and *Scolithus* have been reported from the Monkton by Kindle and Tasch (1948).

Winooski Dolomite

Name: The name Winooski was first used by Hitchcock (1861) for all of the red and variegated dolomites of the "Red Sand Rock series" (Figure 1). Cady (1945) modified the name to apply only to those beds lying between the Monkton quartzite and the Danby formation.

Distribution: The Winooski dolomite forms a continuous north-south belt close to the western margin of the East Middlebury quadrangle (Plate 1). The map pattern is modified from Cady (1945).

Description: The Winooski dolomite immediately above the Monkton quartzite is pink on a fresh surface, but higher in the formation it is buff or gray (Cady, 1945). The beds range from four inches to a foot in thickness and are separated by thin siliceous shale partings.

Thickness: The thickness of the Winooski is difficult to measure because of lack of outcrop, but Cady (1945) estimates that it is about 800 feet thick at Brandon.

Age: No fossils have been reported from the Winooski dolomite, but because it interfingers with the Monkton, Cady (1945) believes it is of Lower Cambrian age. There is a possibility that the upper part of the formation represents Middle Cambrian time.

Danby Formation

Name: The names Danby and Wallingford were originally proposed by Keith (1932) to describe dolomitic units of rock in the townships of Danby and Wallingford (Figure 1), but were revised by Cady (1945), who made the Wallingford a member of the Danby.

Distribution: Cady (1945) shows the formation as a narrow north-south belt just west of Mount Pleasant and East Middlebury (Plate 1).

Description: As defined by Cady (1945) the Danby consists of resistant, differentially weathered beds of gray to white sandstones 1 to 2 feet thick, separated by buff to gray dolomite 10 to 12 feet thick. The lower sandstones are massive and pass upward into more dolomitic sandstones. The upper member, the Wallingford member, is described as a "dark iron-gray magnesium limestone" in which the sandstone beds are thin and highly dolomitic. The sandstone beds show abundant crossbedding throughout the formation.

Thickness: Cady (1945) estimates its thickness to be 800 feet at Leicester, and Rodgers (1937) gives a thickness for the equivalents of the Danby as 400 feet.

Age: Rodgers (1937) reports the occurrence of a Franconian fauna in the equivalent of the upper part of the Danby and suggests that the lower part is possibly Dresbachian. Cady (1945) accepted an Upper Cambrian age for the Danby formation.

Clarendon Springs Dolomite

Name: Keith (1932) applied the name Clarendon Springs to all the dolomitic beds lying between the arenaceous beds of the Danby and the non-dolomitic beds of the overlying Shelburne marble (Figure 1). However, Rodgers (1937) on faunal evidence indicates a stratigraphic break in the upper part of the dolomites. On this basis he suggested that the dolomitic beds lying above this break belong to higher strata. Cady (1945) found no such relationship in west-central Vermont and postulates that the "dolomite facies grade laterally into the limestone and marbles at several places and thus are contemporaneous with the limestone and marbles." The name is used here according to the definition of Cady.

Distribution: The Clarendon Springs dolomite as mapped by Cady (1945) is exposed in a narrow, north-south band in the western part of the East Middlebury quadrangle (Plate 1).

Description: This formation consists of "rather uniform, massive smooth-weathering gray dolomite characterized by numerous geodes and knots of white quartz. . . . On the fresh surface the dolomite is gray" (Cady, 1945). The upper part contains sandy beds and irregular nodules of flint.

Thickness: Cady (1945) measured 80 feet of dolomite at Middlebury and 120 feet at Brandon. A thickness in the vicinity of 200 feet is not improbable in the East Middlebury sheet.

Age: Rodgers (1937) inferred a Trempealeauan age for the Clarendon Springs and correlates it with all but the upper few feet of Division A of the Beekmantown of Brainerd and Seely (1890).

Cambrian Rocks of the Eastern Sequence

Monastery Formation

Name: The Monastery formation is named for exposures on the north-east slopes of Monastery Mountain (Figure 1). It includes the rocks lying between the Mount Holly complex and the overlying Granville formation. The basal conglomerate and schistose sandstone unit is called the Tyson member of the Monastery formation (Thompson, 1950). An intermediate black graphitic quartz-muscovite schist is named the Battell member from exposures in Hancock Branch on the southeast slopes of Battell Mountain.

Distribution: The Monastery formation is exposed over a wide belt that trends in a general north-south direction (Plate 1). It crops out between Monastery Mountain and Pine Gap and between the western slopes of Gillespie Mountain and the eastern slopes of Burnt Hill. The schist underlying Worth Mountain is also assigned to the Monastery.

The Tyson member is represented by a series of lenses at the base of the Monastery formation. The largest belt of exposures is in the valley of West Branch one mile east of Bingo Camp (Figure 2.) A smaller lense is exposed on Hat Crown and extends as a narrow belt northward toward Burnt Hill. Between these localities the Tyson is very thin or absent. A shallow synclinal area of Tyson is exposed along the ridge culminating in Worth Mountain.

The Battell member is well developed on the southeast slopes of Battell Mountain and in Hancock Branch. Elsewhere it is thin or absent.

Description: The Monastery formation consists dominantly of pale-green quartz-chlorite-muscovite schist (Plate 10, Figure 2). This schist contains fairly equidimensional and slightly strained quartz. Muscovite, commonly forms a large portion of the rock. Porphyroblasts of albite, 1 to 2 millimeters in diameter, and chlorite are more abundant in some beds than others. An analysis of chlorite from the Monastery formation is given in Table 11. The optics of this chlorite are (–) 2v small; $\alpha = 1.633$, $\beta = 1.638$, $\gamma = 1.638$; pleochroism X = colorless, Y = Z = green. Garnet porphyroblasts and sporadic grains of chloritoid are also present (Plate 9, Figure 1). The index of the garnet is 1.810. This index is similar to that of garnets found to be almandite with small proportions of grossularite and spessartite (Weiss, 1949). A few thin beds of schistose quartzite and arenaceous, porphyroblastic albite-quartz-muscovite schist are interbedded with the upper part of this formation.

Schistose conglomerate, silvery quartz-muscovite schist, gray porphyroblastic albite-quartz-biotite-muscovite schist and micaceous quartzite predominate in the Tyson member. The conglomerate is in lenses at the base of the member. Quartz is the chief constituent of the coarse fragments, which range from 2 mm. to 5 cm. in largest dimension. Locally the pebbles are flattened in the plane of the schistosity and in some instances are folded. The matrix is mainly fine-grained quartz, albite and muscovite. Ankerite and magnetite are megascopically abundant. Silvery quartz-muscovite schist is in places interbedded with the conglomerate. This schist contains much quartz, some of which is badly strained. Muscovite and a small amount of chloritoid and ankerite are the other constituents. The albite-quartz-biotite-muscovite schist contains porphyroblasts of albite, some of which contain inclusions arranged in spiral patterns, up to 1.5 millimeters in diameter. Other constituents are muscovite and small amounts of biotite which have developed layers 0.2 to 0.5 millimeter thick. In places, thin beds of impure, micaceous quartzite form the basal unit of the Tyson. The quartz grains are commonly intergrown with one another, although a few larger grains are probably relicts of original sedimentary grains. The rock is spotted with small ankerite grains, and some albite, muscovite and biotite are present (Plate 8, Figure 1).

The Battell member is black graphitic quartz-muscovite schist which in places contains porphyroblasts of albite 1 to 2 millimeters in diameter. A few lenses of dolomitic marble form the basal beds of this unit. These marbles are discontinuous and are never over 70 feet thick. They are white to gray on a fresh surface and contain isolated grains of quartz and more rarely biotite. Beds of porphyroblastic albite-quartz-chlorite-muscovite schist are also associated with this unit. The albite forms grains 2 to 3 millimeters in diameter and contains many inclusions of graphite.

Characteristic modes of the Monastery formation are given in Tables 8, 9, and 10.

Thickness: The thickness of the Monastery formation is difficult to measure, because of the gentle average dip of the beds (see page 85) and the complexity of minor folding. Estimates range from 1700 feet measured between Bingo Camp and Philadelphia Peak to 1400 feet measured from Battell Mountain to Texas Gap.

The Tyson member is believed to thin from a maximum thickness of 600 feet to a feather edge. It forms the basal portion of the Monastery formation and in many places is present as a bed ten to fifteen feet thick.

The Battell member probably never exceeds a thickness of 80 feet. It thins to a feather edge in the southern part of the Rochester-East Middlebury area. Stratigraphically, this member is about 750 feet above the base of the formation.

Relations with the Mount Holly complex: The relations between the

Figure 1. Feldspar-quartz-biotite-muscovite schist. Quartz grain is probably detrital. Tyson member of the Monastery formation. Crossed nicols. X 50.

Figute 2. Porphyroblastic garnet-oligoclase-quartz-chloritoid-chlorite-muscovite schist. Monastery formation. Nicols are not crossed. X 35.

PLATE 8

TABLE 8

Modes of the Monastery Formation

	1	*2*	*3*	*4*
Porphyroblasts				
Albite	..	..	25	..
Groundmass or grains				
Quartz	68	69	17	81
Albite	5	..	..	5
Muscovite	24	21	43	7
Biotite	..	..	13	4
Chlorite	..	..	2	2
Chloritoid	..	8	..	..
Ankerite	1	2	..	..
Apatite	tr	tr	tr	tr
Zircon	tr	tr	..	tr
Tourmaline	..	..	..	tr
Opaques	2	tr	tr	1
Grain size in mm.				
Porphyroblasts			1.0–2.0	
Groundmass or grains	0.2–0.4	0.2–1.0	0.2–0.6	0.2–1.0
Palimpsest grains	8.0–16.0			
Texture	S*	G**	S	G***

1. Schistose conglomerate (Tyson member) (0.3 mile W. of Bingo Camp)
2. Quartz-muscovite schist (Tyson member) (0.5 mile W. of Bingo Camp)
3. Porphyroblastic albite-quartz-biotite-muscovite schist (Tyson member) (0.6 mile N60°E from Kirby Peak)
4. Micaceous quartzite (Tyson member) (0.2 S30°E from Hat Crown)

*—schistose
**—gneissose
***—granulose

Mount Holly complex and the Cambrian rocks are not distinct on the east limb of the Green Mountain anticlinorium. In this region the difference in metamorphism is not a sound criterion for distinguishing the two systems, because rocks of both systems have reached metamorphic equilibrium under the same physical conditions.

Regional mapping indicates that the Monastery formation truncates several different lithologic types. In the vicinity of Bingo Camp the

TABLE 9

Modes of the Monastery Formation

	5	6	7	8
Porphyroblasts				
Albite	5	. .	35	. .
Groundmass or grains				
Quartz	25	49	31	79
Albite	. .	3	. .	9
Muscovite	54	44	31	7
Biotite	. .	tr	tr	4
Chlorite	13	. .	tr	. .
Ankerite	. .	tr	. .	1
Apatite	tr	. .	tr	tr
Epidote	. .	. .	tr	. .
Tourmaline	tr	. .	. .	tr
Zircon	. .	. .	. .	tr
Opaques	3	4	3	tr
Grain size in mm.				
Porphyroblasts	1.0–2.0		0.6–0.9	
Groundmass or grains . .	0.2–0.8	0.1–0.5	0.1–0.5	0.1–0.7
Texture	S*	S	S	Gr**

5. Quartz-chlorite-muscovite schist (1.5 miles S20°W of Philadelphia Peak)
6. Graphitic quartz-muscovite schist (Battell member) (0.5 mile S15°E of Battell Mountain)
7. Porphyroblastic albite-quartz-muscovite schist (1.2 miles E. of Monastery Mountain)
8. Micaceous quartzite (0.8 mile N17°W of Philadelphia Peak)

*—schistose
**—granulose

Monastery rests on porphyroblastic garnet-quartz-chlorite-muscovite schist, whereas in the vicinity of Middlebury Gap it rests on quartz-feldspar-muscovite gneiss.

The contact between the Mount Holly complex and the Monastery formation is exposed in West Branch 0.3 mile west of Bingo Camp. Gray quartz-albite-muscovite schist containing beds of quartz pebbles rests on microcline augen gneiss. The actual contact can be located only within a foot or two because the exposure is covered with moss and

TABLE 10

Modes of the Monastery Formation

	9	*10*
Porphyroblasts		
Albite	12	..
Garnet	1	2
Groundmass or grains		
Quartz	21	36
Muscovite	45	41
Chlorite	20	20
Chloritoid	..	tr
Apatite	tr	..
Epidote	..	tr
Tourmaline	tr	tr
Opaques	1	1
Grain size in mm.		
Porphyroblasts	0.8–1.0	0.5–2.3
Groundmass or grains	0.1–0.5	0.05–0.5
Texture	S*	S

9. Porphyroblastic garnet-albite-quartz-chlorite-muscovite schist (1 mile N25°E from Battell Mountain)
10. Porphyroblastic garnet-quartz-chlorite-muscovite schist (0.6 mile N80°E from Kirby Peak)

*—schistose

water. A second exposure occurs in a brook 0.3 mile N60°E from the summit of Worth Mountain. Here, quartz-muscovite schist lies on quartz-feldspar-biotite-muscovite gneiss. Although the difference in lithology is apparent, no structural discontinuity was observed.

Age: No fossils have been found in the Monastery formation. Consequently the determination of its age is based on its stratigraphic position relative to the Pre-Cambrian rocks and on its lithologic similarity with certain units belonging to the western sequence. The Monastery lies directly on the Pre-Cambrian rocks. The Tyson member is lithologically similar to strata in the lower part of the Mendon formation. Also the relation between the intermediate graphitic quartz-muscovite schist and the associated dolomitic marble of the Monastery

TABLE 11

*Chemical Analysis of Chlorite from the Monastery Formation**

SiO_2	27.78
TiO_2	0.32
Al_2O_3	24.67
Fe_2O_3	2.08
FeO	24.91
MnO	0.13
MgO	8.97
CaO	none
Na_2O ..	0.13
K_2O ..	
H_2O (total)	10.80
	99.79

Chlorite from albite-quartz-chlorite-muscovite schist (1.3 miles N67°W of Granville)

*—Analysis by F. A. Gonyer

is similar to the relation between the Moosalamoo member and the Forestdale member of the western sequence. If this correlation is correct, it suggests a Lower Cambrian age of the Monastery formation.

GRANVILLE FORMATION

Name: The name Granville is proposed for the graphitic quartz-muscovite schist lying above the Monastery formation. It receives its name from exposures in the township of Granville (Figure 1). Although an unfolded section of the Granville has not been found in the Rochester quadrangle, its lithology is well displayed in outcrops in the White River 0.3 mile N25°W of the village of Granville.

Distribution: The Granville formation is exposed in a narrow, sinuous belt which extends from the vicinity of Granville southward to Philadelphia Peak (Plate 1). Because of duplication by folding, it is exposed over a wide area between Granville and Texas Gap. Toward the south it crops out along the west slope of Gillespie Mountain and can be traced to Philadelphia Peak, which it completely encircles. From Philadelphia Peak the belt of Granville trends in a southeast direction out of the area. Small isolated patches are exposed on the east shoulder of Battell Mountain.

Description: The Granville formation is typically a gray to black

Figure 1. Porphyroblastic albite-quartz-chlorite-muscovite schist. Monastery formation. Nicols are not crossed. X 50.

Figure 2. Porphyroblastic albite-quartz-chlorite-muscovite schist. Monastery formation. Nicols crossed. X 50.

PLATE 9

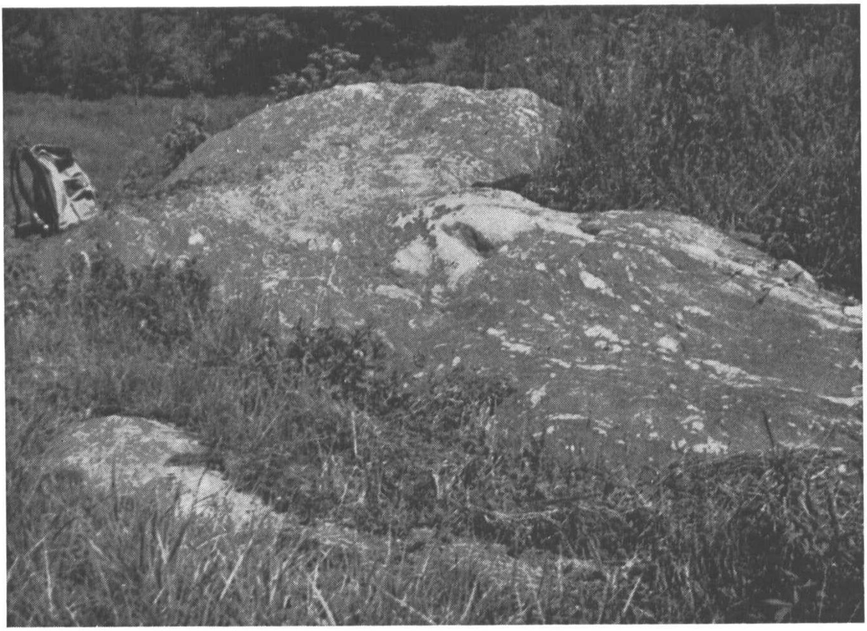


Figure 1. Porphyroblastic albite-quartz-chlorite-muscovite schist containing thin lenses of quartzite. Monastery formation, 3.2 miles S67°E from Bingo Camp.

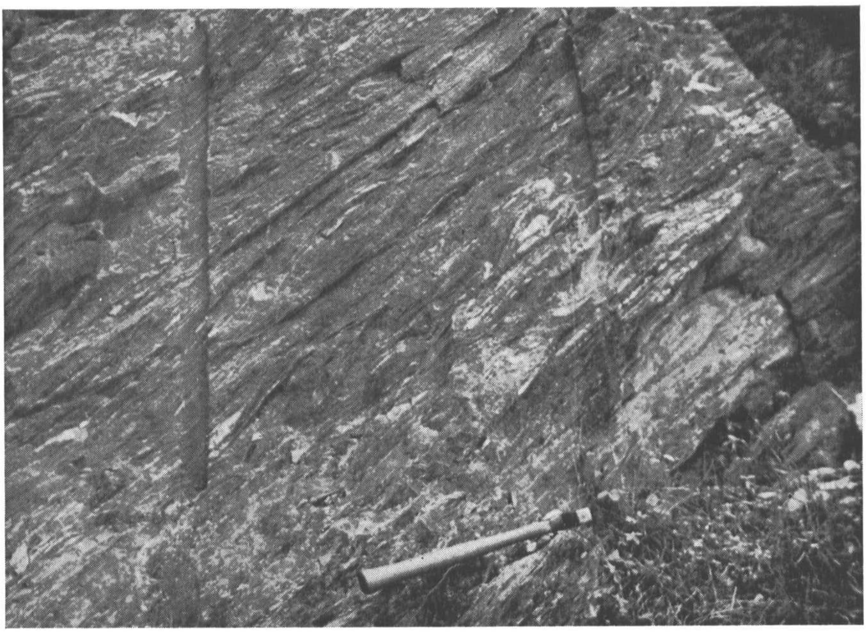


Figure 2. Porphyroblastic albite-quartz-chlorite-muscovite-schist. Note plications. Monastery formation, 3 miles west of Hancock.

PLATE 10

Figure 1. Graphitic quartz-muscovite schist. Note relation of plications to schistosity. Granville formation. Nicols are not crossed. X 50.

PLATE 11

TABLE 12

Modes of the Granville Formation

	1	*2*
Porphyroblasts		
Albite	13	12
Pyrite	tr	..
Groundmass		
Quartz	11	66
Muscovite	57	16
Chlorite	12	3
Ankerite	..	tr
Apatite	tr	tr
Opaques	7	3
Grain size in mm.		
Porphyroblasts	1.0–1.5	1.0 –1.2
Groundmass	0.1–0.6	0.08–0.4
Texture	S*	G**

1. Graphitic albite-quartz-chlorite-muscovite schist (0.6 mile W. of Texas Gap)
2. Graphitic albite-quartz-chlorite-muscovite gneiss (in White River 0.3 mile W. of West Hill School)

*—schistose
**—gneissose

graphitic albite-quartz-chlorite-muscovite schist (Plate 11, Figure 1). Modes are given in Table 12. It is rusty-weathering and contains sporadic porphyroblasts of pyrite. In places beds of blue-gray quartzite, usually only one to three inches thick, are interbedded with the graphitic albite-quartz-chlorite-muscovite schist. Beds of dark-green albite-epidote-calcite-chlorite schist 10 to 50 feet thick are part of this formation in the White River just west of Granville and similar beds crop out near Texas Gap. Buff-colored, dolomitic marble in lenses one foot thick and 5 to 6 feet long were observed within this formation three miles west of Rochester.

Thickness: The Granville formation is estimated to be 350 to 400 feet thick in the Rochester quadrangle. As nearly as it could be determined its thickness remains about constant throughout the area, although it may thicken slightly toward the northern border of the quadrangle.

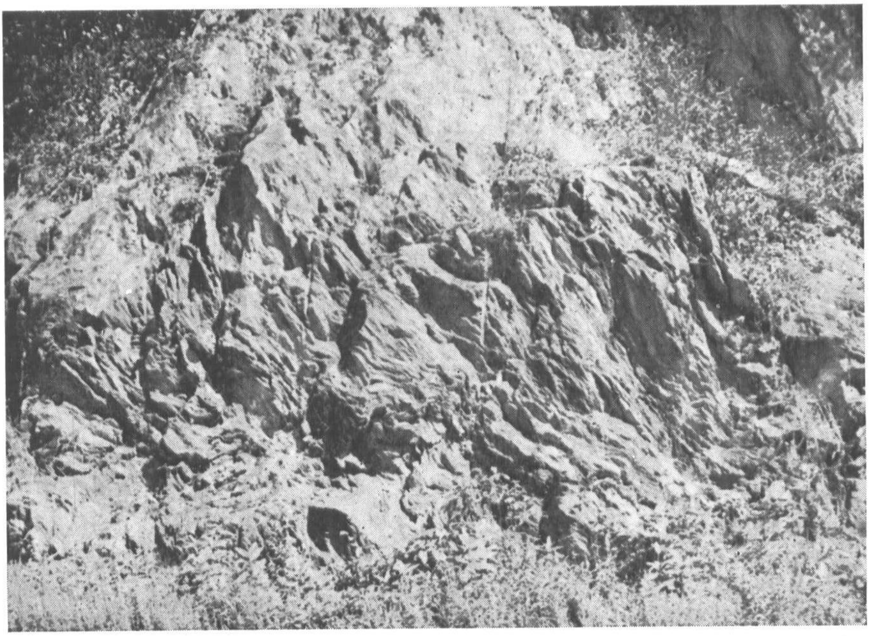


Figure 1. Quartz-chlorite-muscovite schist. Pinney Hollow formation, 1 mile south of Rochester.

Figure 2. Albite-epidote-calcite-chlorite schist showing compositional banding. Pinney Hollow formation, 0.4 mile N50°E from Hancock.

PLATE 12

Age: The age of the Granville is difficult to determine. No fossils have been found within this formation, and it bears little lithologic similarity to the rocks of the western sequence. However, the few lenses of dolomitic marble suggest that it may be the equivalent of a marble unit on the west flank of the Green Mountain anticlinorium. Tentatively it is correlated with at least part of the Dunham, and hence it is assigned to the Lower Cambrian.

Pinney Hollow Formation

Name: The Pinney Hollow was first named from exposures in Pinney Hollow, Vermont (Perry, 1928) (Figure 1). Perry defines these rocks as ". . . . pale green and thinly laminated, often with minute crenulations, in the general plane of the schistosity, passing into lines of slip cleavage. Distinct layers of quartz and micaceous minerals, chlorite and sericite, are visible and magnetite grains are often of sufficient size to be recognized."

Although much of the rock assigned to the Pinney Hollow formation in the Rochester-East Middlebury area is similar to that described by Perry (1928), a thick middle member consists of dark-green albite-epidote-calcite-chlorite schist. Because of the extent and the thickness, it is named the Hancock member of the Pinney Hollow formation and is named from exposures in Hancock Tunnel.

Distribution: The Pinney Hollow formation crops out in a wide band that follows the White River Valley (Plate 1). The western contact skirts the rim of Hancock Tunnel, encloses Gillespie Mountain and Childs Mountain, makes a deep indentation on the northeast slopes of Childs Mountain, and follows the White River northward through Granville. A small outlier caps Philadelphia Peak. The eastern contact is at the top of the steep slope just east of Hancock and Rochester.

Description: In the Rochester-East Middlebury area, the Pinney Hollow is typically pale-green quartz-chlorite-muscovite schist and dark-green albite-epidote-calcite-chlorite schist. The basal unit consists of pale-green quartz-chlorite-muscovite schist, some beds of which contain albite (Plate 12, Figure 1). This lower unit is overlain by the Hancock member consisting of dark-green albite-epidote-calcite-chlorite schist which thins toward the northern boundary of the area (Plate 12, Figure 2). Thin beds of quartzite and lenses of dolomitic marble are intercalated with these rocks. The upper unit of this formation is

TABLE 13

Modes of the Pinney Hollow Formation

	1	*2*	*3*	*4*	*5*
Porphyroblasts					
Albite . . .	5	11	20	. .	26
Chloritoid .	. .	. .	. .	20	. .
Groundmass					
Quartz . .	40	48	32	37	. .
Muscovite .	40	29	32	35	6
Chlorite . .	14	9	13	3	27
Epidote . .	tr	2	tr	. .	29
Calcite . .	. .	. .	. .	. .	9
Apatite . .	tr	tr	tr	. .	. .
Tourmaline	. .	. .	tr	. .	. .
Zircon . .	. .	tr	tr	. .	. .
Opaques .	1	1	3	5	3
Grain size in mm.					
Porphyroblasts	1.0	1.0–1.5	1.0–1.6	0.6–1.3	0.7–1.0
Groundmass	0.1–0.8	0.2–0.5	0.1–0.3	0.1–0.3	0.1–0.6
Texture . . .	S*	S	S	G**	S

1. Quartz-chlorite-muscovite schist (1.3 miles S28°E of Lower Granville)
2. Porphyroblastic albite-quartz-chlorite-muscovite schist (0.2 mile N. of summit of Hancock Mountain)
3. Porphyroblastic albite-quartz-chlorite-muscovite schist (0.3 mile S45°W of summit of Hancock Mountain)
4. Porphyroblastic chloritoid-quartz-muscovite schist (0.4 mile E. of Philadelphia Peak)
5. Albite-epidote-calcite-chlorite gneiss (Hancock member) (1.6 miles S76°E of Philadelphia Peak)

*—schistose
**—gneissose

similar to the basal unit except that it contains a few thin beds of dark-green albite-epidote-calcite-chlorite schist essentially similar to the Hancock member. Modes of the Pinney Hollow are given in Tables 13 and 14.

