UNITED STATES DEPARTMENT OF THE INTERIOR

GEOLOGICAL SURVEY

GEOLOGY OF THE WINDSOR QUADRANGLE, MASSACHUSETTS

By

Stephen Allen Norton

This report is preliminary and
has not been edited or reviewed
for conformity with Geological
Survey standards or nomenclature.

OPEN-FILE
67-170

CONTENTS

LIST OF FIGURES

LIST OF TABLES

LIST OF PLATES

ABSTRACT

The Windsor quadrangle lies on the boundary between the eugeosynclinal and miogeosynclinal rocks of the Appalachian geosyncline on the western flank of the metamorphic high in western New England.

Precambrian rocks crop out in a north-trending belt in the central part of the quadrangle. They have been classified into 2 formations. The Stamford Granite Gneiss crops out in the eastern half of the Precambrian terrane. It is a microcline-quartz-biotite augen gneiss. Stratified Precambrian rocks (the Hinsdale Gneiss) crop out entirely to the west of the Stamford Granite Gneiss. Thet are predominantly highly metamorphosed felsic gneisses and quartzites with minor calc-silicate rock, amphibolite, and graphitic gneiss.

Eugeosynclinal rocks (the Hoosac Formation and the Rowe Schist), ranging in age from Lower Cambrian to Lower Ordovician, crop out in a north-trending belt east of the Precambrian terrane. They are composed predominantly of albite schist and muscovite-chlorite schist with minor garnet schist, quartz-muscovite-calcite schist, felsic granulite and gneiss, quartzite, greenschist, and carbonaceous phyllite and schist.

West of the Precambrian rocks, the Hoosac Formation is overlain by a miogeosynclinal sequence (the Dalton Formation, Cheshire Quartzite, Kitchen Brook Dolomite, Clarendon Springs Dolomite, Shelburne Marble, and the Bascom Formation) ranging in age from Lower Cambrian to Lower Ordovician. These

rocks are unconformably overlain by the Berkshire Schist of Middle Ordovician age that is composed of carbonaceous schist, phyllite, and quartzite.

The relationships in the zone of transition between the miogeosynclinal and eugeosynclinal rocks are unknown because the rocks of this zone are no longer present. The contact between the eugeosynclinal Hoosac Formation and the Dalton Formation is conformable and apparently represents continuous deposition.

The dominant structure is a large recumbent, north-plunging, west-facing anticline (the Hoosac nappe) with a Precambrian core. The miogeosynclinal rocks are inverted in the northwestern part of the quadrangle and upright in the southwestern part of the quadrangle. A later generation of open, post-metamorphic folds has folded the recumbent folds in the miogeosynclinal rocks.

The eugeosynclinal rocks show 3 phases of folding. The earliest folds are isoclinal, have steep plunges, were syn-metamorphic, and have a strong axial plane schistosity. Two post-metamorphic generations of folds are more open and have axial plane cleavage.

The development of the Hoosac nappe and the isoclinal folds was accompanied by regional metamorphism of the garnet zone. The pressure exceeded the pressure for the triple point of the Al_2SiO_5 polymorphs. The composition of the paragonite coexisting with muscovite suggests a period of retrograde metamorphism for the Paleozoic rocks as well as the Pre-Cambrian rocks that were originally of higher grade (sillimanite?).

Later events include high-angle faulting (Triassic?), erosion, and Pleistocene glaciation.

CHAPTER I

INTRODUCTION

Location

The Windsor quadrangle, in Berkshire County, northwestern Massachusetts, is bounded by north latitudes 40°37'30" and 42°30'00" and by west longitudes 73°07'30" and 73°07'30". The quadrangle is 57 square miles in area and includes portions of the towns of Windsor, Cheshire, Dalton, Savoy, and Adams. It is in the Green Mountain section of the New England-Maritime physiographic province (Lobeck, 1948) in the northern part of the Berkshire Highlands and includes part of the Hoosac Range that extends into southern Vermont.

Topography

Less than one percent of the area is covered with water and less than 5 percent is swamp. The lowest point, less than 800 feet above sea level, is in the vicinity of the Hoosic River is Adams. The highest point, 2,450 feet above sea level, is Borden Mountain in Savoy Mountain State Forest in the northeast part of the quadrangle. The relief in the central part of the area is less than 200 feet, in the western part it is about 1000 feet, and in the eastern part the local relief is as much as 500 feet.

The topography of the area is that of a rejuvenated peneplain carved on Precambrian rocks in a central core flanked by Paleozoic metasedimentary rocks on the west and east. The central area is characterized by low rolling hills with slight relief, the surface being above 1500 feet above sea level, with a drop-off on each side at the contact of the Precambrian and Cambrian rocks.

The drainage divide is approximately a north-south line through the center of the quadrangle. Rivers draining to the northwest, west, and southwest, are part of the Housatonic River drainage basin; those flowing northeast are part of the Deerfield River basin, and those draining east and southeast are part of the Westfield River basin. The Deerfield and Westfield Rivers are tributaries of the Connecticut River.

Extensive kame terrace sand deposits, resulting from Pleistocene glaciation, occur in the extreme northwestern corner of the quadrangle. Elsewhere, the effects of glaciation are less notable. South-facing slopes are steep owing to plucking, whereas north-facing slopes are gentle, have poor exposures, and are covered with a veneer of till.

Method of Study

The bedrock geology of the Windsor quadrangle was mapped in 29 weeks during the summers of 1963, 1964, 1965, and 1966. Mapping was carried out on a 1:24,000 topographic base, with 20 foot contour intervals, published by the United States Geological Survey in 1960. Larger scale maps of extensive outcrop areas with complex structure were made at a scale of 1:1200. Traverses were conducted by pace and compass methods, supplemented with altimeter readings in rough terrain.

Two hundred and fifty thin sections and about 100 metamorphic mineral assemblages were studied in detail. Indices of refraction were determined by oil immersion and the basal spacings of mica were determined by X-ray diffraction methods.

Acknowledgements

The field and laboratory work was financially supported by the United States Geological Survey in cooperation with the Commonwealth of Massachusetts Department of Public Works. The writer was employed by the U. S. Geological Survey during the summers of 1963, 1964, 1965, and 1966, and during the academic year 1966-67 while at Harvard University. The writer was under the direct supervision of Lincoln R. Page, Chief, Branch of Regional Geology of New England, U. S. Geological Survey.

Laboratory and other research facilities were furnished by Harvard University, Department of Geological Sciences. Thin and polished sections were provided by the U. S. Geological Survey.

Field work and laboratory studies were performed under the general supervision of Lincoln R. Page of the U. S. Geological Survey, and Professors Marland P. Billings and James B. Thompson Jr. of the Department of Geological Sciences at Harvard University. They have read this manuscript and their helpful suggestions are greatly appreciated.

The author has benefited from guidance and counsel of Normal L. Hatch, Alfred H. Chidester, and Philip H. Osberg of the U. S. Geological Survey,

Aeromagnetic surveys have been flown, by the U. S. Geological Survey, for all the quadrangles in the northwestern part of Massachusetts, including the Windsor quadrangle. They have been helpful in reconnaissance work and in indicating structural trends where bedrock exposures are poor.

Previous Work

The first significant step toward understanding the complex geology in the vicinity of the Windsor quadrangle was taken in 1832 with the publication of a geologic map with brief notes on the geology of Massachusetts (Hitchcock, 1832). A more complete report by Hitchcock (1833) immediately followed and in 1841 a revised map with expanded text was published.

Emmons (1844) discussed the distribution and significance of Taconic rocks west of the quadrangle.

In 1894, Pumpelly and others published a map with detailed petrologic and structural data for the northern part of the Berkshire Highlands. Their structural interpretation for the quadrangle has been only slightly modified by later work.

Emerson (1898a) described the Paleozoic rocks east of the Precambrian rocks in the quadrangle.

Stratigraphic revisions (Emerson, 1898b and 1899) for the rocks in the western half of the quadrangle allowed minor reinterpretation of the structural and stratigraphic relationships These appeared on the map published by Emerson in 1917.

Dale (1932) described in detail the distribution and petrology of the carbonate units west of the Hoosac Range in

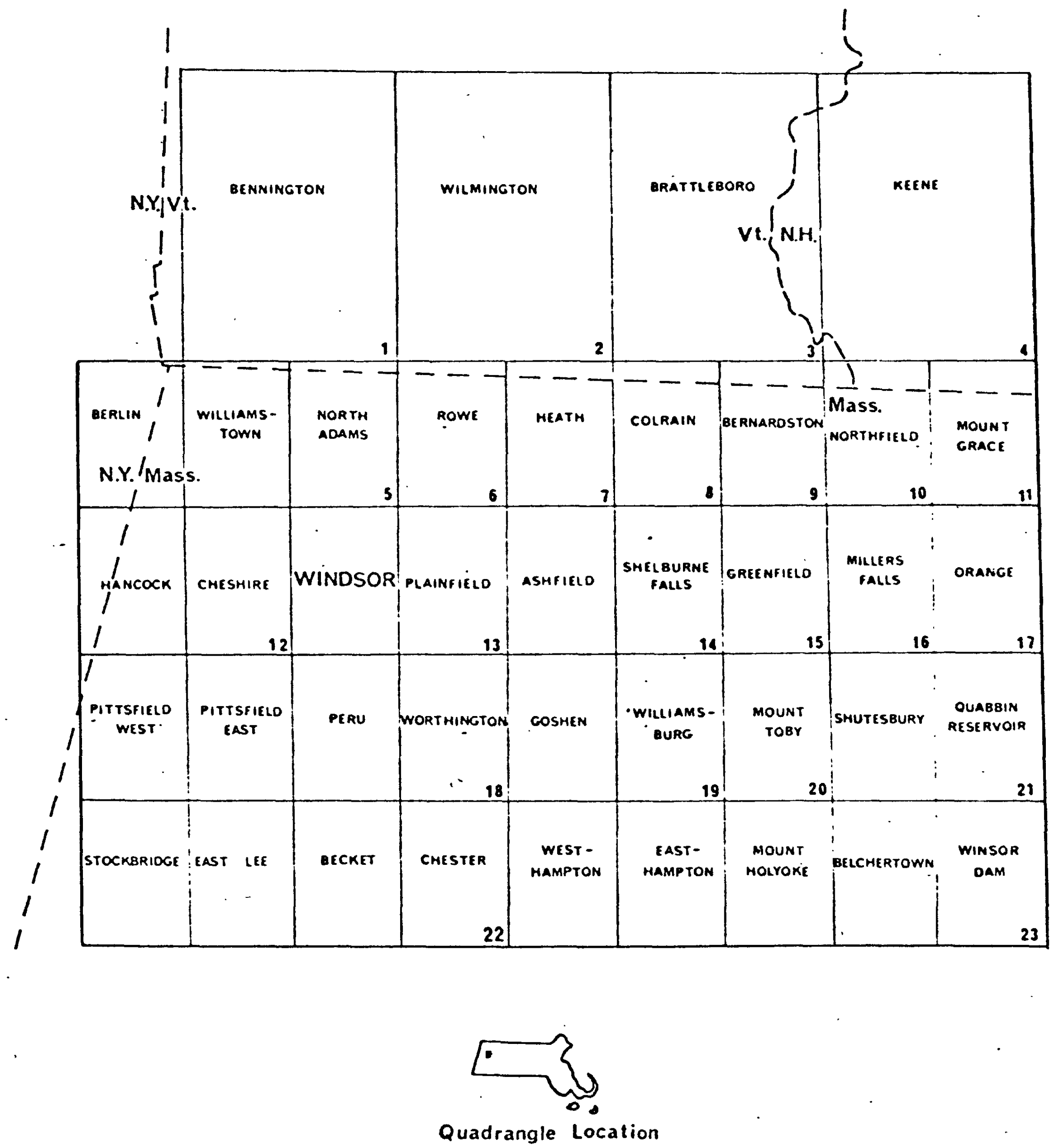
Massachusetts and the Green Mountain anticlinorium in Vermont.

Prindle and Knopf (1932) made the first major attempt to revise the structural and stratigraphic interpretation of the area.

In the early 1950's Herz (1958, 1961) mapped the structure and stratigraphy of the Cheshire and North Adams quadrangles west and north, respectively, of the Windsor quadrangle. Christensen (1963) made a detailed structural analysis of the Hoosac nappe, the major structural feature in the area, and slightly modified the interpretation of Pumpelly and others and Prindle and Knopf.

Recent bedrock mapping has been carried out in the nearby Rowe, Heath, and Plainfield quadrangles by Chidester, A. H., Osberg, P. H., Hatch, N. Jr., and Norton S. A. (figure 1).

Figure 1. Index to recent geologic mapping in northwestern Massachusetts, southern Vermont, and southwestern New Hampshire.



References for recent bedrock mapping in the vicinity of the Windsor quadrangle

Numbers correspond with areas on figure 1.

1. MacFayden, John A., Jr., 1956, The Geology of the Bennington area, Vermont: Vermont Geol. Survey Bull. 7, 72 p.

2. Skehan, James W., S.J., 1961, The Green Mountain anticlinorium in the vicinity of Wilmington and Woodford, Vermont: Vermont Geol. Survey Bull. 17, 159 p.

3,4. Portions of these two quadrangles were mapped by Trask, Newell J., Jr., 1964, Stratigraphy and structure in the Vernon-Chesterfield area, Massachusetts, New Hampshire, and Vermont: unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, 99 p. plus appendix.

5. Herz, Norman, 1961, Bedrock Geology of the North Adams quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ 139.

6. Chidester, Alfred H., and others, in press, Bedrock and Surficial Geology of the Rowe quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ.

7. Hatch, Norman, and Hartshorne, Joseph, in preparation, Geology of the Heath quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ.

8. Segerstrom, Kenneth, 1956, Bedrock geology of the Colrain quadrangle, Massachusetts-Vermont: U. S. Geol. Survey Geol. Quad. Map GQ-86.

9. Balk, Robert, 1956, Bedrock geology of the Massachusetts portion of the Bernardston quadrangle, Massachusetts-Vermont: U. S. Geol. Survey Geol. Quad. Map GQ-90.

Trask, Newell J., Jr., 1964, Stratigraphy and structure in the Vernon-Chesterfield area, Massachusetts, New Hampshire, Vermont: unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, 99 p. plus appendix.

10. Trask, Ibid.

Robinson, Peter, 1963, Gneiss domes of the Orange area, Massachusetts and New Hampshire: unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, V. 1., 253 p.

Balk, Robert, 1956, Bedrock geology of the Massachusetts portion of the Northfield quadrangle, Massachusetts-NewHampshire: U. S. Geol. Survey Geol. Quad. Map GQ-91.

11. Robinson, Ibid.

Hadley, J. B., 1949, Bedrock geology of the Mount Grace quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ-3.

12. Herz, Norman, 1958, Bedrock geology of the Cheshire quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ-108.

13. Osberg, Philip H., and others, in preparation, Bedrock geology of the Plainfield quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ.

14. Segerstrom, Kenneth, 1956, Bedrock Geology of the Shelburne Falls quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ-87.

15. Willard, M. E., 1952, Bedrock geologic map of the Greenfield quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ-20.

16. Balk, Robert, 1956, Bedrock Geology of the Millers Falls quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ-92.

Robinson, Ibid.

17. Robinson, Ibid.

18. Hatch, Norman L., Jr., mapping in progress.

19. Segerstrom, Kenneth, 1956, Bedrock geology of the Williamsburg quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ-85.

20. Willard, M. E., 1951, Bedrock geologic map of the Mount Toby quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ-8.

21. Robinson, Peter, mapping in progress.

22. Norton, Stephen A., 1962, Geology of Chester, Massachusetts with special reference to the Chester Emery Deposits: unpublished A. B. Thesis, Department of Geology, Princeton University, 131 p.

23. Robinson, Ibid.

CHAPTER 2

STRATIGRAPHY AND PETROLOGY

General Statement

The Windsor quadrangle straddles the tectonic axis of the Berkshire Highlands, This axis approximates the boundary between the Cambro-Ordovician quartzite-carbonate sedimentary sequence typical of the miogeosyncline to the west and the graywacke-shale-volcanic rock sequence typical of the eugeosyncline to the east (figure 2). Stratigraphic interpretation is particularly important here for it is possible to trace eastern rocks across the Berkshire Highlands and establish their relationship to the lower part of the western sequence.

About 8,000 feet of Cambro-Ordovician eugeosynclinal sedimentary rocks are present within the Windsor quadrangle. The miogeosynclinal sequence spans the Lower Cambrian to Middle Ordovician epochs and totals 2,000 feet. All the sediments have been subjected to regional metamorphism in the garnet zone.

Because of the distinctly different rock sequences east and west of the Berkshire Highlands, the two stratigraphic sequences will be treated separately.

Precambrian Rocks

General Statement

Precambrian rocks occupy the central part of the quad-

Explanation

Dw	Waits River Formation (Devonian)
DSn	Northfield Formation (Siluro-Devonian)
Oh	Hawley Formation (Ordovician)
Obs	Berkshire Schist and equivalent rocks in southern Vermont (Ordovician)
Om	Moretown Formation (Ordovician)
O€r	Rowe Schist (Cambro-Ordovician)
O€m	Autochthonous miogeosynclinal rocks (Cambro-Ordovician)
O€t	Allochthonous Taconic rocks (Cambro-Ordovician)
€h	Hoosac Formation (Cambrian)
p€	Precambrian rocks

from recent mapping of

A. H. Chidester
N. L. Hatch
N. Herz
S. A. Norton
P. H. Osberg

and from older maps of

T. N. Dale
B. K. Emerson
L. M. Prindle and E. B. Knopf

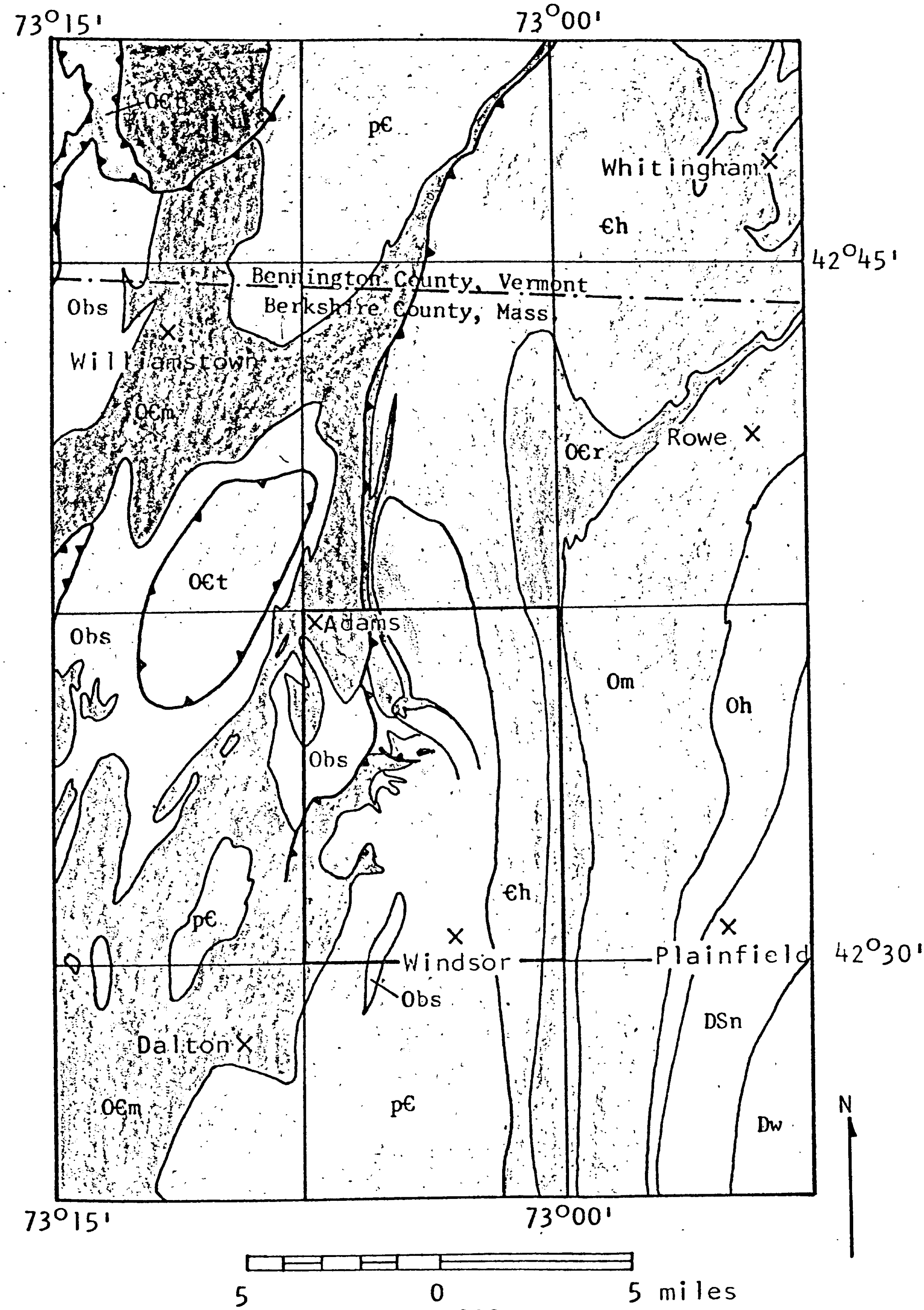


Figure 2. Regional geologic map of southern Vermont and northwestern Massachusetts.

rangle and crop out in a belt about 4 miles wide that approximately bisects the quadrangle along a north-south axis.

Two formations have been distinguished, the Stamford Granite Gneiss and the Hinsdale Gneiss. The former has been mapped as a continuous formation from the north to the south boundary of the quadrangle. It crops out entirely to the east of the Hinsdale Gneiss.

Stamford Granite Gneiss

The Stamford Granite Gneiss of Precambrian age, first defined by Hitchcock (1861, p. 601) at localities in Stamford, Vermont, has been described in the Hoosac Range by Pumpelly and others (1894).

The rocks mapped as Stamford Granite Gneiss in the Windsor quadrangle and immediately to the north in the North Adams quadrangle are not continuous with those of the type locality. The rock is also different and may not be equivalent to the Stamford Granite Gneiss of Hitchcock. The name is here retained to conform with the usage of Herz (1961).

The rock crops out in a broad belt in the north-central part of the quadrangle. It is bordered on the west by the layered Hinsdale Gneiss of Precambrian age (Herz, 1961) and to the east by granulite, gneiss and minor conglomerate at the base of the Hoosac Formation.

Prindle and Knopf (1932, p. 270) suggest that the Stamford Granite Gneiss intrudes the Hinsdale Gneiss. In the Windsor quadrangle the contact between these two units is not exposed and the age relationships are not clear.

The breadth of outcrop decreases from a maximum of 2 miles at the northern border of the quadrangle to less than 300 feet (if indeed present) in the east-central part of the quadrangle. The body appears to thicken to the south, attaining a breadth of outcrop of at least 1,500 feet.

Outcrops are typically rather rusty and show an undulating foliation caused by biotite and rare muscovite wrapping around augen of microcline that commonly exceed 1/2 inch in length. There also are augen of blue quartz except in the southern part of the quadrangle where the rock is strongly granulated. Here most feldspar augen are drawn out to several inches and the blue quartz loses its color except in some of the larger uncrushed cores.

The rock is composed of nearly equal parts of quartz and microcline totaling between 80 and 90 percent of the rock. Quartz may compose as much as 40 percent of the rock. Muscovite and biotite make up essentially the rest of the rock. Accessory minerals are plagioclase, clinozoisite, hematite, magnetite, sphene, chlorite (after biotite), and garnet (only in the southern outcrops), (table 1).

The rock is remarkably uniform throughout the quadrangle, the only difference being the presence of rare garnet in the southern outcrops. The Rapakivi texture that is diagnostic of this rock at the type locality according to Skehan (1961)

TABLE 1

Estimated Modes

	Stamford Granite Gneiss		Hinsdale Gneiss			
Specimen	110	869	37-g	450	1430-a	1540-b
Quartz	40	20	60	45	15	tr
Muscovite	10	2	1	2		
Albite	present	present		36-P	5	tr
Microcline	40-M	70-M	8	11-P		
Chlorite	tr					
Garnet		1	5			
Biotite	5	5	10	2	25	1
Epidote group	5	2	12	2	30	
Diopside					20	
Hornblende						99
Magnetite†			1	1		
Hematite			2			tr
Calcite + dolomite					5	
Zircon			tr			
Sphene		tr				

P = porphyroblastic

M = megacrysts

† = includes ilmenite in all tables except table 17

TABLE 1 - Continued

Hinsdale Gneiss

Specimen	1645	1862	2491	[illegible]58-d	2590
Quartz	40	60	38	10	80
Muscovite	tr		tr		2
Albite			50		10
Microcline	60-P		5		
Chlorite		2			
Garnet					2
Biotite	tr	8	5		5
Epidote group	tr	4	1	15	
Diopside				73	
Hornblende		25			
Tremolite				1	
Apatite	tr	tr			
Zircon			tr		
Sphene		tr			
Pyrite + pyrrhotite		1		1	
Graphite					1

P = porphyroblastic

Location of Specimens[1]

Stamford Granite Gneiss

110	Granulated augen gneiss: intersection of New State Road and Adams Road, (2).[2]
869	Granulated augen gneiss: elevation 2,050 feet on Hume Brook, (9).

Hinsdale Gneiss

37-g	Grayish-brown gneiss: roadcut west of Savoy, at intersection of Main Road and Windsor Road, (6).
450	Light buff gneiss: elevation 2,050 feet, west side of knoll on Harrington Road, (5).
1430-a	Green gneiss: elevation 1,900 feet, N.35°W. from Heat △ 2,061 feet, (4).
1540-b	Massive amphibolite: elevation 1,910 feet, N.73°W. 1,200 feet from Heat △ 2,061 feet, (4).
1645	White, gneissic granulite: elevation 1,600 feet on Weston Brook, (7).
1862	Dark gray granulite: Hill, 2,068 feet, (5).
2491	Gray, albite gneiss: elevation 2,030 feet in brook, 1,900 feet east-northeast of hill X 2,094 feet, (2).
2558-d	Pyrrhotite-diopside granulite: elevation 2,000 feet, southwest side of knob 1,300 feet north-northeast of hill X 2,043 feet, (5).
2590	Grayish-brown equigranular gneiss: elevation 1,882 feet, on Drowned Land Brook, (5).

[1] See figure 3 for location of specimens.

[2] Number in parentheses indicates the ninth of the quadrangle (figure 3).

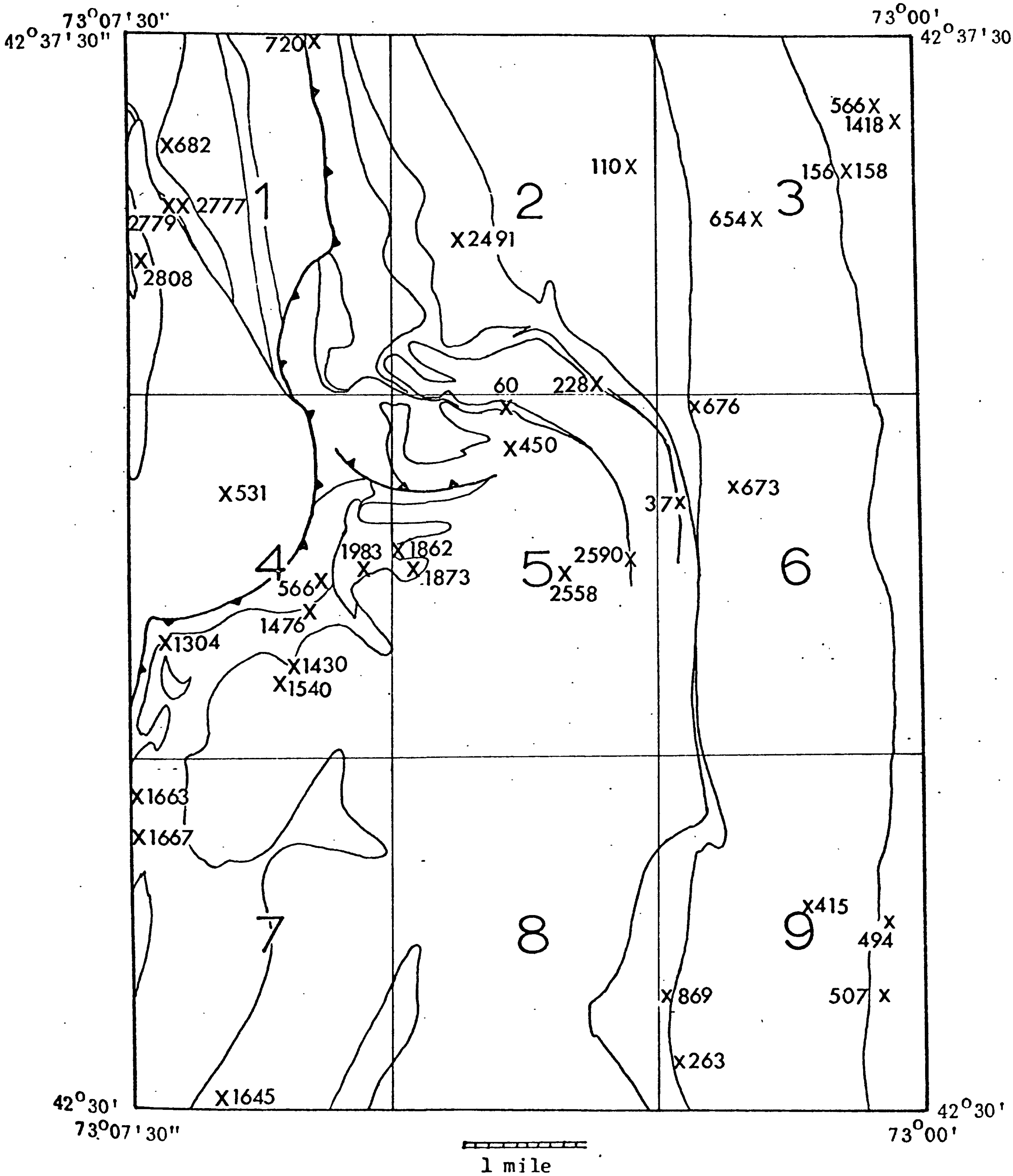


Figure 3. Location of specimens for which modes are given in tables. In text, localities are indicated thus ( ). Each sub-area is 2 and 1/2' square and covers approximately 6.3 square miles.

is not present in the Windsor quadrangle. Furthermore, little perthite was seen as reported by Skehan (1963, p. 35) for the Vermont localities.