The schist overlying the Granville formation is pale-green porphyroblastic albite-quartz-chlorite-muscovite schist (Plate 13, Figure 1). The albite porphyroblasts commonly average 1 to 2 millimeters in size,

TABLE 14

Modes of the Pinney Hollow Formation

	6	*7*	*8*	*9*
Porphyroblasts				
Albite	65	38	11	9
Magnetite	..	..	..	4
Groundmass				
Quartz	..	..	13	..
Muscovite	tr	9	..	tr
Chlorite	18	15	48	72
Biotite	tr	..	..	..
Epidote	13	29	3	9
Calcite	..	8	23	5
Opaques	4	1	2	1
Grain size in mm.				
Porphyroblasts	0.5–0.7	0.5 –0.6	0.4 –0.6	0.5 –2.5
Groundmass	0.1–0.4	0.03–0.4	0.05–0.3	0.06–0.4
Texture	G*	G	S**	S

6. Porphyroblastic albite-epidote-chlorite gneiss (Hancock member) (Hancock Branch, village of Hancock)
7. Porphyroblastic albite-epidote-calcite-chlorite gneiss (Hancock member) (2 miles N10°E from village of Hancock)
8. Porphyroblastic albite-quartz-epidote-calcite-chlorite schist (Hancock member) (1 mile N50°E from village of Granville)
9. Porphyroblastic albite-epidote-calcite-chlorite schist (Hancock member) (0.8 mile N63°E from village of Granville)

*—gneissose
**—schistose

and many show evidence of rotation. In some beds, chloritoid is associated with these schists (Plate 14, Figure 1). Muscovite from these chloritoid-bearing schists was chemically analyzed. Its analysis appears in Table 15. The optical properties of this muscovite are (–) $2v = 38°$, $\alpha = 1.561$, $\beta = 1.600$, and $\gamma = 1.601$. Locally magnetite porphyroblasts up to six millimeters in size are present. Quartz lenses are ubiquitous and tend to follow the contortions of the foliation.

This basal schist is overlain by the Hancock member. It is a dark-green albite-epidote-calcite-chlorite schist characterized by an abundance of chlorite and small megascopic crystals of epidote. Locally calcite is

Figure 1. Porphyroblastic albite-quartz-chlorite-muscovite schist. The albite porphyroblast shows evidence of rotation. Pinney Hollow formation. Nicols are crossed. X 50.

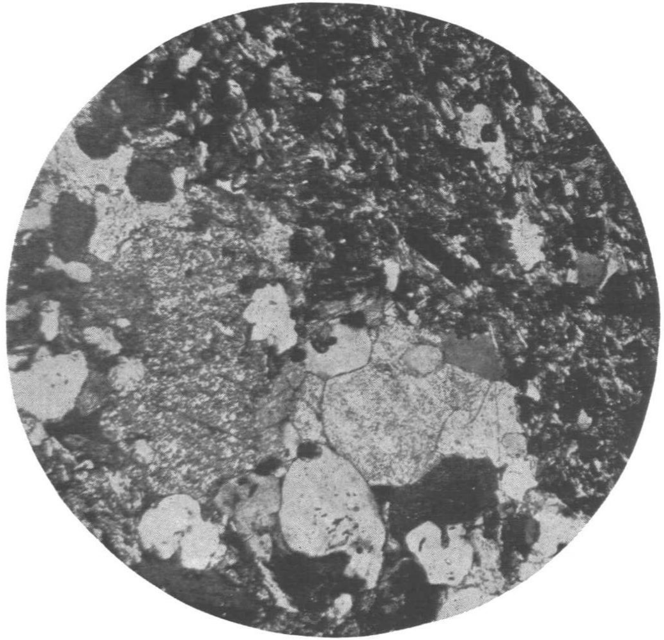


Figure 2. Albite-epidote-calcite-chlorite schist. Large grains are calcite. Pinney Hollow formation. Nicols crossed. X 50.

PLATE 13

TABLE 15

*Chemical Analyses of Minerals from the Pinney Hollow Formation**

	1	*2*
SiO_2	46.10	26.07
TiO_2	0.44	0.19
Al_2O_3	35.00	22.60
Fe_2O_3	2.36	0.72
FeO	0.70	24.79
MnO	none	0.12
MgO	0.18	13.84
CaO	none	0.06
Na_2O	1.48	none
K_2O	8.42	none
H_2O (total)	5.02	11.43
F	0.13	none
	99.83	99.82

1. Muscovite from quartz-chloritoid-muscovite schist (0.4 mile E. of Philadelphia Peak)
2. Chlorite from albite-epidote-calcite-chlorite schist (0.1 mile W. of village of Hancock)

*—Analyses made by F. A. Gonyer

abundant (Plate 13, Figure 2). This member is generally stratified, with the chlorite segregated into bands 1 to 2 centimeters thick separated by equally thick layers rich in albite and calcite. Chlorite from this member near the village of Hancock was analyzed. Its analysis appears in Table 15. The optical properties of this chlorite are (–) 2V small, $\alpha = 1.625$, $\beta = 1.628$, $\gamma = 1.628$, pleochroism X = colorless, Y = Z = green. Epidote knots 6 to 8 inches in largest dimension are locally present along the stratification planes. West of Hancock thin beds of white quartz-tremolite-calcite granulite are interstratified with the dark-green albite-epidote-calcite-chlorite schist, and locally thin discontinuous beds of buff-weathering, dolomitic marble are present.

The schist above the Hancock member is similar to that of the basal unit. These rocks consist of pale-green quartz-chlorite-muscovite schist with a few interbeds of dark-green albite-epidote-calcite-chlorite schist. In places the beds contain abundant albite porphyroblasts. Toward the top of the formation the chlorite becomes less abundant and the number of albite porphyroblasts increases. These upper beds contain local layers

Figure 1. Quartz-chlorite-muscovite schist. Compositional banding is present. Stowe formation. Nicols are not crossed. X 50.

Figure 2. Porphyroblastic chloritoid-quartz-muscovite schist. Pinney Hollow formation. Nicols are not crossed. X 50.

PLATE 14

of black to gray quartz-muscovite schist and locally thin muscovitic quartzites. The top of the Pinney Hollow is gradational into the Ottauquechee formation through a stratigraphic interval of about 50 feet.

Thickness: Estimates of the thickness of the Pinney Hollow are difficult to make because of the intense crumpling. Crude measurements indicate that the Pinney Hollow is about 1500 feet thick at the southern margin of the area and thins to about 1000 feet at the northern border of the area. The lower unit consisting of pale-green albite-quartz-chlorite-muscovite schist is from 400 to 600 feet thick. The Hancock member is est mated to be about 500 feet thick in the vicinity of Rochester and Hancock, but it thins toward the northern boundary of the quadrangle to about 200 feet. The upper pale-green quartz-chlorite-muscovite schist has a thickness which ranges from 400 to 600 feet.

Age: No fossils have been found in the Pinney Hollow formation. But because of its stratigraphic position between the equivalents of the Cheshire quartzite and the Monkton formation, it is probably of Lower Cambrian age.

Ottauquechee Formation

Name: The Ottauquechee formation was first described in the Ottauquechee River Valley "about half way between West Bridgewater and Bridgewater Corners" (Perry, 1928) (Figure 1). The formation is characterized by massive, slate-gray quartzite and black schist. Rocks of similar lithology are assigned to the Ottauquechee in the Rochester-East Middlebury area.

Distribution: The Ottauquechee formation within this area crops out in a belt one and one-half miles wide between the White River and Braintree Mountain (Plate 1).

Description: Typically the Ottauquechee is a gray to black graphitic quartz-muscovite schist (Plate 15, Figure 1 and Plate 16, Figure 1). Albite porphyroblasts 1 to 2 millimeters in diameter along with small amounts of chlorite are commonly present. An analysis for carbon indicates that the rock contains 3 to 4 percent by weight. This carbon is present as inclusions of graphite in the muscovite. Beds 5 to 40 feet thick of dark-green albite-epidote-calcite-chlorite schist similar to those in the Pinney Hollow formation and pale-green albite-quartz-chlorite-muscovite schist, some beds of which contain crossed biotite, are inter-

Figure 1. Interbedded blue-gray quartzite and graphitic quartz-muscovite schist. Ottauquechee formation, 1.6 miles S79°E from Lower Granville.

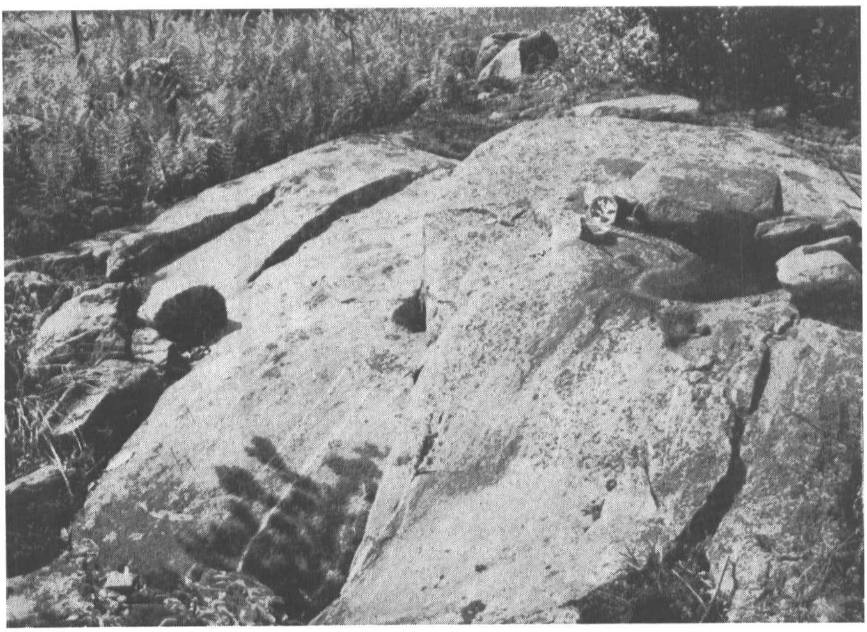


Figure 2. Massive, blue-gray quartzite. Ottauquechee formation, 1.4 miles N80°E from Lower Granville.

PLATE 15

Figure 1. Graphitic quartz-muscovite schist. Ottauquechee formation. Note boudinage in small quartz stringer. Nicols are not crossed. X 50.

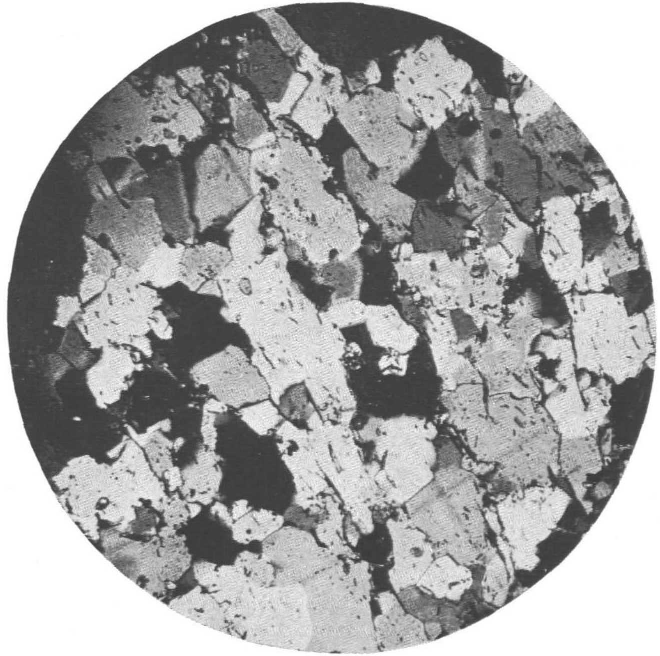


Figure 2. Graphitic quartzite. Ottauquechee formation. Nicols crossed. X 50.

PLATE 16

TABLE 16

Modes of the Ottauquechee Formation

	1	2	3	4	5	6
Porphyroblasts						
Albite	..	10	..	..	..	..
Groundmass or grains						
Quartz	42	24	..	5	47	91
Albite	..	..	27	12	21	..
Muscovite	50	40	..	..	16	4
Biotite	..	..	tr	..	tr	..
Chlorite	..	20	36	49	14	..
Epidote	..	..	35	30	tr	..
Ankerite	1	..	..	..	..	..
Apatite	..	tr	..	..	..	..
Calcite	..	tr	tr	1	..	tr
Tourmaline . . .	..	tr	..	..	..	..
Zircon	..	tr	..	..	..	..
Opaques	7	6	2	3	2	5
Grain size in mm.						
Porphyroblasts . .		1.0–2.0				
Groundmass or grains	0.1–1.0	0.1–0.5	0.2–0.8	0.1–0.7	0.1–1.2	0.1–0.4
Texture	S*	S	S	S	S	Gr**

1. Graphitic quartz-muscovite schist (2.3 miles S84°E of Lower Granville)
2. Porphyroblastic albite-quartz-chlorite-muscovite schist (2.3 miles S83°E of Hancock)
3. Dark-green albite-epidote-calcite-chlorite schist (0.8 mile S61°E of Granville)
4. Albite-quartz-epidote-chlorite schist (1 mile N75°E from South Hollow School)
5. Gray-green, banded albite-quartz-chlorite-muscovite schist (1.5 miles N75°W of Mount Cushman)
6. Blue-gray quartzite (1.8 miles S81°E of Lower Granville)

*—schistose

**—granulose

bedded with the graphitic quartz-muscovite schist. Near the base of the formation the rock consists of gray-green, banded albite-quartz-chlorite-muscovite schist with the quartz and albite segregated into layers 1 to 3 millimeters thick separated by muscovite, chlorite and more rarely biotite. Gray quartzite in beds and lenses from a fraction of an inch to 30 feet thick are also associated with the base of the formation (Plate

15, Figure 2 and Plate 16, Figure 2). Typical modes are given in Table 16.

Gray-green, banded albite-quartz-chlorite-muscovite schist interbedded with beds of gray muscovitic quartzite is gradational with the underlying Pinney Hollow formation. This contact is placed at a thin bed of dark-green albite-epidote-calcite-chlorite schist in the northern part of the map. In the southern part of the map, the contact is gradational through a stratigraphic interval of 50 feet. Higher in the sequence the muscovitic quartzite becomes more graphitic and thus takes on a blue-gray color. Above this transition the formation consists dominantly of graphitic quartz-muscovite schist, although local areas of pale-green albite-quartz-chlorite-muscovite schist are commonly encountered. Several discontinuous, thin beds of dark-green albite-epidote-calcite-chlorite schist occur throughout the Ottauquechee. The upper contact is gradational with the Stowe formation and consists of "small beds composed alternately of phyllite and chlorite schists" (Perry, 1928).

Thickness: The thickness of the Ottauquechee is difficult to determine because of minor folding and cleavage. Assuming an average dip of 30 to 40° its approximate thickness ranges from 1800 to 2500 feet. These thicknesses correspond reasonably well with estimates of Chang (1950) and Thompson (1950).

Age: No fossils have been found in the Ottauquechee, and subsequently its age is not definitely known. However, because the Ottauquechee lies between the eastern equivalents of the Cheshire quartzite and the Monkton formation, it is tentatively assigned to the Lower Cambrian.

Stowe Formation

Name: Cady (personal communication, 1948) has suggested the name Stowe for the schists immediately overlying the Ottauquechee formation in the Montpelier quadrangle (Figure 1). Similar schists crop out in the Rochester-East Middlebury area.

Distribution: The Stowe formation caps the summits of Braintree Mountain for its entire length and extends to the eastern border of the area (Plate 1). Its eastern contact is not within the Rochester quadrangle.

Description: The schists of the Stowe formation are in many respects similar to those of the Pinney Hollow (Plate 14, Figure 2). Pale-green albite-quartz-chlorite-muscovite schist, the most abundant rock, contains poorly developed porphyroblasts of albite. Locally the rock consists of bands, 1 to 2 millimeters thick, of quartz and albite separated by chlorite and muscovite. In most localities quartz is segregated into discontinuous bands and lenses 2 mm. to 1 cm. thick. These layers and lenses probably follow the bedding planes.

Dark-green albite-epidote-calcite-chlorite schist, called the Brackett member, is present along the eastern border of the area. This rock is similar in lithology to the Hancock member of the Pinney Hollow formation. It is named from exposures in Brackett Brook. Modes of the Stowe formation appear in Table 17.

Thickness: The thickness of the Stowe cannot be measured within the Rochester-East Middlebury area. Chang (1950) estimates an average thickness of 2000 feet in the Woodstock area, but it may be of the order of 3000 feet thick in the Rochester quadrangle.

Age: The exact age is unknown since no fossils have been found. The Stowe is overlain by schists that are in part lithologically similar to the Monkton quartzite (Thompson, personal communication, 1952), and consequently the Stowe is probably Lower Cambrian.

Correlation between the Eastern & Western Sequences

Two lithologically different Cambrian successions have been established in the Rochester-East Middlebury area. The eastern and western sequences are roughly equivalent in age but represent rocks of two contrasting environments of deposition. The eastern sequence consists dominantly of meta-sediments of detrital origin. These sediments are thought to have been derived from an eastern source because of the difficulty of transporting large quantities of detritus across an area composed of limestones. The western sequence consists mainly of marbles and quartzites which represent sediments of a tectonically stable area. The detrital portion of the western sequence was derived from the west (Cady, 1945).

Correlation between these two sequences is difficult. However, both sequences lie unconformably on the Pre-Cambrian complex. A stratigraphically higher unconformity occurs in the eastern sequence above the rocks exposed in the Rochester-East Middlebury area (Thompson, 1950; Chang, 1950). Cady (1950) reports poorly preserved corals

TABLE 17

Modes of the Stowe Formation

	1	*2*	*3*
Grains			
Quartz	31	13	22
Albite	11	25	9
Muscovite	34	14	..
Biotite	..	..	3
Chlorite	22	25	12
Epidote	..	..	52
Calcite	..	..	1
Apatite	tr	tr	..
Opaques	2	3	1
Grain size in mm.	0.1–0.3	0.1–0.7	0.1–0.8
Texture	S*	S	G**

1. Albite-quartz-chlorite-muscovite schist (2.5 miles S78°E of Lower Granville)
2. Albite-quartz-chlorite-muscovite schist (2.2 miles N71°E of Braintree Gap)
3. Albite-quartz-epidote-chlorite gneiss (Brackett member) (2.5 miles N59°E of Braintree Gap)

*—schistose
**—gneissose

tentatively identified as *Streptelasma* in rocks above this unconformity. If these corals are correctly identified, they suggest that the underlying unconformity is Lower to Middle Ordovician. In the western sequence a similar unconformity exists that separates the Orwell limestone from the Whipple marble and the Hortonville slate. This unconformity is believed to be of Middle Ordovician age (Cady, 1945). If both the Lower Cambrian and the Middle Ordovician unconformities are correctly identified on both flanks of the Green Mountain anticlinorium, the rocks bounded by them are equivalent.

A more detailed correlation between the smaller units of the eastern and western sequences is hazardous because lithologic similarity and stratigraphic position are the main criteria for determining equivalency. The Mendon formation is represented on the east flank of the Green Mountain anticlinorium by the Tyson member of the Monastery formation. The Moosalamoo member and the Forestdale member are lithologically similar to the lenses of marble and the graphitic quartz-muscovite schist unit of the Monastery formation. Thompson (personal communication, 1952) reports that the Monkton quartzite is at least in part lithologically similar to rocks stratigraphically above the Stowe formation in the Ludlow quadrangle. Other rock units of the eastern sequence bear little resemblance to those of the western sequence.

Figure 6 is a stratigraphic diagram indicating a tentative correlation of the various lithologic units. The Granville and the Ottauquechee formations are interpreted as lenses of graphitic quartz-muscovite schist that represent a sedimentary facies between a dominantly carbonate facies and a dominantly detrital facies. The Pinney Hollow and the Stowe are interpreted as tongues of a dominantly detrital facies which pinch out toward the west.

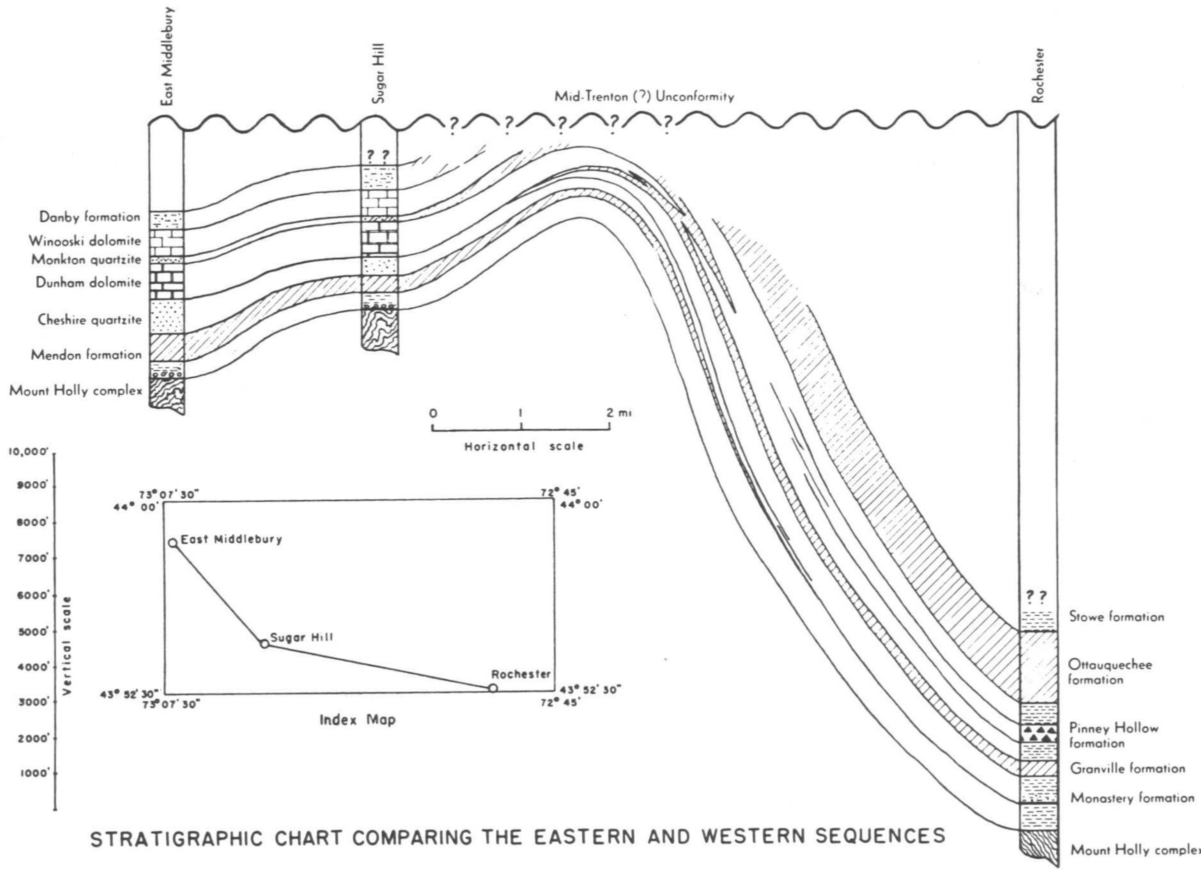
Both the East Middlebury and the Rochester sections are compiled from measurements made in the vicinity of these two localities. Data for the lower part of the Sugar Hill section are from the vicinity of Sugar Hill, but the data for the thicknesses above the Moosalamoo member are those of Fowler (1950) for the Pittsford area.

Unstratified Rocks

Ultramafic Rocks

Metamorphosed ultramafic rocks are common in the Rochester quadrangle. Most of these bodies are small pod-shaped masses a few tens of feet across and essentially conformable to the surrounding rocks. Within this area, they are located almost wholly in the Ottauquechee formation, but elsewhere in Vermont the ultramafic rocks occur in all of the Paleozoic strata beneath the Shaw Mountain formation of Middle Ordovician (?) age (Thompson, 1950).

In general, the ultramafic rocks have been completely altered to talc-magnesite rock (Figure 7), although a relatively large body two miles S65°E from Hancock is composed mainly of serpentine. In this body there is a zonal distribution of the rocks. The central portion is serpentine containing many small veinlets of carbonate, whereas the outer margin is composed of a thin, discontinuous talc-magnesite zone. Even within some of the smaller masses isolated remnants of serpentine are completely surrounded by talc and magnesite. In some cases, a discontinuous, narrow belt of gray chlorite schist occurs at the contact between



STRATIGRAPHIC CHART COMPARING THE EASTERN AND WESTERN SEQUENCES

Figure 6

the ultramafic body and the wall rock. This chlorite schist is called blackwall.

The process of serpentinization utilizes water, and because in this region all the ultramafic masses have been converted to serpentine, water must have been an ubiquitous constituent. On the other hand, the formation of talc and magnesite invo ved the addition of carbon dioxide. Because this process has gone on only to a limited extent, the supply of carbon dioxide must have been limited or the reaction must have been very slow.

The age of the ultramafic rocks have long been a problem. All of them appear to have been involved in folding (Chidester, personal communication, 1948), and none have been reported from above the Middle Trenton unconformity in central Vermont. Consequently, from the information at hand an Early Ordovician age is postulated.

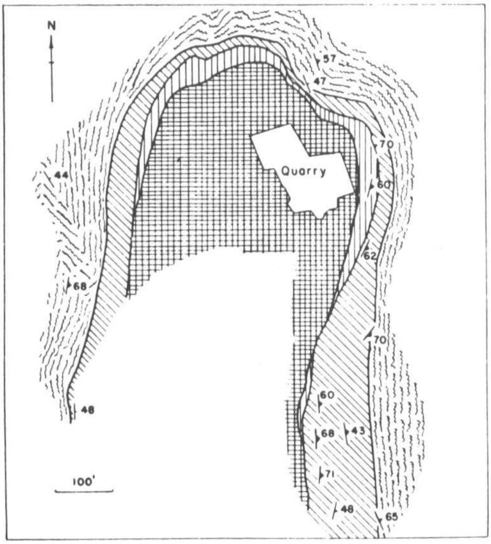
Additional information in Quebec supports an Ordovician age. Clark and Fairbairn (1936) report serpentine pebbles in a basal Silurian breccia near Lake Memphremagog. MacKay (1921), Tolman (1936), and Cooke (1937) found exposures of ultramafic rocks in the base of the Beauceville series of Middle Ordovician (?) age in the vicinity of the Thetford Mines.

Dikes

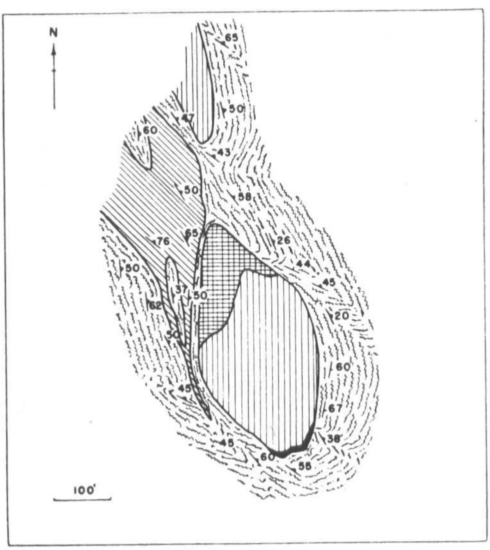
Numerous small dikes intrude the rocks of the Rochester-East Middlebury area. Two kersantite dikes were observed in the Pre-Cambrian gneisses. Both dikes have been metamorphosed although some indication of their cross-cutting relations have been preserved. These dikes are composed dominantly of biotite, epidote, calcite, and a little quartz. Their calculated chemical compositions indicate their original nature.

A swarm of fine-grained albite porphyry dikes cuts the Pinney Hollow formation between Rochester and Hancock These dikes trend approximately north-south, parallel to one of the two prominent joint sets, and are all cross-cutting and slightly metamorphosed. Albite is by far the most abundant mineral, forming both the fine-grained matrix and the phenocrysts. All of the feldspar contains small inclusions of sericite. Quartz is not abundant and is distinctly anhedral. Chlorite is present in large porphyroblasts and appears to be an alteration product of biotite.

Diabase dikes are the latest intrusives in the area. They are generally nearly vertical, following either the east-west or the north-south joint



Metamorphosed ultramafic body ½ mile S40°W from North Hollow School, Rochester, Vermont



Metamorphosed ultramafic body at Talc Ledge, Rochester, Vermont

Graphitic quartz-muscovite schist

Albite-epidote-carbonate - chlorite schist

Serpentine

Blackwall (chlorite)

Talc - carbonate

DETAILED MAPS OF ULTRAMAFIC BODIES

Figure 7

set. Most are fine grained, but under the microscope they exhibit an ophitic texture. Small euhedral grains of feldspar are randomly oriented in a matrix that has been altered to chlorite and calcite. Octahedra of magnetite are common.

STRUCTURAL GEOLOGY

General Statement

The Rochester-East Middlebury area is underlain by the Green Mountain anticlinorium, which in this vicinity trends slightly west of north (Plate 1). It consists of three parallel anticlines that are overturned toward the west and form a major axis-culmination in the latitude of Ripton. Reversals in plunge form a series of minor culminations and depressions.

The east flank of the anticlinorium has a gentle average dip that in detail is buckled by large folds. These folds are commonly overturned toward the west and in many cases plunge toward the south, although minor reversals in plunge occur.

A large synclinorium exposed west of the Green Mountain anticlinorium is called the Middlebury synclinorium (Cady, 1945). The east limb of this fold is exposed west of East Middlebury. The minor folds on this limb are commonly overturned toward the west, and some are isoclinal. The Lake Dunmore thrust (Cady, 1945) is a small break thrust developed on one of these isoclinal folds.

The major structures must be inferred from observations of lithology and minor structures obtained at scattered exposures. In most regions the minor structures greatly facilitate the interpretation of the major structures, but in regions of complex folding the minor structures may be amenable to more than one interpretation. Consequently, their use in deciphering the major structures is open to question. In these complexly folded areas, although minor structures are important, the most significant factor in solving the major structures is lithologic distribution. It follows that thorough geologic mapping is essential.

The major folds have many smaller structures superimposed on them. Drag folds, slip-cleavage, and various lineations are the most prominent minor structures and occur throughout the Rochester-East Middlebury area. The majority of these structures on the east flank of the Green Mountain anticlinorium are incongruous with the major structures, but minor structures on the west flank of the anticlinorium are congruous.

Bedding

Clues to Bedding

Bedding is not easy to observe in the schistose rocks of the Rochester-East Middlebury area. The most conspicuous indications of bedding are lenses of dolomitic marble interbedded with graphitic quartz-muscovite schist in the Granville formation. These dolomitic lenses are from 10 to 15 feet long and from 1 to 2 feet thick. These lenses, however, are not common and are usually not persistent along the strike.

Locally, thin, highly contorted beds of quartzite are fairly persistent within the limits of an outcrop. These small beds of quartzite range from 0.5 to 2 inches thick and commonly contain a limited amount of micaceous minerals and some iron oxides. These mineral impurities distinguish these beds from quartz veins and pods.

Bedding in many of the schistose rocks is shown by concentrations of albite porphyroblasts along certain layers. These alternations of layers containing different amounts of albite reflect a difference in composition in the original beds. The layers rich in albite are commonly only a fraction of an inch thick although locally they may be 2 or 3 inches thick.

Many of the dark-green albite-epidote-calcite-chlorite schists contain a high concentration of calcite. Some layers are more calcareous than the rest of the schist and upon weathering are etched out, leaving lines of pits that parallel the bedding.

Bedding in the conglomerates and arenaceous quartz-muscovite schists of the Mendon formation is relatively easy to discern. Textural variation is common and demarcates the stratification planes. In many places lenses of cobbles indicate the bedding planes.

Although stratification in the Cheshire quartzite is obvious, in places it is difficult to distinguish from jointing. However, true bedding planes usually have a pseudo-rippled surface that is never present on a joint plane. Commonly stratification planes are indicated by thin partings of graphitic quartz-muscovite schist that are present in even the most massively bedded phases of the quartzite.

Bedding Trend and Average Dip

The bedding is highly contorted in the schistose rocks, and although in detail the stratification is everywhere steeply inclined, the average

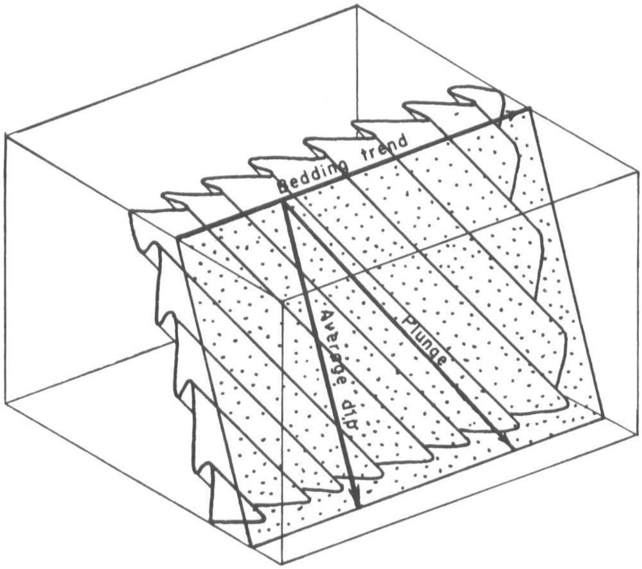


Figure 1. Bedding trend and average dip of plicated beds. (After White and Billings, 1951)

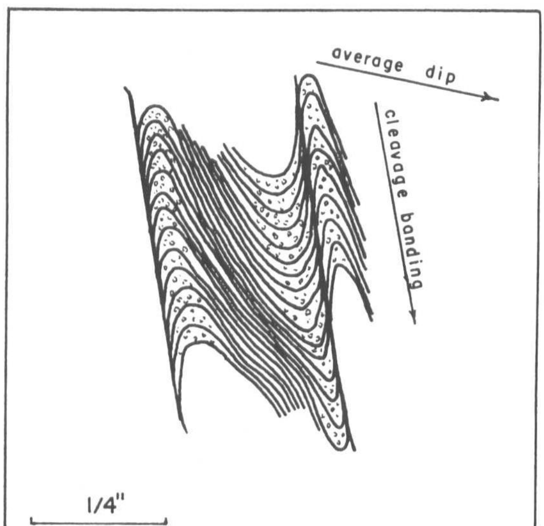


Figure 2. Cleavage banding produced by crystallization of albite and quartz in crests and troughs of plications.

PLATE 17

dip of a given bed is more gently inclined (Plate 17, Figure 1). Thus the bedding is moderately dipping with many small, sharply crested, nearly isoclinal folds superimposed on it.

Because in one outcrop the continuity of a bed is not easy to see, an observation on only one limb of a small fold would lead to erroneous conceptions. Instead, if measurements are taken on an imaginary plane tangential to the crests of a series of folds, a more correct idea is obtained regarding the true dip and strike. An observation taken in this way measures the bedding trend (strike) and average dip (dip).

Schistose rocks that contain an abundance of non-platy minerals usually display compositional bending that reflects the original bedding. In these rocks tracing the continuity of a bed is relatively easy, and the small-scale contortion of the bedding is readily observed. Consequently, the bedding trend and the average dip can be measured with a fair degree of accuracy.

In schistose rocks with an abundance of platy minerals, compositional banding is commonly absent, although locally thin beds of quartzite indicate the bedding. Where the beds of quartzite are also lacking, the determination of bedding trend and average dip is difficult.

East of the White River the bedding trends are slightly northwest and the average dips are 35 to 45° toward the east (Plate 2). From the White River valley westward to the easternmost anticline of the Green Mountain anticlinorium the bedding trends are dominantly north-south, and the average dips range from 35° to the east adjacent to the White River to 10 or 20° east as the crest of the anticline is approached. Local differences in average dips are common. Two and three-quarter miles east of Cobble Hill the average dip is about 20° west. Locally the bedding trend becomes east-west on the noses of folds.

Major Folds

Green Mountain Anticlinorium

Ripton anticline: The Ripton anticline is the westernmost of the major folds that forms the Green Mountain anticlinorium. This anticline has a culmination in the vicinity of the villages of Ripton from which the fold plunges north and south.

The west limb of the Ripton anticline is buckled by a series of sharp folds overturned toward the west. In these sharp folds the bedding alternately is overturned steeply to the east and dips gently toward the east. Although the beds at any one place may be steeply dipping or overturned, the average dips are from 60 to 30° to the west. The steepest part of the west flank is north of the Middlebury River (Plate 1). From the Middlebury River to the southern boundary of the area, the west limb becomes gently dipping to the west and highly contorted by small folds.

The east limb of the Ripton anticline is more difficult to interpret because schistosity is in many cases the only structure present in the Pre-Cambrian rocks. This schistosity generally dips steeply toward the east. However, the distribution of the Mendon formation clearly indicates an anticline in the vicinity of Ripton (Plate 1).

The structures in the Pre-Cambrian rocks of the core of the Ripton anticline are difficult to interpret. In many cases the only structure present is a crude schistosity oriented approximately parallel to the schistosity in the younger rocks. West of Ripton a few exposures showing bedding indicate that it dips steeply to the east and dragfolds indicate that it is overturned. The bedding, however, flattens out 0.3 mile west of Ripton where it dips 30° toward the east.

Bread Loaf Anticline: A second anticline lies east of and parallel to the Ripton anticline. This anticline is widest in the vicinity of Bread Loaf and consequently derives its name from this village. The Bread Loaf anticline is topographically expressed by the valley between the Sugar Hill-Hogback Mountain ridge and Worth Mountain. Farther south, however, this anticline forms the topographic ridge dominated by Romance Mountain.

The evidence for the Bread Loaf anticline is based mainly on the distribution of the Mendon and Monastery formations (Plate 1). The area of Mendon underlying the Sugar Hill-Hogback Mountain ridge and the two areas of Mendon 1.4 miles N40°E from Ripton demarcate the western flank of this anticline. The Monastery exposed on Worth Mountain indicates the position of its east flank, and the absence of Mendon in the latitude of Bread Loaf is proof of a culmination in this vicinity.

Structural details are meager because exposures are lacking in the low-lying area around Bread Loaf and in the valley of Goshen Brook. Throughout the Pre-Cambrian rocks of the core of this anticline the schistosity dips steeply east. This schistosity parallels the schistosity in the overlying Mendon formation. Bedding in an exposure 1.2 miles

N20°E from Bread Loaf dips steeply east (Plate 2). In a few exposures on the west slope of Romance Mountain the average dip of the bedding is 30 to 35° toward the east.

Burnt Hill Anticline: The Burnt Hill anticline is well exposed on Burnt Hill. This anticline lies east of and parallel to the Bread Loaf anticline. It underlies the Kirby Peak-Burnt Hill ridge, Middlebury Gap, Monastery Gap, and Bingo Camp where it extends beyond the limits of the area.

Structurally the Burnt Hill anticline has a dominant southerly plunge throughout its extent in the Rochester-East Middlebury area, and south of Middlebury Gap it forms only a low anticline. The bedding dips 10 to 30° east and west from the crest, but a half mile farther west, the bedding becomes vertical or is steeply overturned toward the west. These dips suggest that this anticline is recumbent. Moreover, minor folds that contort the general bedding trends of both limbs are commonly recumbent or sharply overturned toward the west (Plate 2).

Within the Pre-Cambrian where schistosity and minor folds are exposed in the same outcrop, the schistosity wraps around the noses of the folds. However, when no folding or bedding is evident, the schistosity commonly dips steeply toward the east. This schistosity is interpreted in the same manner as the schistosity in the younger rocks.

Sugar Hill Syncline: The Sugar Hill syncline that separates the Ripton and the Bread Loaf anticlines forms a topographic ridge including Sugar Hill and Hogback Mountain. It plunges northward and southward from the latitude of Ripton (Plate 1). To the north the Sugar Hill syncline is coextensive with the Starksboro syncline (Cady, 1945).

The Sugar Hill syncline is sharply overturned toward the west (Plate 2). The normal limb dips from 30 to 50° to the east, whereas the overturned limb dips from 50 to 80° east. The syncline is complicated in detail by smaller folds that are overturned and plunge in the same direction as the major fold.

A small patch of Mendon formation 1.4 miles N40°E of Ripton (Dale, 1910) is completely surrounded by Pre-Cambrian gneiss and is consequently interpreted as an outlier in a local depression within the synclinal trough.

Worth Mountain Syncline: Worth Mountain is underlain by a shallow, complex syncline. This fold is formed by a depression lying between the slightly more elevated Burnt Hill anticline and the Bread Loaf anticline. The wide expanse of Tyson is due, in part, to topographic control. The Worth Mountain syncline has a gentle southerly plunge and has many smaller folds superimposed upon it. The plunges of these minor folds do not always coincide with the plunge of the major structure (Plate 2). The west limb of the Worth Mountain syncline dips 20 to 60° east, and its east limb dips 60 to 70° west. However, the east limb is locally overturned and dips about 80° toward the east.

East Flank of the Green Mountain Anticlinorium

In general the beds of the east flank of the Green Mountain anticlinorium dip gently to the east, although in detail they are involved in considerable folding (Plate 1). The average dip ranges between 20 and 50° east. In places gentle dips to the west and steep dips to the east are observed. Thus the eastern flank of the anticlinorium is a broadly undulating sequence of rocks that locally undergoes sharp folding.

The lithologic distribution of the Pinney Hollow and the Ottauquechee formations indicates that the area in the vicinity of the valley of the White River is underlain by a shallow synclinorium. This structure consists of many large folds that undergo a series of axis-culminations and depressions. However, the syncline 1.6 miles N60°E from Rochester has an overall southerly plunge. These folds are complicated by a second generation of minor folds that in many cases obscure the older structures.

Considerable sharp folding may be observed on the knob 1.2 miles N40°W of Hancock (Figure 8). Here a bed of dark green albite-epidote-calcite-chlorite schist is exposed in an anticline that plunges about 45°S. beneath the associated quartz-chlorite-muscovite schist. The top of the knob is located on the crest of the anticline. The beds on its west limb dip steeply westward, and the beds on its east limb are thrown into a series of sharp folds with overturned limbs dipping from 50 to 65° east and normal limbs dipping 40 to 45° east.

Similar folding is exposed in Allbee Brook between the elevations of 990 and 1320 feet (Plate 22, Figure 1). A bed of dark-green albite-epidote-calcite-chlorite schist is contorted into many sharp folds which plunge both north and south at steep angles. The bedding likewise dips steeply toward the east, and its strike alternates from northeast to northwest.

The west limb of the Green Mountain anticlinorium forms the east limb of the Middlebury synclinorium. In general the beds dip steeply to the west, but in detail they are much folded. These folds are all either asymmetrical or overturned toward the west, and many of them are isoclinal.

The strata above the Falls of Lana in Sucker Brook are involved in folds that have a minor culmination in this vicinity (Figure 8). This culmination exposes a large area of Forestdale marble that is crumpled into a series of minor folds. The Cheshire quartzite at the Falls of Lana dips about 70° west and is right-side up. A short distance east of the dolomite-quartzite contact the dips pass through the horizontal, becoming 20 to 30° east. In the break above this point the dips are variable.

An overturned, isoclinal anticline of Cheshire quartzite underlies Bryant Mountain and Sunset Hill. The east limb of this fold dips 35 to 55° east into the syncline underlying Lake Dunmore (Plate 2). The bedding passes through the horizontal on the northwest slope of the ridge between Sunset Hill and Bryant Mountain and becomes overturned toward the west. This west limb has been broken by a break-thrust which Cady (1945) has named the Lake Dunmore thrust. The thrust plane was not observed, but the fault was inferred because the Moosalamoo member lies directly on the Dunham on Sunset Hill. On the south slopes of Sunset Hill the Cheshire is again in normal contact with the Dunham, and the Moosalamoo plunges south beneath the Cheshire in an isoclinal fold. Consequently, the Lake Dunmore thrust is interpreted to be of small displacement because it merely moves the Moosalamoo westward over the outcrop width of the Cheshire which in this vicinity is of the order of 1000 feet.

Minor Folds

Folds that are too small to be shown on the geologic map are classified as minor folds (Plate 18). These folds range in size from 1 centimeter to as much as 100 feet in wave length. The tightness of these folds depends on the nature of the rock. A high percentage of platy minerals decreases the competency of the beds, and therefore isoclinal folds are common. In rocks with a high proportion of non-platy minerals, the folds are more open.

Where the bedding trends are constant most folds have limbs that

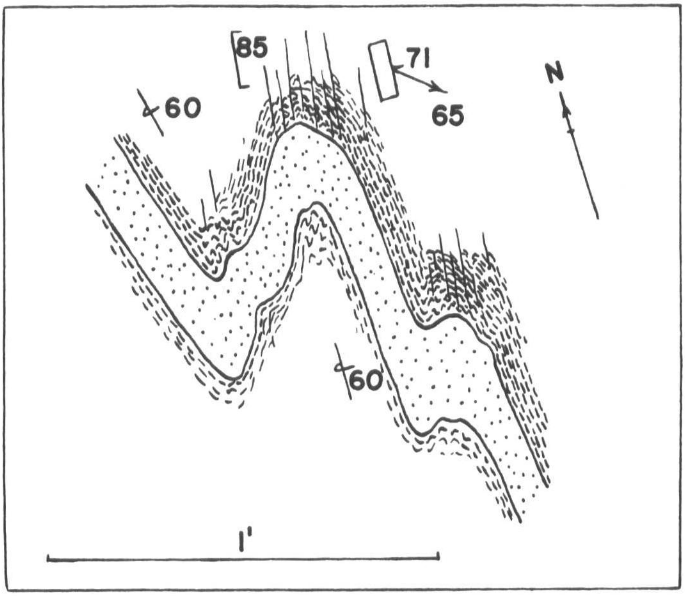


Figure 1. Fold in thin quartzite bed, Granville formation. 1.1 miles N. of Granville.

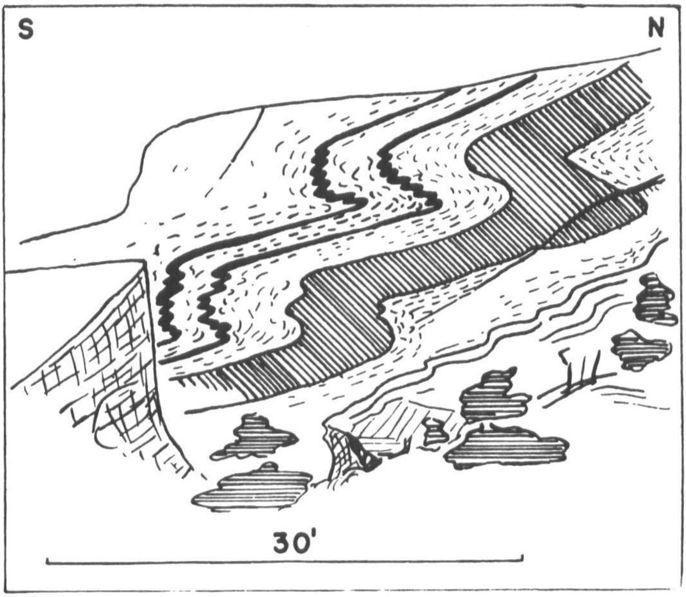


Figure 2. Folds in exposure of Pre-Cambrian gneiss. Silent Cliff.

MINOR FOLDS

PLATE 18

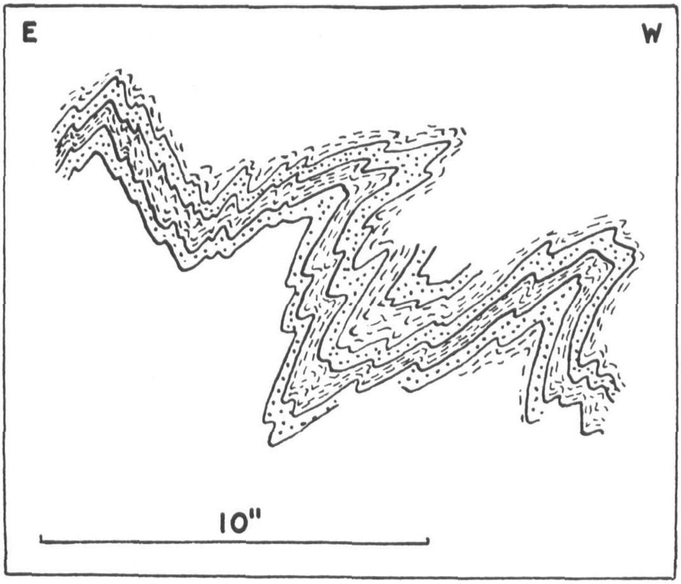


Figure 3. Folds in Pre-Cambrian schist. Burnt Hill.

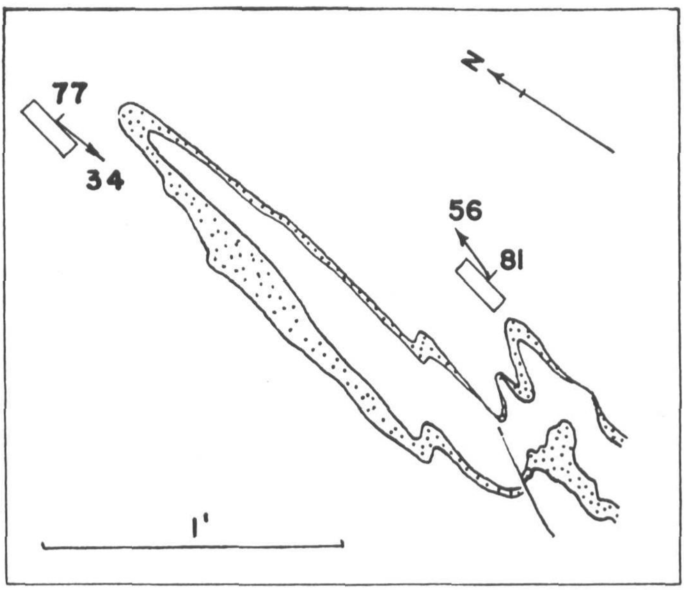
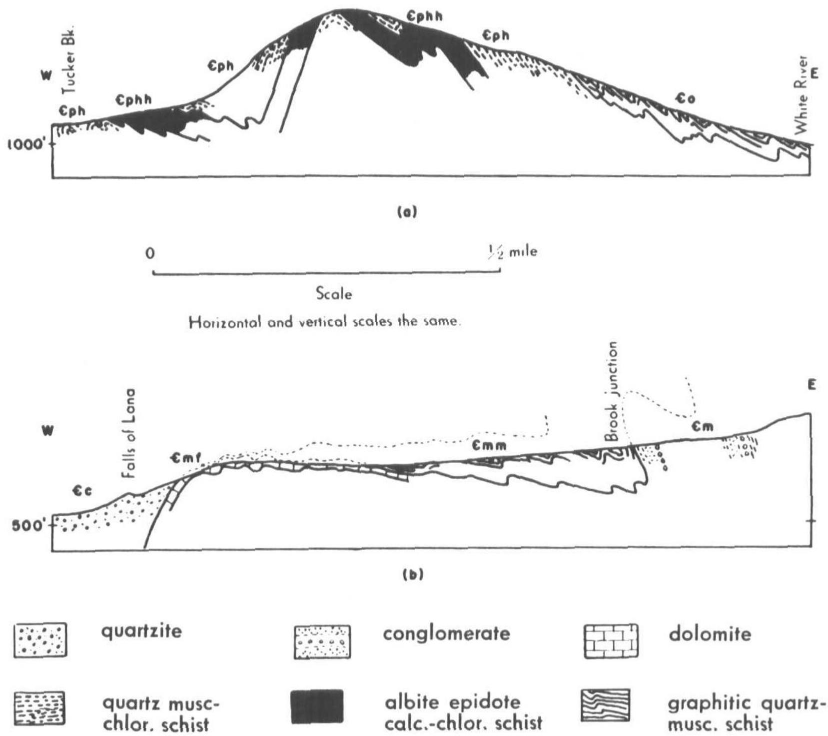


Figure 4. Map of folded fold, Pinney Hollow formation. Thatcher Brook.