The microcline augen containing abundant anhedral quartz inclusions have a cataclastic texture. Quartz fills the cracks with optical continuity.

At the outcrop in the large roadcut east of the intersection of Main Road and Windsor Road west of Savoy and on the north side of the Berkshire Trail, 500 feet west of where Hume Brook crosses the highway, there are two polymineralic "inclusions" of fine grained aplitic felsic rock with a weak foliation or bedding parallel to the regional foliation. These "inclusions" may have a metamorphic or an igneous origin. The remarkable uniformity of composition suggests an igneous origin, but there are no cross-cutting contacts, reaction skarns, or assimilation features. The numerous inclusions in the microcline may indicate recrystallization in a fine-grained matrix. Certainly the present textures are no older than the regional Paleozoic metamorphism.

The western contact of the Stamford Granite Gneiss with the Hinsdale Gneiss is essentially north-south and is subparallel to the eastern erosional contact with the Cambrian clastic member at the base of the Hoosac Formation. This suggests that the Stamford Granite Gneiss may cross-cut the Hinsdale Gneiss on a regional scale within the quadrangle. Locally, the rocks near the western contact are structurally concordant. With the lack of traceable rock types in the Hinsdale Gneiss, it is not possible to determine if the two rocks are just locally structurally parallel or if the Hinsdale Gneiss rock types are distributed in bands parallel to the contact.

Hinsdale Gneiss

General Statement:

The Hinsdale Gneiss of Precambrian age is named for occurrences at Hinsdale, Massachusetts (Emerson, 1892) some four miles south of the Windsor quadrangle. Herz (1958, 1961) used the term Hinsdale Gneiss for those rocks occurring west of the Stamford Granite Gneiss in the North Adams quadrangle and for the banded gneisses on North Mountain in the Cheshire quadrangle.

The structural and stratigraphic relationships between the eastern and western exposures of the Hinsdale Gneiss in the Windsor quadrangle are not clear. Thus there seems to be no justification or basis for division of the Hinsdale Gneiss as previous writers have done (e.g. Emerson, 1892).

It crops out in a belt west of the Stamford Granite Gneiss in the northern part of the quadrangle, in a southeasterly-trending belt in the central part, and in a broad belt trending north through the south-central part of the Windsor quadrangle. There are very distinctive rocks in the Hinsdale Gneiss, but the outcrops are so discontinuous and the structure is so complex that no stratigraphy could be established.

The formation consists predominantly of quartz-feldspar-biotite gneiss, minor quartzite, and very minor amounts of calc-silicate rock, amphibolite, and graphitic gneiss.

Petrology:

The original sedimentary character of the Hinsdale Gneiss is quite apparent east of Adams on the west slope of the Hoosac Range. There layering is common and pronounced. The rocks range from quartzites, 1 to 6 inches thick, to quartz-feldspar-clinozoisite (or epidote)-biotite gneiss (table 1). Muscovite is present in amounts ranging from 0 to 2 percent. The feldspathic gneisses comprise about 90 percent of the rocks present. The felsic gneisses are equigranular rocks with grain size ranging from 1/2 to 2 mm. Porphyroblastic textures are best developed in the rare calc-silicate gneiss and amphibolite.

Rarely the more feldspathic gneisses are accompanied by quartz-feldspar-(muscovite) pegmatites. They are discontinuous and commonly occur within a single bed. They are as large as several feet long and one foot wide. The pegmatites are the result of recrystallization in place during the regional Precambrian metamorphism and are not intrusive.

In the southeast part of the quadrangle, the felsic gneisses of the Hinsdale Gneiss (Becket Gneiss of Emerson, 1892) have the same mineralogical composition as the felsic gneisses east of Adams. However, they are more massive; layering is poorly developed. The different structures may be a result of different style of deformation in the two areas or possibly original differences in the primary (sedimentary) structures.

In the southwest part of the quadrangle the rocks in the Hinsdale Gneiss are more varied. In addition to the feldspathic gneisses and granulites of the northern and eastern areas, amphibolite, calc-silicate rocks, rusty graphitic gneisses, and very coarse-grained microcline granulites are common.

The distribution of the various rock types is summarized below:

1. Well-layered gneisses are best developed on the slopes east of Adams.
2. The gneisses in the eastern and southeastern exposures of the Hinsdale Gneiss are more granular and more homogeneous, layering being poorly developed.
3. Amphibolite and other mafic rocks are most common in the southwestern part of the quadrangle.
4. Rocks containing porphyroblasts of microcline (exclusive of the Stamford Granite Gneiss) are most common in the southwestern part of the quadrangle.
5. Graphitic and calc-silicate gneisses are most common in the western exposures of the Hinsdale Gneiss and commonly occur together.

Amphibolite has a patchy distribution. Although foliation is poorly developed faint layering is present; it is caused by a variation in the proportions of hornblende and plagioclase.

Very massive amphibolite is also present (e.g. No. 1540-b in table 1). Commonly interbedded hornblende-plagioclase rock and quartz-feldspar granulite are seen with no cross-cutting contacts. It could not be established whether all of these are pre-metamorphic dikes, or sedimentary rocks because structural relationships between outcrops with similar rocks are unclear.

Unlike the mafic rocks in the Paleozoic metasedimentary sequence, they have very little or no chlorite and calcite. They are generally equigranular rocks with grains as much as 5 mm. in diameter. Larger grain size is related to a higher proportion of hornblende in the rock.

Very coarse-grained feldspathic gneisses form a small percentage of the gneisses but are spectacular and useful for distinguishing between Precambrian and Cambrian gneisses. Their texture is nearly pegmatitic and their mineralogy is very simple. Generally alkali feldspar (predominantly microcline) and quartz compose greater than 95 percent of the rock (No. 1945, table 1).

The bands typically are 2 to 3 inches thick and are separated by thin (1 to 5 mm.) bands of biotite-rich rock.

This porphyroblastic texture is in sharp contrast with the textures in the Dalton Formation. Thus the presence of the microcline porphyroblasts is useful in distinguishing these two rock units when their mineralogy is similar.

Calc-silicate gneisses compose less than 1 percent of the Hinsdale Gneiss. They are most notable for the spectacular textures developed in them. Commonly, diopside and tremolite are as much as 10 mm. in diameter and knots of diopside in calcite as much as 6 inches in diameter are present at the graphite prospect north of Jackson Road. Graphite in flakes as large as 5 mm. is commonly associated with the calc-silicates. Scapolite is commonly present but usually highly altered to muscovite and paragonite.

Rare rusty-weathering feldspar-quartz gneisses and granulites are generally associated with the calc-silicate gneisses. They commonly contain as much as 1 percent highly crystalline graphite.

Correlation:

With the exception of the western exposures of the Hoosac Formation and the Stamford Granite Gneiss, Pumpelly and others (1894) assigned all rocks below the limestones west of the Precambrian terrane and all rocks beneath the Hoosac Formation east of the Precambrian terrane to the Vermont Formation.

Emerson (1892, 1917) divided the Precambrian rocks south of Main Road into two units, the Hinsdale Gneiss and the Becket Gneiss on the west and east respectively. This distinction was based on the lithologic heterogeneity and well banded nature of the Hinsdale Gneiss. These contrasted with the granitic and apparently more homogeneous Becket Gneiss.

Prindle and Knopf (1932) equated the Mount Holly Gneiss of Vermont with the Hinsdale Gneiss of Emerson and retained the term Becket Gneiss for the more massive granitic gneisses. The Mount Holly Gneiss was thought to be a highly metamorphosed sedimentary sequence which in part was of migmatitic origin.

Christensen (1963) although interested only in the structural interpretation of the Hoosac nappe, put all Precambrian rocks except the Stamford Granite Geniss in the "Vermont Formation" although he did mention different lithologic types.

Paleozoic Metasedimentary Rocks

Eastern Sequence

General Statement

The eastern sequence of the Cambro-Ordovician metasedimentary rocks (figure 4) lies east of the Stamford Granite Gneiss and consists of the following units in order of decreasing age: Hoosac Formation, Rowe Schist, Moretown Formation, Hawley Formation. The sequence is composed of schists, quartzites, phyllites, feldspathic granulites and gneisses, and metavolcanic rocks which are the eugeosynclinal facies of the western miogeosynclinal sequence lying west of the Hinsdale Gneiss.

Only the Hoosac Formation and Rowe Schist crop out in

	Western Sequence	Lithology	Thickness	Thickness	Lithology	Eastern Sequence
Middle Ordovician	Berkshire Schist	carbonaceous schist and phyllite, black quartzite calcareous schist, carbonaceous schistose marble	1,000'	1,000'	Not exposed in map area	Hawley Formation
				9,000'	Not exposed in map area	Moretown Formation
Lower Ordovician	Bascom Formation	dolomitic, schistose, and feldspathic marble	100'			
	Shelburne Marble	white calcite marble	100'			
Upper Cambrian	Clarendon Springs Dolomite	white to buff, dolomite marble with quartz stringers and minor dolomitic marble	300'	5,000'	muscovite-chlorite schist carbonaceous gray phyllite, black quartzite, greenschist, amphibole gneiss	Rowe Schist
Middle Cambrian	Kitchen Brook Dolomite	buff or white to blue-gray mottled dolomite marble	300'			
Lower Cambrian	Cheshire Quartzite	massive quartzite	2-100'	3,000'-4,000'	albite schist, garnet schist, granulite and gneiss, quartz-muscovite-calcite schist, greenschist	Hoosac Formation
	Dalton Formation	granulite, quartzite, gneiss, schist, conglom.	500'			
	Hoosac Formation	garnet schist, albite schist, conglom., qtzite.	300'			
Precambrian	Hinsdale Gneiss	felsic gneiss, quartzite, amphibolite, calc-silicate gneiss, graphitic gneiss, minor coarse-grained felsic gneiss			microcline-quartz-biotite augen gneiss	Stamford Granite Gneiss

Figure 4. Stratigraphic column for the Windsor quadrangle, Massachusetts.

the Windsor quadrangle.

The stratigraphic sequence for the rocks east of the Precambrian terrane is given as follows by Emerson (1917):

Savoy Schist (Emerson, 1892)
Chester Amphibolite (Emerson, 1892)
Rowe Schist (Emerson, 1892)
Hoosac Schist (Emerson, 1892)
unconformity
Becket Gneiss (Emerson, 1892)

Because of the reconnaissance nature of the original mapping, Emerson did not recognize that the Chester Amphibolite in the Windsor area was not continuous with the large amphibolite body at Chester, Massachusetts. With the absence of this formation, it is impossible to distinguish the Rowe Schist from the basal portion of the Savoy Schist. Consequently the stratigraphy has been revised (Hatch and others, in Cohee, 1966). The stratigraphy as established by the writer and other workers in the area is given in figure 4. This column differs from Emerson's in the following respects:

1. The Savoy Schist has been dropped as a formational name. The basal chlorite-sericite schists of the Savoy Schist have been placed in the Rowe Schist as redefined by Hatch and others (1966).
2. The Chester Amphibolite has been dropped as a formation name.

Hoosac Formation

General Statement:

The Hoosac Formation (Hoosac Schist of Wolff in Pumpelly and others, 1894) of Lower Cambrian age was first described from a type locality on the eastern flank of the Hoosac Range in the old Hawley quadrangle (Emerson, 1892). As mapped here the Hoosac Formation includes all rocks between the Stamford Granite Gneiss and the Rowe Schist. In the Windsor quadrangle the Hoosac Formation crops out both east and west of the Precambrian rocks. East of the Precambrian terrane the Hoosac Formation crops out in a band 6,000 to 7,000 feet wide for the entire length of the quadrangle. Exposures are poor because the rocks are schistose, dips are low near the Precambrian-Cambrian unconformity in the north, and glacial debris is abundant.

The rocks of the Hoosac Formation range from relatively massive to thinly bedded. Beds are from less than 1/2 inch to 24 inches in thickness. It is more massive in the southern part of the quadrangle, indicating original differences in sedimentary structures. Bedding is caused by variation in the proportions of albite and quartz; the albite porphyroblasts stand out on weathered surfaces.

The color ranges from gray to brownish-gray to greenish gray. The variation in color throughout the formation can be related to the proportions of chlorite, biotite, and white mica. These mineralogical changes in most places are gradual both vertically and laterally within the formation. Interbedding of, for example, chlorite-bearing and biotite-bearing schist is rare.

The Hoosac Formation is composed of, in decreasing order of abundance, albite schist, gneiss and granulite, garnet schist, and quartz-muscovite-calcite schist. Of these four rock types, the albite schist comprises most of the formation. The other three, although distinctive and mappable, are thin and do not persist along strike outside the quadrangle. In the Windsor quadrangle these three types are considered as members of the Hoosac Formation.

Modes are given in table 2.

Albite Schist:

Seventy-five percent of the formation is a remarkably uniform albite schist. The variations are subtle difference in bedding characteristics, the color of the fresh rock, and mineralogic content.

The albite schist is the best exposed rock in the Hoosac Formation. Good exposures may be seen in many of the brooks flowing eastward into the Westfield River and along the Westfield River.

Several specimens show graded beds 1/2 inch thick. The graded beds are composed of quartzose layers grading into muscovite-paragonite-albite assemblages. More typically, bedding ranges from a fraction of an inch to several feet. Beds are

TABLE 2

Estimated Modes

Hoosac Formation - Eastern Sequence

	Granulite and gneiss member		Garnet schist member	Albite schist		Quartz-muscovite-calcite schist member
Specimen	263	676-b	673	33a	654	415
Quartz	60	25	55	10	60	60
Muscovite	2	2	30[1]	3	15	25
Paragonite						
Albite	1	61	5-P	82	20-P	
Microcline	30	5				
Chlorite	tr		3	tr	tr[2]	
Garnet		2	5-P	2		
Chloritoid			tr*			
Biotite	1	3		3	2	
Epidote group			1	tr	1	
Tourmaline			tr		1	
Magnetite	2	1	tr			2
Hematite						7[2]
Apatite					tr	
Calcite + dolomite	2			tr		5
Zircon	tr		tr			
Sphene	tr	tr	tr			

* = armored with garnet
1 = most abundant white mica but absolute amount undetermined
2 = secondary
P = porphyroblastic

Location of Specimens[1]

Hoosac Formation - Eastern Sequence

Granulite and Gneiss Member

263 Gray fine-grained granulite: 100 feet east of intersection of Hume Brook and the Berkshire Trail, (9).

676-b Feldspar gneiss: in Center Brook, 500 feet upstream from dirt road across stream, (6).

Garnet Schist Member

673 Silvery-green schist: 1,500 feet north-northeast from Savoy, on knob, (6).

Albite Schist

33a Gray, massive, albite schist: intersection of Griffin Hill Road and Main Road, (6).

654 Gray, albite granular schist: 850 feet N.45°E. of Borden Mountain, (3).

Quartz-muscovite-calcite Schist Member

415 Silvery-green, rust-spotted, granular schist: elevation 1,830 feet on Steep Bank Brook, (9).

1 See figure 3.

marked by quartz-rich continuous layers as much as 1 inch thick.

Quartz, albite, muscovite, and paragonite usually compose more than 80 percent of the rock. Minor but ubiquitous minerals include some combination of chlorite, biotite, and garnet. Accessory minerals include clinozoisite (or epidote), tourmaline, magnetite, and carbonaceous material ("graphite"). The carbonaceous dust occurs in the matrix and commonly forms sigmoid inclusion patterns in the porphyroblasts where it is far more abundant than in the matrix of the rock.

Porphyroblastic albite ranges in size from less than a millimeter to more than a centimeter and is commonly 5 to 10 times the size of the minerals forming the matrix. The porphyroblasts are typically poikiloblastic with inclusions of euhedral garnet, magnetite, quartz, and muscovite or paragonite. Carbonaceous dust is common in the albite, but the albite is rarely black as it is at the type locality of the Hoosac Formation on the western flank of the Hoosac Range in the North Adams quadrangle The albite content of the plagioclase is greater than 95 percent and albite may make up greater than 50 percent of the rock.[1]

---

[1] Compositions of metamorphic minerals reported here are applicable to the same phases in all the Paleozoic schists. The determination of the composition of these minerals is treated in the mineralogy chapter.

Muscovite, with as much as 20 percent paragonite in solid solution, and paragonite, with as much as 10 percent muscovite in solid solution, together comprise from 10 to greater than 50 percent of the rock.

The almandite-rich garnet is found throughout the albite schist but is more typical and abundant in the southern half of the quadrangle. A comparable mineralogic change in the Rowe Schist and the Moretown Formation to the east suggests that this change in mineralogy may be related to the metamorphic gradient in this area. The staurolite isograd is only 3 to 4 miles to the east of the eastern border of the quadrangle.

Chlorite and biotite generally do not exceed 5 percent of the rock. Their composition is highly variable depending on the mineral assemblage and unknown factors.

The $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios for biotites are greater than 0.5 while the same ratio for chlorite ranges from 0.35 to 0.7.

Gneiss and Granulite Member:

A gneiss and granulite unit lies at the base of the Hoosac Formation. It is a sequence of rather heterogeneous rocks cropping out immediately to the east of and on the Stamford Granite Gneiss for nearly the entire length of the quadrangle; an outcrop width of 100 feet is inferred at the north edge of the quadrangle, but at the south edge the outcrop width is approximately 1,000 feet. The lower contact may be observed in Center Brook 6,000 feet north-northwest of Savoy just north of where the dirt road crosses the brook. Here granulites and gneisses lie directly on the coarse-grained Stamford Granite Gneiss; one pebbly bed contains pebbles of the granite gneiss as much as 1 inch in maximum dimension. Pebbly beds also have been found in the stream about 2,000 feet southwest of Borden Mountain. A rather distinctive bed about 20 feet thick containing about 4 percent magnetite has been traced for nearly a mile in the southern part of the quadrangle. Associated with this bed are beds with what appear to be pebbles stretched out into pencils down the dip.

Petrologically the rocks are highly variable. They are quartzite, quartz-feldspar-mica schist, and quartz-feldspar-biotite granulite and gneiss. Accessory minerals include magnetite, muscovite, and garnet.

Albite, where present, contains abundant inclusions of quartz. Microcline has rounded outlines and is not porphryoblastic. Some of the gneisses low in the section contain large grains of microcline. This suggests that the rocks may have been derived from the Stamford Granite Gneiss.

Because of the conformable relationship with the overlying garnet schist member of the Hoosac Formation and the unconformity (erosional at least) with the underlying Stamford Granite Gneiss, this rock is placed at the base of and assigned to the Paleozoic sequence of sediments.

Garnet Schist Member

A very distinctive coarse-grained garnet schist occurs directly beneath the albite schist.

It crops out as ridges and knobs in the otherwise flat topography in the lower Paleozoic section. In the southern part of the quadrangle, outcrops are continuous for as much as a quarter of a mile. To the north, the outcrops become discontinuous, but this unit has been traced to within half a mile of the quadrangle boundary. The outcrop width ranges from 100 to 300 feet.

The schistosity is highly contorted. A fresh specimen has a greenish-gray pearly luster. The weathered surfaces have a knobby surface due to porphyroblasts of almandite that range from 1/4 to 1 inch in diameter. Sigmoid structures in the almandite may be seen on the more weathered surfaces.

The rock is composed of quartz, white micas, chlorite, and almandite. Accessory minerals include chloritoid and albite. In the southern half of the quadrangle the schist is richer in chloritoid that commonly gives the rock a bluish cast. Albite is more common in the northern exposures, comprising as much as 25 percent of the rock. The rock is coarse-grained where the phyllosilicates predominate. Where albite is present, the rock is more granular and appreciably finer-grained.

Although Herz (1961) did not distinguish this unit in the North Adams quadrangle, Pumpelly and others (1894, p. 61) mention the presence of a garnetiferous schist at the base of the Hoosac Formation in this area. The same rock type has been traced south through the Peru quadrangle. The rock appears identical to the Heartwellville Schist (Skehan, 1963) in southern Vermont and the Gassetts Schist (Rosenfeld, 1954; Thompson, 1952) in east-central Vermont to which it is presumed to be equivalent.

Quartz-muscovite-calcite Schist Member:

This member is stratigraphically in the middle of the albite schist. Outcrops are widely spaced for the entire length of the quadrangle but the rock has not been traced outside the quadrangle. The best exposures are in the brooks that flow eastward into the Westfield River in the southeast part of the quadrangle.

This member is about 200 feet thick and typically is not all calcareous. It consists of interbedded albite schist and quartz-muscovite-calcite schist that weathers punky or to brown spots on fresh surfaces where the ferrous iron in the calcite has been leached out and oxidized. The fresh rock has a pearly luster because of the abundance of muscovite. Most of the calcareous beds are composed of quartz, muscovite, and calcite; chloritoid was found in one specimen.

Rocks of this general composition have been recognized north along strike, but have not been separately mapped. Many workers in central Vermont (e.g., Chang and others, 1965) have recognized similar rocks at a similar horizon.

Greenschist:

Very minor greenschists are as much as 10 feet in thickness in the albite schist near the contact with the Rowe Schist. Lithologically they are identical with those found in the Rowe Schist.

Thickness:

The cumulative thickness for all the rocks in the Hoosac Formation is between 3,000 and 4,000 feet.

Age:

No fossils have been found in the Hoosac Formation. A tentative age of Lower Cambrian or earlier was assigned by Wolff (in Pumpelly and others, 1894). This is substantiated by two lines of reasoning:

1. The Hoosac Formation conformably underlies the Rowe Schist, the lowest rocks of which are probably equivalent to and have been correlated with the West Castleton Formation and the Mettawee member of the Bull Formation (Theokritoff, 1964; Chang and others, 1965). The Mettawee member of the Bull Formation has Lower Cambrian fossils (Theokritoff, 1964). The lower member of the Bull Formation, the Bomoseen

Graywacke, although unifossiliferous, is chemically similar to the Hoosac Formation.

2. The Hoosac Formation has been traced around the northern limit of the Stamford Granite Gneiss to the western side of the Precambrian terrane just east of Adams, Massachusetts. Here the garnet schist member and the albite schist of the Hoosac Formation are both present. They are stratigraphically overlain by the Dalton Formation and Cheshire Quartzite. This section can be correlated with the stratigraphic sequence Cheshire Formation, Dalton Formation, Stamford Granite Gneiss found on Clarksburg Mountain in northwestern Massachusetts. There Walcott (1888, p. 235-236) found Olenellus fragments in a pebbly conglomerate about 100 feet above the unconformity and within the Dalton Formation.

Rowe Schist

General Statement:

The name Rowe Schist (Hatch and others, in Cohee, 1966) is applied to the rocks found between the Hoosac Formation and the base of the Moretown Formation. The base of the formation is commonly indistinct since the Hoosac Formation commonly grades vertically into the Rowe Schist. Grading is accomplished by a gradual decrease in the amount of albite and development of chlorite instead of biotite or biotite and chlorite. This is a compositional change rather than a metamorphic change.

Although the top of the Rowe Schist is not exposed within the Windsor quadrangle, it lies less than 1,000 feet east of the quadrangle boundary. The upper contact is placed at the first appearance of beds of pale green or buff quartz-feldspar granulite with accessory clinozoisite, muscovite, and either chlorite or biotite. The granulites (and occasionally schists of the same mineralogical composition) contain continuous quartz-feldspar beds that are 1/8 to 1/4 inch thick (the "pinstripe" of Vermont workers). Both of these rocks are typical of the basal beds of the Moretown Formation.

The Rowe Schist is exposed in a north-trending band along the eastern border of the quadrangle. Although the rock does not crop out well, nearly complete sections can be seen in various brooks. The lower half of the formation is exposed in Baker Brook and the east-flowing unnamed brook 2,500 feet south of Baker Brook. The upper half of the section is exposed in Ross Brook and its continuation into Gulf Brook in the North Adams quadrangle. A nearly complete section of the entire formation is exposed in Horsefords Brook. Both of the east-flowing brooks in the southeast corner of the quadrangle provide a nearly continuous section through the black phyllites and black quartzites of the lowest member of the formation.

The Rowe Schist is composed of, in decreasing order of abundance:

light-green to silvery, muscovite-chlorite schist
carbonaceous phyllite and black quartzite
greenschist
gray phyllite and quartzose granulite

The last three rock types are considered as members of the Rowe Schist.

Muscovite-chlorite Schist:

More than half of the Rowe Schist is composed of light-green to silvery, fine-to medium-grained (as much as 1 mm.), slabby, muscovite-chlorite schist.

Bedding is indistinct because of well developed axial plane schistosity. Beds range from a fraction of a millimeter to 1 inch in thickness and the rock is usually very thinly laminated; the quartzose layers are not continuous as they are in the Moretown Formation.

Mineralogically, the rocks are quartz-muscovite-chlorite schist with accessory albite, almandite, chloritoid, paragonite, clinozoisite, and magnetite. Those schists rich in clinozoisite weather chalky. Typical modes for this rock type are given in table 3.

Quartz and white mica compose 70 to 95 percent of the rock. Garnet, chlorite, chloritoid, magnetite, and clinozoisite never exceed 5 percent each. The magnetite is found as porphyroblasts only in rocks free of carbonaceous material.

Beds are marked by lenses of quartz that are elongated

TABLE 3

Estimated Modes

Rowe Schist

	Muscovite-chlorite schist		Carbonaceous rocks		Quartzites		Greenschist	Amphibole gneiss
Specimen	72	158	156	507-b	494	566-b	175	1418
Quartz	16	15	50	50	99	95	8	25
Muscovite / Paragonite	70 [1]	80 [1]	40 [1]	10	tr	5		
Albite	10	tr-P					10	tr
Chlorite	3	2	9	12			40	10
Garnet		tr						
Biotite				7				
Epidote group	tr						11	25
Hornblende							30	40
Tourmaline	tr		tr					
Magnetite	tr	tr	1			tr		
Hematite			tr			tr [2?]	1	
Calcite + dolomite				20	1			
Sphene								tr
Graphite †			tr	1	tr	1		

1 = most abundant white mica but absolute amount undetermined
2 = secondary
P = porphyroblastic
† = carbonaceous material in the Paleozoic rocks loosely called graphite

Location of Specimens[1]

Rowe Schist

Muscovite-chlorite schist

72 Green, mica schist: on Adams Road where the east-flowing brook crosses east-northeast of Borden Mountain, (3).

158 Grayish-blue, spangled schist: elevation 1,740 feet on brook that flows east into Parker Brook, (3).

Carbonaceous Phyllite and Schist

156 Grayish-blue, phyllite: elevation 1,760 feet on brook that flows east into Parker Brook (3).

507-b Light gray, granular schist: elevation 1,660 feet on southern fork of the brook flowing into the Westfield River, (9).

Black Quartzite

494 Black, thinly-bedded quartzite: in the Westfield River, 1,000 feet south of the Windsor Jambs swimming pool, (9).

566-b Silvery-gray, thinly laminated quartzite: elevation, 1,560 feet on Baker Brook, (9).

Greenschist

175 Massive greenschist: elevation 1,680 feet on Horsefords Brook, (3).

Amphibole Gneiss

1418 Weakly foliated amphibole gneiss: elevation 1,650 feet, 800 feet from point where Tannery Road enters Windsor quadrangle from the east, (3).

[1] See figure 3.

down the dip of the schistosity. The dimensions of these lenses are generally as follows: the longest dimension is down the dip of the schistosity and commonly is 10 to 20 times the thickness (as much as 1 inch) of the lens. Along the strike of the schistosity, the lenses measure 5 to 10 times the thickness. The upper part of the formation has more of these lenses than the lower part, where quartz, although equally abundant, is more uniformly distributed rather than as discreet lenses.