MINOR FOLDS

PLATE 18



€m—Mendon fm. €mf—Forestdale member of the Mendon fm. €mm—Moosalamoo member of the Mendon fm. €c—Cheshire quartzite €ph—Pinney Hollow fm. €phh Hancock member of the Pinney Hollow fm. €o—Ottauquechee fm.

(c) Structure section from Tucker Bk. to the White River, drawn 1.0 mile north of Hancock
(b) Structure section along Sucker Brook.

DETAILED STRUCTURE SECTIONS

Figure 8

dip and strike in two different directions on opposite sides of the crest. Commonly the length of one limb is longer than the length of the other limb. The limb with the shorter length is called the short limb, whereas the limb with the longer length is called the long limb. Furthermore, when viewed in a horizontal plane, minor folds crumple the bedding into zig-zag patterns. If the long limbs are offset to the right, the fold has a dextral pattern; if the bedding trend is offset to the left, the fold has a sinistral pattern (White and Jahns, 1950). Shear sense indicates the direction of sliding of rocks past each other along fractures (Billings, 1942, p. 16). In most cases, the shear sense also can be ascertained from the folded pattern of bedding or foliation. The direction of rotation of the short limb of the plication indicates the shear sense. Thus if the short limb has been rotated in a clockwise direction, the overlying rock-mass has moved to the right relative to the underlying rocks.

There are two generations of minor folds in the Rochester-East Middlebury area. The first generation of folds is generally preserved in rocks containing a low content of platy minerals. These folds are preserved only as remnants of crests and troughs that commonly plunge to the south. The second generation of folds is particularly well preserved in rocks containing an abundance of platy minerals. These folds and associated minor structural features contort bedding planes and schistosity planes alike. The majority of these folds on the east flank of the Green Mountain anticlinorium plunge from 35 to 60° northeast.

A few exposures contain evidence of both generations of minor folds. The most common evidence is found in outcrops that contain the trace of the bedding on a schistosity plane. The trace of the bedding usually plunges from 30 to 60° southeastward. The plane of the schistosity, however, is folded by small crinkles that plunge toward the northeast. An exposure of graphitic quartz-muscovite schist 1.7 miles N15°W from Hancock contains a thin sericitic quartzite bed that is folded into a tight synclinal trough that plunges 52°, S35°E. In the same outcrop the schistosity is folded into small flexures that plunge 11°, N12°E. A so-called folded fold is well displayed at an elevation of 1170 feet in Thatcher Brook (Plate 18, Figure 4). Here a layer of sericitic quartzite is folded into an isoclinal fold. Superimposed on this fold is a second fold that buckles both limbs of the primary structure. Although the plunge of the older fold is difficult to measure, it is thought to plunge to the southeast. The younger fold plunges to the northeast.

Clear-cut examples of older folds are not common east of the Burnt

Hill anticline. When found, these folds usually plunge toward the south, although a few reversals in plunge do occur. The ax al planes of these folds dip at a high angle toward the east. Most of the drag folds in the East Middlebury quadrangle belong to the older generation. These folds have a sinistral pattern and plunge in harmony with the major structures. These minor folds are located on the west flank of the Green Mountain anticlinorium.

The evidence for determining whether these folds in the vicinity of Burnt Hill are older or younger is lacking. They plunge dominantly southward (Plate 2). Their axial planes dip 20 to 70° to the east, and all of the folds are overturned toward the west.

The distribution of the second generation of folds is restricted to the area east of the Burnt Hill anticline. The plunges of the majority of these folds are 20 to 70° to the northeast. Their axial planes dip 60 to 80° east (Plate 2).

Foliation

Foliation is the property of a rock to break along approximately parallel surfaces (Billings, 1942, p. 213). This structure is prominent in most rocks in the Rochester-East Middlebury area and results from the parallelism of platy minerals. In most cases where indications of bedding are present, the foliation and the bedding are sensibly parallel. The foliation even wraps around the noses of folds. This type of foliation has been called mimetic cleavage (Billings, 1942, p. 218).

In rocks composed dominantly of platy minerals, the foliation is generally steeply inclined. The dips change as much as 20 degrees in successive measurements across the strike. This change in dips is accompanied by a slight change in their strikes. These recurring alternations in strike and dip are interpreted as indicating sharply crested plications that are not otherwise visible because of the lack of compositional banding.

In the eastern part of the area the foliation dips 65 to 85° east and has a northerly strike (Plate 2). As the crest of the Burnt Hill anticline is approached, the dips become more gentle. However, even where the dips are steep, the writer believes that the average dips, taken across the crests of the plications, are relatively gentle.

Slip Cleavage

The term slip cleavage was first used by Dale (1896, p. 561) in the Green Mountain region to describe planes of cleavage transverse to the bedding. He states: "The slippage due to cleavage seems to have been mostly confined to certain narrow belts or bands, which recur at intervals of a foot or so. This appears to be the result, in some cases, of the inequality of the limbs of the plications and of the slippage occurring at the apex of the fold and within the shorter limb." In this paper slip cleavage is used with the same connotation as was used by Dale (Plate 20, Figure 1).

Slip cleavage of two types is recognized in the Rochester-East Middlebury area; reverse slip cleavage and normal slip cleavage. Reverse slip cleavage is most common and develops from asymmetrical plications. As the small plications are formed, their short limbs become stretched. Continued rotational deformation results in the development of a stretch-thrust which forms on the short limb. The long limbs between adjacent stretch-thrusts become rotated and tend to parallel the cleavage planes (Plate 19, Figure 1).

If the long limbs bounded by cleavage planes are rotated into parallelism with the cleavage planes, evidence of the original bedding (or foliation) is lost. Such cleavage consists of the parallel alignment of the platy minerals with the cleavage planes and is more properly called schistosity. Schistosity is the property of schistose rocks to break along surfaces of secondary origin (Billings, 1942, p. 219). Of course, schistosity formed in this way is impossible to distinguish from original foliation.

In normal slip cleavage the cleavage planes develop without relation to plications of the foliation. The cleavage planes form on either the short or the long limbs of the plications or in rocks in which plications are absent. The foliated laminae bounded by cleavage planes are rotated in a direction such that the displacement on the cleavage planes is analogous to that of normal or high-angle reverse faults.

In order to facilitate discussion of slip cleavage, the strain in the rock is referred to a coordinate system consisting of three mutually perpendicular axes. One axis of strain (b) lies in the cleavage plane and parallels the trace of the foliation on the cleavage plane. A second axis of strain (c) lies in the cleavage plane and is perpendicular to the b axis. The third axis of strain (a) is perpendicular to the other two axes.

The development of reverse slip cleavage is interesting in that it necessitates an increase in the length of the rock-mass parallel to the c axis, and a decrease in the length of the rock-mass parallel to a. This must mean that reverse slip cleavage develops in response to a stress system that imposes a maximum shortening parallel to a and a minimum

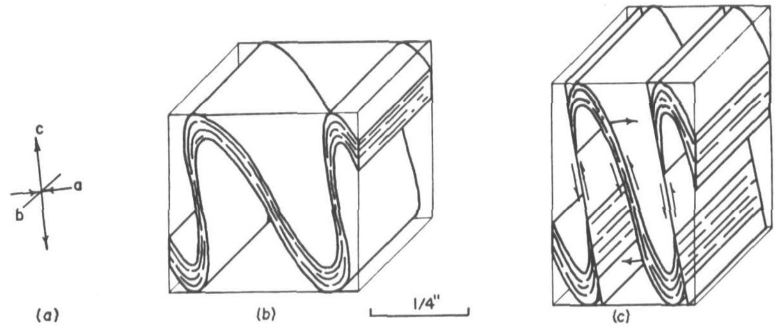


Figure 1. Reverse schistosity. (a) Strain relations in rock. (b) Small flexures. (c) Formation of slip planes and rotation of "long limbs" toward parallelism with slip planes.

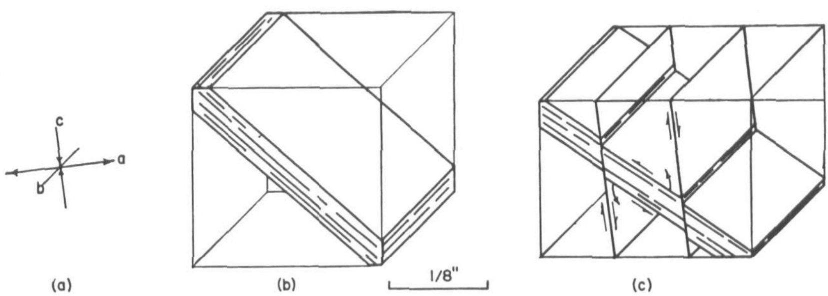


Figure 2. Normal schistosity. (a) Strain relations in rock. (b) Lamella in rock. (c) Formation of slip planes and rotation of the segments of the lamella toward perpendicularity to the slip planes.

SCHISTOSITY

PLATE 19

shortening parallel to *c*. The direction of rotation of the foliation planes bounded by cleavage planes is in a direction which relieves the stressed condition.

In normal slip cleavage the rotation of the foliation planes bounded by cleavage planes extends the rock-mass parallel to *a* and shortens the rock-mass parallel to *c* (Plate 19, Figure 2). Thus normal slip cleavage must result from a stress condition imposed on the rock that necessitates shortening of the rock-mass parallel to *c* and extension of the rock-mass parallel to *a*.

Slip cleavage in the Rochester-East Middlebury area strikes either north or slightly east of north and in general dips from 55 degrees east to the vertical (Plate 2). The schistosity in the Moosalamoo member is actually slip cleavage, because it truncates the bedding (Plate 21, Figures 1 and 2). Good examples of the relationship of slip cleavage to small folds may be observed at an elevation of about 1170 feet in Sucker Brook. These folds have an amplitude of from 4 to 5 feet and are cut by closely spaced cleavage. The shear sense indicates that the cleavage is reverse slip cleavage. It is essentially parallel to the axial planes of the folds.

Cleavage banding is the arrangement of compositional bands of secondary origin parallel to cleavage planes (Dale, 1896, p. 561). Cleavage banding originates during the development of reverse slip cleavage. In the first stages of its development, the individual laminae between planes of foliation are thrown into a series of plications in which the upper layers move toward the crests of the plications. This relative movement of the foliated laminae creates potential voids in the crests and troughs of the small folds and consequently serves to localize the growth of quartz and albite. On the other hand, because the limbs undergo shearing, little room exists here for the development of non-platy minerals. Therefore, it appears that the platy minerals are concentrated along the incipient cleavage surfaces and the non-platy minerals are concentrated in the crests and troughs (Plate 17, Figure 2). Because of these relative concentrations, the rock has a banded appearance of platy minerals alternating with non-platy minerals. These bands, however, are oriented at a large angle to the bedding or to the foliation.

Lineation

Lineation, as used in this paper, includes all linear structures. The principal lineations measured in the Rochester-East Middlebury area

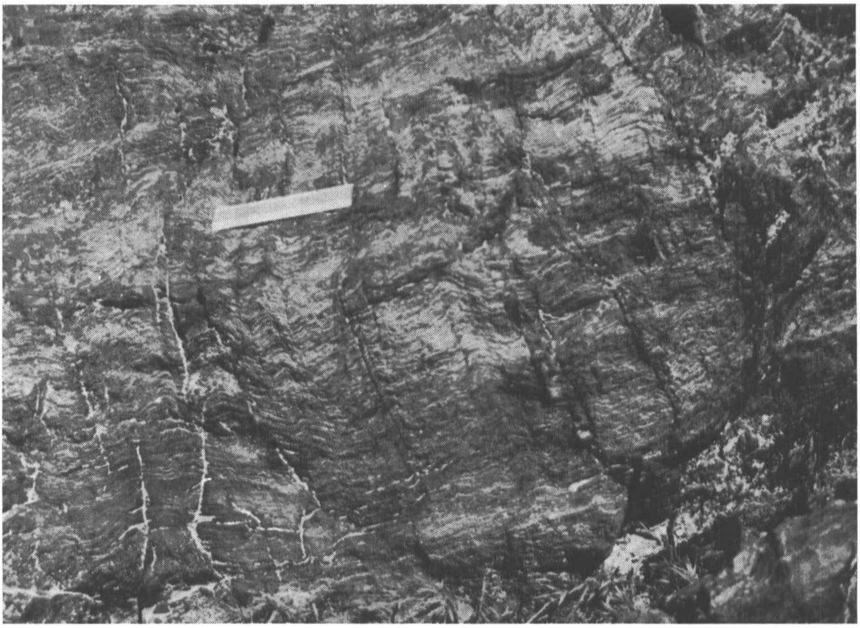


Figure 1. Schistosity in quartz-chlorite-muscovite schist. Foliation dips gently to the left. Ottauquechee formation, 3.2 miles S70°E from Hancock.

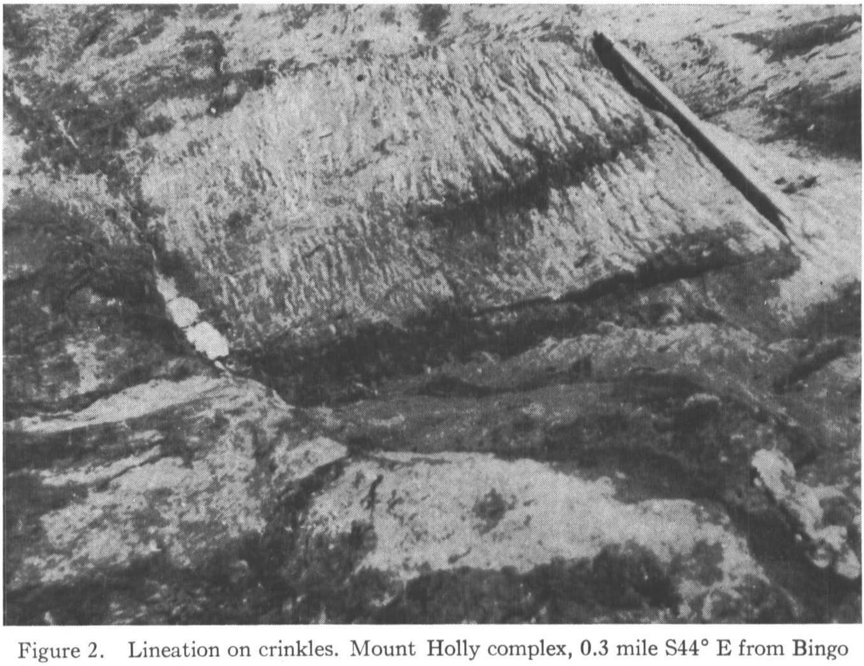


Figure 2. Lineation on crinkles. Mount Holly complex, 0.3 mile S44° E from Bingo Camp.

PLATE 20

Figure 1. Slip Cleavage in the Moosalamoo member of the Mendon formation. Nicols are not crossed. X 50.

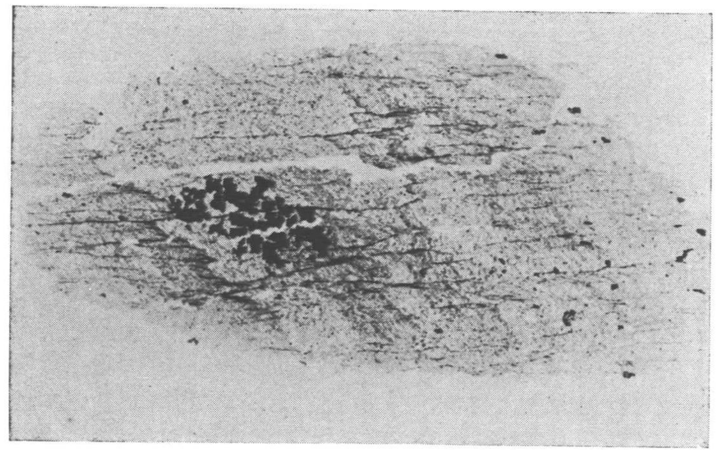


Figure 2. Slip cleavage truncating bedding. Moosalamoo member of the Mendon formation. X 2.

PLATE 21

are intersections of bedding and cleavage, crinkles, mineral streaming, and slickensides. Lineations formed by the intersection of bedding and cleavage are difficult to find in the schists on the east flank of the Green Mountain anticlinorium. Most of those measured plunge parallel to the younger folds. However, a few of these lineations plunge toward the south parallel to the plunges of the primary folds.

Crinkles are really small folds (Plate 20, Figure 2). These folds appear on the foliation planes as small wrinkles which are commonly broken on the short limb by micro-thrusts. The majority of the crinkles plunge in a direction parallel to that of the younger folds.

Mineral streaming is a lineation caused by a shearing-out of platy minerals in the foliation plane. The direction of this lineation indicates the direction of relative movement of the rocks. In this area the micas and the chlorites are smeared into faint streaks that plunge nearly down the dip. The direction of plunge of this streaming is commonly from 70 to 90 degrees of the direction of plunge of the younger folds. The lineations created by mineral streaming are fairly constant throughout the Rochester quadrangle; that is, most plunge nearly east (Plate 2).

Slickensides are striated surfaces that result from the differential movement of the rocks on opposite sides of a planar feature. The striations are linear features that indicate the direction of relative movement of the rocks on opposite sides of the planar feature. Slickensides are generally restricted to quartz lenses parallel to the foliation planes. These become striated by relative movement of the rocks. In this area good agreement exists between the direction of plunge of the mineral streaming and the slickensides. Both plunge toward the east.

Deformed Cobbles

Exposures of conglomerate at widely separated points afford an opportunity to compare their deformation. Unfortunately most of the exposures of conglomerate are west of the Burnt Hill anticline so that a complete analysis across the Green Mountain anticlinorium could not be made. Moreover it cannot be assumed that the pebbles were originally spherical. In the trough of the syncline 0.7 mile east of Mount Mossalamoo, the cobbles are apparently only slightly deformed and here have an average sphericity of 0.76. This value compares favorably with the sphericity of Recent, Pleistocene, and Pre-Cambrian cobbles (Pettijohn, 1943) (Table 18).

Elsewhere, however, the cobbles are deformed into ellipsoids. The conglomerate exposed 1.2 miles S68°W of Ripton is composed of cobbles that range in size from 0.5 to 6 inches in greatest dimension. These cobbles are not greatly deformed, the ratios of their axes ranging from 1:2:4 to 1:2.8:3. Some of the cobbles are not deformed, presumably because they were "protected" by their neighbors.

In well exposed outcrops of conglomerate 3.2 miles S20°E of Ripton the cobbles range in size from 4 to 12 inches in greatest dimension. The ratios of their axes are from 1:2:3 to 1:4:6. Again all of the cobbles are not deformed to the same degree, and a few cobbles are indented by adjacent cobbles.

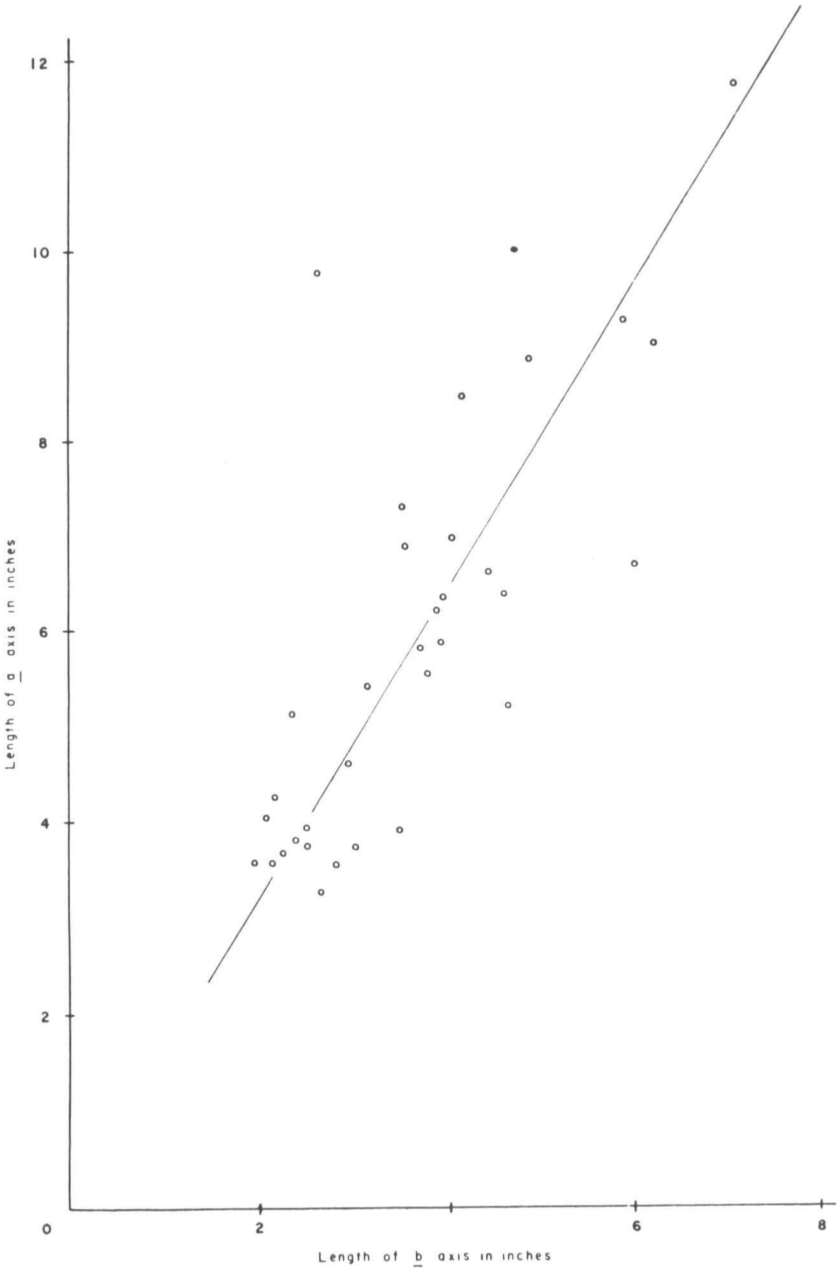
In all exposures of conglomerate in the Rochester-East Middlebury area the plane containing the intermediate and the long axes of the cobbles is parallel to the foliation. The long axes of the cobbles are oriented nearly directly down the dip of the foliation. All of the deformed cobbles are located on the flanks of folds.

The deformation of these cobbles can best be analyzed by using three axes of strain which are mutually perpendicular. The axis of intermediate strain, *b*, is oriented parallel to the axes of the older folds. The axis of maximum extension, *a*, lies in the plane of the foliation and is perpendicular to the intermediate strain axis. The axis of minimum extension, *c*, is perpendicular to the other two.

The deformation of the cobbles may be accounted for in terms of flattening along the *c* axis and a differential extension along the *b* and the *a* axes. The fact that the planes containing the *a* and the *b* axes of the cobbles lies in the plane of the foliation precludes shearing of the cobbles along closely spaced fractures parallel to the foliation. The cobbles apparently acted as resistant components in the rock, and therefore most of the shearing occurred in the matrix.

If the cobbles were originally spherical, the lengths of the *a* and the *b* axes indicate the relative amount of differential extension of the rock-mass in the plane of the foliation. But because the cobbles were not originally spherical, the lengths of the axes of the deformed cobbles represent the differential extension of the rock-mass superimposed on the shape of a non-spherical cobble.

The difference in the length of the *a* and the *b* axes of the deformed cobbles, however, is not inherited from the original shape. This conclusion is suggested because all cobbles have a similar orientation. Furthermore, if the ratios of the lengths of the *a* and the *b* axes are



a AND b AXES OF DEFORMED COBBLES FROM SIX LOCALITIES WEST OF WORTH MTN.

Figure 9

TABLE 18

Comparison of the Sphericity of Cobbles

Type	*Age*	*Locality*	*Sphericity*	*Number Measured*
Fluvial	Recent	San Gabriel Canyon, California	0.71	50
Glacial outwash	Pleist.	Aurora, Illinois	0.72	100
Fluvial	(?)Camb.	Ripton, Vermont	0.76	15
Fluvial	(?) Pre-Camb.	Lobay Bay, Ontario	0.74	38

TABLE 19

Comparison of Undeformed and Deformed Cobbles

	Volume in inches^3	b *in inches*	a *in inches*
Undeformed.	31.0	3.4	3.8
Deformed.	27.2	3.9	5.8
Undeformed.	19.5	2.7	2.9
Deformed.	19.8	2.2	4.5
Undeformed.	22.3	2.9	3.4
Deformed.	23.1	3.1	5.6

compared for deformed and undeformed cobbles having similar volumes, it is evident that the lengths of the *a* axes of the deformed cobbles have been increased whereas the lengths of the *b* axes remain about the same (Table 19). Consequently the extension parallel to the *a* axis represents an actual extension of the rock-mass. This extension is about 60 per cent greater than the original length of the *a* axis in the undeformed cobble. The ratio of the lengths of the *a* and the *b* axes is approximately constant regardless of the size of the cobbles (Figure 9). This means that if *b* remains constant, the rock presumably is flattened a minimum of 60 per cent also.

Rotated Porphyroblasts

In the Rochester-East Middlebury area east of the Burnt Hill anticline inclusions of sericite, epidote, ilmenite, and graphite commonly form S-shaped spirals within porphyroblasts of albite (Plate 22, Figure

1). These spirals have an increasingly smaller radius of curvature from the center towards the periphery of the porphyroblast. The S-shaped patterns are visible in thin section only where the section is cut normal, or nearly so, to the axes of the younger folds. Moreover, the direction of rotation of the porphyroblasts is the same as that of the short limbs of the younger folds.

The development of the S-shaped inclusions is due to simultaneous growth and rotation of the enclosing porphyroblasts (Plate 22, Figure 2). The inclusions were originally a part of the compositional banding of the rock. However, as a growing porphyroblast encloses a portion of the rock, certain components of the original banding remain as inclusions, and with rotation of the porphyroblast, the trains of inclusions are rotated out of parallelism with the compositional banding. Because the inclusions at the center of the crystal were enclosed first, they show greater rotation than the subsequently enclosed inclusions near the periphery of the porphyroblast. The pattern of inclusions depends on the rapidity of growth, the duration of growth, and the amount of rotation of the porphyroblast. In this region the central portions of the spirals have rotated from 40 to 90°.

These S-shaped patterns of inclusions are true rotational features and not merely the trace of a small crinkle that served as a locus for the later development of the porphyroblast. If this were not the case, the crinkle would extend beyond the limits of the porphyroblast; however, this is not the case.

Garnets are much less abundant than albite, and consequently data on their rotation are scarce. However, garnets have undergone some rotation, because in thin section smaller grains of sericite and chloritoid that grew across the boundary of the garnet are consistently broken or bent. On the other hand, the garnets do not show "snowball" patterns of inclusions like those within the albite. Thompson (1950) reports that in the Ludlow quadrangle the garnets contain S-shaped patterns of inclusions, and there the amount of rotation is similar to that obtained from a study of the albite porphyroblasts in the Rochester quadrangle.

Joints

No systematic study of the joints was made, although two sets are relatively prominent throughout the area. The majority can be classified as strike-joints and cross-joints (Terminology of Cloos, 1937). In the schists on the east flank of the anticlinorium the major rock adjustment

Figure 1. Rotated porphyroblast of albite showing well developed S-shaped pattern of inclusions. Monastery formation. Nicols are not crossed. X 50.

(a) (b) (c)

Figure 2. Development of S-curves of inclusions in porphyroblasts, a, b, and c show successive stages where growth of porphyroblast is combined with rotation. (after Fairbairn, 1949)

PLATE 22

was accomplished by movement on the foliation planes, and consequently joints are poorly developed. The strike-joints parallel or nearly parallel the strike of the axes of the minor folds. These joints strike approximately N10°E to N30°E and are nearly vertical. The cross-joints strike from N70°W to N85°W and dip steeply north or south.

In the more competent beds of the west flank of the Green Mountain anticlinorium the strike-joints strike nearly north-south parallel to the axes of the major folds. These joints dip nearly vertically. Cross-joints strike nearly east-west and dip vertically also. In general these are subnormal to the fold axes.

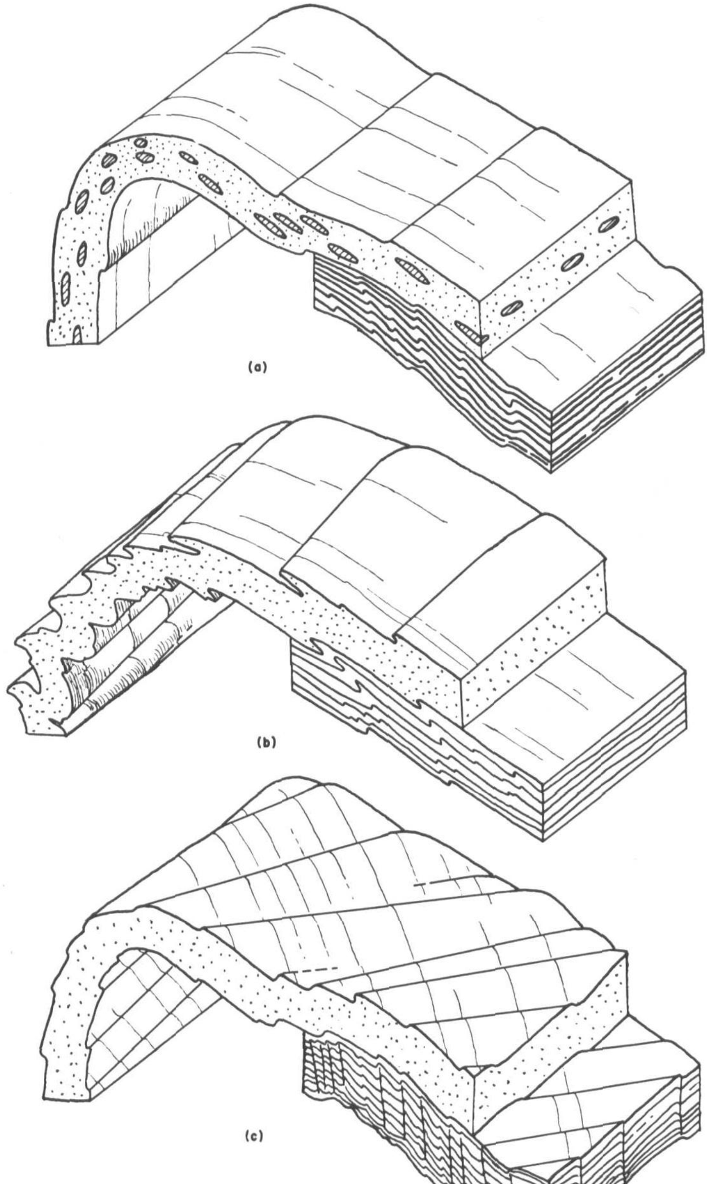
Structural Synthesis

General Statement

The major and minor structures in the Rochester-East Middlebury area are capable of an interpretation that welds them into a unified whole (Figure 10). For the most part, the minor structures east of the Burnt Hill anticline are incongruous with those of the Green Mountain anticlinorium and appear to be of later origin. The minor structures of the Burnt Hill anticline are amenable to two interpretations. They may have been formed contemporaneously with the formation of the Green Mountain anticlinorium, or they may have been formed subsequent to the formation of the anticlinorium. In either case, these structures indicate a large amount of movement of the overlying rocks toward the west. The minor structures west of the Burnt Hill anticline are congruous with the structures of the Green Mountain anticlinorium.

Synthesis East of the Burnt Hill Anticline

Within this area the lithologic distribution indicates a series of folds that generally plunge toward the south. Although this interpretation of the major structure is confirmed by the plunges of a few large folds, it is not supported by the majority of the minor folds. On the contrary, most of the minor folds plunge toward the northeast. However, the field relations indicate that these northeasterly plunging folds were formed late in the tectonic history of the Green Mountain anticlinorium. Moreover these secondary folds were formed by a movement of the overlying rock-mass toward the northwest, whereas the major structures of the Green Mountain anticlinorium were apparently formed by a movement of the overlying rock-mass toward the west.



RELATIONSHIPS BETWEEN MINOR AND MAJOR STRUCTURES
(a) West of Burnt Hill. (b) On Burnt Hill. (c) East of Burnt Hill.
Major folds trend about N-S

Figure 10.

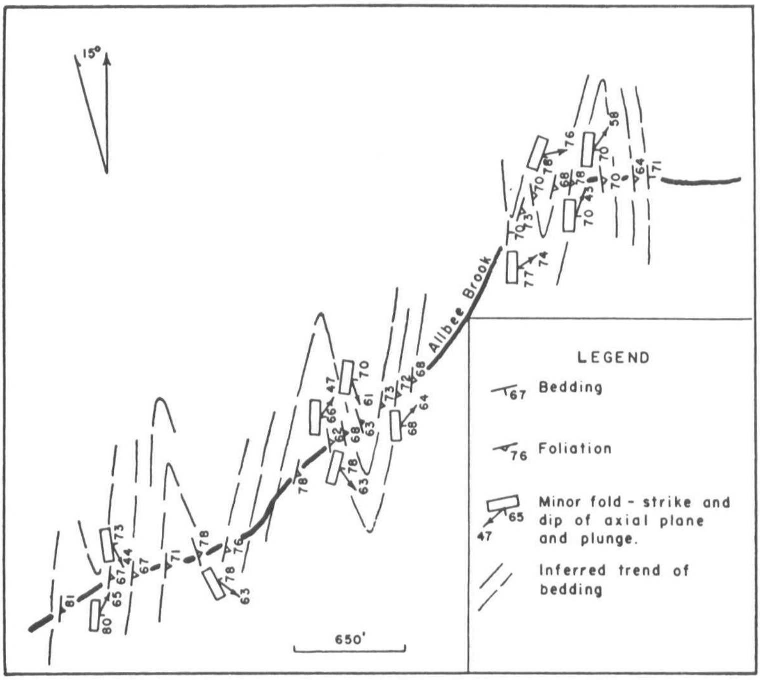


Figure 1. Minor folds between the altitudes of 990 feet and 1320 feet on Allbee Brook, Granville.

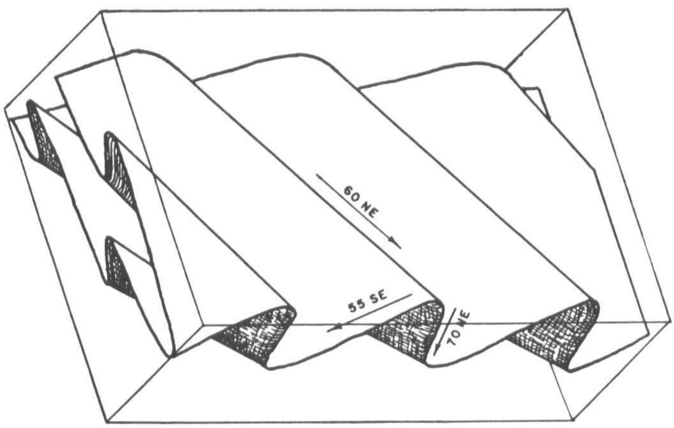


Figure 2. Interpretation of plunges of minor folds: minor syncline folded by northeast plunging minor folds.

PLATE 23

A few minor folds that plunge toward the south have been preserved. In general these folds plunge too steeply to account for the existing lithologic distribution. Nevertheless, these southerly plunging folds are believed to be remnants of folds formed during the folding of the anticlinorium and therefore are called older folds. Initially these older folds plunged dominantly toward the south and were tightly compressed and overturned toward the west. At a later time, however, these folds were folded about northeasterly plunging axes. The plunges of the older folds were thereby rotated to a position where they plunge more steeply than formerly (Plate 23, Figure 2). This mechanism accounts for the fold plunges observed in Allbee Brook (Plate 23, Figure 1).

The linear structures are congruous with the younger minor folds. The plunge of the crinkles is parallel to that of the younger folds. Both mineral streaming and slickensides are oriented at right angles to the axes of the younger folds. Minor folds and linear structures indicate that the direction of movement of the overlying rock-mass was toward the northwest during the second stage of folding.

A semi-quantitative measure of the amount of movement along the foliation planes can be obtained from a study of rotated porphyroblasts. The S-shaped spirals of inclusions in albite indicate a rotation of from 40 to 90°. The shape of the spiral is a function of the time of initiation of growth of the porphyroblast and the frictional resistence which it experiences during rotation. It is improbable that all of the porphyroblasts began to grow at the same instant or that they began to grow at the moment shearing was initiated. Therefore, it seems reasonable to take those porphyroblasts with a maximum amount of rotation as being indicative of the minimum amount of movement by shearing. In this respect probably none of the porphyroblasts record the total amount of movement.

The frictional resistence to rotation will further retard the formation of the S-shaped spirals of inclusions. Moreover, in most of the porphyroblasts the central section of the spiral is straight indicating that there is a certain minimum diameter that is required before the grain is caused to rotate.

The amount of shear may be calculated by assuming that the porphyroblast acts like a "ball-bearing." With a given angle of rotation, the grain rolls along the underlying surface a given distance, and the overlying surface rolls along the top of the grain an equal distance.

Thus the amount of shear is equal to the total lateral distance that the upper layer moved divided by the diameter of the porphyroblast.

$$D = 2R\phi$$

$$S = \phi\ D/2R = \phi$$

D = lateral distance; S = movement due to shear; R = radius of grain; ϕ = angle of rotation in radians.

If 90 degrees is taken for the value of ϕ, then since the radius cancels out, the amount of movement becomes $\pi/2$. This value would give an extension of the rock-mass of about 54 per cent. Estimating the thickness of the sedimentary column from the base of the Cambrian to the Middle Ordovician unconformity as being 24,000 feet, the total shear in this column would be about 37,680 feet, or about 7 miles. This means that a given point on the Middle Ordovician unconformity, now exposed beyond the limits of the area, would be moved about 7 miles westward by a movement of this kind.

This calculation indicates that in a horizon of thickness n, the absolute movement is 1.6 n. Schmidt (1918) has calculated a factor of 3 and Becke (1924) has determined a factor of 5.6 from analyses of rotated porphyroblasts.

Although the axes of rotation of the porphyroblasts nearly coincide with the axes of the younger folds, the time of their rotation is not definitely established. The rotation could have occurred during the formation of the major structures, in which case a small subsequent rotation about an axis only slightly inclined to the first axis would produce a semi-parallelism of the earlier and later axes of rotation. On the other hand, all of the rotation would have occurred during the second stage of folding.

Minor Structures on the Burnt Hill Anticline

The minor folds on the Burnt Hill anticline may belong to either the first or to the second stage of folding. These folds plunge dominantly southward. On the east flank of the major anticline the axial planes of these small folds dip toward the east. However, they become more gently inclined at the crest and on the west limb of the Burnt Hill anticline. All of these minor folds are for the most part sharply overturned toward the west.

If these minor folds were formed at the same time as the major folds, they indicate that the Burnt Hill anticline plunges southward from an axis-culmination a short distance north of the Rochester-East Middlebury area. The minor folds are overturned toward the west even on the crest of the Burnt Hill anticline. Consequently they suggest that the Burnt Hill anticline is overturned or perhaps recumbent toward the west.

On the other hand, these minor folds can be interpreted as belonging to the second stage of folding. The orientation of the axes of the younger folds indicates that they were formed by a relative movement of the overlying rock-mass toward the northwest. Thus as the overlying rock-mass moved up the east limb of the Burnt Hill anticline, the secondary folds would plunge toward the northeast. At the crest of the major fold the secondary folds would plunge nearly horizontally, and on the west imb they would plunge toward the south (Figure 10).

Synthesis West of the Burnt Hill Anticline

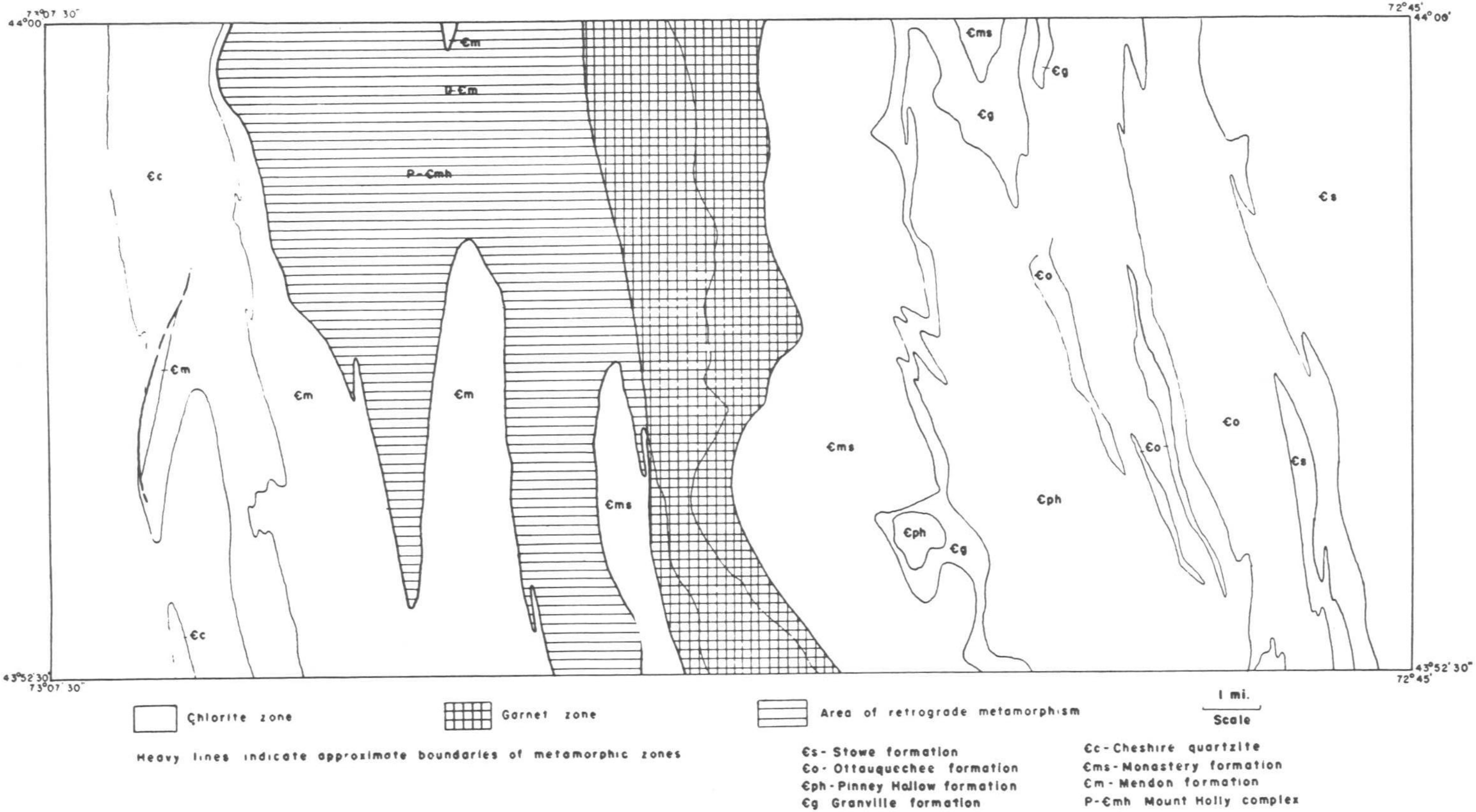
West of the Burnt Hill anticline the minor structures appear congruous with the major structures. The plunges of the axes of the minor folds are parallel to those of the major folds, and these can be used to determine the position of the major anticlines and synclines.

The cobbles west of the Burnt Hill anticline are deformed into ellipsoids. These cobbles are flattened parallel to the foliation and are elongated in a direction perpendicular to the major fold axes. The amount of elongation is about 60 per cent which is in close agreement with the elongation of the rock-mass east of the Burnt Hill anticline as deduced from the amount of rotation of albite porphyroblasts. Because those cobbles that are located in troughs of synclines are not deformed, the deformation of the cobbles and the simultaneous extension of the rock-mass must have occurred during the earlier stage of folding. It is therefore suggested that the rotation of the albite porphyroblasts on the east flank of the Burnt Hill anticline occurred during the earlier stage of folding also. This means that the movement of the rock-mass causing the younger folds was of small extent. Thus the minor folds on the Burnt Hill anticline belong to the older stage of folding.

METAMORPHISM

General Statement

All the rocks of the Rochester-East Middlebury area have suffered some degree of metamorphism. Many of those rocks are amenable to



METAMORPHIC ZONING

Figure 11.

treatment by the zonal concept of metamorphism. In this method the effects of increasing temperature and pressure can be observed in rocks of appropriate composition by the changes in the mineral assemblages. The first appearance of certain diagnostic minerals indicates higher metamorphic intensities. These diagnostic minerals are called index minerals. Barrow (1893) established the following index minerals for "pelitic schists" in Scotland: chlorite, biotite, garnet, staurolite, kyanite, and sillimanite. Rocks of other compositions have different suites of index minerals. However, if rocks of diverse chemical compositions are interbedded with "pelitic schists," some equivalency of the various sets of index minerals can be established.

The rocks of this area are divided into two large groups; the alkali-aluminous rocks and the mafic-aluminous rocks. The zonal minerals in the alkali-aluminous rocks are chlorite and garnet. A biotite zone is not distinguished because in this area the occurrence of biotite is controlled entirely by the chemical composition of the rocks. Chlorite and actinolite are the corresponding index minerals in the mafic-aluminous rocks. No index minerals occur in the carbonate rocks on the west flank of the Green Mountain anticlinorium except for a few scattered localities where phlogopite is present.

A narrow belt of garnetiferous schists extends southward from the vicinity of Battell Mountain through Middlebury Gap to Bingo Camp (Figure 11). The rest of the Paleozoic rocks are less metamorphosed. On the other hand, much of the Pre-Cambrian complex has suffered slightly more intense metamorphism but has since been subjected to retrograde metamorphism.

Alkali-Aluminous Rocks

General Statement

In the alkali-aluminous rocks the relation between bulk chemical composition and the metamorphic mineral assemblage can best be visualized on a tetrahedral diagram. The apices of this diagram are Al_2O_3 − alkalies, K_2O, $FeO - Fe_2O_3 - TiO_2$, and MgO (Barth, 1936). Silica and water are assumed to be in excess. However, a study of the rocks in thin section suggests that the system was not saturated with carbon dioxide, and thus carbon dioxide represents an oxide that cannot be adequately handled on these diagrams. Consequently, if ankerite or dolomite is present in the rock, the appropriate amounts of FeO and

MgO must be subtracted from the total mole per cent of these constituents.

Compositions are difficult to plot on three-dimensional diagrams. Therefore, the positions of the minerals are projected onto the sides of the tetrahedron, and compositions are plotted on the resulting triangular diagrams. Constant reference should be made to the tetrahedral diagram, however, to insure a correct understanding of the mineral relations.

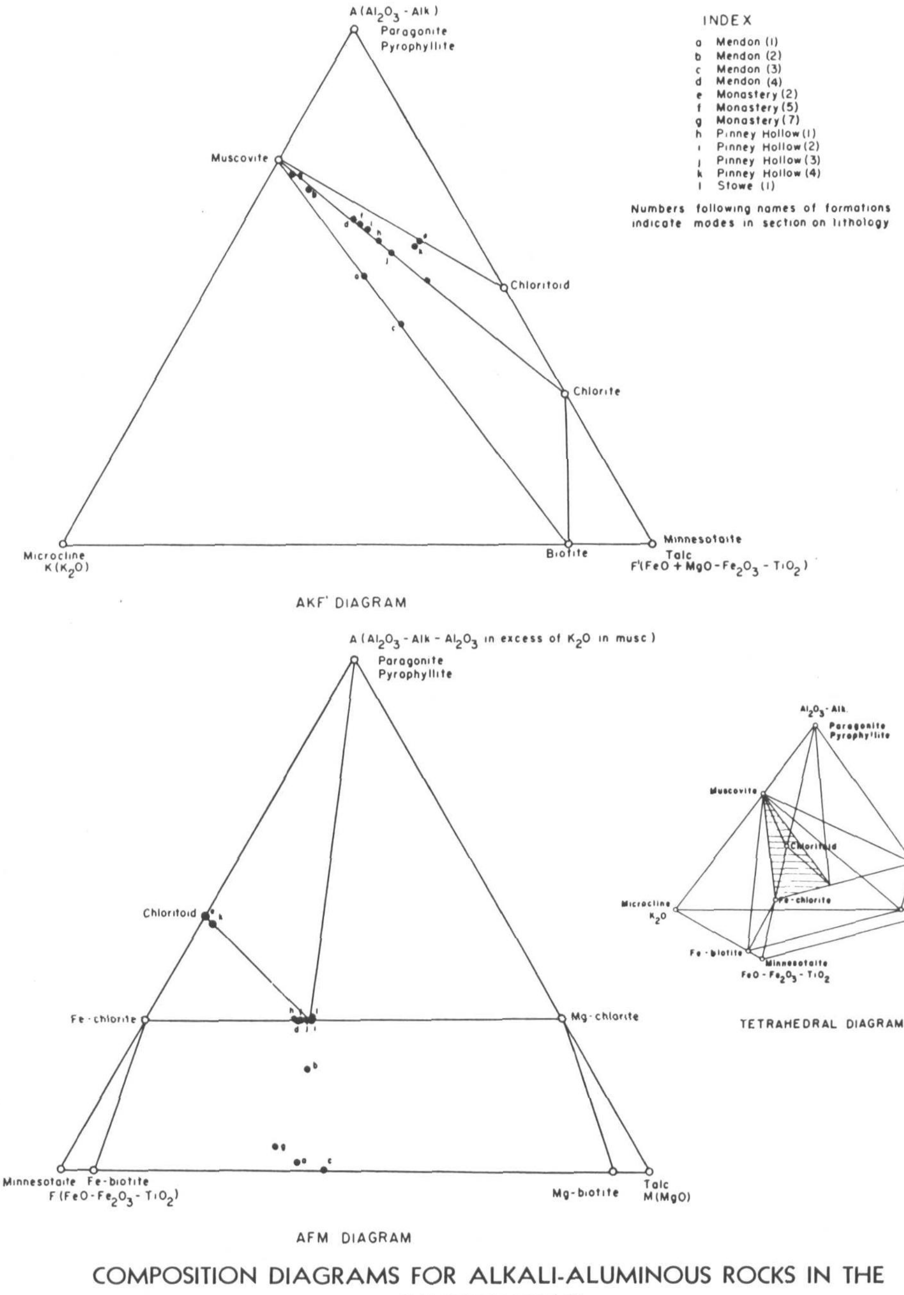
Two triangular diagrams are employed, the AKF′ and the AFM diagrams. In the AKF′ diagram Al_2O_3 − alkalies forms one apex of the triangular diagram; K_2O and MgO + FeO − Fe_2O_3 − TiO_2 form the other apices. However, the AKF' diagram is not suited to explain the relation between minerals in which iron and magnesia substitute for one another in all proportions and minerals in which iron and magnesia do not substitute for one another. The AFM diagram is used to illustrate these relations. In the AFM diagram one corner is represented by total Al_2O_3 − alkalies − Al_2O_3 in excess of alkalies in muscovite. FeO − Fe_2O_3 − TiO_2 and MgO represent the other two corners of the diagram. The amount of alumina in excess of alkalies in muscovite can be obtained by multiplying the weight per cent of muscovite as determined from the mode by 0.202.