The origin of the quartz lenses is problematic. Many explanations have been put forth. The quartz lenses may be

1. Original discontinuous quartz-rich beds or sedimentary lenses.
2. Continuous beds that have been torn apart tectonically.
3. Concentrations of quartz derived by metamorphic mechanism such as metasomatism or differentiation.
4. Original cherty beds that have been pulled apart tectonically or originally discontinuous cherty beds.

Further to the east, at higher metamorphic grade cherty beds in the iron formations of the Hawley Formation have not recrystallized to a grain size equivalent to that found in the Rowe Schist at lower metamorphic grade.

In the Windsor quadrangle, there is abundant evidence for several generations of folding, the first of which was accompanied by metamorphism. The maximum dimension of these lenses is parallel to early syn-metamorphic isoclinal fold axes. The lenses have subsequently been folded by later folds; thus they were either present prior to or were formed during the first folding.

An interesting discussion of this problem is given by Albee (1957a, p. 102-103).

Carbonaceous Phyllite and Black Quartzite Member:

The distinction between the Hoosac Formation and the Rowe Schist is based mainly on texture and structure with a secondary emphasis on mineralogy and color.

The basal beds of the Rowe Schist in the Windsor quadrangle consist of either carbonaceous phyllite or light-green schist. The phyllite and light-green schist are more schistose and thinner-bedded than the Hoosac Formation. The light-gray phyllite is a muscovite-paragonite-quartz rock with accessory chlorite, chloritoid, magnetite, almandite, and carbonaceous dust. The light-green schist differs in having no carbonaceous material, little or no paragonite, and as much as 20 percent albite. Both of these rocks are very fine-grained and bedding is absent or indistinct.

The underlying Hoosac Formation is coarser-grained, more granular (because of a smaller proportion of phyllosilicates) and has beds ranging from 1/2 inch to 12 inches in thickness. It is composed of quartz, albite, muscovite, paragonite, biotite, and/or chlorite. Accessory minerals include almandite, clinozosite, and magnetite.

Within the Windsor quadrangle, a carbonaceous phyllite is commonly present at the base of the Rowe Schist so this distinction is clear and the contact is fairly sharp in most places. This zone of predominantly carbonaceous phyllite is about 1,000 feet thick at the northern boundary of the quadrangle, has a discontinuous distribution and thins to zero in the east-central part, and thickens to 2,000 feet at the southern border of the quadrangle. The zone weathers to a soft gray to rusty or yellow-brown rock.

At the point of maximum outcrop width in the south, the unit is carbonaceous and about 5 percent of the member is composed of massive, vitreous, black quartzites as much as 2 feet in thickness.

The quartzite commonly contains as much as 5 percent dolomite; the phyllite has as much as 20 percent dolomitic cement. This member persists for at least 10 miles south of the quadrangle.

The carbonaceous phyllite on the contact in the northern part of the quadrangle has been traced to the Mohawk Trail (Route 2) north of the quadrangle. Further to the north and east the phyllite is discontinuous or absent and the contact between the two formations is more arbitrary.

The distribution of the carbonaceous rocks within the Rowe Schist is notably uneven. It is interesting that the pinch and swell of the basal carbonaceous member is mimicked by a zone of carbonaceous phyllite in the middle of the Rowe Schist (see below). Apparently the variable thickness of the two members is not the result of tectonic thinning or dip changes. Probably the original distribution of reducing conditions is responsible for the quantity and location of these carbon-rich sediments.

This member is lithologically similar to the Ottauquechee Formation and the West Castleton Formation. The dolomitic quartzites and carbonaceous phyllites are also on the right horizon to be tentatively correlated with the Plymouth member of the Hoosac Formation in Vermont (Foyles and Richardson, 1929; redefined by Thompson, 1950). These rocks are placed in the Rowe Schist because of textural and mineralogical similarity to the Rowe Schist.

Greenschist:

Most of the greenschists are in the upper half of the Rowe Schist. They range widely in both their physical appearance and mineralogically. Although single homogeneous greenschists are as much as 20 feet in thickness it has not been possible to trace any bed more than a few thousand feet with any certainty. They comprise less than 5 percent of the formation. The greenschist ranges from punky weathering, massive, dark green, homogeneous rock to fissile, well-banded, slabby rock with well developed schistosity. Banding is related to the abundance of epidote (or clinozoisite). All the greenschists have a grain size of 1 mm. or less. There are three rock types in this group:

Albite-epidote-chlorite-quartz schist with accessory calcite and opaque minerals is the most abundant type of greenschist in an area of extensive outcrop. It is found in homogeneous layers as much as 20 feet thick.

The second most important type has very thin alternating beds of epidote-rich and epidote-poor rock and in places alternating albite-epidote rock and quartzose beds as much as 1/2 inch thick. Calcite is lacking or sparse. Hornblende is common and ranges up to 20 percent of the rock; hornblende is rare in the more calcareous greenschist.

The third variety is an amphibole gneiss. It is a massive, 1 to 2 mm. grain-size, weakly foliated rock that weathers chalky gray-green. Although it has a weak foliation which is parallel to the regional schistosity, there is no banding or other planar feature. The only outcrops of this rock are on the north slopes of the hill just south of Tannery Falls in the northeast corner of the quadrangle. It is composed of chlorite, clinozoisite, hornblende, and quartz.

Typical modes for these rocks are given in table 3.

The majority of the greenschists are thinly bedded, internally banded and probably represent limey detrital volcanic material. The inability of the writer to trace these rocks for any distance and their composition makes it unlikely that they represent metamorphosed volcanic flows of any kind.

The massive amphibole gneiss may be a dike rock. The contact of the amphibole gneiss with the Rowe Schist was conformable where exposed and could not be followed. The area of exposure has no other rock types exposed within it and may even represent a volcanic pipe.

Gray Phyllite and Quartzose Granulite Member:

A zone of interbedded gray phyllite, quartzose granulites, and light-green muscovite-quartz-chlorite schist occurs mid-way in the Rowe Schist. This zone is less than 1,000 feet thick west of Tannery Falls, thins to the south and is last observed in the dirt road just north of Chickley River at an elevation of 1,900 feet. This zone leaves the quadrangle to the east where it has been mapped by Osberg (in press). It is well exposed in the southwest corner of the Plainfield quadrangle 200 feet east of the Windsor quadrangle boundary.

The gray phyllite is nearly a pure muscovite-paragonite-quartz rock; the granulite is very rich in quartz with minor muscovite. Modes are given in table 3. The gray color in outcrops is due to the carbonaceous dust associated with the muscovite and paragonite. The gray phyllite and granulites compose 30 to 50 percent of the zone; the remainder is muscovite-chlorite schist.

Thickness:

Including the part of the formation within the Plainfield quadrangle, the thickness of the Rowe Schist probably is in excess of 5,000 feet and does not vary appreciably. However, individual members thicken and thin along strike.

Age:

The Rowe Schist, as redefined by Hatch and others (in Cohee, 1966) is equivalent to the Pinney Hollow, Ottauquechee, and Stowe Formations of southern Vermont (Doll and others, 1961). No fossils have been found within the Rowe Schist in the Windsor quadrangle; hence all dating must be indirect by correlation and physical continuity. The best age for the Rowe Schist is based on a tentative age for the Ottauquechee Formation of Vermont.

Thompson (in Chang and others, 1965, p. 42-43) states:

> "There is, however, good evidence that the continuation of the Ottauquechee Formation in Quebec (Osberg, 1956, 1965; Cady, 1960) can be traced north and west around the Sutton Mountains anticlinorium into the Sweetsburg slate which is at least in part Middle Cambrian and into the Scottsmore Formation which is Lower Cambrian (Osberg, 1965, p. 227, quoting G. Theokritoff). The Ottauquechee Formation in eastern Vermont has features in common with the West Castleton and Hatch Hill Formations of the Taconic slate belt (Theokritoff, 1964; Zen, 1961). . . . The West Castleton carries Lower Cambrian fossils and the Hatch Hill carries Upper Cambrian fossils."

On this basis the Rowe Schist spans at least Lower Cambrian to Upper Cambrian time.

The carbonaceous schists of the Hawley Formation in the Plainfield quadrangle immediately to the east of the Windsor quadrangle may be correlated with the interbedded volcanic rocks, schistose granulites, and carbonaceous phyllites of the Missisquoi Formation of central Vermont. These rocks have been traced northward into the Eastern Townships of Quebec near Lake Memphremagog and further north where they are known as the Beauceville Series. Here they contain graptolites of late Middle Ordovician (Trentonian) age.

Thus the top of the Rowe Schist is probably no younger than Lower Ordovician, considering the thickness of the Moretown Formation. The base of the formation would be placed somewhere in the Lower Cambrian based on the correlation of the West Castleton Formation with the Ottauquechee Formation (see above).

Stratigraphic Problems

One of the major objectives of the program of the U. S. Geological Survey in western Massachusetts is to carry the eastern Paleozoic stratigraphy southward from Vermont to Connecticut. The Windsor area is of critical importance because it lies on the Precambrian-Cambrian contact.

The Cambro-Ordovician sequence of central and southern Vermont that has been carried to the Massachusetts-Vermont state line is as follows (Skehan, 1961, p. 25):

Cram Hill Formation (Ordovician)
Moretown Formation (Ordovician)
Stowe Formation (Ordovician ?)

Ottauquechee Formation (Cambrian)

Chester Amphibolite (Cambrian)

Pinney Hollow Formation (Cambrian)

Hoosac Formation (Cambrian)

Heartwellville Schist (Cambrian ?)

Mapping in the Heath quadrangle (Hatch and Hartshorne, in preparation), Rowe quadrangle (Chidester and others, in press), Plainfield quadrangle (Osberg and others, in preparation), and the Windsor quadrangle indicates that the stratigraphy given above does not persist in detail to the south. These workers have shown that the sequence undergoes both appreciable thickening and thinning along strike and that some of the formations are discontinuous.

The Heartwellville Schist (Skehan, 1961) is extensively exposed in southern Vermont and rocks of equivalent type are exposed in the Sherman dome (Chidester and others, in press) on the Vermont-Massachusetts boundary. In the southern part of the North Adams quadrangle, the garnet schist is absent. It is extensively exposed in the Windsor and Peru quadrangles but is not persistent further south.

In many places within the Rowe quadrangle where the basal carbonaceous rocks reported in the Windsor quadrangle are missing, the distinction between the Hoosac Formation and the Rowe Schist (Pinney Hollow type of Vermont) is difficult. The two formations both interfinger laterally and grade vertically into each other.

All of the rock types between the Hoosac Formation and the Moretown Formation are greatly thinned in the northeast corner of the Rowe quadrangle (figure 2) and the northwest corner of the Heath quadrangle. In places the section is less than 1,000 feet thick and is composed predominantly of amphibolite and greenstones. None of the pelitic rocks can be traced through this area. The sequence Stowe Formation, Ottauquechee Formation, Chester Amphibolite, Pinney Hollow Formation reappears in the Rowe quadrangle as a mappable sequence of rock types. But in the southern part of this quadrangle there is apparently more than one band which corresponds to the Ottauquechee lithology (Hatch and others, in Cohee, 1966).

In the Windsor and North Adams quadrangles, the only mappable carbonaceous rocks comparable to the Ottauquechee Formation directly overlie the Hoosac Formation. Either the Pinney Hollow rocks are absent and the Ottauquechee rocks rest directly on the Hoosac Formation or there is another carbonaceous unit in the section. Midway in the Rowe Schist there is another poorly developed and discontinuous unit of gray phyllite and associated quartzose granulites (see above) which may or may not be equivalent to the Ottauquechee Formation, depending on the fate of the Pinney Hollow Formation.

It has been demonstrated by Chidester and others (in press) and the writer's mapping that the Chester Amphibolite of Vermont (which was incorrectly correlated with the Chester

Amphibolite of Chester, Massachusetts) is not physically continuous south of the Vermont-Massachusetts border. Whereas the Chester Amphibolite in Vermont occurs just below the Ottauquechee Formation, the Chester Amphibolite of Massachusetts (Emerson, 1892) occurs just below the Moretown Formation, within the rocks that would be equivalent to the Stowe Formation of Vermont.

In summary, the writer and coworkers in the area map the rock sequence between the Hoosac Formation and the Moretown Formation as one unit, the Rowe Schist. The term Rowe Schist has been redefined from the original usage of Emerson (1898) because field mapping has demonstrated at least two black or gray carbonaceous bands that may be correlated on lithologic bases with the Ottauquechee Formation. The repetition of carbonaceous rocks has clearly not been caused by folding or faulting. It has been demonstrated that the sequence Ottauquechee Formation, Chester Amphibolite, Pinney Hollow Formation is not continuous and although the boundaries can be mapped the formations cannot be traced south from the Vermont border where the stratigraphy is last reported to be complete from the Hoosac Formation through the Moretown Formation. Skehan's section (1961) is actually less complete than he shows. The tectonic thinning around the Sadawga dome may have squeezed out the pelitic rocks. More than one greenstone unit has been found all of which were originally assigned to the Chester Amphibolite that was below the Savoy Schist of Emerson (1898).

These are at several stratigraphic horizons, and none of these rock units can be traced for more than 5 miles along strike.

Paleozoic Metasedimentary Rocks

Western Sequence

General Statement

West of the Precambrian terrane in the Windsor quadrangle the Cambrian and Ordovician sediments are more than 2,000 feet thick. These rocks are equivalent in age tothe rocks in the eastern sequence discussed above and are the miogeosynclinal facies equivalent of them. They are exposed in a belt trending north for the entire length of the quadrangle. No major unconformities are present during the Cambrian and the Lower Ordovician. Basal Lower Cambrian clastic sediments give way to calcite and dolomite marbles with only minor amounts of clastic rocks in the carbonate section. Carbonate deposition continued from upper Lower Cambrian to the upper Lower Ordovician. The Berkshire Schist of Middle Ordovician age lies unconformably on all the older Paleozoic rocks and the Hinsdale Gneiss. The Berkshire Schist is the youngest sedimentary rock present in the quadrangle. Because of the unconformity, most of the carbonate sequence has been removed. Exposures of the carbonates are poor.

The stratigraphic column is given in figure 4. The sequence from the Dalton Formation of Lower Cambrian age to the Berkshire Schist of Middle Ordovician age is consistent with the other studies in this area (Herz, 1958, 1961; MacFayden, 1956). The relationship of the Hoosac Formation to the western miogeosynclinal sequence is of particular interest because this is the only area south of Chittenden, Vermont (Brace, 1953; Doll and others, 1961) where the eastern rocks may be in non-faulted contact with the western sequence. The interpretation adopted by this writer is that the Dalton Formation conformably overlies the Hoosac Formation in the Windsor quadrangle.

The possibility of these two formations being in fault-contact will be discussed at the end of the structural geology chapter.

Hoosac Formation

General Statement:

The Hoosac Formation of Lower Cambrian age has been traced around the northern exposures of the Stamford Granite Gneiss in the southern part of the North Adams quadrangle by Herz (1961) and has been mapped to the south into the Windsor quadrangle. The formation crops out well on the western slopes of the Hoosac Range east of Adams. Commonly at the break in slope at the Precambrian-Cambrian contact continuous outcrops of the Hoosac Formation are present for hundreds of feet.

Only two of the rocks represented in the eastern sequence of the Hoosac Formation are found in the western sequence; the garnet schist and the albite schist.

Garnet Schist Member:

The garnet schist is well developed and extensively exposed in the northwestern and central part of the quadrangle. It lies directly on the Hinsdale Gneiss. South and southwest of Patton Brook, the basal rocks are a few tens of feet of brown-weathering biotitic quartzite. This rock grades abruptly into garnet schist both laterally and vertically. Mineralogically and physically it is somewhat different from the eastern garnet schist. It is typically rusty dark brown to black with a very irregular schistosity and dotted with garnets as much as one third inch in diameter. Bedding is usually indistinct and is brought out only by weathering. Minor conglomeratic lenses are present with channel fillings and graded bedding that indicate the section is structurally overturned.

Muscovite, paragonite, quartz, and almandite form the bulk of the rock. Accessory minerals include chlorite, biotite, albite, and ubiquitous carbonaceous dust. The latter occurs both as inclusions with a sigmoid distribution in porphyroblasts of almandite and albite and in the muscovite-rich portions of the rock (table 4). It is more abundant in the porphyroblasts than in the matrix.

Albite Schist:

Albite schist conformably overlies the garnet schist. It is poorly exposed on the slopes east of Adams.

The contact with the garnet schist is transitional. Both

TABLE 4

Estimated Modes

Hoosac Formation - Western Sequence

	Albite schist			Garnet schist	
Specimen	60-a	228-b	2173	37e	2235
Quartz	70	50	90	10	5
Muscovite	10	5	1	0	80[1]
Paragonite				45	
Albite	10-P	25-P	2	25-P	5-P
Microcline	2	tr	1		
Chlorite	tr	tr		9+1[2]	2
Garnet				5-P	5-P
Biotite	5	15	5	tr	
Epidote group		5	1		
Tourmaline					1-P
Magnetite	tr		tr	tr	tr
Apatite	2				
Zircon			tr	tr	
Graphite				tr	1

[1] = most abundant white mica but absolute amount undetermined

[2] = secondary

P = porphyroblastic

Location of Specimens[1]

Hoosac Formation - Western Sequence

Albite Schist:

60-a	Tan, laminated granular schist: roadcut on Main Road at the height of land approximately 4,600 feet west of Tomb Cemetery, (5).
228-b	Biotite-speckled, laminated schist: 3,200 feet north of Tomb Cemetery on Brown Road, (2).
2173	Gray, massive granulite: elevation 1,840 feet on Patton Brook, (2).

Garnet Schist Member:

37e	Silvery, grayish-green garnet schist: roadcut east of the intersection of Main Road and Windsor Road, (6).
2235	Carbonaceous, garnet schist: at elevation 1,770 feet, due west of hill X 1,843 feet, (1).

1 See figure 3.

interbedding and vertical gradation take place. The lower contact is placed at the first appearance of garnets in the more schistose rock. The upper contact with the Dalton Formation is placed at the first recognizable flaggy quartz-feldspar granulite of any significant thickness. This contact is exposed in the unnamed small brook that flows southwest into Patton Brook at an elevation of 1,510 feet. Bedding is parallel on both sides of the contact and there is no evidence of faulting. There is very little intercalation of the well-bedded Dalton Formation and the fairly massive Hoosac Formation.

The rock is a massive, tan to brown rock with very indistinct bedding. It is richer in quartz and far less schistose than the albite schist in the eastern sequence. Chlorite is present only in trace amounts.

Thickness:

The total maximum thickness of the Hoosac Formation west of the Precambrian terrane is approximately 300 feet thick, both members being equal in thickness. The formation thins to the south and is not present in the southwestern part of the quadrangle. The thinning to the south and southeast may be in part or totally caused by tectonic thinning.

Dalton Formation

General Statement:

The type locality of the Dalton Formation of Lower Cambrian age is just south of the Dalton Railway Station in the

Pittsfield East quadrangle (Emerson, 1899) where the unconformity of the basal conglomerates on the Precambrian rocks may be seen. A nearly complete section is present there.

South of Savoy Road in the Windsor quadrangle the rocks of the Dalton Formation were originally assigned to the Vermont Formation (Pumpelly and others, 1894) and then to the Becket Gneiss (Emerson, 1899) of presumed Cambrian age. They were interpreted as the transition between the basement rocks and the overlying Cheshire Quartzite. These rocks north of Savoy Road were included in the Vermont Formation of Cambrian age (Pumpelly and others, 1894). The later formation included all the felsic granulites west of the Stamford Granite Gneiss. These rocks were called Dalton Formation by Emerson (1917) and then divided into the Cheshire Quartzite and Mount Holly Complex by Prindle and Knopf (1932).

The descriptions of the Dalton Formation vary with the area and the worker. The term Dalton Formation as used by the writer includes the entire sequence of rocks between the Hinsdale Gneiss and the Cheshire Quartzite of Emerson (1892, 1917). Where the Hoosac Formation underlies the Dalton Formation the base of the Dalton is placed at the first flaggy granulites. The Dalton Formation is highly variable both vertically and along strike. Few distinctive strata may be followed for any great distance along strike.

Extensive sections of the Dalton Formation are exposed in the brooks that drain west from the Hoosac Range, on the northwest slopes of Weston Mountain, and in Dry Brook.

Four distinctive rocks are present within the Dalton Formation:

1. Conglomerate.
2. Feldspathic granulite, gneiss, and quartzite.
3. Massive quartzite.
4. Muscovite-quartz schist.

Conglomerate Member:

The conglomerate member is exposed only where the Dalton Formation lies directly on the Hinsdale Gneiss. It is thickest in the vicinity of Weston Mountain. Toward the north, just south of Harrington Road, the basal rock is a granulite containing granules. The maximum size of the pebbles is 6 inches in the south and decreases to the north. The majority of the pebbles are composed of quartz. The distinguishing blue quartz that is characteristic of many of the Precambrian rocks is usually restricted to a grain size less than 1/2 inch in maximum dimension.

The upper contact of the conglomerate member as shown on plate 1 is rather arbitrary because pebbly or granule-containing beds are common throughout the lower half of the Dalton Formation. The upper contact is placed at the last bed of conglomerate that contains pebbles larger than 1/2 inch in diameter.

The rocks range from quartz-cemented quartz-pebble conglomerate to quartz-pebble conglomerate with a muscovite matrix.

The color of the latter is generally a pale gray-green with a greasy appearance.

In the Weston Mountain area, channel fillings, graded beds, and cross beds indicate that the section is upright. The same criteria have been used in the northwestern corner of the quadrangle to establish that the stratigraphic column is inverted.

Estimates of the total thickness of this member are difficult to make because exposures are so poor. It probably is no thicker than 200 feet.

Feldspathic Granulite, Gneiss, and Quartzite:

Feldspathic granulites, gneisses, and quartzites comprise 75 percent of the Dalton Formation. They are thickest in the northwest and southwest parts of the quadrangle. Excellent sections are exposed in the gorges in Tophet Brook at an elevation of 1,450 feet, on Dry Brook between elevations 1,600 and 1,700 feet, on MacDonald Brook, and on South Brook in the vicinity of Weston Mountain.

The rock is typically buff to orange-brown or gray weathering. Beds range from very thin to 6 inches in thickness. The thicker flaggy beds are laterally very persistent. Cross bedding is rare. The rock is equigranular (1/4 to 1/2 mm.) and, because of the large proportion of granular minerals, weathers to a friable sandy surface. The rock is composed of quartz, muscovite, and biotite with variable amounts of microcline, albite, chlorite, and interstitial calcite cement. Accessory minerals include nearly ubiquitous tourmaline, magnetite, apatite, sphene, and zircon. Quartz is commonly in excess of 80 percent of the rock. Representative modes are given in table 5.

The albite commonly has sigmoid patterns, marked by the numerous inclusions of quartz, muscovite, and rarely carbonaceous dust. It rarely exceeds 5 percent. Microcline is generally clear (not poikiloblastic), very fine-grained, and apparently detrital. It ranges from 0 to 10 percent in abundance. The biotite is usually orange-brown to yellow-brown or slightly olive-brown, whereas the biotite in the adjacent Hinsdale Gneiss is typically olive in color.

Individual beds are extremely homogeneous indicating constant conditions of deposition or complete reworking of all sedimentary structures. Rare cross-bedding indicates that the source of the material is to the east.

The protolith of this rock would be classified as an arkose, quartz sandstone, or graywacke depending on the proportions of phyllosilicates and feldspar.

Massive Quartzite Member:

Pure, massive quartzites as much as 20 feet thick occur in the Dalton Formation. These rocks are identical with the Cheshire Quartzite and where they are thick and outcrop is sparce, it is not possible to distinguish the Dalton Formation

TABLE 5

Estimated Modes

	Dalton Formation					Cheshire Quartzite
Specimen	545-a	720	1304	1663	1677	546-b
Quartz	8	85	73	98	75	99
Muscovite	80	5	20	1	2	tr
Albite	2	1			15	
Microcline		5	tr	tr	5	
Chlorite		1		tr		
Biotite	10	1	5	tr	2	
Tourmaline					tr	
Magnetite	tr	1		tr	tr	tr
Hematite						tr
Apatite				tr	tr	
Calcite + dolomite		1			tr	
Zircon		tr				tr
Sphene				tr		
Graphite	tr		2			

Location of Specimens[1]

Dalton Formation

545-a Gray schist: 1,300 feet from the intersection of Windsor Road and Sand Mill Road at elevation 1,600 feet on Windsor Road, (4).

720 Thinly laminated, light gray gneiss: intersection of Hoosac Street and Walling Road in Anthony Brook, (1).

1304 Sooty, rusty, thinly laminated schist: elevation 1,500 feet, on ridge, 1,850 feet southeast of hill X 1,231 feet, (4).

1663 Laminated, gray, quartzite: elevation 1,540 feet on brook flowing into South Brook, (7).

1677 Massive, gray, biotite-speckled granulite: elevation 1,700 feet on South Brook, (7).

Cheshire Quartzite

546-b Massive, white quartzite: 300 feet southeast from the intersection of Windsor Road and Sand Mill Road, in Dry Creek, (4).

[1] See figure 3.

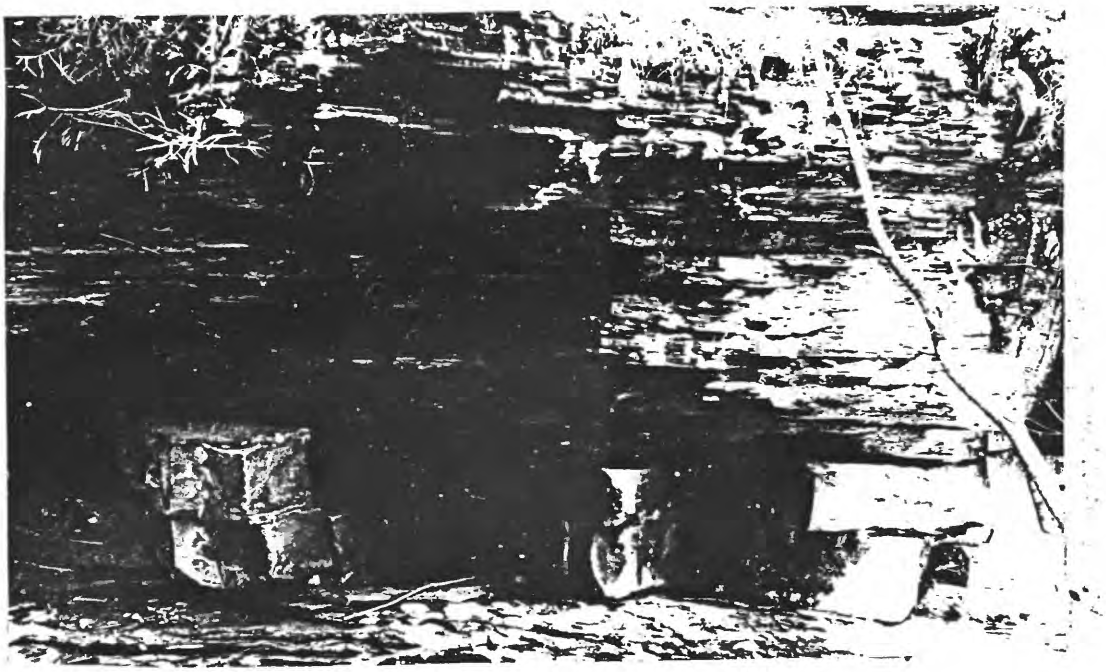


Figure 5. Interbedded flaggy granulite and massive quartzite in the Dalton Formation. The quartzite is 3 feet thick. Elevation 1,580 feet on South Brook.

and the Cheshire Quartzite.

In the northwestern part of the quadrangle, there are at least two of these massive quartzites in the Dalton Formation that are mappable for two miles along strike. Typically they are massive, unbedded quartzites, ranging from 10 to 20 feet in thickness and are interbedded with the thin-bedded rocks of the Dalton Formation. Very few thin beds of quartzite are present in this part of the quadrangle. In the southwestern part of the quadrangle, both thin- and thick-bedded quartzites occur throughout the Dalton Formation. They are typically glassy, weather orange to buff, and are identical with the Cheshire Quartzite.

These massive but discontinuous quartzites make mapping difficult in the Weston Mountain area. Several doubly plunging folds have been shown in this area based on the outcrop pattern of rocks presumed to belong to the Cheshire Quartzite. It is possible that rocks presumed to be part of the Cheshire Quartzite may be interbeds of Cheshire-like rocks in the Dalton Formation. However, analysis of structural features associated with the surrounding rocks supports the interpretation adopted.

Muscovite-quartz Schist:

Muscovite-quartz schist is commonly present at the contact between the Dalton Formation and Cheshire Quartzite. It is exposed because of the resistance of the overlying Cheshire Quartzite. Good exposures are present in Dry Brook at an elevation of 1,560 feet and on the ridge extending south from MacDonald Brook toward Weston Mountain at an elevation of 1,480 feet. Muscovite and paragonite comprise 20 to 80 percent of the rock. Quartz ranged from 10 to 75 percent. Accessory minerals include albite, microcline, biotite, opaque minerals, and as much as 2 percent carbonaceous dust. This rock corresponds very closely with the Moosalamoo member of the Mendon Formation and equivalent rocks in southern Vermont. (Keith, 1932; Prindle and Knopf, 1932, p. 272)

Thickness:

The Dalton Formation is less than 400 feet thick in the northwestern part of the quadrangle and less than 500 feet thick in the southwestern corner.