Chemical compositions of typical rocks calculated from their modes are plotted as molecular proportions on the diagrams.

Chlorite Zone

Within the chlorite zone the typical mineral assemblages observed in the Rochester-East Middlebury area are albite, quartz, chlorite, and muscovite with rare chloritoid; quartz, chloritoid, and muscovite; and albite, quartz, biotite, and muscovite. The albite is close to An_0 and is typically porphyroblastic. The muscovite contains small amounts of ferrous and ferric iron as well as soda. The chlorite is probably aphrosiderite having a magnesia to iron ratio of 1:2.8. In biotite the optical properties indicate that the MgO to FeO ratio approximates 1:1.8. The chloritoid is iron rich.

When the molecular proportions of typical rocks are plotted on the AKF′ diagram, they have the distribution shown in Figure 12. Those mineral assemblages containing biotite, muscovite and quartz lie on the muscovite-biotite join. Biotite is stable throughout the chlorite zone in rocks of this composition. A few rocks are highly aluminous and their calculated molecular proportions lie within the muscovite-chlorite-



COMPOSITION DIAGRAMS FOR ALKALI-ALUMINOUS ROCKS IN THE CHLORITE ZONE

Figure 12.

biotite triangle. These rocks have a high content of muscovite and small amounts of biotite and chlorite, both of which are stable in the chlorite zone. Other rocks with a highly aluminous composition lie near the muscovite-chlorite join. Commonly chlorite is the only mafic mineral although in some cases chloritoid may also be present.

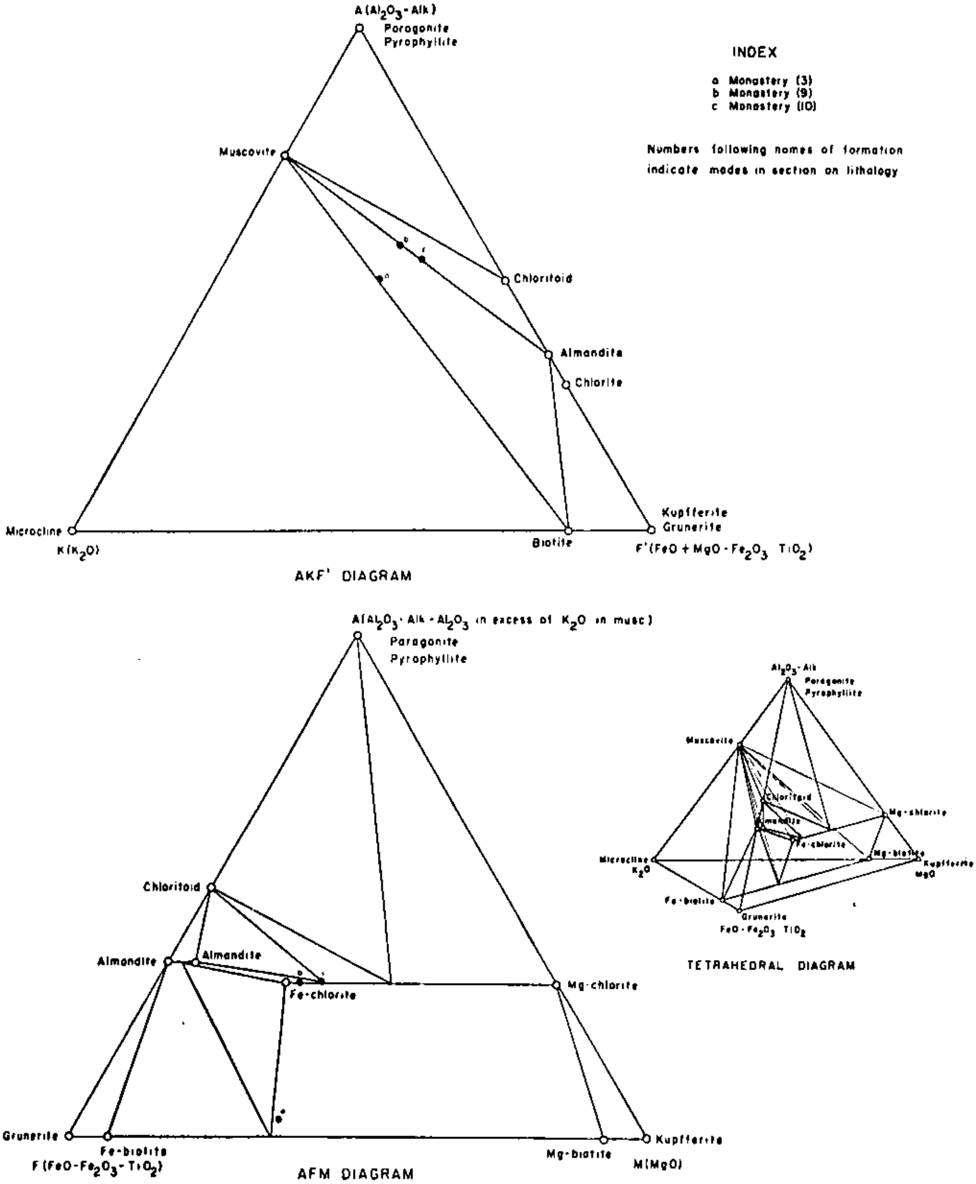
The relations between chlorite, biotite and chloritoid cannot be shown on the AKF′ diagram because magnesia does not enter the chloritoid structure in this metamorphic grade. However, if the molecular proportions are plotted on an AFM diagram, the relations become obvious. The molecular proportions of those mineral assemblages containing chloritoid are either coincident with chloritoid or lie on the chloritoid-chlorite join (Figure 12). Those mineral assemblages containing biotite and chlorite lie within the biotite-chlorite field, and those assemblages containing muscovite and chlorite are concentrated around chlorite.

The compositions of the rocks in the Rochester-East Middlebury area are such that a biotite zone cannot be distinguished. This, however, by no means invalidates biotite as an index mineral in rocks of the appropriate composition. If the molecular proportions of the rock lie on or slightly above the muscovite-chlorite join in the AKF′ diagram, and if the relative proportion of FeO increases with a slight increase in metamorphic intensity, biotite will crystallize because the molecular proportions in effect move across the muscovite-chlorite join into the muscovite-chlorite-biotite triangle. The proportion of iron may be increased by some CaO of epidote entering plagioclase and thereby decreasing the amount of available Al_2O_3. Also some reduction of ferric to ferrous iron may occur in rocks containing graphite. The carbon dioxide evolved during the reaction probably escapes as a gas.

Garnet Zone

In this zone the molecular proportions of the alkali-aluminous rocks are plotted on AKF′ and AFM diagrams similar to those used for the chlorite zone (Figure 13). These rocks are characterized by garnet. The appearance of garnet, however, does not mean the disappearance of chlorite, because the garnet at this grade is almandite and forms at the expense of only the ferrous portion of the chlorite. As a consequence the chlorite of this zone has a slightly higher magnesia content than in the chlorite zone. The feldspars are commonly slightly calcic with compositions ranging from An_5 to An_{10}.

In the Rochester-East Middlebury area the characteristic mineral



COMPOSITION DIAGRAMS FOR ALKALI-ALUMINOUS ROCKS IN THE GARNET ZONE

Figure 13.

assemblages found in the garnet zone are garnet, albite, quartz, chlorite, and muscovite with or without accessory chloritoid; and albite, quartz, biotite, muscovite, and small amounts of chlorite. The molecular proportions of these rocks plotted on the AKF′ diagram have the distribution shown in Figure 13. Rocks containing garnet, chlorite, and muscovite lie on the muscovite-garnet join. Rocks containing biotite, muscovite and a little chlorite lie nearly on the muscovite-biotite join.

The relation between chlorite, chloritoid, garnet, and biotite can be seen to greater advantage on the AFM diagram. Rocks that contain garnet, chlorite, and muscovite lie close to chlorite in the garnet-chlorite field. Rocks containing garnet, chloritoid, chlorite, and muscovite lie within the garnet-chloritoid-chlorite field close to chlorite. The bulk compositions of these rocks approach the composition of those rocks whose distribution is concentrated about chlorite in the AFM diagram for the chlorite zone. Rocks containing biotite and chlorite lie near biotite in the chlorite-biotite field. With the exception of the composition of the feldspar, the mineral assemblage for rocks of this composition is much the same in both the chlorite and the garnet zones.

Mafic-Aluminous Rocks

General Statement

Thin belts of dark-green albite-epidote-calcite-chlorite schist are interbedded with the Paleozoic sequence on the east flank of the Green Mountain anticlinorium. Rocks of similar composition are exposed locally in the Pre-Cambrian complex. These rocks belong to the mafic-aluminous group.

The mafic-aluminous rocks are associated with alkali-aluminous rocks. Consequently the evolution of mineral assemblages in the two compositional types can be compared. The presence of actinolite corresponds roughly with the appearance of garnet in the alkali-aluminous rocks. White and Billings (1951) have noted the same approximate equivalency of garnet and actinolite in the Woodsville quadrange, N. H.—Vt.

The mineral relations of the mafic-aluminous rocks are indicated on a tetrahedral diagram (Figure 14) in which the apices are $Al_2O_3 -$ alkalies, K_2O, $FeO + MgO - Fe_2O_3 - TiO_2$, and CaO. However, in order to facilitate plotting of chemical compositions, an ACF diagram is used. In this diagram one apex is $Al_2O_3 -$ alkalies $-$ excess Al_2O_3 over alkalies in muscovite. Alumina that enters the feldspars, muscovite, and biotite is not available for combination with the other minerals on the diagram, and therefore this alumina must be subtracted from the total alumina. The Al_2O_3 in excess of alkalies in muscovite can be found by multiplying the modal muscovite by 0.202. The other apices are CaO and FeO+ $MgO - Fe_2O_3 - TiO_2$.

Molecular proportions calculated from the modes are plotted on ACF diagrams (Figure 14).

Chlorite Zone

The stable mineral assemblage in this zone is albite, epidote, calcite and chlorite. The composition of the albite is close to An_0. Epidote is probably pistacite with about 10 per cent of the $Ca_2Fe_3Si_3O_{12}OH$ molecule. Chlorite has approximately equal molecular proportions of MgO and FeO and is probably ripidolite. The mafic-aluminous rocks in the Paleozoic sequence lie in the calcite-epidote-chlorite triangle, although considerable compositional variations occur (Figure 14). The joins outside of the calcite-epidote-chlorite triangle are unsubstantiated in as far as this study goes.

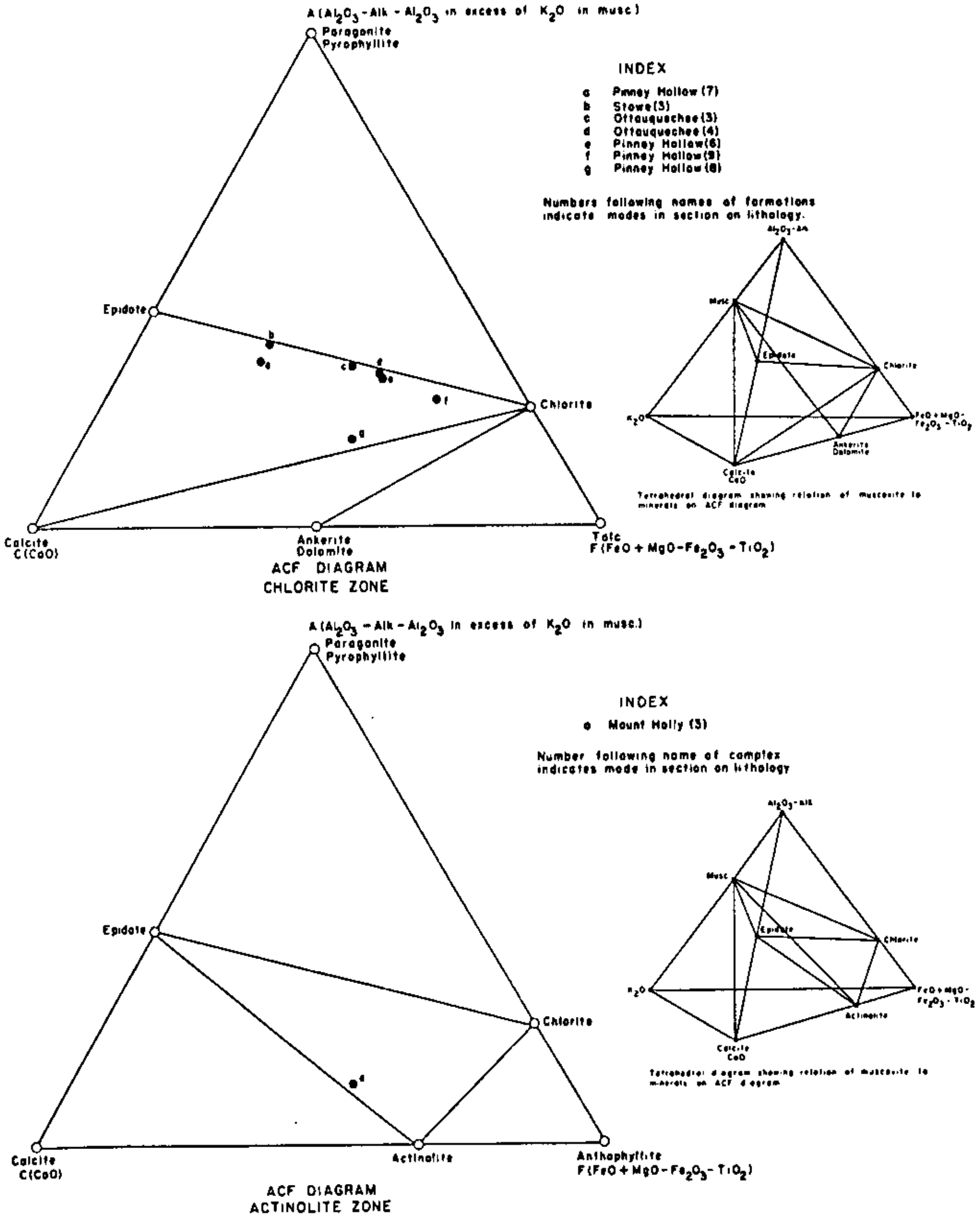
Actinolite Zone

Actinolite, plagioclase, epidote, and chlorite comprise the stable mineral assemblage in the actinolite zone. The plagioclase is An_5 to An_{10}. Epidote is, as far as could be determined, pistacite with about 10 per cent of the $Ca_2Fe_3Si_3O_{12}OH$ molecule. Chlorite is probably ripidolite with nearly equal molecular proportions of FeO and MgO. However, chlorite is not abundant in this zone.

The occurrence of mafic-aluminous rocks is meager in the actinolite zone, and consequently only one rock is plotted on the ACF diagram (Figure 14). The molecular proportions of this rock lie in the epidote-actinolite-chlorite triangle. Because the composition of this rock approximates that of rocks plotted on the ACF diagram for the chlorite zone, actinolite is inferred to have been derived from calcite (or ankerite or dolomite, if they are initially present) and chlorite. This inference is in some measure substantiated by the fact that mafic-aluminous rocks in this zone contain only small amounts of chlorite.

Retrograde Metamorphism

Retrograde metamorphic effects are found throughout the Pre-Cambrian rocks except for a narrow belt exposed along the Battell



COMPOSITION DIAGRAMS FOR MAFIC ROCKS

Figure 14.

Mountain–Worth Mountain ridge. Within the retrograde metamorphic Pre-Cambrian rocks the garnets are typically crushed and altered to chlorite. In some places this alteration has gone to completion. Many of the feldspars also show retrograde effects which give them a greenish appearance. This alteration consists of the formation of epidote and sericite from the original plagioclase. These retrograde features are interpreted as the effects of a later stage of metamorphism superimposed on an earlier and more intensive metamorphism of the Pre-Cambrian rocks.

The presence of these retrograde features suggests that the Pre-Cambrian rocks previously had attained a higher metamorphic grade than is now prevalent in this part of the Green Mountains. A few of the gneisses contain abundant microcline and little or no muscovite. These rocks may represent metamorphism of pre-existing sedimentary rocks extremely rich in K_2O. In other rocks in the Pre-Cambrian muscovite is abundant. These rocks do not appear to have suffered a metamorphism more intense than the upper limit of the garnet zone or perhaps the lower limit of the staurolite zone. Most of these rocks are coarse-grained, but pseudomorphs after garnet are the only evidence of middle-grade metamorphic minerals.

Metasomatism

Evidence for large-scale metasomatism is for the most part lacking in the Rochester-East Middlebury area. The contacts between lenticular marbles and the inclosing schists are sharp and wide-spread induction of alkalies does not appear to have occurred. The calculated chemical compositions of the schists compare favorably with chemical analyses of unmetamorphosed shales.

However, small-scale movement of material is evident. Vein quartz in nodules and stringers along the foliation planes is interpreted as being due to movement of silica out of the rock and into the incipient planes of slipping, particularly in the crests and troughs of folds. Localized movement is also necessitated by the presence of porphyroblasts.

Water and carbon dioxide apparently move readily through the rocks. These constituents seem to be available in sufficient abundance during metamorphism to form carbonates and hydrated silicates. Consequently it is felt that the system is always saturated with these two constituents. With sufficient time and proper conditions, these constituents are usually involved in metamorphic reactions.

Cause of Metamorphism

Heat, confining pressure, stress, and the presence of solutions contribute to the metamorphism of rocks. Among these, however, heat is believed to be most important, and thus the problem of the cause of metamorphism, in large part, becomes one of finding an adequate source of heat. Regional metamorphism is variously attributed to heat emanating from magmatic intrusion, to heat produced by friction resulting from intense deformation of the rocks, and to heat derived from deep burial with the consequent regional rise of temperature in the rocks.

No igneous rocks adequate to accomplish the metamorphism are exposed in this area. Moreover, except for small stocks at Cuttingsville, Vermont, and perhaps on Clarksburg Mountain, Vermont, there are no igneous intrusives exposed in the Green Mountain anticlinorium. The Green Mountain anticlinorium has been extensively eroded, and if the metamorphism was the result of subjacent intrusives, it would be expected that a larger area of igneous rock would now be exposed.

In the Rochester-East Middlebury area the garnet zone coincides with the Burnt Hill anticline. Assuming the garnet zone represents a higher metamorphic intensity than the chlorite zone, its localization suggests that the increase in metamorphism is due to the slightly greater deformation of this anticline relative to the other folds in the area. However, estimates of the heat produced by this mechanism indicate that it is insufficient to produce regional metamorphism.

In the light of the above, the main source of heat must have been a rise in the temperature of the rock because of its depression beneath the surface of the earth. This heat accounts for the broad patterns of metamorphism, but the details in metamorphic zoning may well be the result of localized intense deformation.

PRE-METAMORPHIC NATURE OF THE ROCK

During metamorphism many of the original textures and minerals disappear because of recrystallization. As a result, the original nature of many rocks in the Rochester-East Middlebury area is difficult to ascertain. However, most of the rocks contain remnants of bedding and some contain relict detrital grains. Consequently they are inferred to be of sedimentary origin. Some information regarding the nature of the sediments can be gained from a consideration of their chemical compositions.

TABLE 20

Chemical Composition of Sediments (Analyses Incomplete)

(Pettijohn, 1949)

	Graywacke	*Subgraywacke*	*Quartzose Sandstone*
SiO_2	68.1	74.4	99.4
Al_2O_3	15.4	11.3	none
Fe_2O_3	3.4	0.8	0.3
FeO	3.4	3.9	
MgO	1.8	1.3	tr
CaO	2.3	1.2	none
Na_2O	2.6	1.6	none
K_2O	2.2	1.7	none

TABLE 21

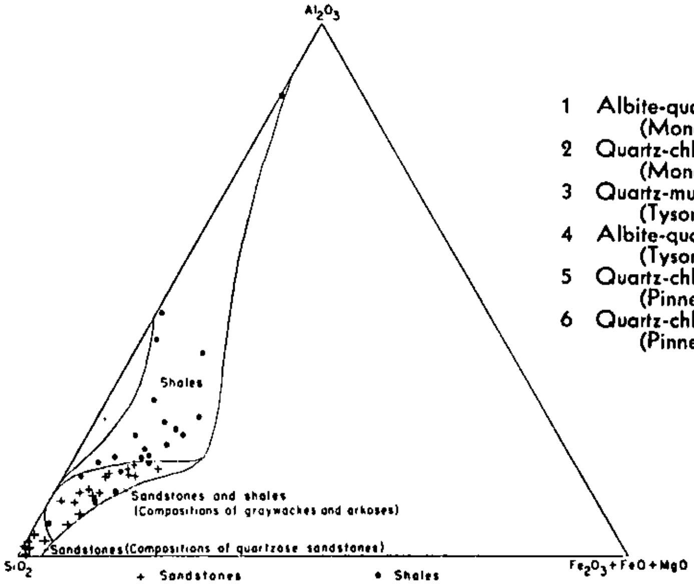
Relation of Chemical Composition to Size of Grain

(Modified from Pettijohn, 1949)
(Analyses Incomplete)

	Fine Sand	*Silt*	*Clay*
SiO_2	71.1	61.3	48.1
Al_2O_3	10.2	13.3	18.8
$FeO+Fe_2O_3$	3.7	3.9	6.9
MgO	1.7	3.3	3.5
CaO	3.6	5.1	4.9
Na_2O	0.8	1.3	1.2
K_2O	2.2	2.3	2.6

The chemical composition of a sediment depends on the abundance and kind of constituent minerals. The mineral assemblage in turn depends on the length of time that the agents of decomposition and disintregration, acting at the surface of the earth, operate on the sediment and on the size of the original sedimentary grains. Sediments continually acted upon become simpler in composition; the sand-sized fraction becoming more quartzose and the smaller sized fractions becoming more aluminous. These relations are shown in Tables 20 and 21. These major chemical modifications of sediments cease when it is buried.

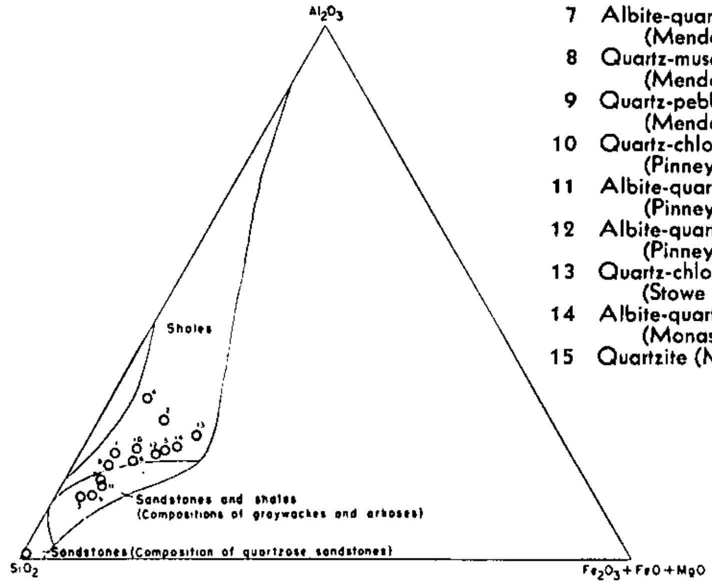
Chemical compositions of detrital sedimentary rocks, obtained from the literature, are plotted on a triangular diagram in which the apices



COMPOSITION DIAGRAM FOR SANDSTONES AND SHALES ANALYSES TAKEN FROM THE LITERATURE

Index

1 Albite-quartz-muscovite schist (Monastery fm.)
2 Quartz-chlorite-muscovite schist (Monastery fm.)
3 Quartz-muscovite schist (Tyson member)
4 Albite-quartz-biotite-muscovite schist (Tyson member)
5 Quartz-chlorite-muscovite schist (Pinney Hollow fm.)
6 Quartz-chlorite-muscovite schist (Pinney Hollow fm.)



COMPOSITION DIAGRAM FOR REPRESENTATIVE ROCKS OF THE ROCHESTER EAST MIDDLEBURY AREA

7 Albite-quartz-biotite-muscovite schist (Mendon fm.)
8 Quartz-muscovite schist (Mendon fm.)
9 Quartz-pebble conglomerate (Mendon fm.)
10 Quartz-chlorite-muscovite schist (Pinney Hollow fm.)
11 Albite-quartz-chlorite-muscovite schist (Pinney Hollow fm.)
12 Albite-quartz-chlorite-muscovite schist (Pinney Hollow fm.)
13 Quartz-chlorite-muscovite schist (Stowe fm.)
14 Albite-quartz-chlorite-muscovite schist (Monastery fm.)
15 Quartzite (Monastery fm.)

Figure 15.

are SiO_2, $Fe_2O_3+FeO+MgO$, and Al_2O_3. The quartzose sandstones, graywackes, and shales have the distribution shown in Figure 15. In this diagram the arkoses and subgraywackes are grouped with the graywackes. Some shales fall within the area of graywackes and associated sandstones. Consequently this area is not indicative of grain size. On the other hand, no sandstones fall in the area of aluminous shales, and the quartzose sandstones also form a distinctive group. Therefore, compositions lying within these areas are definitive concerning the grain-size of the original sediment.