Age:

An age of Lower Cambrian is assigned to the Dalton Formation. Walcott (1888, p. 236) found fragments of a trilobite, apparently the genus Olenellus, in a quartzite bed about 100 feet above the unconformity to the northwest on Clarksburg Mountain. Further north 2 miles east of Bennington he also found Olenellus. Although the horizon in which the Olenellus fragments were found was originally assigned to the Cheshire Quartzite (Whittle, 1894), Prindle and Knopf (1932) placed this horizon in the Dalton Formation since it underlies carbonaceous muscovite schists which grade upward into the Cheshire Quartzite (Osberg, 1952). This is the same sequence seen in central Vermont in the Dalton Formation. The Dalton Formation is equivalent to the Pinnacle Formation (Doll and others, 1961) of northern Vermont.

Cheshire Quartzite

General Statement:

The Cheshire Quartzite of Lower Cambrian age derives its name from localities east of Cheshire, Massachusetts where it forms knobs and ridges on the western dip-slope of North Mountain in the Cheshire quadrangle (Emerson, 1892).

The Cheshire Quartzite crops out poorly in the northern part of the Windsor quadrangle, except for the top of Burlingame Hill on White Road; in the southern part, extensive outcrops form dip-slopes and cap ridges.

The lower contact of the Cheshire Quartzite is placed at the top of the muscovite-quartz schist of the Dalton Formation. The upper contact is apparently sharp although it has not been observed. In the gorge in Tophet Brook just east of Adams, an outcrop of Kitchen Brook Dolomite is on the west side of the brook, whereas a friable clean glass sand typical of the Cheshire Quartzite is on the east side. There is a sharp break from clastic to carbonate sedimentation.

The typical outcrop of Cheshire Quartzite has no discernible bedding except in pits where it has been worked for glass sand. Bedding is marked by very thin layers rich in micaceous minerals. Beds originally rich in feldspar and subsequently kaolinized are rare in the Windsor quadrangle but are reported by Herz (1958, 1961) and Chute (1943). This feature is present in the road cut at the height of land between Adams and Savoy on Main Road. Although Herz (1958, 1961) mentions a kaolinized zone at the top of the Cheshire Quartzite, this writer did not find any such zone within the Windsor quadrangle.

The rock ranges from a massive vitreous rock to a friable disaggregated massive rock.

Petrology:

The fresh and weathered rock is white, slightly yellow-buff, or reddish. The color is a function of the impurities in the siliceous cement and is not controlled by the bedding.

There is a wide range of grain size (very fine- to coarse-grained sand), the average grain size being about 1/2 mm. The grain boundaries are highly sutured. None of these grains have detrital outlines or siliceous overgrowths. Quartz composes more than 99 percent of the rock. The individual grains have abundant tiny crystallite inclusions uniformly disseminated throughout the entire grain.

In the friable portions of the formation, disaggregation of the quartz grains has occurred around crystallographically homogeneous and continuous grains, never across them. This suggests that the disaggregation has not been caused by tectonic crushing of a well-indurated quartzite.

The friable unconsolidated portions of the formation[3] have served as a minor source for glass sand (Chute, 1945). The formation can be traced along strike from friable glass sand into vitreous quartzite in a short distance. The distribution of the glass sand within the formation seems unrelated to structural position.

Evidently when the original sediment was cemented by silica, there was some chemical control (possibly related to sedimentological controls) that prevented the entire unit from being indurated. No calcite cement is present in the indurated rock but "concretions" of glassy quartzite have been found entirely surrounded by friable quartzite. There is no major mineralogical difference between the uncemented rock and the vitreous rock although there is a suggestion that there are more iron oxides in the friable quartzite. It is not clear whether these iron oxides were part of the original sediment or a result of much later infiltration of iron-precipitating solutions (primarily) into the more porous rock with deposition of iron oxides.

Thickness:

Although the upper and lower contacts of the Cheshire Quartzite have not been observed, the formation is not believed to be more than 100 feet thick within the quadrangle. In the central part of the quadrangle, the thickness is less than 50 feet and in the eastern-most exposure it is 2 feet thick.

Age:

No fossils have been found in the Cheshire Quartzite in this area, but Walcott (1888) found fragments of Hyolithes, and Olenellus 2 miles west of Bennington, Vermont in his "Cheshire Formation", a horizon apparently in the transitional zone between the Dalton Formation and the massive quartzites of the Cheshire Quartzite. This would be stratigraphically above the Olenellus horizon on Clarksburg Mountain and below the muscovite-quartz schist of the Dalton Formation. The Cheshire Quartzite is considered to be Lower Cambrian in age.

Kitchen Brook Dolomite

General Statement:

The Kitchen Brook Dolomite of upper Lower and Middle Cambrian age was named by Herz (1958). The type locality is on Kitchen Brook in the Cheshire quadrangle. Herz was unable to locate the contact with the underlying Cheshire Quartzite or the overlying Clarendon Springs Dolomite. He correlated the Kitchen Brook Dolomite with the Dunham Dolomite-Monkton Quartzite-Winooski Dolomite sequence in Vermont (Cady, 1945). Because the counterpart of the Monkton Quartzite was not present in this area Herz mapped the entire unit between the Cheshire Quartzite and the distinctive Clarendon Springs Dolomite as one formation.

The Kitchen Brook Dolomite is exposed only in the northwestern corner of the quadrangle. To the south it is cut out by the Middle Ordovician unconformity and a thrust fault. A large section is exposed in the Tophet Brook gorge just east of Adams.

Petrology:

The rock typically weathers to a light-tan to buff sandy textured surface. In the middle of the formation it weathers white to mottled blue and white. Bedding is indistinct and irregular except where there are thin interbeds of calcareous dolomite marble or closely spaced partings rich in muscovite. Bedding is better defined in the upper part of the formation, 1 to 2 inch beds being the most common.

Sedimentary structures are rare. One questionable cross-bed indicates that the section is upside down.

Mineralogically the rock is a dolomite marble quite low in accessories (table 6). The sandy texture of the weathered samples is the result of the presence of a few percent of quartz. Accessory minerals include quartz, muscovite (concentrated along bedding planes), and microcline (disseminated and rare). The disseminated muscovite gives a discernible non-penetrative foliation in an otherwise structureless equigranular (1/2 to 1 mm.) rock.

Thickness:

The lower contact of the Kitchen Brook Dolomite is well located in Tophet Brook but the upper contact is poorly located. The thickness probably does not exceed 300 feet in

TABLE 6

Estimated Modes

	Kitchen Brook Dolomite	Clarendon Springs Dolomite	Shelburne Marble	Bascom Formation
Specimen	682	2777	2779	2808
Quartz	tr	tr	tr	3
Muscovite	tr	tr	tr	2
Albite			tr	1
Microcline	tr	tr		2
Magnetite				tr
Calcite + dolomite	99	99	99	90
Rutile	tr			
Graphite				tr

Location of Specimens

Kitchen Brook Dolomite

682 White dolomite marble with blue mottling: 3,300 feet north of Leonard Road on Savoy Road, (1).

Clarendon Springs Dolomite

2777 Sugary, yellowish-white dolomite marble: elevation 1,180 feet, due west of Hales Hill summit, (1).

Shelburne Formation

2779 Sugary, light-buff calcite marble: elevation 1,115 feet N.86°W. of Hales Hill summit, (1).

Bascom Formation

2808 Blue-gray, fine-grained dolomitic calcite limestone: elevation, 1,010 feet on Dry Creek, (1).

the quadrangle.

Age:

The Kitchen Brook Dolomite is unfossiliferous in this area. If the correlation with the Vermont sequence is correct, the Kitchen Brook Dolomite is Lower Cambrian and possibly Middle Cambrian (Cady, 1945; Osberg, 1952).

Clarendon Springs Dolomite

General Statement:

The name Clarendon Springs Dolomite was originally applied by Keith (1932) for all the dolomitic beds lying between the arenaceous beds of the Danby Formation and the non-dolomitic Shelburne Marble in west-central Vermont.

The Clarendon Springs Dolomite in this area includes all the rocks between the Kitchen Brook Dolomite and the Shelburne Marble, and is exposed only on the west slope of Hale's Hill in the northwestern corner of the quadrangle. It is equivalent (Herz, 1961) to the Clarendon Springs Dolomite and Danby Formation of Vermont. No rocks similar to the Danby Formation are found in the Windsor quadrangle. The Clarendon Springs Dolomite is cut out by the Middle Ordovician unconformity and the thrust fault southeast of Hale's Hill.

Petrology:

The dolomite marble is typically white to gray, rarely weathering to light buff. It is recognized by the presence of quartzitic stringers and knots that are presumably highly deformed beds. Cellular weathering is common in the lower part of the formation where the quartz-rich layers contain calcite. In the upper part of the formation the bedding is more pronounced because of muscovite along bedding planes; here there are fewer quartzitic layers. Beds range from laminae to 12 inches in thickness in the upper part of the formation. Grain size ranges from 0.1 to 1.0 mm. Dolomite is the dominant mineral; microcline and muscovite are the common accessory phases. Quartz and calcite are locally abundant.

Thickness:

Neither the upper nor the lower boundary is exposed; however, the upper contact is placed at the base of the first bed of massive calcite marble of the Shelburne Marble.

The estimated thickness for this formation is 300 feet. Herz (1961) estimates the thickness to be 800 feet in the North Adams quadrangle and 0 to 800 feet in the Cheshire quadrangle.

Age:

The Danby Formation and Clarendon Springs Dolomite probably span the late Cambrian and early Ordovician epochs, (Cady, 1945).

Shelburne Marble

General Statement:

The Shelburne Marble, named after the type locality in Shelburne, Vermont, is the formation in which all the presently operating lime quarries in the Adams area are located. The Shelburne Marble was originally described by Keith (1932). Bain (1931) had divided the Boardman Formation (equivalent to the Shelburne Marble plus a few feet of overlying rock) into three members. These divisions can not be recognized in the Windsor quadrangle.

The Shelburne Marble crops out on the west slope of Hale's Hill, where a 50 foot section is exposed; further south and north it is cut out by the Middle Ordovician unconformity.

Petrology:

The exposures on Hale's Hill are similar to the Columbian marble member of the Shelburne Marble. The rock weathers to a gray to white sandy surface similar to the dolomite marble in the Kitchen Brook Dolomite. Bedding is pronounced ranging from a few inches to more than a foot. Beds are separated by thin layers rich in muscovite. The individual beds have more than 99 percent calcite. Accessory minerals in the marble include quartz, plagioclase, and muscovite. The grain size is generally less than 1 mm.

Thickness:

The contact between the Clarendon Springs Dolomite and the Shelburne Marble on Hale's Hill is known within 10 feet; the position of the upper contact is less certain. On the basis of mapping, the formation is estimated to be no more than 100 feet thick in the map area.

Age:

The Shelburne Marble is equivalent to the "Beekmantown Division B" and is of Lower Ordovician age (Cady, 1945, p. 545).

Bascom Formation

General Statement:

The Bascom Formation of Lower Ordovician age was first described by Cady (1945) in central Vermont. It crops out in the Windsor quadrangle only in the lower section of Dry Brook.

Petrology:

The exposures of Bascom Formation in Dry Brook consist of massively bedded, fine- to medium-grained (1 mm.) dolomitic calcite marble, with rare, closely spaced micaceous partings. The fresh rock is light gray to white or more commonly mottled blue and gray and weathers to either dusty gray or orange.

The marble contains as much as 10 percent accessory minerals. One mode is given in table 6.

Thickness:

The thickness of the Bascom Formation is unknown because neither the upper nor the lower contact is exposed, and it is overlain unconformably by the Berkshire Schist. It is cut out by the Middle Ordovician unconformity to the south.

Age:

The formation is assigned to the upper part of the Beekmantown Group and is therefore of upper Lower Ordovician age (Cady, 1945).

Berkshire Schist

General Statement:

The Berkshire Schist is named for occurrences in Berkshire County, Massachusetts. As originally defined by Dale (1891, p. 6) it included those schists on the lower slopes of Mount Greylock up to but not including the Bellowspipe Limestone. Herz (1958) redefined the Berkshire Schist to include the Bellowspipe Limestone as a lentil in the upper part of the Berkshire Schist, thus including some of the Greylock Schist (Dale, 1891) in the Berkshire Schist. This sequence in turn is overlain by the chloritoid schist of the Greylock Schist, that is presumably part of the Taconic thrust sheet. Since the equivalent rocks in the Windsor quadrangle do not contain the massive limestone (Bellowspipe equivalents) either definition of the Berkshire Schist will fit the rocks described below.

The Berkshire Schist crops out in several areas in the western part of the Windsor quadrangle, and forms Stafford Hill.

The Berkshire Schist, southwest of Adams, is unconformable on the Bascom Formation and Shelburne Marble. To the southeast this unconformity cuts down through the section. The Berkshire Schist lies directly on the Precambrian rocks just south of Savoy Road. South of this area, the Berkshire Schist lies unconformably on and in fault contact with the Cheshire Quartzite. The unconformity is not exposed. In the south-central part of the quadrangle, there is a syncline of the Berkshire Schist within the Hinsdale Gneiss. Emerson (1899) mapped this as Hoosac Schist after first making a tentative correlation with the Dalton Formation. He later (1917) reinterpreted it as the Berkshire Schist.

The formation has three important rock types:

1. Schist and phyllite
2. Black quartzite
3. Calcareous rocks

These rocks are more or less carbonaceous and some are sulfidic and form sooty black to rusty-brown exposures. Bedding is indistinct in the schistose units; where discernible, it is usually indicated by quartz-rich beds. Disharmonic folding results in highly contorted exposures.

Because the formation has undergone extensive folding, bedding is hard to trace and facies changes, both vertical and horizontal, are difficult to establish.

In general the basal beds of the formation reflect the rock types of the rocks they truncate. Where the Berkshire

Schist lies on rocks of the Dalton Formation, Cheshire Quartzite, or Hinsdale Gneiss the basal beds are commonly quartzites. Highly carbonaceous schistose calcite marble forms the basal beds where the underlying rock is a carbonate. Away from the erosional unconformity, the schistose marble and quartzite are rare. Abundant outcrops of thin black quartzites are present on the top of Stafford Hill.

Schist and Phyllite:

The most common rocks in the Berkshire Schist are schists and phyllites. Bedding is rare; the rock is schistose and highly contorted with usually more than one planar surface developed and several sets of crinkles. The predominant schistosity is parallel to bedding. The schistosity is cut by a later axial plane cleavage along which there is little or no recrystallization.

The schist is a quartz-mica-albite rock with variable amounts of chlorite, chloritoid, almandite, or biotite. The white mica is either muscovite or muscovite and paragonite. Carbonaceous material is ubiquitous and commonly gives the rock a sooty appearance.

The phyllite contains the same minerals but commonly has more paragonite, chloritoid, and garnet. Porphyroblasts of garnet and albite are as much as 5 mm. in diameter. They both have inclusions with a sigmoid distribution, The sigmoid pattern is marked by inclusions of quartz, muscovite, and ubiquitous carbonaceous dust. The albite commonly has a poikiloblastic texture, whereas the garnets are more commonly non-poikilitic. Commonly the garnets show a shattered texture which has been subsequently smeared out by later deformation. The albite in the same rocks appears shredded, shearing having taken place along the (010) cleavage.

Chloritoid and biotite are commonly fine-grained (0.1 mm. or less) and constitute a small fraction of the rock. Chloritoid is common as inclusions in the porphyroblasts of garnet but rare in the matrix of the porphyroblasts.

Quartz grains are commonly highly sutured where they predominate. The muscovite and quartz are usually equigranular and form a matrix with a grain size of 0.5 mm. or less for the porphyroblasts.

Accessory minerals are magnetite, pyrite, clinozoisite, zircon, and microcline. Microcline with albite overgrowths has been observed (No. 1981). Poikilitic calcite is rarely present in the schistose rocks.

The protolith of these rocks was a shale rich in aluminous clays. Some of the specimens are richer in sodium than normal shales, (table 7).

Black Quartzite:

Abundant black quartzite is found at the Hoosic River in the northwest corner of the quadrangle; on Stafford Hill, especially on the northern slopes; and on and south of Savoy

TABLE 7

Estimated Modes

Bérkshire Schist

Specimen	517-a	531	1476	1873	1908
Quartz	2	75	7	1	15
Muscovite[1] / Paragonite	83	2	85	86	5
Albite		tr			2
Microcline			5-P		
Chlorite				2	
Garnet	5	5		5	
Chloritoid	9				
Biotite		15	3-P	5	5
Epidote group					2
Magnetite	tr	1-P		1	
Tourmaline	tr				
Hematite	tr[2]			tr	
Calcite + dolomite					70
Graphite		1			1

1 = relative amounts of muscovite and/or paragonite not determined

2 = secondary

P = porphyroblastic

Location of Specimens[1]

Berkshire Schist

517-a Carbonaceous, garnet phyllite: elevation 1,450 feet on Windsor Road, (4).

531 Dark gray, biotite speckled, fine-grained schist: 500 feet northeast of bend on road on Stafford Hill (1,640 feet knob), (4).

1476 Biotite-speckled, rusty, gray schist: elevation 1,870 feet, 2,850 feet S.40°E. from B.M. 1,487 feet, (4).

1873 Dark gray, biotite-garnet schist: elevation 2,020 feet, 2,400 feet N.76°E. from Johnson △, (4).

1908 Black, carbonaceous, calcite marble: elevation 2,080 feet, 400 feet N.80°E. from Hill 2,103 feet, southwest of Harrington Road, (5).

[1] See figure 3.

Road near the town line between Cheshire and Savoy.

The rock is friable to vitreous. Bedding is well developed. Beds range from laminae to flaggy beds that are 2 to 3 inches thick. The grain size of all the constituents is less than 0.5 mm. and more commonly is 0.1 to 0.2 mm. Quartz grains have highly sutured boundaries. The textures are very similar to those observed in the Cheshire Quartzite.

The rock has between 70 and 95 percent quartz with variable amounts of muscovite, chlorite, and biotite, and very minor amounts of calcite. Albite does not exceed a few percent. Carbonaceous dust is ubiquitous and the rocks are usually sooty. A few beds have as much as 15 percent garnet.

Calcareous Rocks:

Calcareous rocks compose less than 1 percent of the Berkshire Schist. Carbonaceous calcite marble is present where the Berkshire Schist and Bascom Formation are in contact on Dry Brook and southwest of Harrington Road.

Calcite occurs sporadically in the garnet schist and quartzite as an accessory mineral. The association tremolite-talc was observed from calcareous garnet schist in the Hoosic River.

Thickness:

The thickness of the Berkshire Schist is difficult to determine because of the intense deformation. A reasonable thickness would certainly be in excess of 1,000 feet. Herz

(1958), to the west, estimated a minimum of 1,400 feet including the Bellowspipe Limestone and a few hundred feet of schist above this on Mount Greylock.

Age:

The Berkshire Schist was originally assigned to the "Silurian" (Ordovician) by Dale (1891, p. 6) although no fossils were found in this area. This unit correlates with the Hortonville Formation (Zen, 1959, p. 3) and Ira Formation (Zen, 1959, p. 3) which in turn are correlated with the Walloomsac Formation (MacFayden, 1956, p. 27) in southwestern Vermont (figure 2). Cady (1945) has found Middle Ordovician fossils (Trentonian) in the basal beds of the Hortonville Formation. All of these units have the same relationship to the Lower Ordovician and older rocks. They rest unconformably on all units from the Beldens Limestone (upper Lower Ordovician) or younger to the Precambrian. This unconformity is particularly clear in the Windsor quadrangle.

It is possible that the Berkshire Schist correlates with the carbonaceous schists of the Hawley Formation in the Plainfield quadrangle (see discussion below). These schists and the Berkshire Schist are similar in appearance. The former have been correlated with the Missisquoi Formation of Vermont that has been traced northward into the Eastern Townships of Quebec. Here it is known as the Magog Formation and contains graptolites of late Middle Ordovician age (Trentonian) (Ber-

1962). This is the same age as the Hortonville Formation and its equivalents in Vermont (Cady, quoted in Zen, 1964, p. 44).

Stratigraphic Problems

There are problems associated with mapping the Dalton Formation, Berkshire Schist, and Hinsdale Gneiss in the vicinity of Hill 2077 feet east of Fales Road. The Berkshire Schist in this area is only slightly carbonaceous and in places non-carbonaceous. It is also more granular because of an increase in feldspar at the expense of phyllosilicates. Here the Berkshire Schist overlaps the Dalton Formation onto the Hinsdale Gneiss. The basal beds of the Dalton Formation are only sparsely conglomeratic. Thus, this area marks the transition from a basal conglomerate to a pelite. Therefore, the Dalton Formation and Berkshire Schist may not be distinguishable on lithologic bases. Here Berkshire Schist has been mapped on the basis of being more rusty and poorly bedded.

One further problem is that the Dalton Formation and Hinsdale Gneiss are essentially structurally concordant at the contact. This structural parallelism persists for several hundred feet below the unconformity. Where the rocks are similar (quartz-feldspar-biotite-muscovite gneiss and granulite) it may be impossible to differentiate between the two. Where the contact is clearly defined by the basal conglomerate, there is no difference in the biotite and microcline above and below the unconformity and no reliable criteria were discovered by which the rocks could be distinguished.

Thus where the Dalton Formation has the basal conglomerate or is underlain by the Hoosac Formation, the position of the unconformity is certain. Where the basal rocks of the Paleozoic sequence are granulites and gneisses rather than conglomerate or schist, the position of the Precambrian-Cambrian unconformity is uncertain.

The syncline in the south-central part of the quadrangle (plate 1 and plate 2, section D) has caused stratigraphic problems in the past. Emerson (1899, 1917) was unable to demonstrate continuity of the carbonaceous schist in the syncline with the Hoosac Formation to the east. The author has been likewise unable to trace this fold either to the east or the west. Lithologically and texturally the rocks in the syncline are unlike the rocks of the Hoosac Formation to the east. Although the rocks in the syncline are garnetiferous in places, they are very highly carbonaceous and sulfidic. Furthermore the rocks do not correlate with the conglomerates in the Dalton Formation to the west. For these reasons these rocks are correlated with the Berkshire Schist. This would suggest that the area now occupied by the syncline of Berkshire Schist was a positive area by the end of the Lower Ordovician since no Cambrian sediments are present.

Sedimentological Significance of the Western Sequence

The sequence Cheshire Quartzite, Dalton Formation, Hoosac Formation represents a continuous unidirectional change in the sedimentary facies in this area. In the northwest part of the quadrangle sedimentation started with potash- and sodium-rich clays, detrital quartz, and probably illite. The upward trend is toward minerals less rich in sodium, potassium, iron, and magnesium and an increase in quartz. The lack of appreciable clastics (conglomerate has been found in few localities) in the lower member of the Hoosac Formation and the abundance of clay minerals indicate rather quiet near-shore marine sedimentation. The source of the sediments may have been local, perhaps to the southeast where the Hoosac Formation was not deposited. The very minor conglomerate and quartzite in the Hoosac Formation in the northwest suggests a local source for some of these rocks.

At the same time, to the southwest and southeast conglomerate was deposited.

With continued deposition the feldspathic sands of the Dalton Formation covered the Dalton conglomerate and Hoosac pelites. At places these sands were reworked and became well sorted and thin beds of pure quartz sand formed locally. Finally, the Cheshire Quartzite was deposited as a blanket over the area.

The sequence of quartzose sediments suggests that the source rocks (Precambrian) had been deeply saprolitized to furnish the material for the Hoosac pelite. The feldspathic sands were derived from less intensely weathered rocks.

The same sequence of sediments is found in the eastern sequence of rocks. Basal conglomerates thin to the north where pelites were deposited. These conglomerates grade vertically into pelitic and less quartzose rocks, indicating an end of the supply of saprolitized source area. Near the end of the Early Cambrian the source area apparently was reduced in relief and weathering processes were sufficiently effective so that only quartzose material was carried to the west. This period of sedimentation would correspond with the deposition of rocks of the Cheshire Formation to the west and possibly the black quartzites at the base of the Rowe Schist.

In the central part of the quadrangle, the entire stratigraphic section of Hoosac through Cheshire is present but the units are very thin, apparently pinching out to the south. Evidently this south-central part of the quadrangle was a positive area during early Cambrian times.

East-West Correlation of Paleozoic Rocks

A correlation across the northern half of the Windsor quadrangle is given in figure 4. Rocks in the Plainfield quadrangle are included in the correlation.

The Windsor quadrangle is one of the few places along the Green Mountain anticlinorium in Vermont and the Berkshire anticlinorium in Massachusetts where east-west correlations may be based on the two rock sequences being in non-faulted contact. The next area to the north where this may be done is in the vicinity of Chittendon, Vermont (Doll and others, 1961).

In the North Adams quadrangle, Herz (1961) has interpreted the Hoosac Formation west of the Hoosac Range as being older and equivalent to the Dalton Formation. Whether or not the Dalton Formation interfingers with the Hoosac Formation or is a lateral equivalent of it could not be ascertained. The interpretation adopted by the writer is that the Dalton Formation overlies the Hoosac Formation in the northwestern part of the quadrangle. This conclusion depends on the structural interpretation adopted. This will be discussed later.

Further north in the Wilmington, Vermont area (Skehan, 1963), the Hoosac Formation is in fault contact with the Cheshire Quartzite which in turn unconformably overlies the Stamford Granite Gneiss on Clarksburg Mountain. The Cheshire Quartzite in this area is interbedded flaggy granulites (Dalton type of the writer) and vitreous quartzite (Cheshire type of the writer) and its relationship to the Hoosac Formation is indeterminate.

North of Rochester, Vermont in the Lincoln Mountain area (Cady, 1962) where the Precambrian rocks plunge beneath the

Lower Paleozoic rocks, it is possible to trace units across the Green Mountain anticlinorium. Although the stratigraphic units differ somewhat from those in the Windsor area, there are marked similarities. The basal unit on both the east and west side of the anticlinorium is albite schist with the rocks grading west into the Cheshire Quartzite and east into the Camels Hump Group, the lower part of which can be correlated with the garnet schist and the albite schist of the Hoosac Formation.

The late Middle Ordovician Berkshire Schist crops out within 7 miles of the Cram Hill member of the Hawley Formation (late Middle Ordovician) in the Plainfield quadrangle. They are correlated on the basis of age and rock type. Because of crustal shortening, it is impossible to determine how far apart these two rocks were when they were deposited.

The Berkshire Schist is unconformable on the entire Cambrian and Lower Ordovician sequence at the edge of the positive Precambrian area while the Cram Hill member is part of a conformable sequence of clastics well to the east of the positive area.

Regional Correlation of Paleozoic Rocks

A correlation chart (figure 6) shows relationships between the stratigraphy of the Windsor area, Vermont, and southwestern Massachusetts. The most obvious differences between the Vermont and the Windsor sequences are as follows:

	EASTERN SEQUENCE				
	This report	East-central Vt. Chang, et al., 1965	This report	Southwest Massachusetts Zen, 196~~8~~6	West-central Vt. Doll, et al., 1961
MIDDLE ORDOVICIAN	Hawley Fm. Moretown Fm.	Missisquoi Fm.: Cram Hill mem Barnard Volc. Moretown mem	Berkshire Schist	Walloomsac Schist	Hortonville Fm.
LOWER ORDOVICIAN	Rowe Schist	Stowe Fm.	Bascom Fm. Shelburne Mar.	Stockbridge Group: Unit G Unit F Unit E Unit D	Beldens Ls. Bascom Fm. Cutting Dol. Shelburne Mar.
UPPER CAMBRIAN		Ottauquechee Fm.	Clarendon Springs Dol.	Unit C	Clarendon Springs Dol. Danby Fm.
MIDDLE CAMBRIAN		Pinney Hollow Fm.	Kitchen Brook Dol.	Unit B Unit A	Winooski Dol. Monkton Qtzite. Dunham Dol.
LOWER CAMBRIAN	Hoosac Fm.	Hoosac Fm. Tyson Fm.	Cheshire Qtzite. Dalton Fm. Hoosac Fm.	Cheshire Qtzite. Dalton Fm.	Cheshire Qtzite. Dalton Fm.
PRECAMBRIAN	Stamford Granite Gneiss	Mount Holly Complex	Hinsdale Gneiss	Berkshire Massif	Mount Holly Complex

Figure 6. Correlation of metamorphosed sedimentary rocks.