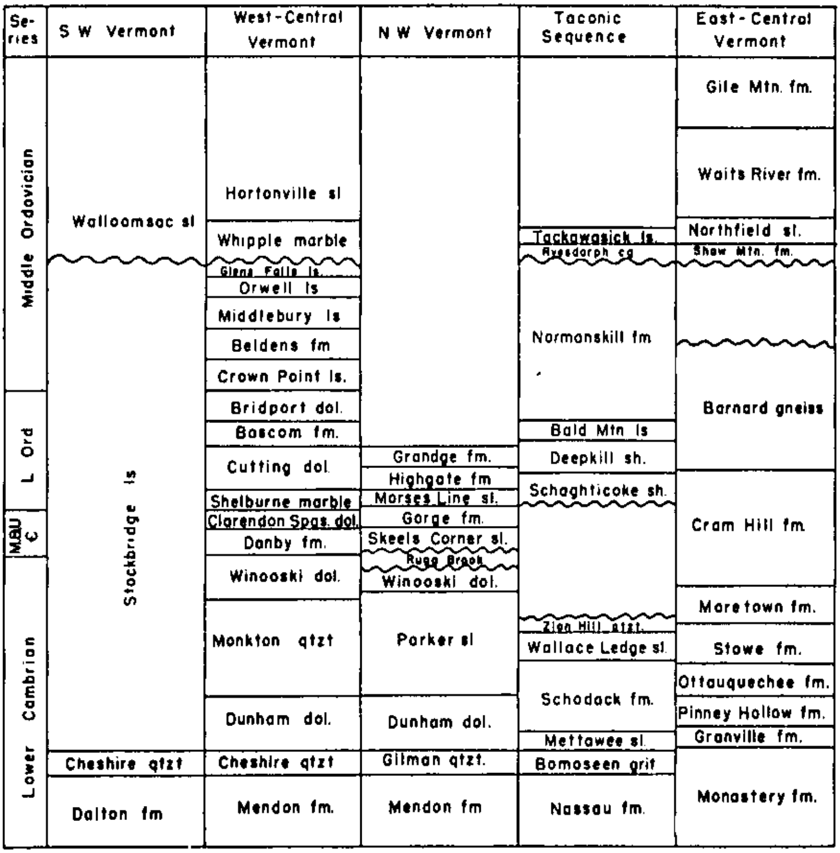
Calculated chemical compositions of rocks from the Rochester-East Middlebury area are plotted on the triangular diagram in Figure 15. All the compositions lie within the sedimentary boundaries. Those lying in the area of shales are interpreted as being originally shale. Those lying in the intermediate area have compositions that approach graywackes, but may have been either sandstones or shales. The single composition lying in the area indicating high silica content is interpreted as a quartzose sandstone.

The interpretation of the albite-epidote-calcite-chlorite schists is not conclusive in the Rochester-East Middlebury area. Their compositions approach those of basalts. However, their compositions could also be reproduced by combining appropriate proportions of dolomite and shale. Most of these schists show bedding, and many are gradational into the adjacent schists. Thompson (personal communication, 1950) reports that rocks of similar composition in southern Vermont contain amygdaloidal relicts suggesting that some of these rocks are flows.

REGIONAL RELATIONS

The Rochester-East Middlebury area covers but a small portion of the Green Mountain anticlinorium. In order to produce a more complete picture of the regional geology, the stratigraphy and structure of this area are integrated with the stratigraphy and structure of other areas in Vermont.

The correlation of the stratigraphy of the Rochester-East Middlebury area with that of other parts of Vermont is summarized in Figure 16. The certainty of correlation in most of the rocks is reduced because of lack of fossils. Correlation is based mainly on lithologic similarity and equivalence of stratigraphic position. Whenever possible these correlations are supplemented by paleontological evidence.



CORRELATION CHART FOR VERMONT

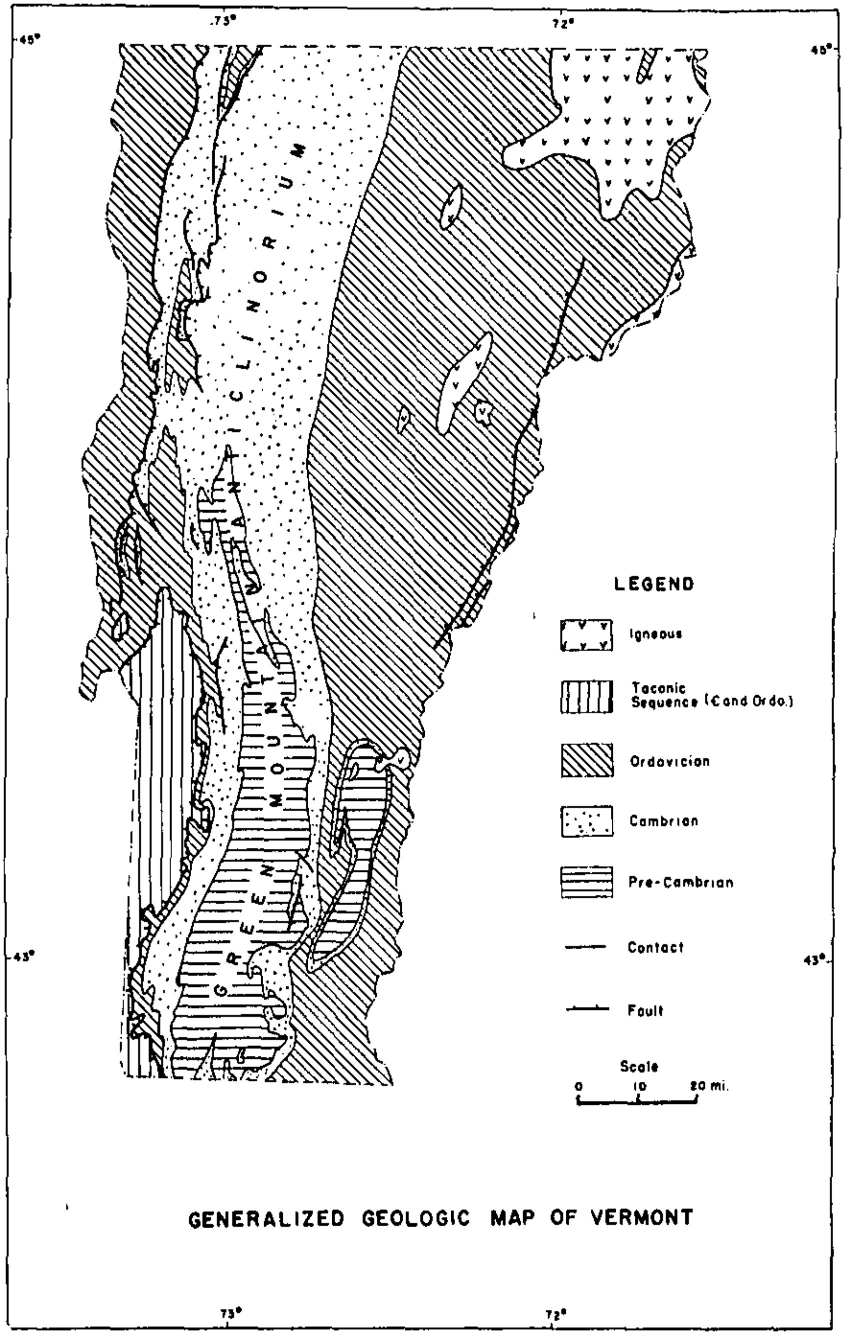
Figure 16.

The stratigraphic sequence on the east flank of the Green Mountain anticlinorium rests directly on the Pre-Cambrian complex. The formations established in central Vermont can be traced into southern Vermont (Chang, 1950; Thompson, 1950). The sequence has not been adequately studied in north-central Vermont.

The lithologic units established on the west flank of the Green Mountain anticlinorium also rest on the Pre-Cambrian complex and can be traced with minor variations in lithology the entire length of Vermont. The stratigraphic units in southwestern Vermont (Prindle and Knopf, 1932) bear a striking resemblance to units in west-central Vermont. However, in northwestern Vermont many of these units grade laterally into units of different lithology (Shaw, 1949; Stone, 1951). Thus the Winooski dolomite thins toward the north, and its place is taken by the Parker slate and the Rugg Brook. The Danby formation thickens northward and becomes the Skeels Corner slate, and the Clarendon Springs dolomite is equivalent to the Gorge formation. This succession exposed in northwest Vermont is intermediate in lithologic character between the eastern and western sequences of central Vermont.

Although correlation along strike is substantiated in large measure, correlation across the Green Mountain anticlinorium is much more difficult. Correlation of the rock units in the Rochester-East Middlebury area has been discussed in a previous section (Figure 6). The Mendon formation and the Cheshire quartzite are believed roughly equivalent to the Monastery formation. The Moretown formation is lithologically similar to the Monkton quartzite in west-central Vermont and to the Parker slate in west-central Vermont. (Thompson, personal communication, 1952.) The Waits River formation on the east flank of the anticlinorium is believed equivalent to the Hortonville slate on the west flank.

Correlation between the rocks of the Rochester-East Middlebury area and the Taconic sequence is even more hazardous than the correlation between the eastern and western sequence exposed on the flanks of the Green Mountain anticlinorium. The Taconic sequence occupies a range that extends southward from Brandon, Vermont, into Massachusetts and Connecticut (Figure 17). These rocks are of Cambro-Ordovician age and lie with structural discordance on the Cambro-Ordovician sequence of the west flank of the Green Mountain anticlinorium. The Nassau beds are somewhat similar in lithology to the lower part of the Monastery formation as mapped in the Rochester-



73°
72°
45°
45°
M U I R O N I L C I T N A
N I A T N U O M N E E R G
LEGEND
Igneous
Taconic Sequence (C and Ordo.)
Ordovician
Cambrian
Pre-Cambrian
Contact
Fault
Scale
0 10 20 mi.
43°
43°
GENERALIZED GEOLOGIC MAP OF VERMONT
73°
72°

Figure 17.

East Middlebury area. The Bomoseen, although less metamorphosed, is remarkably similar to the upper part of the Monastery and is probably equivalent to the Cheshire quartzite. The Schodack represents the graphitic quartz-muscovite schist of the Granville and Ottauquechee formations, and the Wallace Ledge and the Zion Hill probably, in part, are equivalent to the Stowe formation. Fowler (1949) reports an unconformity separating the Zion Hill quartzite from the overlying "undoubted Ordovician slate." Prindle and Knopf (1932) recognize a disconformity at a similar stratigraphic position in the Taconic quadrangle, and they, therefore, believe that the Middle and Upper Cambrian are missing or are very thin. The overlying slates are of Lower and Middle Ordovician age. The Rhysdorf conglomerate may represent the horizon of the Middle Ordovician unconformity.

Structurally the Rochester-East Middlebury area lies on the northern limit of the known outcrop of Pre-Cambrian rocks in the Green Mountain anticlinorium. This anticlinorium extends the length of Vermont, but Pre-Cambrian rocks are exposed at its core only from the vicinity of the Rochester-East Middlebury area to the southern border of the state (Figure 17). The Green Mountain anticlinorium is overturned and in the northern part of Vermont is overthrust toward the west.

The Taconic sequence is reported to be a klippe that has been thrust westward presumably from the Green Mountain anticlinorium. However, the root zone has not been found within the limits of the Rochester-East Middlebury area. If the thrust does root in the Green Mountain anticlinorium, the thrust plane has been removed from the Green Mountain area by erosion, and the frontal part of the nappe remains as an erosional remnant (Figure 19).

GEOLOGIC HISTORY

The geologic record of the Pre-Cambrian is obscure, but most of these rocks appear to be of sedimentary origin. These sediments were folded, metamorphosed and eroded by the beginning of Cambrian time. The eastern portion of this Pre-Cambrian erosion surface may have been inundated in Late Pre-Cambrian time and some sediments derived from the east may have accumulated in this sea.

In Cambrian time the entire area was submerged except for a low landmass in the vicinity of the Adirondack Mountains and a larger landmass somewhere to the east of the Green Mountains. The low-

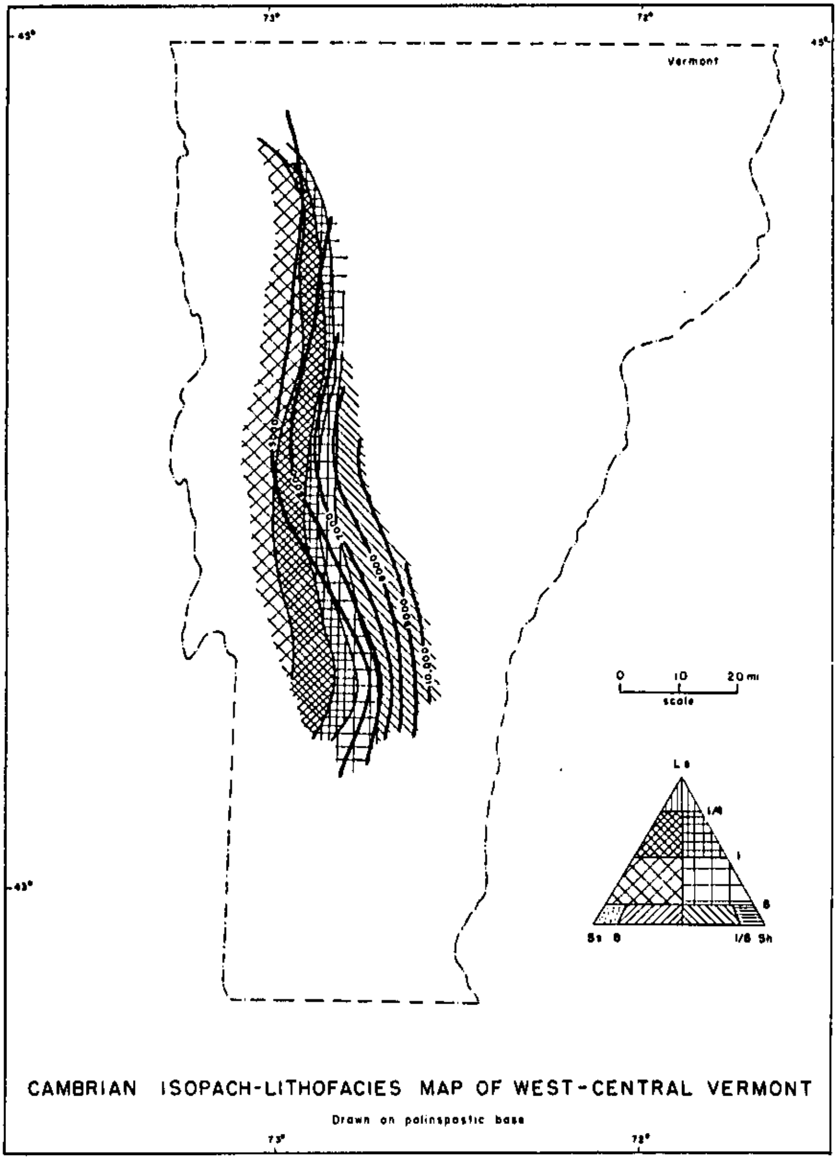
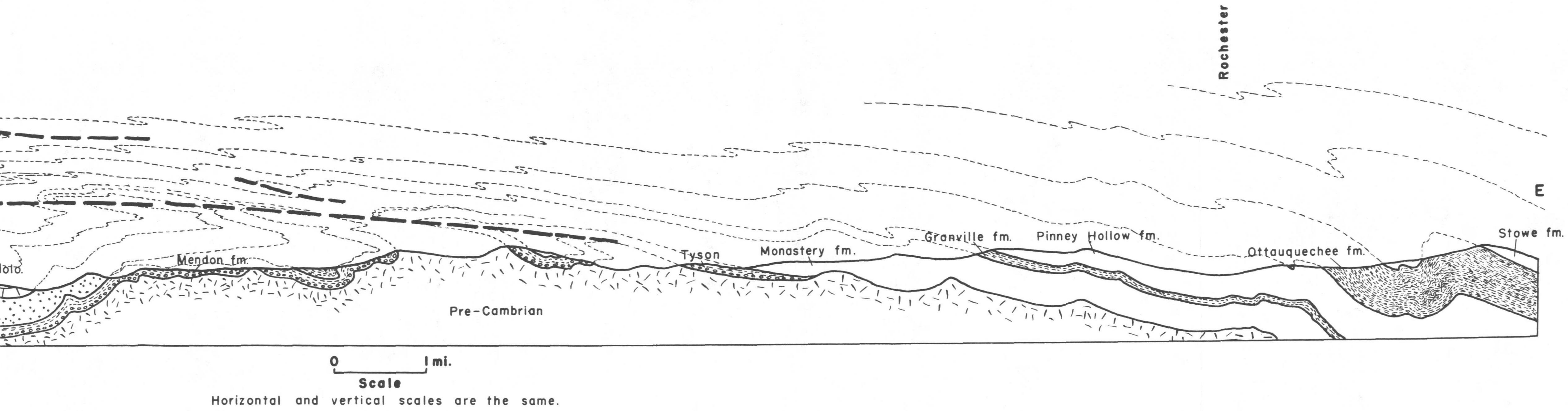

73°
72°
45°
43°
Vermont
0 10 20 mi
scale
Ls
CAMBRIAN ISOPACH-LITHOFACIES MAP OF WEST-CENTRAL VERMONT
Drawn on palinspastic base

Figure 18.



REEN MOUNTAINS SHOWING PROBABLE POSITION OF TACONIC THRUST

(Western portion from Cady, 1945)

lying landmass in the vicinity of the present Adirondack Mountains supplied only limited amounts of detritus, and these were readily winnowed by waves and currents to form quartzose sandstones. The environmental conditions in the area west of the Green Mountains promoted the deposition of thick limestones. The eastern landmass, on the other hand, was undergoing vigorous erosion and supplied large quantities of sediments to the area east of the present Green Mountains. The sediments in this region were mostly aluminous shales, although some of these shales have compositions that suggest a volcanic origin. These shales interfinger with the western quartzose sandstones and limestones in the vicinity of the Green Mountains. This zone of interfingering may have been a factor in localizing later folding.

The inferred sedimentary relations for the Cambrian are shown on an isopach-lithofacies map in Figure 18. The map is constructed on a palinspastic base, but because the shortening of the rock-mass by folding is in some measure compensated by stretching of the rock-mass as indicated by stretched cobbles, the configuration of Vermont is only slightly changed.

In the Cambrian the conditions of sedimentation were characterized by a series of off-lap and on-lap relations. The sequence to the west was thin and consisted of limestones and quartzose sandstones which interfingered to the east with a thick sequence of aluminous shales. In a general way, the limestone-quartzose sandstone sequence is separated from the aluminous shale sequence by a facies of carbonaceous shales. Some pyroclastic beds were associated with the aluminous shale.

Essentially similar conditions continued into the Lower and Middle Ordovician. However, in Mid-Trenton time the Green Mountain region was buckled upward into a broad arch. The crest of this arch probably was subjected to extensive erosion. The unconformity, however, becomes less pronounced toward the flanks of the arch. In late Middle Ordovician time this erosion surface was inundated and covered by several thousands of feet of sediment.

At the end of Ordovician time the Green Mountain area was once again disturbed by orogenic processes. In this disturbance the rocks were again folded, but this time the folding occurred on a large scale, and the rocks were arched into a large anticlinorium. This anticlinorium was overturned toward the west. The rocks of this anticlinorium were pushed westward by distributed movement on closely-spaced foliation planes.

The Taconic klippe can be accounted for by continued deformation causing a thrust plane to develop near the arch-bend of the fold. The upper thrust plate could have moved westward a distance of about seven miles, and the resulting nappe could have formed a large, continuous sheet which has subsequently been eroded to form the Taconic Mountain range (Figure 19).

Proof of the Ordovician orogeny does not occur within the confines of the Rochester-East Middlebury area, and one is forced to look afield before evidence concerning the age of the orogeny is found. However, the Taconic thrust (?) involves rocks of Middle Ordovician age, and the overthrust rocks are overlain by Upper Silurian rocks at Becraft Mountain, New York (Ruedemann, 1942). Thus, the thrusting must be younger than the Middle Ordovician and older than the Upper Silurian. Moreover, rocks which are essentially in the same depositional and orogenic belt are found in New Jersey (Bayley, Kummel, and Salisbury, 1914). There, a well defined Late Ordovician disturbance is indicated with Silurian conglomerate and sandstone resting unconformably on Pre-Cambrian, Cambrian, and Ordovician rocks.

A younger orogeny is also suggested by the regional relations. Relatively undeformed Silurian limestones overlay Ordovician rocks in New York, and a relatively thin calcareous formation of Silurian age overlies the Ordovician rocks in New Hampshire. If these two rock units were originally coextensive, the surface between must have been of low relief at the time of deposition. Because the area between has been differentially elevated above the two areas of Silurian rocks, the Green Mountain region has suffered a younger orogeny—possibly Acadian. The fact that the Taconic mass has been folded with the Middlebury synclinorium tends to corroborate this later folding.

The minor structural features mapped in the Rochester-East Middlebury area and the metamorphism may have been produced in either the Taconic or Acadian orogeny, because the relative tectonic affects of the two orogenies are not known in this area.

BIBLIOGRAPHY

Anderson, E. M., (1948) On lineation and petrofabric structure, and the shearing movement by which they have been produced, Quart. Jour. of the Geol. Soc. of London, vol. CIV, Pt. 1, pp. 99–125.

Bain, G. W., (1926) Geologic history of the Green Mountain front, Rept. of Vt. State Geol., pp. 222–241.

———, (1931) Flowage folding, Amer. Jour. Sci. (5), vol. 22, pp. 503–530.

———, (1936) Serpentization of Vermont ultrabasics, Geol. Soc. Amer. Bull., vol. 47, pp. 1961–1980.

Balk, Robert, (1936) Structural and petrologic studies in Dutchess County, N. Y. Part I. Geologic structures of sedimentary rocks, Geol. Soc. Amer. Bull., vol. 47, pp. 685–774.

———, (1946) Gneiss dome at Shelburne Falls, Mass., Geol. Soc. Amer. Bull., vol. 57, pp. 125–160.

Barrow, G., (1893) On the intrusion of muscovite-biotite gneiss in the south-east Highlands of Scotland, Geol. Soc. London, Quart. Jour., vol. 49, pp. 330–358.

Barth, T. F. W., (1936) Structural and petrologic studies in Dutchess County, N. Y. Part II. Petrology and metamorphism of the Paleozoic rocks, Geol. Soc. Amer. Bull., vol. 47, pp. 775–850.

Bayley, W. S., Kummel, H. B., and Salisbury, R. D., (1914) Description of the Raritan quadrangle, U. S. Geol. Survey Geol. Atlas Folio 191.

Becke, F., (1924) Struktur und Klüftung, Fort. Miner. Band 9, 196 pp.

Billings, M. P., (1937) Regional metamorphism of the Littleton-Moosilauke area, N. H., Geol. Soc. Amer. Bull., vol. 48, pp. 463–566.

———, (1942) Structural Geology, Prentice-Hall, New York, N. Y., 473 pp.

Billings, M. P., and White, W. S., (1950) Metamorphosed mafic dikes of the Woodsville quadrangle, Vt.–N. H., Amer. Miner., vol. 35 (Larsen Volume), pp. 629–643.

Booth, V. H., (1950) Stratigraphy and structure of the Oak Hill succession in Vt., Geol. Soc. Amer. Bull., vol. 61, pp. 1131–1168.

Bowen, N. L., and Tuttle, O. F., (1949) The system $MgO - SiO_2 - H_2O$, Geol. Soc. Amer. Bull., vol. 60, pp. 434–460.

Brainerd, E., and Seely, H. M., (1890) The Calciferous formation in the Champlain Valley, Amer. Mus. Nat. Hist. Bull., vol. 3, pp. 1–23.

Broughton, J. G., (1946) An example of the development of cleavages, Jour. Geol., vol. LIV, pp. 1–18.

Cady, W. M., (1945) Stratigraphy and structure of west-central Vermont, Geol. Soc. Amer. Bull., vol. 56, pp. 515–558.

———, (1950) Fossil cup corals from the metamorphic rocks of central Vermont, Amer. Jour. Sci., vol. 248, pp. 488–497.

Chang, P. H., (1950) Structure and metamorphism of the Bridgewater-Woodstock area, Vt., Ph.D. Thesis, Harvard Univ., 76 pp.

Chidester, A. H., Billings, M. P., and Cady, W. M., (1951) Talc investigations in Vermont, Preliminary Report, U. S. Geol. Survey, Circular 95, 33 pp.

Clark, T. H., (1934) Structure and stratigraphy of southern Quebec, Geol. Soc. Amer. Bull., vol. 45, pp. 1–20.

———, (1936) A lower Cambrian series from southern Quebec, Royal Can. Inst. Trans., vol. 21, Pt. 1, pp. 135–515.

Clark, T. H., and Fairbairn, H. W., (1936) The Bolton igneous group of southern Quebec, Royal Soc. Can. Trans., (3), Sect. IV, vol. 30, pp. 13–18.

Cloos, Ernst, (1937) The application of recent structural methods in the interpretation of the crystalline rocks of Maryland, Md. Geol. Survey, vol. 13, pp. 27–105.

———, (1946) Lineation: a critical review and annotated bibliography, Geol. Soc. Amer. Memoir 18, 122 pp.

———, (1947) Oolite deformation in the South Mountain fold, Md., Geol. Soc. Amer. Bull., vol. 58, pp. 843–918.

Cooke, H. C., (1937) Thetford, Disraeli and eastern half of Warwick map-areas, Quebec, Can. Geol. Sur. Mem. 211, 160 pp.

Currier, L. W., and Jahns, R. H., (1941) Ordovician stratigraphy of central Vermont, Geol. Soc. Amer. Bull., vol. 52, pp. 1487–1512.

Dale, T. N., (1892) On the structure and age of the Stockbridge limestone in the Vermont valley, Geol. Soc. Amer. Bull., vol. 3, pp. 514–519.

———, (1894) On the structure of the ridge between the Taconic and Green Mountain ranges in Vermont, U. S. Geol. Survey, 14th Ann. Rept., Pt. 2, pp. 525–549.

———, (1896) Structural details in the Green Mtn. region and in eastern New York, U. S. Geol. Survey, 16th Ann. Rept., Pt. 1, pp. 543–570.

———, (1899) The slate belt of eastern New York and western Vermont, U. S. Geol. Survey, 19th Ann. Rept., Pt. 3, pp. 159–306.

———, (1904) The geology of the north end of the Taconic range, Amer. Jour. Sci. (4), vol. 17, pp. 185–190.

———, (1904) Geology of the Hudson Valley between the Hoosick and the Kinderhook, U. S. Geol. Survey, Bull. 242, 63 pp.

———, (1910) The Cambrian conglomerate of Ripton in Vermont, Amer. Jour. Sci. (4), vol. 30, pp. 267–270.

———, (1912) The commercial marbles of western Vermont, U. S. Geol. Survey, Bull. 521, 170 pp.

———, (1915) The calcite marble and dolomite of eastern Vermont, U. S. Geol. Survey, Bull. 589, 67 pp.

———, (1916) Algonkian-Cambrian boundary east of the Green Mountain axis in Vermont, Amer. Jour. Sci. (4), vol. 42, pp. 120–124.

———, (1920) Notes on the areal and structural geology of a portion of the west flank of the Green Mountain range, Vermont State Geol., pp. 43–56.

Dapples, E. C., Krumbein, W. C., and Sloss, L. L., (1948) Tectonic control of lithologic associations, A. A. P. G., Bull., vol. 32, pp. 1924–1947.

Doll, C. G., (1943) A paleozoic revision in Vermont, Am. Jour. Sci., vol. 241, pp. 57–64.

———, (1944) A preliminary report on the geology of the Strafford quadrangle, Vt., Vermont State Geol., 24th Rept., pp. 14–28.

Emerson, B. K., (1892) Outlines on the geology of the Green Mountain region in Mass., U. S. Geol. Survey, Geol. Atlas, Hawley Sheet.

———, (1898) Geology of Old Hampshire County, Mass., U. S. Geol. Survey, Mon. 29, 790 pp.

Eric, J. H., (1942) Geology of the Vermont portion of the Littleton quadrangle, Ph.D. Thesis, Harvard Univ., 99 pp.

Eskola, P. E., (1920) The mineral facies of rocks, Norsk. Geol. Tudsskr., vol. 6, pp. 143–194.

———, (1949) The problem of mantled gneiss domes, Quart. Jour. Geol. Soc. London, vol. 104, Pt. 4, pp. 461–476.

Fairbairn, H. W., (1935) Notes on the mechanics of rock foliation, Jour. Geol., vol. 43, pp. 591–608.

———, (1936) Elongation in deformed rocks, Jour. Geol., vol. 44, pp. 670–680.

———, (1949) Structural Petrology of Deformed Rocks, Addison-Wesley Press, Inc., 344 pp.

Fowler, P., (1950) Stratigraphy and structure of the Castleton area, Vt., Vt. Geol. Survey, Bull. 2, 82 pp.

Foye, W. C., (1919) A report on the geological work within the Rochester, Vt. quadrangle, Vermont State Geol., 11th Rept., pp. 76–98.

Foyles, E. J., (1928) Rock-correlation studies in west-central Vermont, Vermont State Geol., 16th Rept., pp. 281–284.

———, (1930) The geology of East Mountain, Mendon, Vt., Vermont State Geol., 17th Rept., pp. 238–251.

Gillson, J. L., (1927) Origin of the Vermont talc deposits, Econ. Geol., vol. 22, pp. 246–287.

Gordon, C. E., (1920) Studies in the geology of western Vermont, Vermont State Geol., 12th Rept., pp. 188–200.

Hadley, J. B., (1942) Stratigraphy, structure and petrology of Mount Cube area, N. H., Geol. Soc. Amer. Bull., vol. 53, pp. 113–176.

———, (1950) Geology of the Bradford-Thetford area, Orange County, Vt., Vt. Geol. Survey, Bull. 1, 36 pp.

Harker, A., (1932) Metamorphism, London, Methuen & Co., Ltd., 362 pp.

Hawkes, H. E., (1941) Roots of the Taconic fault in west-central Vermont, Geol. Soc. Amer. Bull., vol. 52, pp. 649–660.

Hess, H. H., (1933) The problem of serpentinization and the origin of certain chrysotile asbestos, talc, and soapstone deposits, Econ. Geol., vol. 28, pp. 634–657.

Hitchcock, E., Hitchcock, E. J., Hager, A. D., Hitchcock, C. H., (1861) Report on the Geology of Vermont, Claremont, N. H., 982 pp.

Kaiser, E. P., (1945) Northern end of the Taconic thrust-sheet in western Vermont, Geol. Soc. Amer. Bull., vol. 56, pp. 1079–1098.

Kay, M., (1947) Geological nomenclature and the craton, A. A. P. G., vol. 31, pp. 1289–93.

Keith, A., (1923) Cambrian succession of northwestern Vermont, Amer. Jour. Sci. (5), vol. 5, pp. 97–139.

———, (1932) Stratigraphy and structure of northwestern Vermont, Wash. Acad. Sci. Jour., vol. 22, pp. 357–379, 393–406.

Kindle, C. H., and Tasch, P., (1948) Lower Cambrian fauna of the Monkton formation of Vt., Can. Field Nat., vol. 62, 133 pp.

KRUGER, F. C., (1946) Structure and metamorphism of the Bellows Falls quadrangle of N. H. and Vt., Geol. Soc. Amer. Bull., vol. 57, pp. 161–206.

KRUMBEIN, W. C., (1948) Lithofacies maps and regional sedimentary-stratigraphic analysis, A. A. P. G., vol. 32, pp. 1909–1923.

KRYNINE, P. D., (1948) The megascopic study and field classification of sedimentary rocks, The Penn. State Coll. Min. Ind. Ex. Sta. Tech. Paper 130, pp. 130–165.

LEITH, C. K., (1905) Rock cleavage, U. S. Geol. Survey Bull. 239, 216 pp.

LOVERING, T. S., (1928) The fracturing of incompetent beds, Jour. Geol., vol. 36, pp. 709–712.

MACKAY, B. R., (1921) Beauceville map-area, Quebec, Canada Geol. Surv. Mem. 127, 105 pp.

MEAD, W. J., (1940) Folding, rock flowage and foliate structures, Jour. Geol., vol. 48, pp. 1007–1021.

MOORE, G. E., (1949) Structure and metamorphism of the Keene-Brattleboro area, New Hampshire-Vermont, Geol. Soc. Amer. Bull., vol. 60, pp. 1613–1670.

OFTEDAHL, C., (1948) Deformation of quartz conglomerates in central Norway, Jour. Geol., vol. 56, pp. 476–487.

PERRY, R. L., (1928) The geology of Bridgewater and Plymouth townships, Vermont, Vt. State Geol., 16th Rept., pp. 1–64.

PETTIJOHN, F. J., (1943) Archean sedimentation, Geol. Soc. Amer. Bull., vol. 54, pp. 925–972.

———, (1949) Sedimentary rocks, Harper and Bros., New York, 525 pp.

PHILLIPS, F. C., (1947) Lineation in the Northwest Highlands of Scotland, Geol. Mag., vol. 84, pp. 58–59.

PRINDLE, L. M., and KNOPF, E. B., (1932) Geology of the Taconic quadrangle, Am. Jour. Sci. (5), vol. 29, pp. 257–302.

PUMPELLY, R., WOLFF, J. E., and DALE, T. N., (1894) Geology of the Green Mountains in Massachusetts, U. S. Geol. Surv. Mon. 23, 206 pp.

RESSER, C. E., and HOWELL, B. F., (1938) Lower Cambrian Olenellus zone of the Appalachians, Geol. Soc. Amer. Bull., vol. 49, pp. 195–248.

RICHARDSON, C. H., (1902) The terranes of Orange County, Vermont, Vt. State Geol., 3rd Rept., pp. 61–101.

RICHARDSON, C. H., and CAMP, S. H., (1918) The terranes of Northfield, Vt., Vt. State Geol., 11th Rept., pp. 99–119.

RICHARDSON, C. H., (1918) The terranes of Roxbury, Vt., Vt. State Geol., 11th Rept., pp. 120–140.

———, (1920) Geology and mineralogy of Braintree, Vt., Vt. State Geol., 12th Rept., pp. 56–75.

———, (1926) Geology and petrography of Barnard, Pomfret, and Woodstock, Vt., Vt. State Geol., 15th Rept., pp. 127–159.

RODGERS, JOHN, (1937) Stratigraphy and structure in the Upper Champlain Valley, Geol. Soc. Amer. Bull., vol. 48, pp. 1573–1588.

RUEDEMANN, R., (1930) Geology of the Capital district, N. Y. State Mus. Bull. 285, 218 pp.

RUEDEMANN, R., and CUSHING, H. P., (1914) Geology of Saratoga Springs and vicinity, N. Y. State Mus. Bull. 169, 177 pp.

RUEDEMANN, R., (1942) Geology of the Catskill and Kaaterskill quadrangles, N. Y. State Mus. Bull. 331, 251 pp.

SCHMIDT, W., (1918) Bewegungsspuren in Porphyroblasten, Sitz. Kaiserl. Akad. Wiss. Wien Math.-nat. Kl. Abt. 1, 126 pp.

SCHUCHERT, C., (1937) Cambrian and Ordovician of northwestern Vermont, Geol. Soc. Amer. Bull., vol. 48, pp. 1001–1078.

SEELY, H. M., (1910) Preliminary report of the geology of Addison County, Vt. State Geol., 7th Rept., pp. 257–313.

SHAW, A. B., (1949) Stratigraphy and structure of the St. Albans area, Vt., Ph.D. Thesis, Harvard Univ.

SPENCER, A. C., KÜMMELL, H. B., WOLFF, J. E., SALISBURY, R. D., and PALACHE, C., (1908) Franklin Furnace quadrangle, U. S. Geol. Survey, Atlas Folio 161, 27 pp.

STONE, SOLON, (1951) Geology of the Milton, Vt. area, Ph.D. Thesis, Harvard Univ.

SWANSON, C. O., (1927) Notes on stress, strain, and joints, Jour. Geol., vol. 35, pp. 193–223.

THOMPSON, J. B., (1950) A gneiss dome in southeastern Vermont, Ph.D. Thesis, M.I.T., 149 pp.

TILLEY, C. B., (1926) Some mineralogical transformations in crystalline schists, Min. Mag., vol. 21, pp. 34–46.

TOLMAN, C., (1936) Lake Etchemin map-area, Quebec, Can. Geol. Surv. Mem. 199, 20 pp.

TURNER, F. J., (1948) Evolution of the metamorphic rocks, Geol. Soc. Amer. Mem. 30, 342 pp.

VAN HISE, C. R., (1896) Principles of North American Pre-Cambrian geology, U. S. Geol. Survey, 16th Ann. Rept., pp. 571–874.

WALCOTT, C. D., (1888) The Taconic system of Emmons, and the name in geologic nomenclature, Amer. Jour. Sci. (3), vol. 35, pp. 229–242, 307–327, 394–401.

WEISS, J., (1949) Wissahickon schist at Philadelphia, Pa., Geol. Soc. Amer. Bull., vol. 60, pp. 1689–1726.

WHITE, W. S., and BILLINGS, M. P., (1951) Geology of the Woodsville quadrangle, Vt.-N. H., Geol. Soc. Amer. Bull., vol. 62, pp. 647–696.

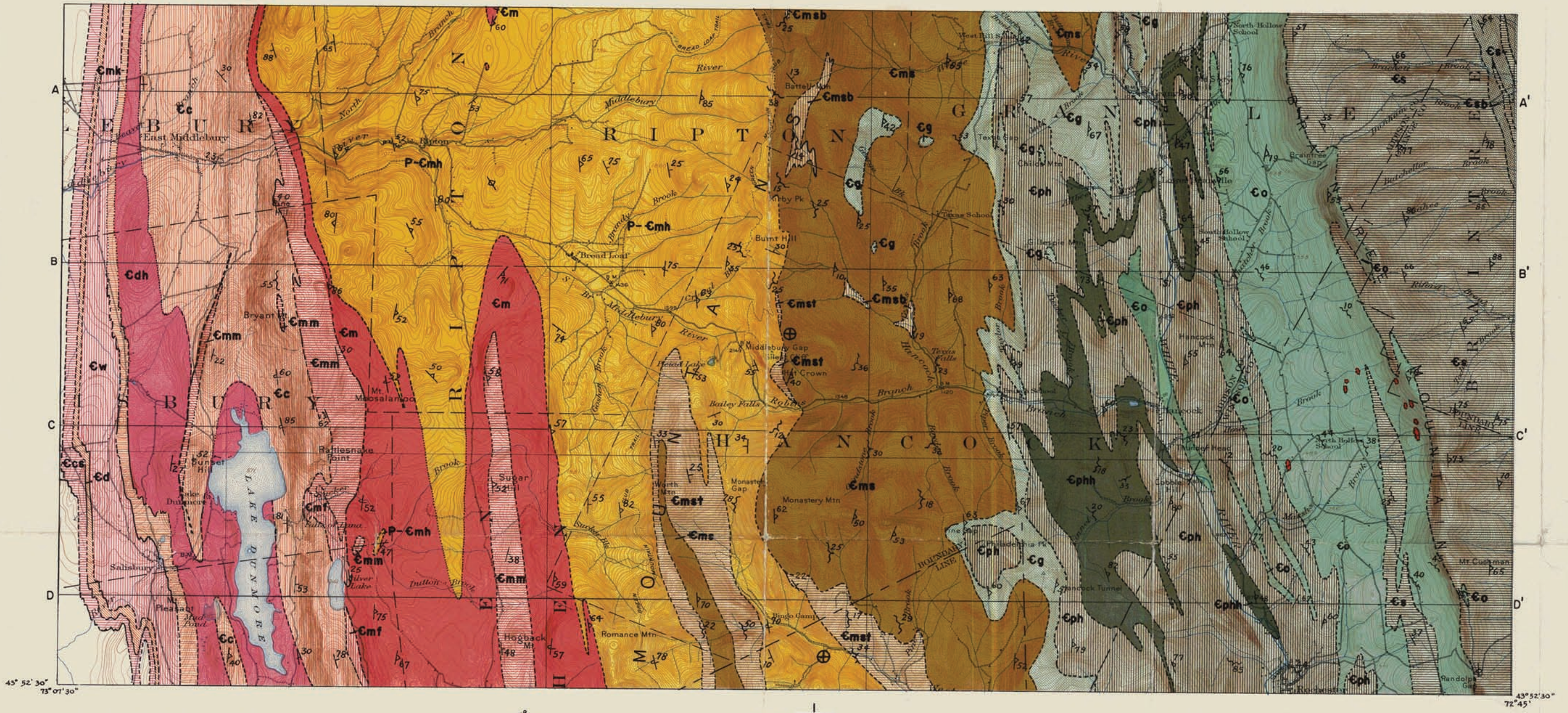
WHITE, W. S., and JAHNS, R. H., (1950) Structure of central and east-central Vermont, Jour. Geol., vol. 58, pp. 179–220.

WHITTLE, C. L., (1894) The general structure of the main axis of the Green Mountains, Amer. Jour. Sci. (3), vol. 47, pp. 347–355.

———, (1894) The occurrence of Algonkian rocks in Vermont and the evidence for their subdivision, Jour. Geol., vol. 2, pp. 396–429.

WOLFF, J. E., (1891) On the Lower Cambrian age of the Stockbridge limestone, Geol. Soc. Amer. Bull., vol. 2, pp. 331–337.

———, (1891) Metamorphism of clastic feldspar in conglomerate schist, Bull. M.C.Z., Harvard Univ., vol. XVI, no. 10, pp. 173–183.



Topography by U.S.G.S. 1902 and 1915
Control by U.S. Coast and Geodetic Survey.

APPROXIMATE MEAN DECLINATION 1944

Scale $\frac{1}{62500}$

1 1/2 0 1 2 3 4 Miles

Contour interval 20 feet

Datum is mean sea level

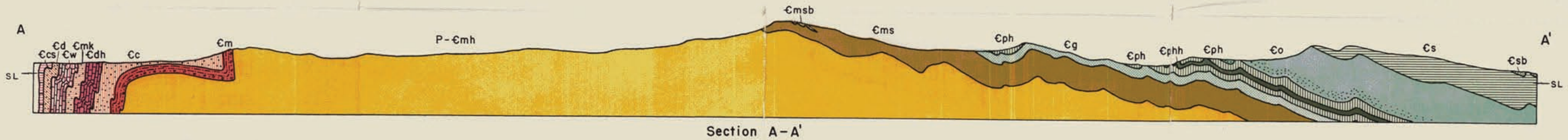
Cady	Osberg

Geology by Philip H. Osberg, assisted by Bruce Nelson and Bruce Bryant. Western portion modified from Wallace M. Cady. Geology surveyed in 1948, 1949, and 1950.

GEOLOGIC MAP OF THE EAST MIDDLEBURY-ROCHESTER AREA, VT.

STRUCTURE SECTIONS

Horizontal and vertical scales the same.



Section A-A'

LEGEND

SEDIMENTARY AND METAMORPHIC ROCKS

Western Sequence | **Eastern Sequence**

Middle Ordo.(?)

Ultramafic rocks (U)
(Serpentine, talc, carbonate)

Upper Cambrian

Western Sequence:

€cs — **Clarendon Springs dolomite**
(Massive gray dolomite)

€d — **Danby formation**
(Interbedded gray quartzite and dolomite)

Eastern Sequence:

€sb, €s — **Stowe formation**
(Quartz-chlorite-muscovite schist. Brackett member, €sb, albite-epidote-carbonate-chlorite schist.)

Lower Cambrian

Western Sequence:

€w — **Winooski dolomite**
(Four to twelve inch beds of dolomite separated by thin dark siliceous partings)

€mk — **Monkton quartzite**
(Red quartzite interbedded with pink to gray dolomite)

€dh — **Dunham dolomite**
(Buff weathering siliceous dolomite)

€c — **Cheshire quartzite**
(Massive white quartzite, twelve to eight inch bedded brown quartzite, thin bedded graphitic quartzite)

€mm, €m, €mf — **Mendon formation**
(Conglomerate, albite-quartz-biotite-muscovite schist. Forestdale member, €mf, white to gray dolomite. Moosalamoo member, €mm, graphitic quartz-muscovite schist)

Eastern Sequence:

€o — **Ottauquechee formation**
(Graphitic quartz-muscovite schist, gray quartzite)

€ph, €phh — **Pinney Hollow formation**
(Albite-quartz-chlorite-muscovite schist. Hancock member, €phh, albite-epidote-carbonate-chlorite schist.)

€g — **Granville formation**
(Graphitic quartz-muscovite schist)

€ms, €msb, €mst — **Monastery formation**
(Albite-quartz-chlorite-muscovite schist. Tyson member, €mst, albite-quartz-biotite-muscovite schist, conglomerate. Battell member, €msb, graphitic quartz-muscovite schist.)

UNCONFORMITY

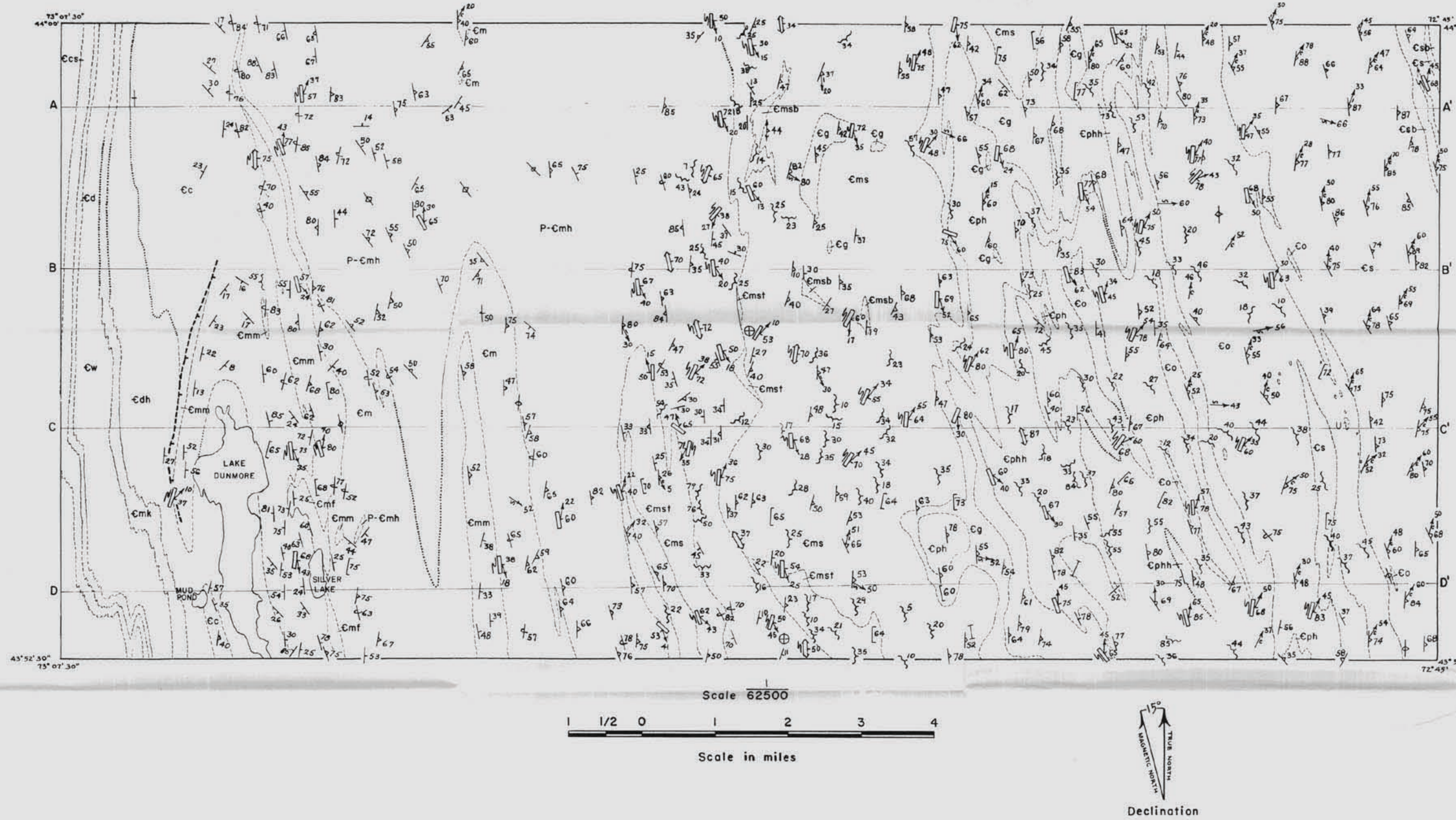
Pre-Cambrian

P-€mh — **Mount Holly complex**
Microcline-quartz-biotite-muscovite gneiss, albite-quartz-chlorite-muscovite schist, epidote amphibolite)

STRUCTURAL SYMBOLS

- Accurate contact
- Approximate contact
- Indefinite or concealed contact
- Thrust fault
- Bedding
- Bedding where vertical
- Bedding where horizontal
- Bedding where overturned
- Generalized bedding of plicated beds
- Foliation
- Foliation where vertical

Plate I.
VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist



LEGEND

STRUCTURAL SYMBOLS

- Accurate contact
- Approximate contact
- Indefinite or concealed contact
- Thrust fault
- Bedding
- Bedding where vertical
- Bedding where horizontal
- Bedding where overturned
- Generalized strike and dip of plicated beds
- Foliation
- Foliation where vertical
- Foliation and lineation
- Foliation and horizontal lineation
- Lineation on crinkles
- Lineation on streaming or slickensides
- Lineation on intersection of cleavage and bedding
- Fold - strike and dip of axial plane and horizontal projection and value of plunge
- Fold - strike and dip of axial plane with horizontal plunge
- Plan of fold
- Slip cleavage
- Slip cleavage where vertical

FORMATION SYMBOLS

WESTERN SEQUENCE

- Єcs - Clarendon Springs dolomite
- Єd - Danby formation
- Єw - Winooski dolomite
- Єmk - Monkton quartzite
- Єdh - Dunham dolomite
- Єc - Cheshire quartzite
- Єmm - Moosalamoo member of the Mendon formation
- Єmm - Forestdale member of the Mendon formation
- Єm - Mendon formation

EASTERN SEQUENCE

- Єsb - Brackett member of the Stowe formation
- Єs - Stowe formation
- Єo - Ottauquechee formation
- Єph - Pinney Hollow formation
- Єphh - Hancock member of the Pinney Hollow formation
- Єg - Granville formation
- Єmsb - Battell member of the Monastery formation
- Єmst - Tyson member of the Monastery formation
- Єms - Monastery formation

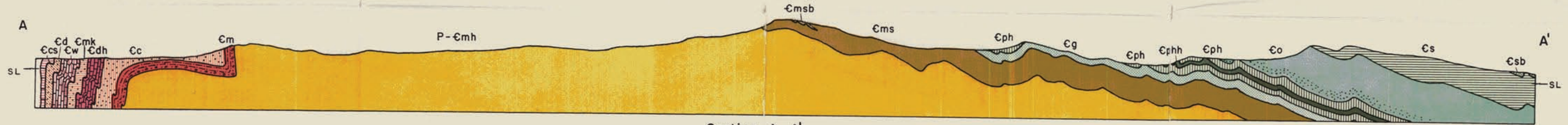
P-Єmh - Mount Holly complex

TECTONIC MAP OF THE EAST MIDDLEBURY-ROCHESTER AREA

BASED ON REPRESENTATIVE DATA

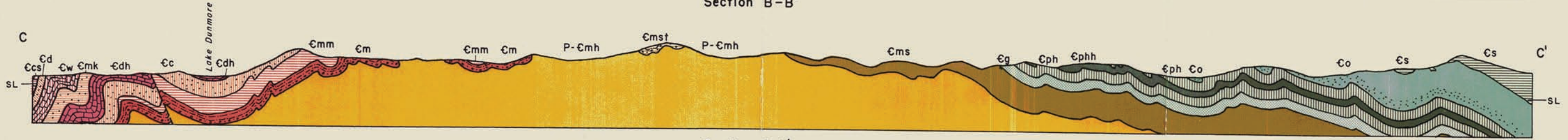
STRUCTURE SECTIONS

Horizontal and vertical scales the same.

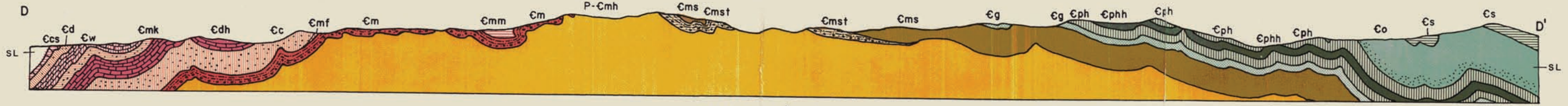


Section A–A'

Section B–B'



Section C–C'



Section D–D'