<u>Miogeosynclinal Sequence</u>:

1. Many of the formations found in central Vermont do not persist as mappable units to the south. These include the Monkton Quartzite, the Danby Formation, parts of the Shelburne Marble, and the Beldens Marble. These rocks may have been tectonically thinned or eliminated. To the south, Zen (1965) reports all of the units of central Vermont.
2. Those formations that persist south to the Windsor quadrangle appear to be much thinner. Tectonic thinning may account for much of this. Apparently this area was not subjected to the subsidence and thick accumulations of carbonate rocks that occurred in Vermont.
3. Except for the Berkshire Schist there are only minor amounts of clastics represented in the post-Cheshire section, indicating that this area must have been farther from the source area than were central and western Vermont. Cady (1945, p. 533, 537, and 546) summarized evidence that indicates that the source of most detritus was probably to the west.

<u>Eugeosynclinal Sequence</u>:

1. Nearly all the rocks beneath the Moretown member of the Mississquoi Formation of east-central Vermont are represented within the Windsor quadrangle. How-

ever, the Rowe Schist, equivalent to the Pinney Hollow, Chester Amphibolite, Ottauquechee, and Stowe Formations is much thinner.

2. Many of the metamorphosed volcanic materials in central Vermont probably represent lava flows, but comparable units in the Windsor area are missing. The greenschists are thin, discontinuous, and compositionally layered; they probably represent detrital volcanic material reworked by sedimentary processes. Evidently volcanic activity was more common in southern and central Vermont during the late Cambrian.

3. The conditions under which the protoliths of carbonaceous schists and quartzites in the Vermont sequence formed persisted at approximately the same time in Massachusetts, but euxinic conditions appear to have been less continuous along the strike of the eugeosyncline in Massachusetts, and were repeated several times. The one exception to this is the occurrence of carbonaceous phyllite and quartzite directly overlying the Hoosac Formation in the Windsor quadrangle. Carbonaceous phyllite, dolomites, and quartzites have been found at the top of the Hoosac Formation in central Vermont, especially in the Woodstock quadrangle (Chang and others, 1965), and these may be correlative. Hatch and Hartshorne

(in preparation) and Chidester and others (in press) mapping in the Heath and Rowe quadrangles respectively have not been able to map the Pinney Hollow Formation, the amphibolite beneath the Ottauquechee Formation, (Skehan, 1963), nor the Ottauquechee Formation.

Intrusive Mafic Rocks

Mappable serpentine and talc bodies occur at two localities in the quadrangle. The largest is the zoned ultramafic body in the center of the quadrangle. The maximum extent of the body is not known, but outcrops suggest it is at least 800 feet (north-south). A talc-carbonate zone at least 40 feet wide occurs around the serpentine core. Variable amounts of chlorite and carbonate occur with the talc and serpentine. Although this body is in the Hinsdale Gneiss it is thought to be Ordovician, consistent with the rest of the ultramafic bodies of western New England.

A talc body in the Rowe Schist in the east-central part of the quadrangle is discussed in the chapter on economic geology. It is not exposed at the surface but is shown on the map as an ultramafic because it has been mined. Apparently this body is entirely steatitized as the dumps contain no serpentine.

CHAPTER 3

STRUCTURAL GEOLOGY

General Statement

The Windsor quadrangle lies on the axis of the Berkshire anticlinorium in the northernmost part of the Berkshire Highlands. Within the quadrangle, the predominant regional strike is north. The Paleozoic rocks have undergone at least three phases of deformation.

The main structural feature in the quadrangle is the Hoosac nappe, a large recumbent anticline, facing west, the axis of which bisects the quadrangle and is coincident with the division of the Paleozoic rocks into miogeosynclinal and eugeosynclinal facies. The term Hoosac nappe is applied to the rocks in, north, and northwest of the Savoy Hollow Brook syncline (plate 5) and the entire homoclinal sequence of Paleozoic rocks east of the Stamford Granite Gneiss. Rocks and structures in the southwestern part of the quadrangle may also be part of the Hoosac nappe but have not been included because of complications to be discussed below.

The eastern Paleozoic homoclinal sequence is part of the upper normal limb; the Paleozoic rocks in the northwestern corner of the quadrangle form the inverted limb of the major recumbent anticline. Several recumbent anticlines and synclines are present in the inverted limb (plates 1, 2, and 5). Both the major and minor structures related to the nappe plunge gently to the north.

This regional structure has been only slightly altered by later tectonic events.

The crystalline core of the Hoosac nappe plunges beneath the present erosional surface 2 1/2 miles north of the quadrangle. The Cambrian-Precambrian contact reappears to the northwest on Clarksburg Mountain, some 5 miles away and to the north-northeast in the Sadawga Pond dome in Wilmington, Vermont (Skehan, 1961).

To the east of the structural axis, at least as far as the eastern border of the Plainfield quadrangle, the structure is essentially a steep homocline with minor isoclinal folds. The Hallockville Pond dome in the center of the Plainfield quadrangle is an intrusive rock (Osberg and others, in preparation) and interrupts the easterly dip of the beds. To the west of this axis in the Cheshire quadrangle, typical miogeosynclinal clastic and carbonate rocks (the Stockbridge sequence, Herz, 1958) are involved in the nappe to a limited extent. They show tight recumbent folds, thrusts, and then refolding.

Precambrian Structure

Poor exposures and the lack of any traceable distinctive strata in the Hinsdale Gneiss make it impossible to deduce any regional Precambrian structures. In the southeast part of the quadrangle, east of a line from Windsor Brook to the town of Savoy the bedding, foliation, and compositional banding in the Precambrian rocks are sub-parallel to the same features in the eastern Paleozoic metasediments (plate 3).

Folds in the Hinsdale Gneiss are either isoclinal with axial surfaces parallel to the local strike of the rocks or the folds are broad open warps with axial surfaces trending towards an east-west strike. The amplitude and wave length of these open folds are as much as 10 feet. They are probably Paleozoic in age.

These later folds are later than the folds responsible for the distribution of the Precambrian rocks. They fold both bedding and schistosity neither of which can be shown to be divergent from the compositional banding. These folds have neither an axial plane schistosity nor an axial plane cleavage, reinforcing the genetic relationship of these folds and the last Paleozoic folding.

The rocks west of the line from Windsor Brook to the town of Savoy and east of the Dalton Formation have unsystematic structures (plates 1, 3, and 4) and without stratigraphy it is impossible to distinguish antiforms or synforms. Only the Precambrian rocks directly beneath the Dalton Formation-Hinsdale Gneiss contact have structures parallel to the Paleozoic structures. Elsewhere in this area structure in the Hinsdale Gneiss appears unrelated to that in the Paleozoic rocks.

The unsystematic dips and strikes suggest either a highly refolded or faulted terrane or both.

In the northwest part of the quadrangle (plate 1), the bedding and foliation in the Hinsdale Gneiss are parallel to the schistosity in the Hoosac Formation near the contact of the two rocks. Away from the contact the strike of the Hinsdale Gneiss is sub-parallel to the Paleozoic rocks but the dip is highly varied. There it is not possible to trace a distinctive bed for more than a few tens of feet even with extensive outcrop. This suggests that these beds have been folded, non-cylindrically, more than once and there has been extreme deformation of the rocks in the core of the nappe, shearing off early structures and bringing the Precambrian structures into structural conformity with the mantling Paleozoic sediments.

The Stamford Granite Gneiss (plate 1) has a schistosity that is probably secondary, and imposed during the Paleozoic deformation responsible for the Hoosac nappe.

Paleozoic Structure - Folding

Hoosac Nappe

General Statement:

The Hoosac nappe is the dominant structure in the quadrangle. The original interpretation for the regional structure in the Windsor area was put forth by Pumpelly and others (1894). They postulated a regional anticline overturned to the west. They correlated the western exposures of the Hoosac Formation with the Berkshire Schist and did not recognize the erosional unconformity between the western Paleozoic sequence up through the Bascom Formation and the Middle Ordovician black schists, phyllites, and quartzites of the Berkshire Schist. The writers believed that the carbonate rocks and the Vermont Formation (the Cheshire Quartzite, Dalton Formation, and the Hinsdale Gneiss) graded laterally into detrital pelites.

Prindle and Knopf (1932) made the first attempt to revise the structural interpretation of the area. They postulated an overturned anticline to explain the distribution of the rock units and several low angle thrust faults striking north to bring the core of the nappe over the miogeosynclinal rocks. The most significant fault was the one placed at the contact of the Hoosac Formation and the Hinsdale Gneiss. Additional faults were postulated which followed the east-west trends of the synclines of the Hoosac Formation within the Hinsdale Gneiss on the inverted limb of the nappe (plates 1 and 5). The recognition of the structural and metamorphic discontinuity within the Vermont Formation was significant. The conglomerate of the Dalton Formation in the southwestern part of the quadrangle and the garnet schist of the Hoosac Formation in the northwestern part of the quadrangle were dated as Lower Cambrian or Late Precambrian. Older rocks were assigned to the Precambrian Mount Holly Gneiss (Hinsdale Gneiss of this writer), Becket Gneiss, and Stamford Granite Gneiss. These stratigraphic interpretations made possible a simpler structural interpretation.

Herz (1958) mapped the Berkshire Schist as a Middle

Ordovician clastic unit unconformably overlying the Cambro-Ordovician rocks of the miogeosynclinal sequence. In the North Adams quadrangle, Herz (1961), on the basis of stratigraphic throw and mineralized fault breccia in the Cheshire Quartzite in the Windsor quadrangle, retained the Hoosic thrust (plate 5). He also postulated a thrust fault to explain the relationships between the garnet schist of the Hoosac Formation and the structurally overlying Hinsdale Gneiss.

The stratigraphic and structural picture was confused by Herz's (1961) use of the term Dalton Formation for the conglomerate at the northern end of the exposures of the Stamford Granite Gneiss in the south part of the North Adams quadrangle. Here the conglomerate is sub-Hoosac and further to the southwest the Dalton Formation is post-Hoosac. They are pebbly conglomerates derived from the Stamford Granite Gneiss with many of the pebbles containing augen of microcline and lenses of cataclastic blue quartz. The pebbles are in an albitic matrix. The rock grades vertically into the albite schist of the Hoosac Formation with no stratigraphic break.

Christensen (1963), in his study of the nappe, used the geologic map of Prindle and Knopf (1932) with slight modifications. He eliminated the low angle thrust faults which Prindle and Knopf introduced to explain the east-west exposures of the Hoosac Formation and the low angle thrust fault between the Hinsdale Gneiss and the Hoosac Formation that Prindle and Knopf (1932) postulated and Herz (1961) retained; he retained the stratigraphic interpretation of Pumpelly and others (1894), who had grouped the Hinsdale Gneiss, Dalton Formation, and the conglomerate at the base of the Hoosac Formation at the northern end of the Stamford Granite Gneiss exposures in the North Adams quadrangle in the Vermont Formation.

Pumpelly and others and Christensen have a sandwich of Hoosac Formation with continuous deposition of the units above and below, but not including it. Christensen constructed an orthographic projection of the nappe by projecting the entire nappe structure down the axis of plunge, neglecting the fact that some of the structure that he was projecting was not cylindrically folded.

Geometry:

The rocks involved in the Hoosac nappe include the Stamford Granite Gneiss, the Hinsdale Gneiss, and the Lower Paleozoic rocks to the east and west (plate 1). The Berkshire Schist is the youngest rock involved in the nappe and is in the core of the major recumbent syncline.

The eastern limb is a normal homoclinal limb, the dip of which flattens out to the west as the axis of the nappe is approached.

The Stamford Granite Gneiss and Hinsdale Gneiss form the crystalline core of the nappe.

In the northwestern part of the quadrangle, the stratigraphic sequence from the Hoosac Formation to the Berkshire schist is inverted and forms the inverted limb of the major recumbent anticline in the nappe. Here the structure of the nappe is essentially cylindrical (figure 7). The axial surface is flat-lying, or nearly so, except where it has been locally warped and folded about northerly axes. The whole comple~~s~~x plunges slightly to the north (plates 2, 3, and 4).

Continuing to the south within the inverted limb the structure of the nappe remains essentially constant. Dips for bedding and schistosity are horizontal or gentle to the east. Minor folding related to the nappe is seen in the Hinsdale Gneiss, the garnet schist of the Hoosac Formation, the Shelburne Marble, and the garnet schist of the Berkshire Schist.

There are three recumbent synclines (plates 1 and 5) in the inverted limb of the nappe. They are all exposed on the south slopes of the hills immediately north of Savoy Road near the Cheshire-Savoy town line. Here the outcrops of the Hoosac Formation and axial traces of the recumbent folds take a sharp swing to the east, then northeasterly, then southeasterly, and then southerly, just west of Savoy. The dips in this area are gentle to the north and outcrop patterns are determined by the topography.

In the upper syncline (the Savoy syncline) where the axial trace trends easterly, the albite schist and the garnet schist of the Hoosac Formation are present in the fold. The former cannot be traced continuously to the east and west as

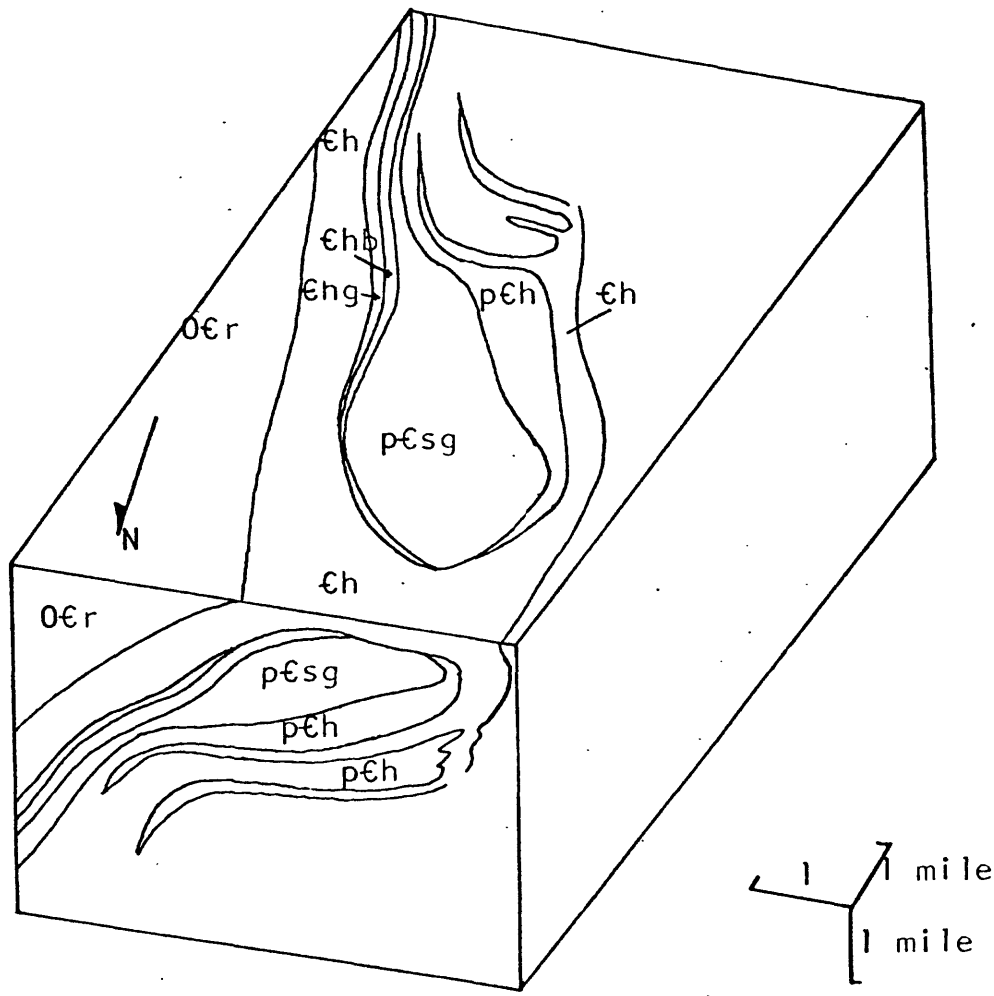


Figure 7. Orthographic block diagram of the structure of Hoosac Mountain.

can the garnet schist. However, measurements on minor recumbent folds indicate that the syncline does project down the northerly plunge of the nappe. The albite schist has apparently been either pulled apart because of flowage or it did not have a continuous distribution prior to the formation of the nappe. The former is more plausible. South of Patton Brook this syncline is stratigraphically symmetrical with biotite quartzites at the base of the garnet schist on the inverted and normal limb with garnet schist of the Hoosac Formation in the middle. Where the Savoy syncline swings to a northerly strike, the dip of the axial surfaces steepens to nearly 70° east.

The middle syncline involves only the garnet schist of the Hoosac Formation. This fold has been closed to the east.

The map pattern of the lowest syncline (Savoy Hollow Brook syncline) is almost identical with that of the Savoy syncline. In the middle and upper synclines the only rock present is the Hoosac Formation but the Paleozoic stratigraphic section up through the Cheshire Quartzite is present in the inverted limb of the Savoy Hollow Brook syncline. The normal limb is occupied by the Berkshire Schist. The Middle Ordovician unconformity is south of the axial trace of the syncline. Here quartzite and feldspathic granulite compose greater than 50 percent of the Berkshire Schist. This locality also coincides with the southern limit of outcrops of the Hoosac Formation on the west side of the

Berkshire Highlands. The relatively thin section (probably less than 200 feet) is apparently continuous in the fold for more than 3 1/2 miles. The last place where this syncline is seen is in Drowned Land Brook southwest of Savoy where the fold is apparently not symmetrical; the relationships are not clear. The normal limb may be sheared out. Here the stratigraphic section including the Hoosac and Dalton Formations and Cheshire Quartzite is less than 15 feet thick. The Cheshire Quartzite maintains a thickness of about 3 feet and can be traced with continuous outcrop for 2,000 feet.

Prindle and Knopf (1932) proposed low angle faults with east-west traces to explain the easterly trend of the Hoosac Formation. Although they do not state the evidence seen in the field for these faults, they are presumably based on the apparent lack of stratigraphic symmetry in the Savoy Hollow Brook syncline.

Near the glass sand outcrop 5,000 feet west of Tomb Cemetery on Main Road, the rocks of the Dalton and Hoosac Formations interfinger and grade laterally into each other. The formations in this area also have either been tectonically thinned or were originally thin. Therefore it seems unreasonable to expect the folds to be stratigraphically symmetrical across the axial surfaces if there are rapid facies changes within several hundred feet. In fact it is in this area that the garnet schist of the Hoosac Formation lenses out into the conglomerate of the Dalton Formation.

Other Major Folds

South of Savoy Road and east of Fales Road there are several recumbent folds, less spectacular than the recumbent synclines north of Savoy Road. They are probably related to the nappe because they are similar in attitude to the synclines north of Savoy Road.

The first of these is a recumbent syncline about 1 mile south of Savoy Road. A thrust fault with an easterly trend has sheared out part of the normal limb. The axial trace of the syncline has an easterly trend; the axial plane dips 15 to 20^{o} north.

South of this fold are two smaller recumbent synclines. The axial planes strike east and dip more steeply to the north than any of the other axial surfaces of the other recumbent folds. The interfolding of the Hinsdale Gneiss and Berkshire Schist that occupies the southernmost small syncline is well exposed on the southern slopes of the hill 3,000 feet northeast of the cornerpost for Cheshire, Windsor, and Savoy. Bedding attitudes and the outcrop pattern for the Berkshire Schist suggest the plunge of the syncline south of hill 2,068 feet is nearly northerly, down the dip of the axial surface.

The southernmost recumbent syncline (Dry Brook syncline, plate 5) is exposed in Dry Brook. It is occupied by the Dalton Formation, Cheshire Quartzite, and the Berkshire Schist and is stratigraphically symmetrical. Conglomerate in the Dalton Formation is exposed on the normal limb just north of Dry Brook. One thousand feet to the north the conglomerate is exposed and dips under the Hinsdale Gneiss. The axis of the fold must be a little north of the Middle Ordovician unconformity in the normal time.

To the south the structure is relatively simple; all beds are upright (plate 1) on a gently dipping homocline which may be the lower normal limb of the westward facing regional recumbent syncline.

West of B. M. 2,106 on North Street the distribution of rocks of the Dalton Formation suggests a refolded structure. Control is poor and the structure is not clearly delineated.

In the south-central part of the quadrangle the Berkshire Schist occupies the core of an isoclinal overturned syncline facing west. The shape of the fold is not well known. Outcrops of the Berkshire Schist are found 1,500 feet northwest of B. M. 1,851 in the unnamed brook, in the unnamed brook southeast of the intersection of Back Dalton Road and the Berkshire Trail, and on the small knob east of Back Dalton Road. The axial plane strikes about N. 30° E. and dips 30° (at the southern edge of the quadrangle) to 85° E. (at the northernmost exposure of Berkshire Schist).

An inverted anticline (Dry Brook anticline, plate 5) is present in the western-most part of the quadrangle in Dry Brook. Here, the Bascom Formation in the inverted limb of the nappe is infolded with the Berkshire Schist that occupies the core of the major recumbent syncline (sections A and B, plate 2). The axial surface of this fold dips easterly. The fold faces to the east. This fold has the opposite shear sense for a drag fold on the inverted limb of a syncline facing toward the west. It must be a cylindrically refolded fold.

Minor Folds - Eastern Sequence

General Statement:

East of the Stamford Granite Gneiss in the Windsor quadrangle, axial surfaces and fold plunges with three orientations and two styles of folding have been recognized.

First Generation Folds:

The oldest folds that can be recognized are the most common and conspicuous type of deformation in the rocks of the Hoosac Formation and Rowe Schist. Folds are typically isoclinal with small amplitude ranging from a few inches to several feet (figure 8). Axial planes strike northerly and dip steeply to the east or are vertical (plate 3). Plunges are generally down the dip (these are the early folds of plates 2 and 3). The steeper the dip of the bedding or axial plane schistosity, the more precisely the values of plunges approximate the dip. The folds fold bedding but the axial plane schistosity clearly cuts across bedding in the noses of folds. The intersection of this axial plane schistosity and the earlier bedding schistosity yields a strong lineation

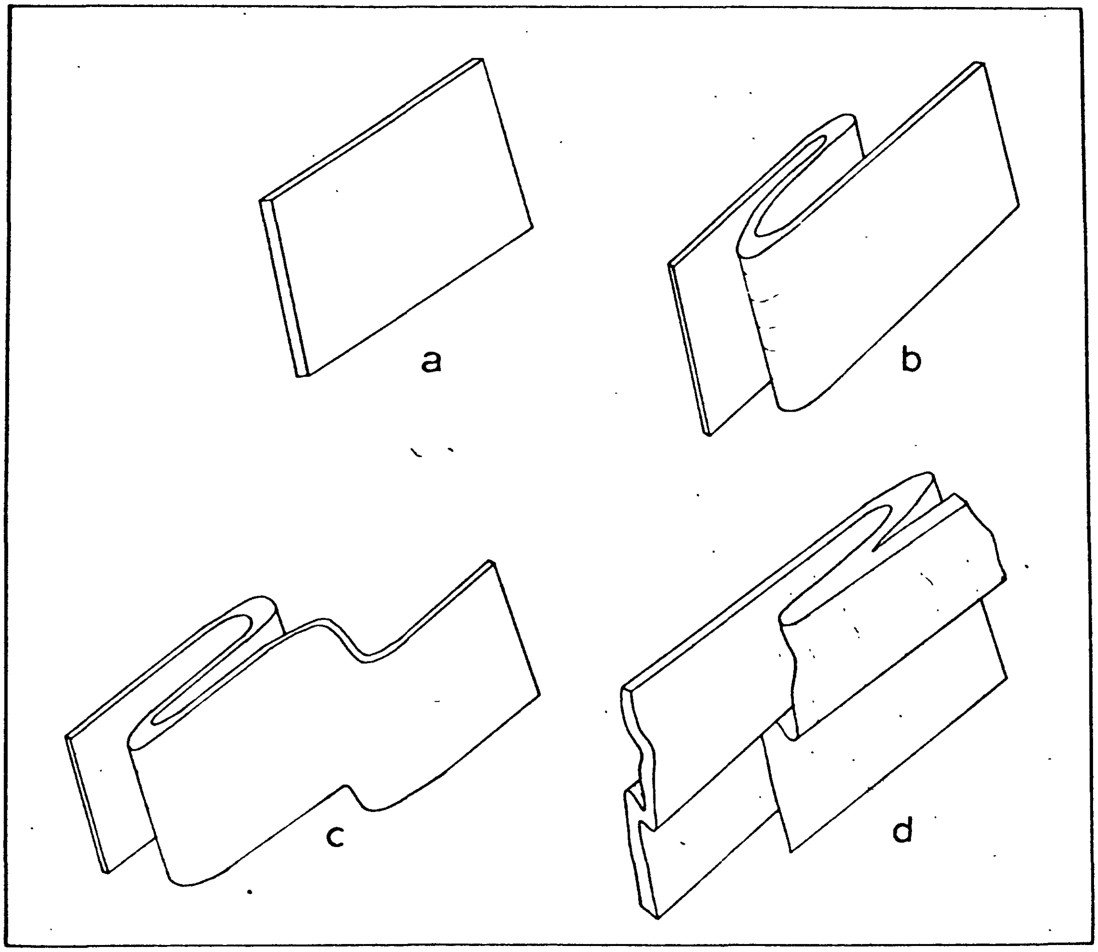


Figure 8. Development of minor folds in the Windsor quadrangle.(a)Initial attitude - steeply dipping beds. (b)Development of isoclinal folds with steep plunges. (c)Development of open folds with steep plunges. (d)Development of open folds with sub-horizontal plunges.

that plunges down the dip.

The first generation folds are syn-metamorphic. The later deformations develop no axial plane schistosity and no prograde metamorphic minerals are later than the first folds. The noses of these early folds are commonly occupied by quartz lenses and pencils.

Second Generation Folds:

Typically, second generation folds (late folds of plates 3 and 4) have axial planes that strike northeast through east to southeast with steep dips and plunges that are nearly down the dip of the axial plane schistosity of the first generation folds. These folds are characteristically open, of small amplitude, and fold the earlier schistosity and bedding (figure 8). No schistosity was developed by this later folding. An axial plane cleavage caused by rotation of micaceous minerals is commonly the only new structure developed. There is a little retrograde chlorite and recrystalized quartz along the cleavage planes. These folds are coaxial but not coplanar with the first generation folds.

This phase of the deformation represents tectonic stresses comparable in magnitude and with similar orientation to those that created the first generation folds; the difference in the orientation of the axial surfaces of these 2 generations of folds probably reflects non-uniform competency of the rock at the time of stress, the later folds being post-metamorphic when the rock was less plastic.

Neither the first nor the second generation folds affect the map pattern of the formational boundaries and they are a minor feature in the totality of the tectonic history.

Third Generation Folds:

Third generation folds (late folds of plates 3 and 4) are most obvious in the more schistose rocks where they appear as crinkles in the schistosity. On large outcrops, they commonly appear as open warps or folds with wave lengths as much as several tens of feet. The amplitude is usually less than 1 foot. Commonly they have a reverse drag sense. (figure 8). The axial planes strike north and are subvertical; the axes are subhorizontal with either a north or south plunge. Typically the first generation fold axes wrap around the noses of the third generation folds, and both bedding and schistosity are folded. No schistosity is developed parallel to the axial surface but a weak axial plane cleavage is rarely developed.

This generation of folding is responsible for the doubly plunging syncline of the garnet schist member of the Hoosac Formation of Judges Hill (plate 5) and minor kinks in the regional strike of units. There is also a doubly plunging syncline of albite schist of the Hoosac Formation within the Stamford Granite Gneiss west of Judges Hill. This is a somewhat questionable structure as there are albite schists that are probably Precambrian in age.

The genesis of this fold set may be caused by doming or arching along a north-south axis or relaxation of the area after the formation of the nappe. Near the Precambrian-Cambrian unconformity the deformation is indicated by the presence of reverse drag folds presumably indicative of the Paleozoic material cascading off the domed area of the arch (figure 8) or upward movement of the arch.

<u>Summary</u>:

The formation of the Hoosac nappe resulted in east over west shear and westerly transport, that is, up dip. Continued stress resulted in flowage of material up dip and formed folds with axes plunging down the dip (first generation folds). These flow folds would obliterate any previous structures such as folds with vertical axial planes and horizontal plunges.

Either strain rates changed or the rock was less plastic and continued stress produced more open folds (second generation folds) the axial planes and plunges of which were variable in attitude because of inhomogeneities in the rocks.

Later post-metamorphic stress resulted in open folds with north-striking axial planes and subhorizontal axes (third generation folds).

Minor Folds

Northwestern Part of the Quadrangle

At least two periods of folding are recognized in the northwestern part of the quadrangle.

The earliest folds are isoclinal recumbent folds with amplitudes as much as 10 feet. Axial surfaces have a variable attitude but generally dip less than 30^{o} east. Plunges and dips of axial surfaces are unsystematic because of later folding but commonly are gentle to the north. These early folds are syn-metamorphic and have an axial plane schistosity. They are well displayed in the Hinsdale Gneiss between the synclines north of Savoy Road that fold the Hoosac Formation. Here they have a consistent east over west shear sense regardless of their position with respect to the axial plane of the fold. Hence there must have been some transport after the nappe was recumbent.

These early isoclinal folds have been refolded by later more open folds that are tight flowage folds in the carbonate rocks, chevron folds in the Berkshire Schist, and open warps in the Hoosac Formation and Dalton Formation. Axial planes are highly variable in attitude but plunges are generally shallow to the north or northwest (plates 3 and 4). An axial plane cleavage is developed in the schistose rocks. The cleavage cuts both bedding and schistosity.

The random orientation of the axial planes of later folds suggests either that stress orientations were not homogeneous during the second folding or the second generation folds have in turn been refolded about a northerly axis.

The recumbent folds were contemporaneous with the first generation of the eastern sequence of folds. The later open folds were contemporaneous with either the second, or more probably the third generation of eastern folds.

The detailed picture and the significance of minor folds has been obscured because the earlier isoclinal folds had sub-horizontal axial planes; the plunges of later folds are unsystematic and the orientation of later axial surfaces would be highly variable due to inhomogeneities within a formation, between formations, and across the Middle Ordovician erosional unconformity.

Minor Folds

Southwestern Part of the Quadrangle

Minor folds in the southwestern part of the quadrangle are not common because of the massive rocks present there.

Just south of MacDonald Brook, recumbent isoclinal folds are present with nearly flat-lying axial planes and north- or west-plunging axes (plates 3 and 4). These folds have amplitudes as much as 5 feet.

A second generation of open folds has folded these early folds. They have north-striking axial planes with dips that range from horizontal to vertical; easterly dips are more common. Plunges are sub-horizontal to northerly and shallow.

These represent fan folds or originally coplanar, coaxial folds that have been later openly folded about a north-south axis. Measurement of folds at several outcrops with 2 generations of open folds suggest tha there are 2 sets of nearly coplanar-coaxial folds.

The latest folds are probably equivalent to the third generation folds found in the eastern sequence. Most of the axial surfaces of late folds dip easterly and have sub-horizontal plunges. The shear sense for these late folds is east over west. This would be consistent with movement away from the axis of the anticlinorium. The cylindrically folded beds (second generation folds) have been refolded yielding doubly plunging structures (e.g. the Weston Mountain Syncline, plate 5). The axial surfaces do not appear to be related to the orientation of the beds.

Linear Features

Lineations (plate 4) are common throughout the entire quadrangle. East of the Precambrian terrane, noses of folds, crinkles (third generation folds), the intersection of bedding and schistosity (first generation folds), and the intersection of bedding or schistosity and axial plane cleavage form strong lineations, the orientation of which is always parallel to the axes of the folds that caused them (b-lineations). Rarely, most commonly in greenschists, mineral streaming is present and is parallel to the noses of early (first generation) folds.

In the northwestern part of the quadrangle the same types of lineations are present. Fold noses are commonly observed only in the Hinsdale Gneiss, Hoosac Formation, and Berkshire Schist. Mineral streaming, especially muscovite and biotite, is common in the Dalton Formation.

In the southwestern part of the quadrangle, pebble stretching is common in the conglomerate member of the Dalton Formation. The direction of maximum elongation is east-west, parallel to the axes of early isoclinal folds. Mineral streaming is common in the muscovite-quartz schist of the Dalton Formation. Open fold axes are common in the flaggy granulites of the Dalton Formation (figure 11).

Rolled Porphyroblasts

Rolled porphyroblasts are a common feature in the Berkshire Schist, the Dalton Formation, and the albite schist and garnet schist of the Hoosac Formation on both the east and west side of the anticlinorium.

Albite commonly shows sigmoid structures in thin-section indicating "roll" as much as 180°; the inclusions are predominantly carbonaceous material and smaller amounts of quartz. Where the albite is in a more schistose matrix, "roll" is usually poorly developed or not at all. Geographic position with respect to the nappe does not seem to be reflected in the amount of roll. Albite from the inverted limb of the nappe and in the recumbent synclines indicates a similar amount of rotation. Some albite indicates a complex period of tectonic transport and growth that was not coincident at all times (figure 9).

Garnet also shows as much as 180° of "roll" with a complicated history like that of albite. The inclusions are either quartz or carbonaceous material.

No reversals of shear sense in the roll were found as reported by Rosenfeld (1954). However, the rare occurrence of a rolled garnet porphyroblast with one shear sense in a matrix of quartz and phyllosilicates with the opposite shear sense indicates the reversal of shear sense after the garnet had stopped growing. This later shear is marked by the development of incipient axial plane cleavage by the rotation of phyllosilicates rather than the growth of new minerals along new schistosity surfaces.

No carefully oriented specimens containing rolled porphyroblasts were obtained from the field and thus it is not possible to relate the precise direction and amount of roll observed in thin section to the regional tectonic picture.

Generation of S-Surfaces and Their Preservation

In the eastern and western sequences of rocks one phase of deformation caused the development of an axial plane schistosity resulting from the development of new minerals

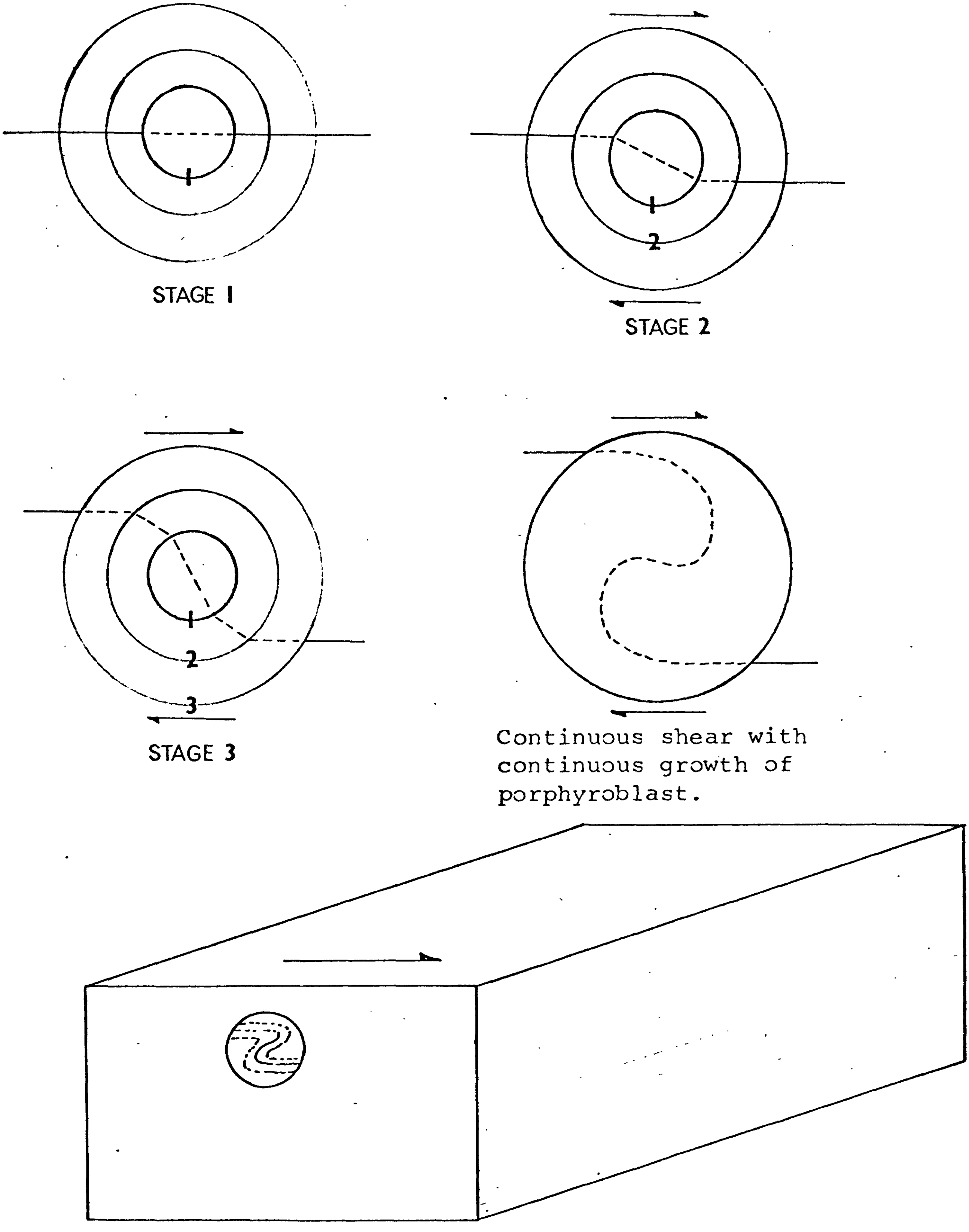


Figure 9. Development of sigmoid distribution of inclusions in porphyroblasts.

and two later generations of folds developed an axial plane cleavage. The latter two are post-metamorphic.

The early isoclinal folding has obliterated any earlier structure. All that remains is the intersection of bedding schistosity and the axial plane schistosity.

The second phase of folding rarely developed any axial plane structures. It folded the first axial surface but was not a very penetrative deformation. The last phase of deformation developed locally an axial plane cleavage with little or no recrystallization. Chlorite after biotite or garnet is commonly associated with this late S-surface and is the only textural evidence of retrograde metamorphism.

Joints and Cleavage

No systematic regional study of joint patterns was made, but the writer measured all planar surfaces in the Cheshire Quartzite between South Brook and the Precambrian-Cambrian unconformity. A plot of joints and cleavage measured in this area is given in figure 10. The joints can be correlated with axial-plane S-surfaces related to folding around a north-south axis and later refolding coaxial with the second generation folds.

Paleozoic Structure - Faults

Hoosic Thrust

A ferruginous cemented breccia composed of fragments of Cheshire rocks has been traced four miles south from the

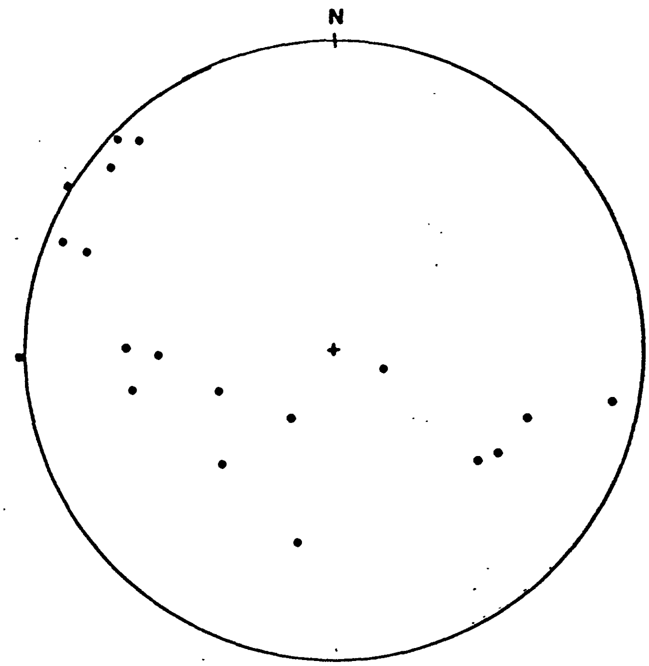


Figure 10. Poles of joints and cleavage measured in the Cheshire Quartzite and the Dalton Formation. Measurements taken east of South Brook and west of the Precambrian-Cambrian contact. Compare with figure 11.

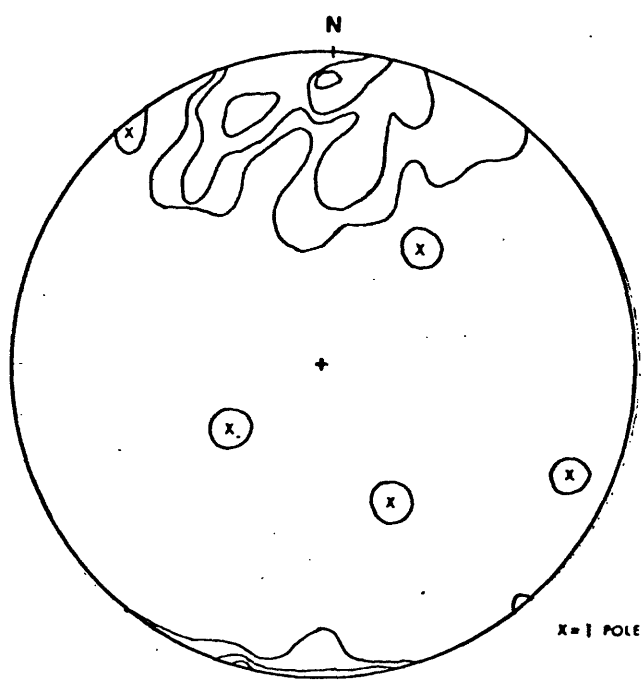


Figure 11. Contours for 46 lineations. Measurements taken in the same area as those for figure 10. Contours drawn at 2, 4, 6, and 8 percent.

boundary of the North Adams quadrangle. Outcrops of this breccia have been found at the following localities:

1. Anthony Creek at an elevation of 1,120 feet.
2. Reed Brook at an elevation of 1,200 feet.
3. For a distance of 2,000 feet in a direction north-northwest along approximately the 1,400 foot contour, just north of Savoy Road.
4. In the unnamed brook just south of Savoy Road that flows west into Dry Brook at an elevation of 1,410 feet. Here the position of the fault is indicated by a large bull-quartz vein.
5. Dry Brook at an elevation of 1,420 feet.

There is a repetition in the Dalton stratigraphy (plates 1 and 2) in the north part of the quadrangle. The stratigraphic throw is about 200 feet. To the south, east of Burlingame Hill, both the Berkshire Schist and the Dalton Formation are cut out.

This evidence suggests the presence of a thrust fault. The trace of the fault is roughly north-south from Anthony Creek to the MacDonald Brook area where the trend is westerly.

In the vicinity of Reed Brook the dip of beds is about 35^{o} east. Therefore the fault must have a steeper dip, perhaps 40^{o}. West and south of Burlingame Hill the dip of the thrust plane must flatten out to 10 to 20^{o} since topographic control on the trace is very strong.

This fault is a continuation of the Hoosic thrust (Herz, 1961) and may connect with the thrust west of North Mountain in the Cheshire quadrangle (Herz, 1958).

Minor Faults

In the recumbent syncline between Dry Brook and the Savoy Hollow Brook syncline there is a small thrust. Its existence there is demanded by the stratigraphic relationships (plate 1). Its dip is 20 to 30° northerly. The eastward and westward extensions are not known; it may connect with the Hoosic thrust.

Two minor faults are exposed where the west-flowing brook drains into South Brook. At an elevation of 1,480 feet in the west-flowing brook a fault plane is exposed. The fault surface strikes N. 35° W. and dips 25° northeast. A layer of bull-quartz 6 inches thick is present on the fault plane. The fault can be followed in the brook for about 40 feet. Stratigraphic relationships indicate that the upper block moved east. This is a low angle normal fault.

About 50 feet east of the junction with South Brook on the same unnamed brook there is another fault. It strikes N. 30° E. and is vertical. The vertical displacement cannot be estimated since the same thin-bedded quartzites are present on both sides of the fault. A fault breccia about 2 feet thick resembles the breccia associated with the Hoosic thrust described above. This same fault cuts across South Brook about 150 feet south of the brook juncture but again the offset could not be estimated.

In the large roadcut just east of the intersection of Main Road and Windsor Road west of Savoy, there are a series of north-striking normal faults that dip 80 to 90° east. Mineralization has occurred on the faults; pyrite is disseminated through the gneisses and the outcrop is deeply weathered because of the sulfides. The displacement on any one fault does not appear to be more than several feet. This fault zone has not been observed along strike to the north or south.

Alternative Interpretation for the Structure

The lack of stratigraphic symmetry in the Savoy Hollow syncline (plate 5) may be caused by a thrust fault that has sheared out the normal limb. The trace of the fault plane could be at the contact of the Cheshire Quartzite and the Berkshire Schist or within the Berkshire Schist. In these two cases the structural interpretation for the rest of the area is unchanged and the age relationship between the Dalton and Hoosac Formations is unchanged. If, however, the hypothetical fault is at the contact of the Dalton and Hoosac Formations, the age relationship between the two formations is ambiguous and the Dalton Formation may be ~~be~~ younger than the Hoosac Formation.

If it could be demonstrated that the axial surfaces of the Savoy Hollow Brook syncline and the Savoy syncline (plate 5) extend into the Dalton Formation then the age relationship between the two formations is unchanged. However, if the axial plane of the Savoy syncline is entirely within the Hoosac Formation, this would permit the possibility of a thrust fault bringing the Hoosac and Dalton Formations into fault contact instead of their being part of a conformable sequence of sediments. These relationships are shown in figure 12.

The existence or non-existence of these faults has great bearing on the Taconic problem.

First, one of these thrusts could serve as the root for the Taconic thrust that probably is present several miles to the west on Mount Greylock where the allochthonous rocks (Greylock Schist) overlie the autochthonous rocks (Berkshire Schist).

The structural interpretations in the southwestern part of the North Adams quadrangle are simplified by the utilization of a low angle thrust (figure 12).

If these two faults bring eugeosynclinal rocks into fault contact with miogeosynclinal rocks the problem of the source area for the Taconic rocks to the west remains unsolved. With the writer's interpretation, the Taconic rocks must all be post-Hoosac since the Hoosac was deposited on both the east and west sides of the present Precambrian terrane.

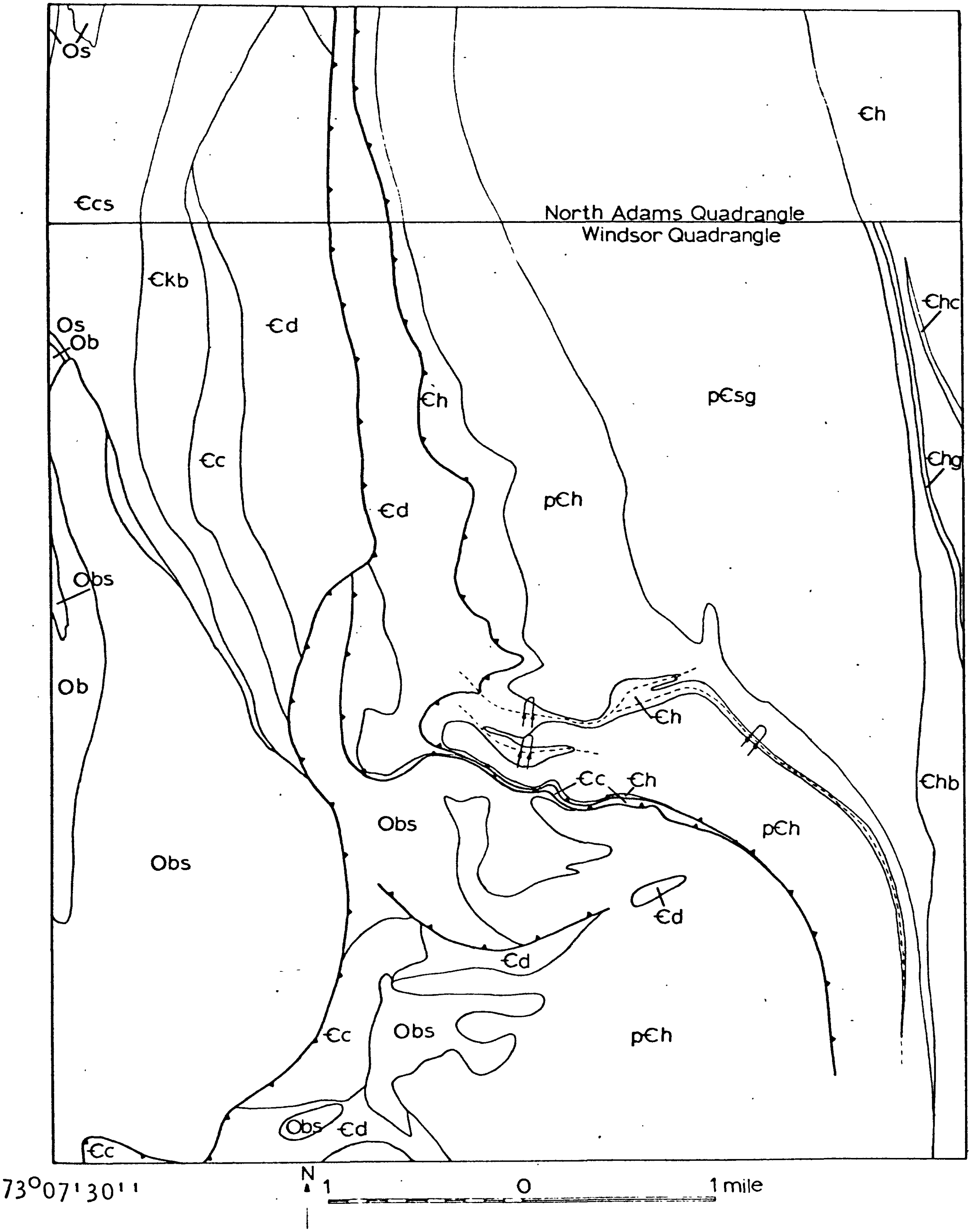


Figure 12. Geologic map of the northwestern part of the quadrangle based on alternative structural interpretations.

Unfolding the recumbent synclines still does not allow enough space for the derivation of the Taconic rocks found to the west unless they are all post-Hoosac. Several thrust slices, that would be part of the reinterpretation, would allow the covering up of the root zone for the Taconic rocks.

The writer does not favor the hypothesis of the existence of several thrust slices in the area for the following reasons:

1. The thickness of the Hoosac Formation is remarkably uniform west of the Hinsdale Gneiss.
2. There is no evidence of faulting at the contact observed north or Patton Brook.

The Savoy and Savoy Hollow Brook synclines have been interpreted as recumbent synclines that plunge beneath the Precambrian crystalline core of the Hoosac nappe (figure 7).

It is possible that these two synclines are cross-folds in the inverted limb of the nappe and do not project down the plunge of the major structure. The axes of these folds would be nearly coincident with the axial traces. The reinterpretation is given in figure 12.

The writer does not favor the hypothesis of cross-folding for the following reasons:

1. From evidence in the Hoosac Tunnel in the North Adams quadrangle (Pumpelly and others, 1894, p. 69-72) and from measurement of lineations at the north end of the exposures of the Stamford

Granite Gneiss (Herz, 1961), the core of the nappe plunges approximately 15° to the north.

2. Minor recumbent fold axes in the northwest part of Windsor quadrangle plunge gently to the north.
3. The thickness of the Hoosac Formation is remarkably uniform in the Savoy syncline. If this fold is a cross-fold, the uniform thickness suggests that the keel of the fold is all that remains. This is highly unlikely.
4. The Savoy Hollow Brook and Savoy synclines have nearly identical outcrop patterns which would seem unlikely for cross-folds.

Age of the Deformation

The Berkshire Schist is the youngest formation involved in the nappe structure. The formation of the nappe is then clearly post-Middle Ordovician.

Although there were significant tectonic events in this area and to the north from the Cambrian to the late Middle Ordovician (Cady, 1945; Shaw, 1958) and possibly in the late Silurian, these tectonic events did not produce significant folding and metamorphism in the Windsor area.

To the north in Vermont, the entire Paleozoic section from the Cambrian to the Devonian has undergone a sequence of deformations and a major episode of metamorphism together.

This has been dated as Middle to Late Devonian (Faul and others, 1963, p. 7-10).

East of the Windsor quadrangle the Silurian and Devonian rocks show the same evidence of deformation and metamorphism as in the eastern homoclinal limb of the Hoosac nappe. Therefore, the age of the Hoosac nappe is placed at the Middle to Late Devonian or Acadian. The first generation of folds in the Windsor area is syn-metamorphic which is therefore Middle to Upper Devonian. Further evidence for the synchroneity of the folding of the rocks and the main metamorphism is in the form of rolled porphyroblasts of garnet and albite that demonstrate a large amount of roll (shear).

The second generation folds appear related to the first and are also assigned to the Middle to Late Devonian. The third generation of folds can be dated only as post-Devonian and pre-Triassic. Faul and others (1963, p. 7-10) have radiometric dates in Vermont in the Carboniferous indicating a pulse of metamorphic activity during this time. This might be the time of the formation of the latest folds.

CHAPTER 4

MINERALOGY

General Statement

The metamorphic rocks east and west of the Precambrian terrane in the Windsor quadrangle are largely pelitic in nature. Thus the metamorphic assemblages may be represented graphically by using a system such as (1) Al_2O_3-K_2O-Na_2O-SiO_2-H_2O, or (2) Al_2O_3-FeO-MgO-K_2O-SiO_2-H_2O.

The following minerals pertinent to these systems are present in the rocks in the Windsor quadrangle:

System 1	Muscovite
	Paragonite
	Microcline
	Albite
	Quartz
System 2	Muscovite
	Microcline
	Quartz
	Chloritoid
	Garnet (almandite)
	Chlorite
	Biotite

Muscovite, paragonite, garnet, chlorite, and biotite were separated from the rocks and investigated both optically and by X-ray diffraction procedures. This enabled an estimate to be made of $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios for the ferromagnesian phases and Na^{+}/K^{+} ratios for coexisting muscovite and paragonite.

Garnet

Seventeen garnet smaples were separated from various metamorphic assemblages and from various formations and localities within the quadrangle (figure 13). It was hoped to establish the $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratio for various assemblages in Thompson's A-F-M projection (Thompson, 1957). These assemblages include:

a) garnet

b) garnet-chloritoid

c) garnet-chloritoid-chlorite

d) garnet-chlorite

e) garnet-chlorite-biotite

f) garnet-biotite

The composition of garnets may usually be expressed in terms of at most 5 components:

1) Almandite (Alm) $Fe_3Al_2(SiO_4)_3$

2) Pyrope (Pyr) $Mg_3Al_2(SiO_4)_3$

3) Spessartite (Spe) $Mn_3Al_2(SiO_4)_3$

4) Grossularite (Gro) $Ca_3Al_2(SiO_4)_3$

5) Andradite (And) $Ca_3Fe_2(SiO_4)_3$

If the physical properties of a garnet are linearly

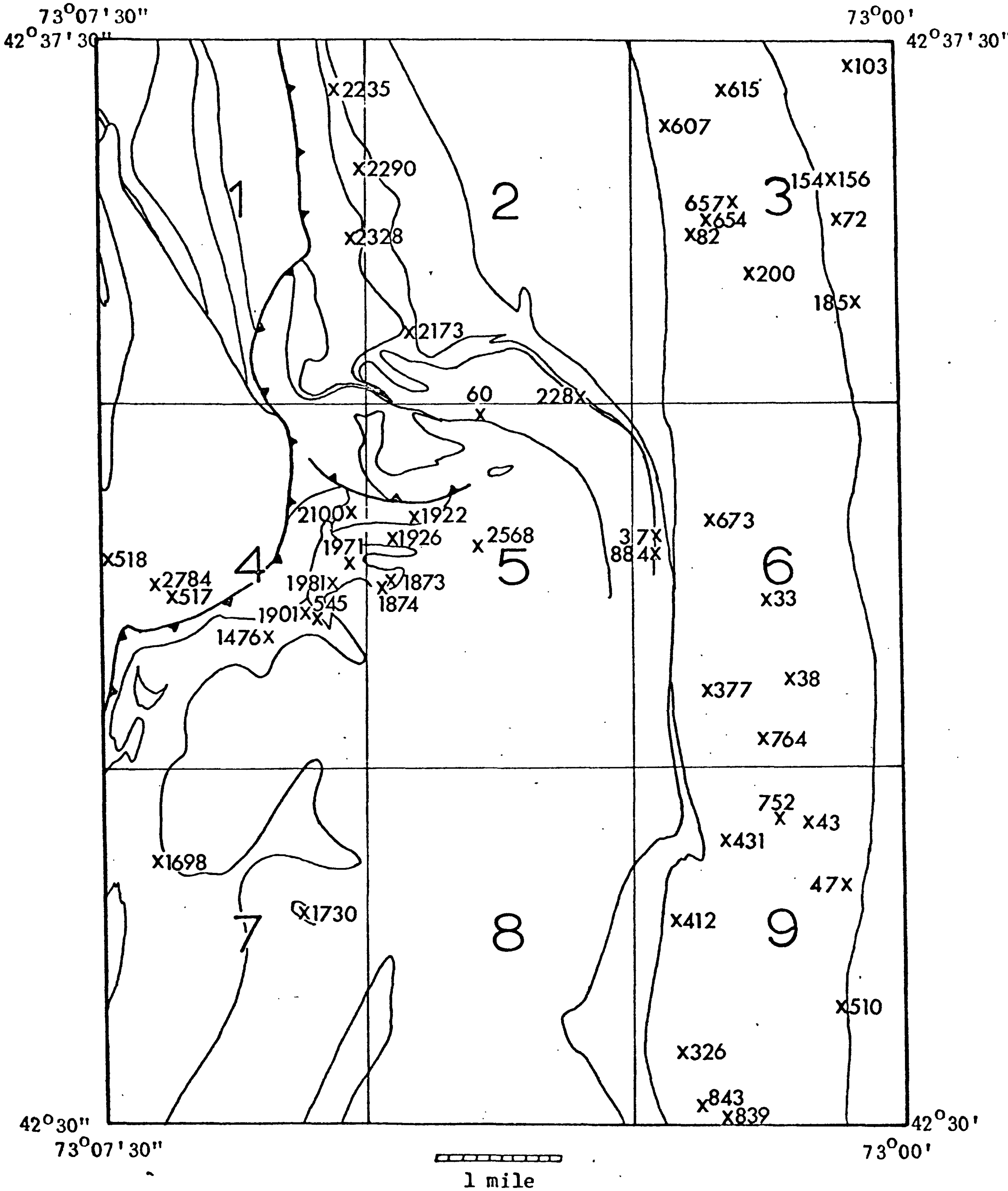


Figure 13. Location of specimens for garnet, biotite, chlorite, muscovite, and paragonite determinations.

related to the mole fractions of the 5 components, then measurement of 4 physical properties will uniquely determine the composition of the garnet.

Index of refraction (n) and cell edge measurements (a_o) are readily attainable but other physical measurements are difficult to perform, unreliable, and may not help in determining the composition.

With only 2 physical parameters known for a garnet, there are 2 degrees of compositional freedom remaining.

If one assumes that the mole fraction of andradite is insignificant, which is reasonable, there is only 1 degree of compositional freedom in the garnet if 2 physical properties are measured.

Figure 14 is modified from Winchell (1958, p. 595). If the mole fraction of andradite is insignificant, then the figure reduces to a compositional tetrahedron (Alm-Pyr-Spe-Gro) superimposed on a graph of n and a_o. Determination of n and a_o for a garnet yields a point on the graph which also gives the range of composition that the garnet might have and still have the same physical properties. In other words, measurement of 1 physical property (e.g., n) will yield a plane within the compositional tetrahedron. All garnet compositions on the plane will have the same value for the measured physical property. Measurement of a second physical property (e.g., a_o) will yield another plane. The intersection of the two planes yields a line (binary solid

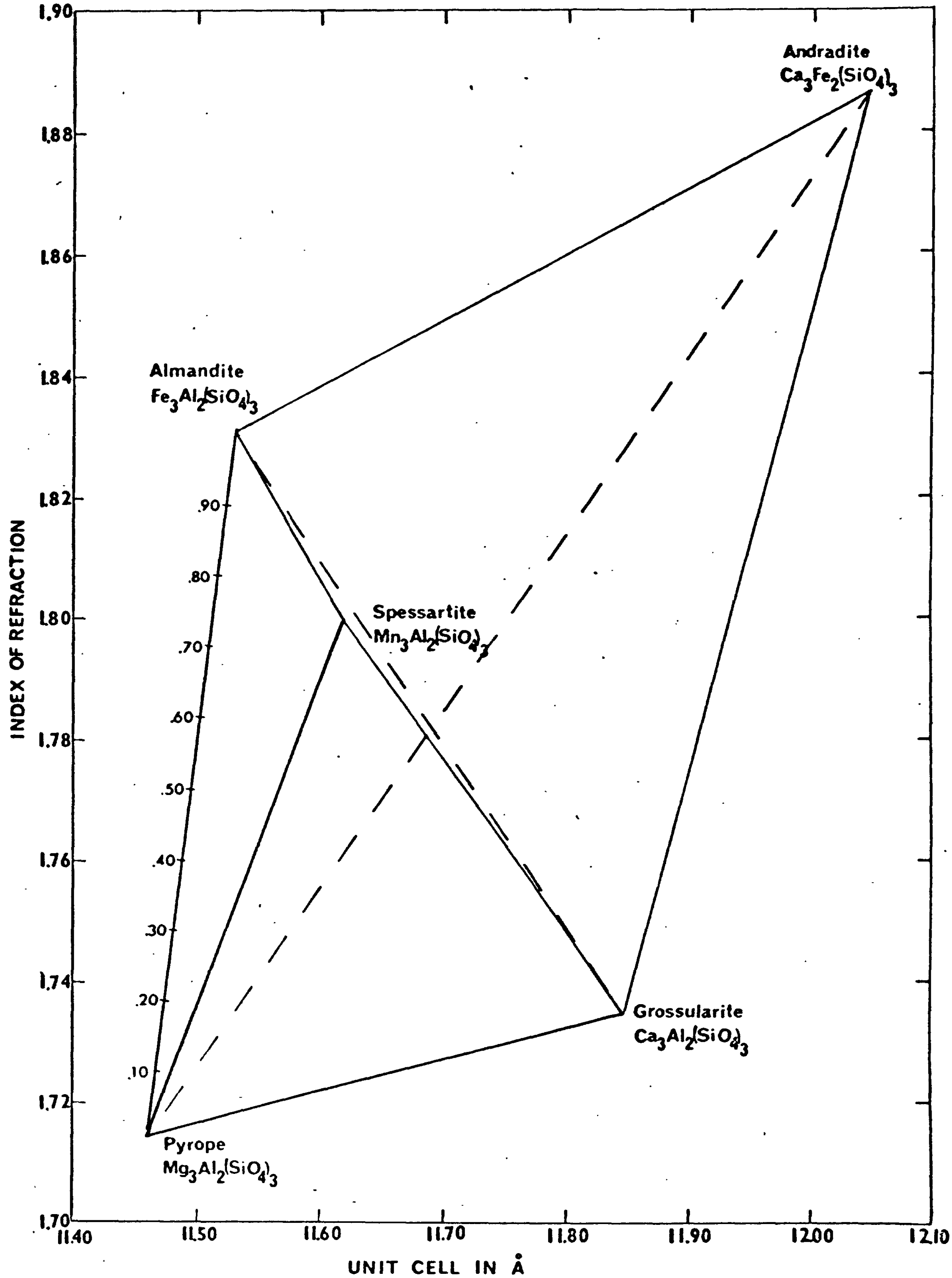


Figure 14. Relation between composition, index of refraction, and unit cell dimensions of garnet. After Winchell, 1958, p. 595.

solution). The end-members of the solid solution series are given by the intersection of this line (projected to a point in the figure) and the bounding faces of the tetrahedron. All garnets with compositions lying on this line will have the same values for the 2 measured physical properties. Further refinement of the composition of the garnet must be done on geologic bases.

Indices of the garnets were determined using sodium light and oils. The oils, calibrated to $\pm$ 0.003, were checked with a Leitz-Jelley Microrefractometer. The precision of the determinations is $\pm$ 0.002. Most of the garnets showed a small variation in $\underline{n}$, on the order of $\pm$ 0.002 to $\pm$ 0.003. This is probably caused by compositional variation within a single garnet.

Appreciable zoning might not be detected if the components that were zoned are (Alm) and (Gro) because they have compensating effects on the index of refraction.

Unit cell edges ($\underline{a_o}$) were determined by measuring d(420) on X-ray diffractometer charts. $Cu_{K\alpha}$ radiation was used with an internal Si standard. Scanning speed was $1/4°2\theta$/minute.

Chemical zoning is obvious in the garnets. The (420) peak on diffractometer charts is typically broad and/or multiple. This would be consistent with Ca and Mn zonation together.

All the data for the garnets are summarized on table 9. Both end members in the last column of table 9 are probably

TABLE 8 - Abbreviations

The following abbreviations are employed in reporting assemblages for the Mineralogy and Metamorphism chapters,

Qtz	-	quartz
Mus	-	muscovite*
Par	-	paragonite
Alb	-	albite
Mic	-	microcline
Ctd	-	chloritoid
Alm	-	garnet (almandite)
Chl	-	chlorite
Bio	-	biotite
Clz	-	clinozoisite, zoisite, epidote-undifferentiated
Mag	-	magnetite (includes ilmenite where ilmenite is not distinguished)
Ilm	-	ilmenite
Hem	-	hematite
Gra	-	graphite or carbonaceous material
Apa	-	apatite
Tou	-	tourmaline
Hbl	-	hornblende
Cal	-	calcite or dolomite
Zir	-	zircon
Sph	-	sphene

NOTE:

Minerals are reported in order of decreasing abundance. Bracketed minerals are present in amounts less than 1 percent.

1 = indicates the mineral is armored by garnet.
2 = indicates the mineral is secondary or retrograde.
* For coexisting muscovite and paragonite, the predominant phase, when both were detected, was judged by relative peak height for (006) on powder diffraction patterns.

67.170

TABLE 9

Garnet Data

Specimen	Other Phases Investigated	Assemblage	Locality	n	a_o(Å) Quality*	$Fe^{+2}/(Fe^{+2} + Mg^{+2})$	Composition
Hoosac Formation							
33-b	Bio, Par, Mus	Par-Mus-Alb-Qtz-Bio-(Cal-Alm-Chl^2-Mag-Clz)	(6)	1.802 ± .002	11.61_3 F	0.96 to 0.87	$Alm_{71}Pyr_3Gro_{26}$ to $Alm_7Pyr_1Spe_{92}$
37e	Chl, Par	Par-Alb-Qtz-Alm-Chl-(Mag-Gra-Zir-Bio)	(6)	1.802 ± .002	11.58_2 F	0.90 to 0.77	$Alm_{72}Pyr_8Gro_{20}$ to $Alm_{27}Pyr_8Spe_{65}$
377	Mus, Par	Qtz-Mus-Par-Alm-Chl-Ctd	(6)	1.803 ± .002	11.58_6 F	0.91 to 0.82	$Alm_{73}Pyr_7Gro_{20}$ to $Alm_{27}Pyr_6Spe_{67}$
412	Chl, Mus, Par	Qtz-Mus-Par-Alm-(Chl-Mag-Ctd)	(9)	1.801 ± .002	11.59_1 P	0.90 to 0.73	$Alm_{71}Pyr_8Gro_{21}$ to $Alm_{19}Pyr_7Spe_{74}$
607	Chl, Mus, Par	Qtz-Mus-Par-Alb-Alm-Chl-(Tou-Mag-Ctd')	(3)	1.802 ± .002	11.59_1 F	0.91 to 0.81	$Alm_{72}Pyr_7Gro_{21}$ to $Alm_{22}Pyr_5Spe_{73}$
673	Chl, Mus, Par	Qtz-Mus-Par-Alb-Alm-Chl-(Mag-Tou-Clz-Sph-Ctd')	(6)	1.800 ± .002	11.60_4 G	0.93 to 0.73	$Alm_{69}Pyr_5Gro_{26}$ to $Alm_{11}Pyr_4Spe_{85}$
839	Bio	Qtz-Mus-Alb-Bio-Alm-(Mag)	(9)	1.795 ± .002			
843	Mus, Par	Mus-Mar-Alm-Qtz-Alb-Mag-(Ctd')	(9)	1.806 ± .002	11.57_8 G	0.91 to 0.83	$Alm_{75}Pyr_7Gro_{18}$ to $Alm_{35}Pyr_7Spe_{58}$
884-b	Mus, Par	Mus-Par-Gar-Alb-Qtz-Ctd-(Mag-Gra-Chl)	(6)	1.804 ± .002	11.59_1 F	0.92 to 0.82	$Alm_{73}Pyr_6Gro_{21}$ to $Alm_{23}Pyr_5Spe_{72}$
884-c	Bio, Par	Alb-Par-Bio-Qtz-Alm-(Mag-Sph-Apa-Zir)	(6)	1.800 ± .002			
2235	Mus, Par	Mus-Par-Qtz-Alm-Chl-Alb-(Gra-Mag-Tou-Chl^2)	(1)	1.805 ± .002	11.59_1 F	0.94 to 0.83	$Alm_{74}Pyr_5Gro_{21}$ to $Alm_{24}Pyr_5Spe_{71}$
2290	Chl, Mus, Par	Mus-Par-Qtz-Alm-Chl-Mag-Alb-(Gra)	(1)	1.800 ± .002	11.58_6 F	0.87 to 0.74	$Alm_{70}Pyr_{10}Gro_{20}$ to $Alm_{23}Pyr_8Spe_{69}$
Berkshire Schist							
517-a	Mus, Par	Mus-Par-Ctd-Alm-Qtz-(Mag-Mag'-Tou-Hem^2)	(4)	1.805 ± .002	11.56_4 F	0.88 to 0.76	$Alm_{77}Pyr_{10}Gro_{13}$ to $Alm_{31}Pyr_5Spe_{64}$
518-a	Chl, Mus, Par	Qtz-Par-Mus-Ctd-Chl-Alm-Tou-Mag-(Hem^2)	(4)	1.802 ± .002	11.57_3 F	0.88 to 0.77	$Alm_{72}Pyr_{10}Gro_{18}$ to $Alm_{33}Pyr_{10}Spe_{57}$
518-b	Bio	Qtz-Mus-Alb-Bio-Alm-Chl^2-(Tou-Mag)	(4)	1.804 ± .002	11.61_3 P	0.99 to 1.00	$Alm_{73}Pyr_1Gro_{26}$ to $Alm_9Spe_{90}Gro_1$
1873	none	Mus-Bio-Alm-Chl^2-(Mag-Hem^2)	(5)	1.800 ± .002			
1971	none	Alb-Hbl-Qtz-Chl-Bio-Alm-Clz-(Chl^2-Mag)	(4)	1.791 ± .002	11.62_2 F	0.88 to 0.00	$Alm_{61}Pyr_8Gro_{31}$ to $Gre_6Pyr_6Spe_{88}$

*G = (420) peak sharp, $K_{\alpha 1}$ and $K_{\alpha 2}$ resolved

F = (420) peak not sharp, $K_{\alpha 1}$ and $K_{\alpha 2}$ not resolved

P = (420) peak broad or multiple

Location of Specimens[1]

Garnet Determinations

Hoosac Formation

33-b	Gray, albite schist: North side of Main Road, opposite △ 1,702 feet, (6).
37-e	Grayish-green, garnet schist: on Main Road, 500 feet east of the intersection of Main Road and Windsor Road, (6).
377	Light gray, rusty, garnet schist: opposite house at end of passable Griffin Hill Road, (6).
412	Rusty, grayish-blue to silvery schist: elevation 2,120 feet on Steep Bank Brook, (9).
607	Gray, garnet schist: elevation 2,030 feet on unnamed brook flowing north into Gulf Brook, (3).
673	Silvery-green schist: 1,500 feet north-northeast of Savoy, on knob, (6).
839	Grayish-green, albite schist: elevation 2,000 feet, 600 feet east of Bumpus Road, on trail, (9).
843	Bluish-gray, garnet schist: elevation 2,030 feet, on trail leading west from Bumpus Road, on south slope, (9).
884-b	Rusty, gray, garnet schist: 1,000 feet south-southeast of intersection of Main Road and Windsor Road, north of old road, (6).
884-c	Albite schist: same as 884-b.
2235	Carbonaceous, garnet schist: elevation 1,770 feet, west of hill 1,843 feet, (1).
2290	Grayish-green, garnet schist: 50 feet west of hill X 1,952 feet, (1).

Berkshire Schist

517-a	Carbonaceous, garnet phyllite: elevation 1,450 feet on Windsor Road, (4).

1 See figure 13.

518-a Silvery-green, crenulated, thinly laminated schist: 200 feet east of the point where Jenks Road leaves the quadrangle to the southwest, (4).

518-b Gray, albite-garnet schist: same as 518-a.

1873 Dark gray, biotite-garnet schist: elevation 2,020 feet, 2,400 feet N.76°E. from Johnson △, (4).

1971 Gray, equigranular granulite: elevation 1,970 feet, 1,600 feet N.27°E. from Johnson △, (4).

not realistic. For the purposes of plotting the garnet compositions on A-F-M projections (see Metamorphism), a composition midway between the end members has been arbitrarily selected.

Inspection of the ratios reveals:

1) For pelitic rocks, $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ranges from 0.80 to 0.99.
2) There is no sensible shift in garnet composition for the 3-phase assemblage garnet-chloritoid-chlorite in an east-west direction. However, the number of samples is small.
3) The garnets are spessartitic almandites with small but significant amounts of grossularite and pyrope. For the pelitic rocks in the quadrangle, grossularite has a possible range from 0 to 26 percent; pyrope has a possible range from 0 to 10 percent.

Biotite

Twenty-one biotites were separated and the γ index of refraction of biotite was determined using sodium light. Indices of the oils were checked on an Abbé Refractometer. Because of edge effects, determinations are difficult and probably only precise to $\pm$ 0.003.

Determination of $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios in biotites from indices of refraction is difficult because of the large numbers of possible elements in solid solution.

Important interlayer cations include Na^{+1}, K^{+1}, and Ca^{+2}. Octahedral cations include Mg^{+2}, Fe^{+2}, Mn^{+2}, Fe^{+3}, Ti^{+4}, and Al^{+3}. Si^{+4}, Al^{+3}, and possible Fe^{+3} are found in tetrahedral positions.

In general, refractive indices are increased by Fe, Mn, Ti, and possibly Al. All the relationships between chemical composition and refractive indices have not been worked out.

Wones (1963) gives curves for the γ index of refraction of biotite versus $\Sigma Fe/(\Sigma Fe + Mg^{+2})$ for several artificial buffer systems. The resulting curves are nearly coincident (Wones, 1963, figure 3, p. 1304). The buffers that most closely approximate the conditions of metamorphism in the Windsor quadrangle are $Fe_{(1-x)}O-Fe_3O_4$ and $Fe_2SiO_4-SiO_2-Fe_3O_4$. The rocks in the Windsor quadrangle apparently never reached equilibrium with a buffer as oxidizing as $Fe_3O_4-Fe_2O_3$. The biotites studied were buffered by either Fe_3O_4 or Fe^{+2}-Silicate-C.

Thus, Wones would predict a linear relationship between γ of biotite and $\Sigma Fe/(\Sigma Fe + Mg^{+2})$ from 1.580 to 1.700 when the ratio is 0.0 to 1.0 respectively. This relationship is shown in figure 15. Wones's experiments, however, were not representative of the actual conditions of metamorphism in the Windsor quadrangle because the amount of Al_2O_3 in the experiments was stoichiometric for $K(Fe, Mg)_3\ AlSi_3O_{10}(OH)_2$.

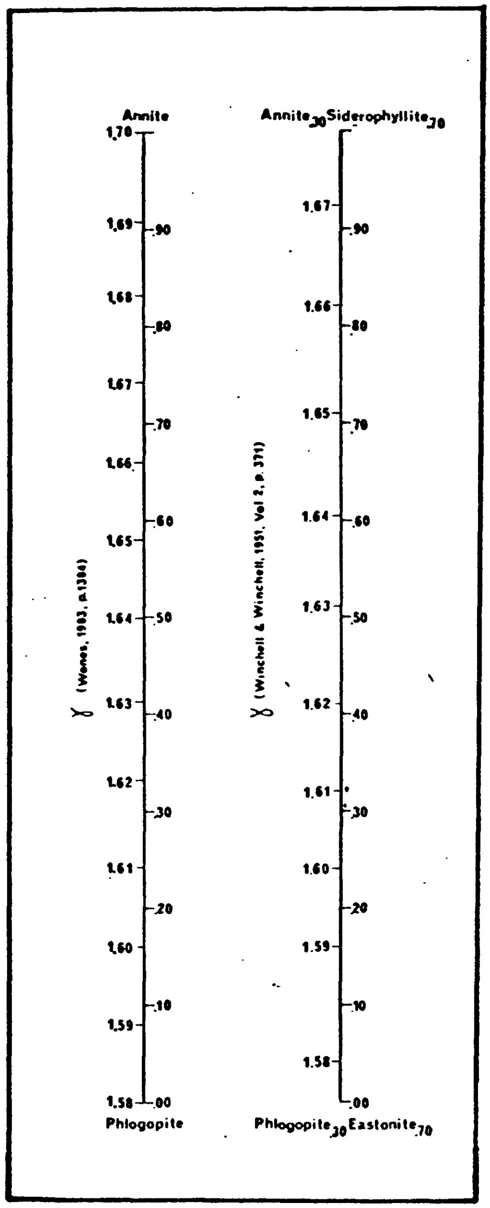


Figure 15. Relation between index of refraction and composition of biotite.

67-170

TABLE 10

Biotite Data

Specimen	Other Phases Investigated	Assemblage	Locality	γ biotite	$Fe^{+2}/(Fe^{+2} + Mg^{+2})$
Hoosac Formation					
33-a	none	Alb-Qtz-Bio-Mus-Alm-(Cal-Chl-Clz)	(6)	1.644 ± .003	0.65 ± .03
33-b	Alm, Par, Mus	Par-Mus-Alb-Qtz-Bio-(Cal-Alm-Chl^2-Mag-Clz-Hem^2)	(6)	1.639 ± .003	0.59 ± .03
43	Chl	Alb-Qtz-Mus-Chl-Cal-Clz-(Tou-Bio-Mag)	(9)	1.638 ± .003	0.59 ± .03
60-a	none	Qtz-Mus-Alb-Bio-Mic-(Mag-Chl^2-Apa)	(5)	1.654 ± .003	0.75 ± .03
82	none	Qtz-Alb-Mus-Bio-Clz-Alm-(Mag)	(3)	1.635 ± .003	0.56 ± .03
228-b	none	Qtz-Alb-Bio-Mus-Clz-(Mic-Chl)	(2)	1.636 ± .003	0.57 ± .03
654	none	Qtz-Alb-Mus-Bio-Clz-(Tou-Apa-Chl^2)	(3)	1.653 ± .003	0.74 ± .03
657	none	Alb-Qtz-Mus-Bio-Chl-Clz-(Apa-Sph-Zir)	(3)	1.654 ± .003	0.75 ± .03
766	none	Qtz-Alb-Mus-Bio-(Mag-Zir-Apa)	(6)	1.635 ± .003	0.56 ± .03
839	Alm	Qtz-Mus-Alb-Bio-Alm-(Mag)	(9)	1.636 ± .003	0.57 ± .03
884-c	Alm, Par	Alb-Par-Bio-Qtz-Alm-(Mag-Sph-Apa-Zir)	(6)	1.632 ± .003	0.53 ± .03
2173	none	Qtz-Bio-Alb-Mic-Mus-(Clz-Zir-Mag)	(2)	1.658 ± .003	0.80 ± .03
Dalton Formation					
1730	none	Qtz-Mus-Bio-(Mag-Zir-Apa)	(7)	1.662 ± .003	0.84 ± .03
1901-b	none	Mus-Atz-Bio-(Mic-Tou)	(4)	1.665 ± .003	0.87 ± .03
1922	none	Qtz-Mic-Alb-Mus-Bio-(Apa-Mag)	(5)	1.653 ± .003	0.74 ± .03
2100	none	Qtz-Mus-Alb-Bio-(Mic-Gra)	(4)	1.655 ± .003	0.77 ± .03
2328	none	Qtz-Alb-Mic-Mus-Bio-(Mag-Sph-Apa-Chl^2)	(1)	1.650 ± .003	0.71 ± .03
Berkshire Schist					
518-b	Alm	Qtz-Mus-Alm-Alb-Bio-Chl^2-(Tou-Mag)	(4)	1.643 ± .003	0.64 ± .03
1476	none	Mus-Qtz-Mic-Bio	(4)	1.653 ± .003	0.74 ± .03
1874	Chl^2	Qtz-Mus-Chl^2-Bio-Alm-(Mag)	(5)	1.639 ± .003	0.59 ± .03
1981	none	Qtz-Alb-Mus-Bio-Mag-(Mic-Zir)	(4)	1.652 ± .003	0.73 ± .03

That is, there were 2 Al for each 24(O, OH). Thus, although Wones's curves account for tetrahedral and octahedral Fe^{+3}, they do not account for Al^{+3} in tetrahedral and octahedral positions in excess of 2 Al/24(O, OH). This is important because all the biotites studied by this writer were associated with either muscovite or paragonite. Furthermore, the oxyannite component is probably insignificant. Thus these curves of Wones (1963) cannot be used for determining $\Sigma Fe/(\Sigma Fe + Mg^{+2})$ ratios.

Winchell and Winchell (1951, p. 371) give a graphical representation for biotites with variable Fe^{+2}/Mg^{+2} ratios and variable Si^{+4}/Al^{+3} ratios versus γ biotite. In this investigation, a value of 3.4Al/24(O, OH) was assumed. This is equivalent to a biotite series from annite$_{.30}$ siderophyllite$_{.70}$ to phlogopite$_{.30}$ eastonite$_{.70}$ (figure 15). Although the Al content of biotite will vary according to the assemblage with which it is in equilibrium, the variation will be small since muscovite is present in excess. For example, biotite in equilibrium with microcline would be less aluminous than biotite in equilibrium with garnet and/or chlorite. For biotites with Al less than 3.4/24(O, OH) the assigned $\Sigma Fe/(\Sigma Fe + Mg^{+2})$ ratio will be too high.

The biotite data are tabulated in table 10. The $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios range from 0.52 to 0.87 (figure 18). The ratios determined for biotite coexisting with either garnet or chlorite are given in figure 17. This figure shows

Location of Specimens[1]

Biotite Determinations

Hoosac Formation

33-a	Gray, albite schist: north side of Main Road, opposite Δ 1,702 feet, (6).
33-b	Gray, albite schist: same as 33-a.
43	Grayish-green, albite schist: west side of River Road, 1,000 feet north of B.M. 1,498 feet, (9).
60-a	Massive, tan, laminated schist: at height of land between Savoy and Adams on Main Road, (5).
82	Silvery-green, albite schist: on trail, 900 feet west of Borden Mountain, (3).
228-b	Biotite-speckled, laminated granulite: elevation 2,045 feet, on Brown Road extension, 3,300 feet north-northwest of Tomb Cemetery, (2).
654	Gray, granular, albite schist: 1,400 feet N.40°E. from Borden Mountain, east side of spur, (3).
657	Grayish-green, albite schist: 250 feet S.80°W. from intersection of Adams Road and Bannis Road, (3).
766	Gray, albite schist: elevation 1,900 feet, east side of knob 2,600 feet east of X 2,012 feet, (6).
839	Grayish-green, albite schist: elevation 2,000 feet, 600 feet east of Bumpus Road, on trail, (9).
884-c	Albite schist: 1,000 feet south-southeast of intersection of Main Road and Windsor Road, north of old road, (6).
2173	Gray, massive, schist: elevation 1,840 feet on Patton Brook, (2).

[1] See figure 13.

Dalton Formation

1730 Gray, thinly laminated, schistose granulite: on hill X 2,029 feet, (7).

1901-b Dark gray, spangled schist: elevation 1,535 feet in Dry Brook, (4).

1922 Thinly laminated quartzite: 400 feet southwest of hill X 2,077 feet, (5).

2100 Biotite-speckled, gray granulite: elevation 1,940 feet, in gut just south of Card △ 2,058 feet, (4).

2328 Thinly laminated quartzite: elevation 1,420 feet on Tophet Brook, (1).

Berkshire Schist

518-b Gray, albite-garnet schist: 200 feet east of point where Jenks Road leaves the quadrangle to the southwest, (4).

1476 Biotite-speckled, rusty, gray granulite: 700 feet west-southwest of hill X 1,917 feet, (4).

1874 Biotite-speckled, gray, laminated granulite: elevation 2,000 feet, 2,200 feet N.77°E. of Johnson △ , (5).

1981 Gray, granular schist: 900 feet N.10°E. of Johnson △ , (4).

that $Fe^{+2}/(Fe^{+2} + Mg^{+2})_{garnet} > Fe^{+2}/(Fe^{+2} + Mg^{+2})_{biotite} >$ $Fe^{+2}/(Fe^{+2} + Mg^{+2})_{chlorite}$.

Chlorite

Fourteen samples of chlorite were separated and the β index of refraction of chlorite was determined in sodium light. The refractive indices of the oils were checked on an Abbé Refractometer. Measurements are probably no better than $\pm$ 0.003 because of edge effects from the chlorite grains. This results in a precision error of $\pm$ 0.03 in the $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratio.

Using Eckstrand's curves (1963) (figure 16) $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios were determined, (see Albee, 1962 for similar results.) The data are tabulated on table 11.

The $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios range from 0.35 to 0.61. Those chlorites coexisting with garnet and/or chloritoid range from 0.47 to 0.61; most are from 0.50 to 0.56; those coexisting with biotite and no garnet have low ratios.

Muscovite - Paragonite

Thirty specimens were selected from the pelitic parts of the Berkshire Schist, Dalton Formation, Hoosac Formation and Rowe Schist to investigate the possible coexistence of muscovite and paragonite. It was also hoped to use the muscovite-paragonite solid solution data as a geologic thermometer. Only specimens with a preponderance of white micas and no microcline were selected. All but 3 showed

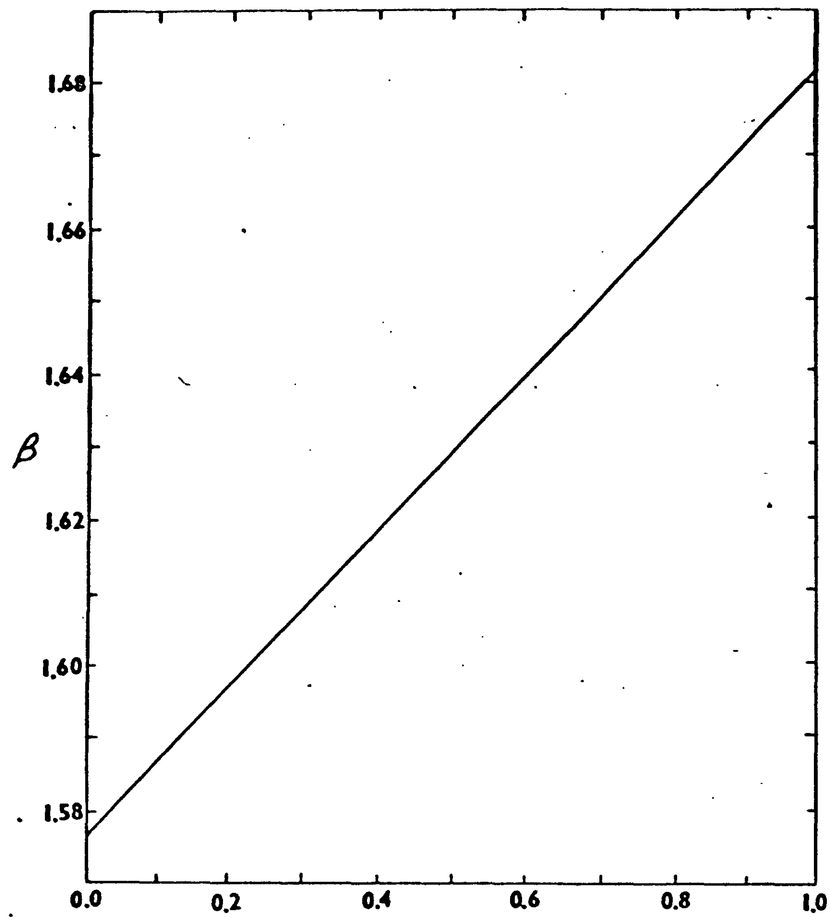


Figure 16. Relation between index of refraction and $Fe^{+2}/(Fe^{+2}+Mg^{+2})$ ratios for chlorite. After Eckstrand, 1963, figure 4, p. 69.

67-170

TABLE 11

Chlorite Data

Specimen	Other Phases Investigated	Assemblage	Locality	β chlorite	$Fe^{+2}/(Fe^{+2} + Mg^{+2})$
Hoosac Formation					
37e	Par, Par², Alm	Par-Alb-Qtz-Chl²-Chl-Alm-(Mag-Gra-Zir-Bio)	(6)	1.635 ± .003	0.55 ± .03
39	Mus, Par	Qtz-Mus-Par-Chl-Alb-(Mag-Alm-Clz)	(6)	1.633 ± .003	0.54 ± .03
43	Bio	Alb-Mus-Chl-Qtz-Cal-Clz-(Tou-Bio-Mag)	(9)	1.623 ± .003	0.43 ± .03
47	none	Qtz-Mus-Chl-(Alm-Tou-Mag)	(9)	1.635 ± .003	0.55 ± .03
412	Mus, Par, Alm	Mus-Par-Qtz-Alm-(Chl²-Mag-Ctd-Ctd')	(9)	1.634 ± .003	0.54 ± .03
510	none	Alb-Qtz-Mus-Chl-Alm-Cal-(Bio-Gra-Mag-Tou)	(9)	1.632 ± .003	0.53 ± .03
607	Mus, Par, Par², Alm	Qtz-Mus-Par-Alb-Alm-Chl-(Ctd'-Tou-Mag)	(3)	1.640 ± .003	0.61 ± .03
615	none	Qtz-Alb-Mus-Chl-Clz-(Tou-Cal-Zir-Bio-Mag)	(3)	1.618 ± .003	0.39 ± .03
673	Mus, Par, Alm	Qtz-Mus-Par-Alb-Alm-Chl-(Mag-Tou-Clz-Sph-Ctd')	(6)	1.629 ± .003	0.50 ± .03
752	Mus	Mus-Qtz-Ctd-Chl-(Mag-Hem²)	(9)	1.636 ± .003	0.56 ± .03
2290	Mus, Par, Alm	Mus-Par-Qtz-Alm-(Chl-Mag-Alb-Gra)	(1)	1.630 ± .003	0.52 ± .03
Berkshire Schist					
518-a	Mus, Par, Alm	Qtz-Mus-Par-Chl-Ctd-Alm-(Tou-Mag)	(4)	1.626 ± .003	0.47 ± .03
1874	Bio	Qtz-Mus-Chl²-Bio-Alm-(Mag)	(5)	1.635 ± .003	0.55 ± .03
1926-6	none	Qtz-Alb-Chl-Mus-Bio-(Gra-Zir-Clz)	(5)	1.614 ± .003	0.35 ± .03

1 = armored with garnet

2 = secondary

Location of Specimens[1]

Chlorite Determinations

Hoosac Formation

37e	Grayish-green, garnet schist: on Main Road, 500 feet east of the intersection of Main Road and Windsor Road, (6).
39	Light green, albite schist: west side of River Road, 500 feet north of Savoy town line, (6).
43	Grayish-green albite schist: west side of River Road, 1,000 feet north of B.M. 1,498 feet, (9).
47	Light green, fine-grained, albite schist: east side of River Road, 600 feet north of Windsor Jambs State Park boundary, (9).
412	Rusty-grayish-blue to silvery schist: elevation 2,120 feet on Steep Bank Brook, (9).
510	Grayish-green, granular, albite schist: elevation 1,770 feet on south fork of brook flowing east into the Westfield River, (9).
607	Gray, garnet schist: elevation 2,030 feet on unnamed brook flowing north into Gulf Brook, (3).
615	Greenish-gray, albite schist: 700 feet northeast of Lewis Hill, (3).
673	Silvery-green schist: 1,500 feet north-northeast of Savoy, on knob, (6).
752	Rust-spotten, granular schist: elevation 1,870 feet on ridge west-northwest of B.M. 1,498 feet, (9).
2290	Grayish-green, garnet schist: 50 feet west of hill X 1,952 feet, (1).

Berkshire Schist

518-a	Silvery-green, crenulated, thinly laminated schist: 200 feet east of point where Jenks Road leaves the quadrangle to the southwest, (4).

1 See figure 13.

1874 Biotite-speckled, gray, laminated granulite: elevation 2,000 feet, 2,200 feet N.77°E. of Johnson △ , (5).

1926-b Biotite-speckled, rusty schist: 1,400 feet N.50°E. of hill 2,068 feet, (5).

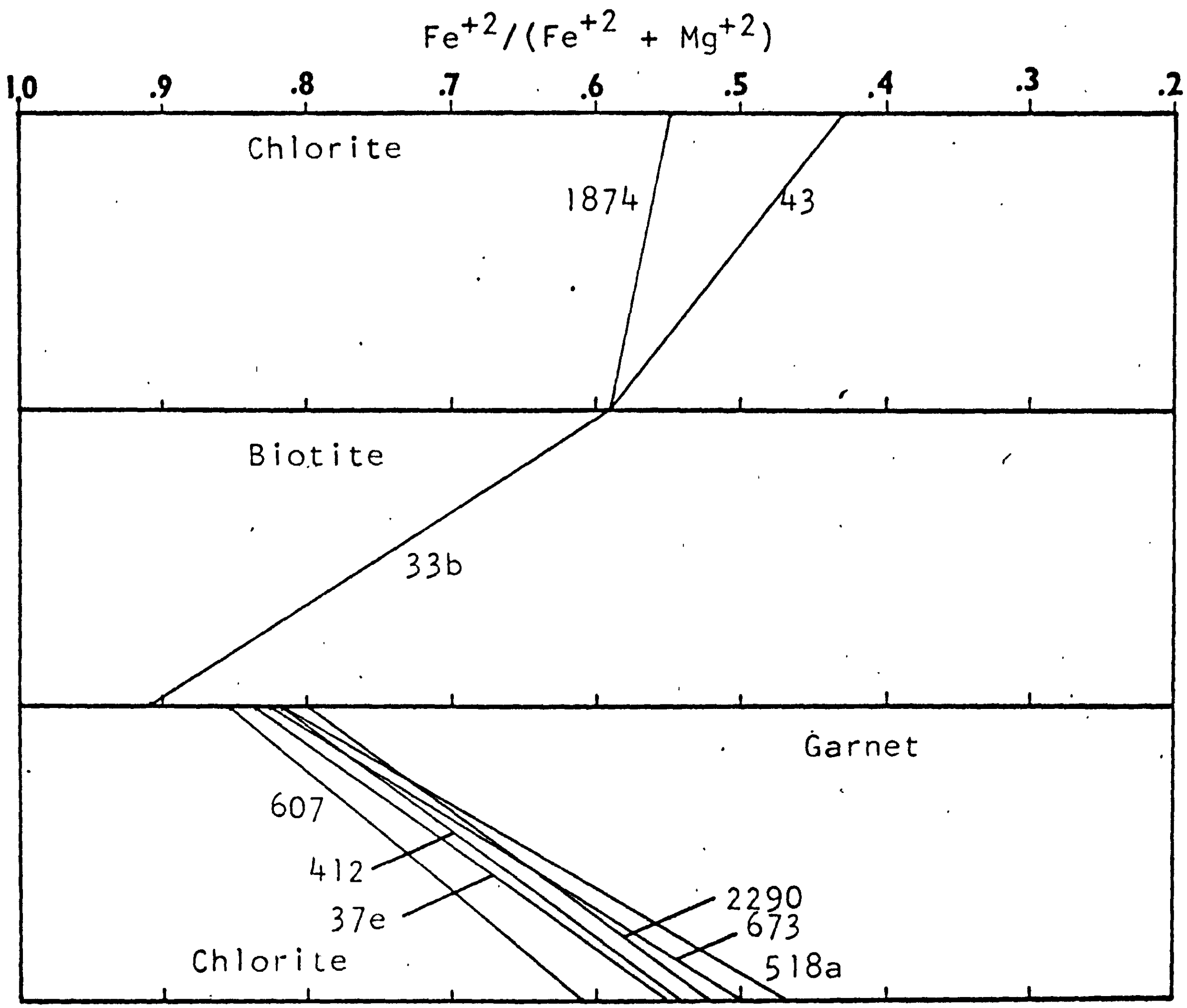


Figure 17. Approximate $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios for coexisting phases.

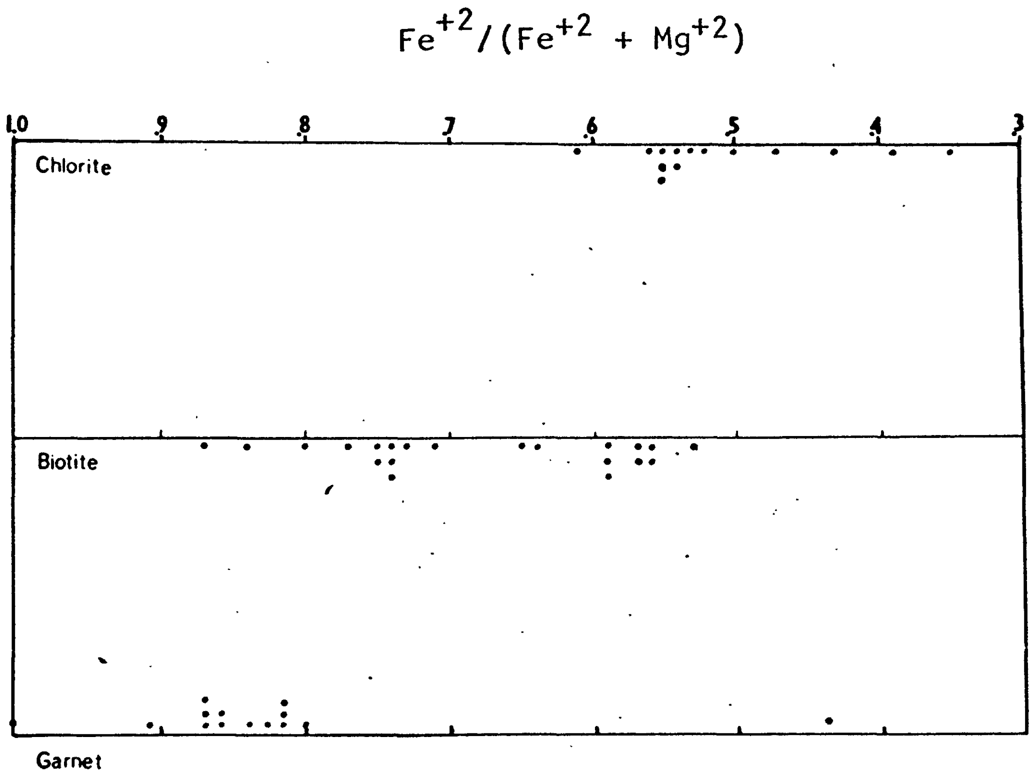


Figure 18; $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios for all measured garnets, chlorite, and biotite. The ratios for the garnets are the mean value for the range given in Table 9.

coexisting muscovite and paragonite.

D-spacings were determined by measuring d(006) peaks on Norelco X-ray diffractometer charts. Cu radiation was used with $K_{\alpha 1}$ = 1.54050Å.

Both silicon and quartz were used as internal standards. For silicon the following data were used:

For (111) $K_{\alpha 1} = 28.422^\circ\ 2\theta$

$K_{\alpha 2} = 28.514^\circ\ 2\theta$

$K_{\alpha} = 28.466^\circ\ 2\theta$

d(111) = 3.136Å

For quartz the following data were used:

For $(10\bar{1}1)$ $K_{\alpha 1} = 26.640^\circ\ 2\theta$

$K_{\alpha 2} = 26.707^\circ\ 2\theta$

$K_{\alpha} = 26.662^\circ\ 2\theta$

$d(10\bar{1}1)$ = 3.344Å

Muscovite and paragonite both have their (006) peaks between these two standards.

Scanning speeds were either $1/4^\circ$ or $1/8^\circ$ 2θ/minute.

The d-values for (006) are measurable to 0.001Å at either scanning speed. Thus d(002) is precise to ± 0.003Å. The error in measurement of the d(standard) and $d(006)_{\text{muscovite or paragonite}}$ gives an error of ± 0.006Å.

Repeated determinations on a single sample established the following:

1. Reproducibility was achieved using either internal standard.

2. Reproducibility was achieved using either scanning speed.
3. Reproducivility was achieved using either pressed powder mounts or permanent slides with a binder.

However, it is apparent that values for $d(002)_{paragonite}$ commonly have a range on the order of $\pm$ 0.150Å in a single hand specimen. Specimen 156 had three distinct paragonite peaks, the difference between the extremes being 0.55Å.

Intensities of peaks on X-ray diffractometer charts indicate that the proportions of paragonite and muscovite are related to the abundance of albite. Where albite greatly exceeded the white micas in abundance, the only white mica was paragonite.

The data are tabulated and summarized in table 12. Results from individual determinations are tabulated in tables 13 and 14. These summarized results are graphically represented in figure 19.

The d(002) values of paragonite range from 9.635 to 9.570Å with a maximum from 9.58 to 9.59Å. Muscovite values range from 9.935 to 9.977Å with a maximum from 9.950 to 9.965Å.

Muscovite values are less variable in the same specimen or multiple samples (103a & b) from the same outcrop.

There is no relationship between geographic or strati-

(Page 162 follows
TABLES 13/14
IN BACK

TABLE 12

d-spacing (002) for Muscovite-Paragonite Pairs*

Hoosac Formation	Muscovite	Paragonite
33-b	9.966	9.582
39	9.957 ± .002	9.583 ± .006
200	9.960 ± .006	9.585 ± .000
326-2	9.936	9.627
377	9.948 ± .003	9.624 ± .000
412	9.945 ± .002	9.618 ± .002
431	9.963 ± .009	9.590 ± .007
607	9.957 ± .003	9.583 ± .004
673	9.955 ± .005	9.595 ± .005
843	9.955 ± .002	9.582 ± .002
884-b	9.961 ± .005	9.589 ± .014
2235	9.957	9.573
2290	9.950 ± .005	9.582 ± .003
Rowe Schist		
72	9.951 ± .006	9.585 ± .005
103-a	9.973 ± .006	9.607 ± .006
103-b	9.977 ± .009	9.598 ± .006
138	9.966	9.633
154	9.961 ± .003	9.587 ± .010
156	9.951 ± .000	9.570 ± .000
		9.600
		9.624 ± .000
158	9.956 ± .004	9.573
185	9.951 ± .005	9.612 ± .005
Dalton Formation		
545-a	9.948	9.579
1698	9.960	9.623 ± .003
Berkshire Schist		
517-a	9.943 ± .004	9.625 ± .004
518-a	9.964 ± .007	9.628 ± .004
2784	9.964 ± .004	9.573 ± .005
Hinsdale Gneiss		
2568	9.966 ± .004	9.583 ± .002
Hoosac Formation		
37-e	not present	9.603 ± .005
752	9.945	not present
884-c	not present	9.570 ± .004

* Standard deviations are calculated from the standard equation $s = \sqrt{\frac{\Sigma d^2}{n}}$ where d is the deviation from the mean and n is the number of determinations.

Values with no standard deviation are based on only one determination.

Location of Specimens[1]

Muscovite-Paragonite Pairs

Hoosac Formation

33-b	Gray, albite schist: intersection of Griffin Hill Road and Main Road, (6).
39	Green, albite schist: west side of River Road, 3,800 feet south on River Road from B.M. 1,702 feet, (6).
200	Silvery, rust spotted, laminated schist: elevation 2,250 feet on Chickley River, (3).
326-2	Crenulated, thinly laminated, gray schist: on High Street, south-southeast of Hill 2,187 feet, (9).
377	Light gray, rusty, garnet schist: at the end of passable Griffin Hill Road opposite the house, (6).
412	Rusty, gray-blue to silvery schist: elevation 2,120 feet on Steep Bank Brook, (9).
431	Gray, albite schist: 600 feet southeast of a point on Savoy Hollow Road 3,400 feet from the intersection of Bates Road and Savoy Hollow Road, (9).
607	Gray, garnet schist: elevation 2,030 feet on north-flowing brook that crosses New State Road and joins Gulf Brook, (3).
673	Silvery-green schist: on knob 1,500 feet north-northeast from Savoy, (6).
843	Blue-gray, garnet schist: 1,200 feet N.11°W. from point where old Route 9 leaves the quadrangle to the south, (9).
884b	Rusty, gray, garnet schist: 900 feet S.30°E. from the intersection of Main Road and Windsor Road, (6).
2235	Carbonaceous garnet schist: elevation 1,770 feet, due west of hill X 1,843 feet, (1).
2290	Grayish-green, garnet schist: elevation 1,940 feet, due west of hill X 1,952 feet, (1).

1 See Figure 13.

Rowe Schist

72 Gray, fine-grained schist: on Adams Road where the east-flowing brook crosses east-northeast of Borden Mountain (about 5,000 feet), (3).

103a Blue-gray phyllite: south side of Tannery Road, 500 feet northwest of the intersection of Ross Brook and Tannery Road, (3).

103b Thinly laminated, silvery-gray phyllite: same as 103a.

154 Silvery gray-blue, fine-grained schist: elevation 1,800 feet on the brook flowing east into Parker Brook, (3).

156 Grayish-blue, fine-grained schist: elevation 1,760 feet on the brook flowing east into Parker Brook, (3).

158 Grayish-blue, spangled schist: elevation 1,740 feet on brook that flows east into Parker Brook, (3).

185 Silvery-gray phyllite: elevation 1,830 feet on Horsefords Brook, (3).

Dalton Formation

545a Gray, biotite granular schist: 1,300 feet from the intersection of Windsor Road and Sand Mill Road at an elevation of 1,600 feet on Windsor Road, (4).

1698 Spangly, muscovite schist: elevation 2,030 feet, 3,300 feet S.83°W. from Hill X 2,147 feet, (7).

Berkshire Schist

517a Carbonaceous, garnet phyllite: elevation 1,450 feet on Windsor Road, (4).

518a Gray, albite schist: 100 feet southeast of Jenks Road where it leaves the quadrangle to the south, (4).

2784 Gray, garnet granulite: elevation 1,420 feet, 3,350 feet west of B.M. 1,487 feet, (4).

Hinsdale Gneiss

2568 Gray, gneissic granulite: elevation 2,070 feet, 750 feet S.50°E. of Hill 2,156 feet, (5).

Location of Specimens[1]

Muscovite or Paragonite-bearing Rocks

Hoosac Formation

37e	Grayish-green, garnet schist: roadcut 300 feet east of the intersection of Main Road and Windsor Road, west of Savoy, (6).
752	Rust-spotted schist: 2,000 feet N.70°W. from B.M. 1,498 feet, at an elevation of 1,870 feet, (9).
884c	Albite schist: 900 feet S.30°E. from the intersection of Main Road and Windsor Road, (6).

1 See figure 13.

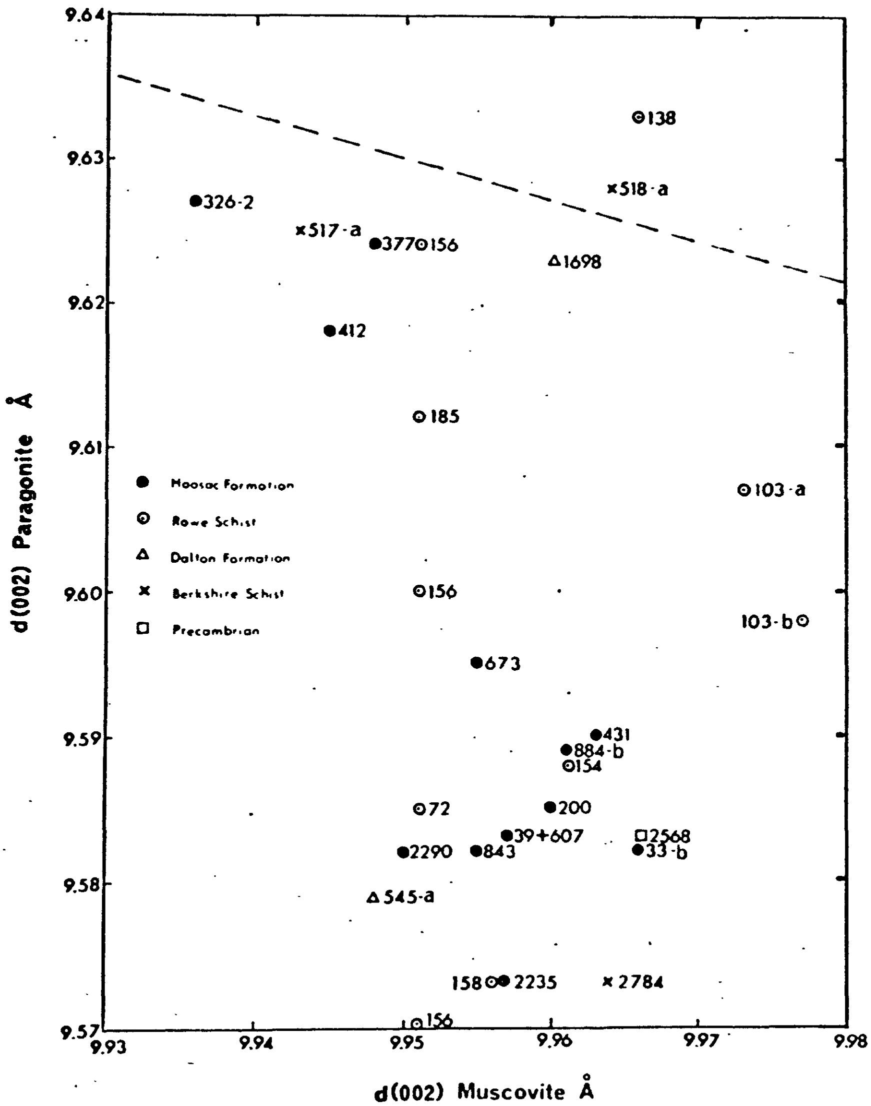


Figure 19. d(002)$_{muscovite}$ and d(002)$_{paragonite}$ for investigated pairs. Dashed line is Zen and Albee's curve for coexisting muscovite and paragonite (1964, p. 909).

graphic position and d(002) values for either muscovite or paragonite.

Muscovite and paragonite can take an appreciable amount of margarite component into their structure. The amount may or may not be significant, but is certainly variable in the various assemblages in which paragonite and muscovite were found because only a few of the assemblages were saturated with respect to a calcium component. Margarite component would not appreciably affect the d-spacings of (002) in paragonite because margarite and paragonite have essentially the same d(002) values (figure 20). However, the effect on muscovite may be significant.

To estimate the amount of margarite component in the muscovite and paragonite the γ index of refraction and d(002) values for the three components were contoured on figure 20. It was assumed that variation in these physical properties was ideally related to mole fraction of the components. Theoretically, measurement of these two physical properties should yield a unique composition for the phase.

Deer and others (1962, v. III, p. 96) indicate a linear relationship between the γ index of refraction and the ratio $\frac{Ca}{Ca+Na}$ for the binary paragonite-margarite. The relationship between γ index of refraction and $\frac{Na}{Na+K}$ is apparently controlled more by components other than margarite, paragonite, and muscovite. Assuming that Deer's relationship is essentially correct, the amount of margarite component in the

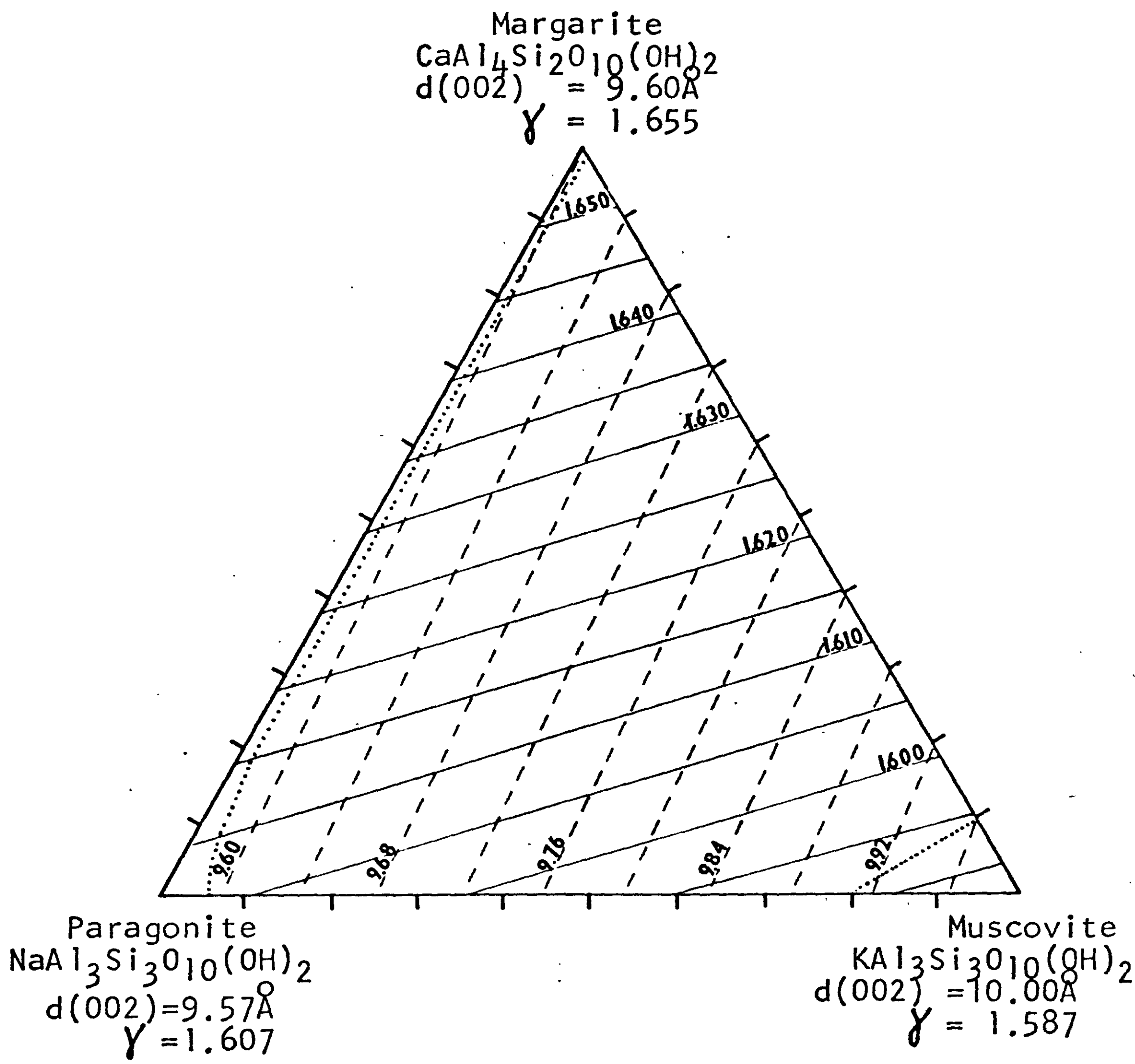


Figure 20. Contours for d(002) and indices in the ternary margarite ($CaAl_4Si_2O_{10}(OH)_2$)-muscovite ($KAl_3Si_3O_{10}(OH)_2$)-paragonite ($NaAl_3Si_3O_{10}(OH)_2$). Dotted lines are schematic solvi at 600°C. Data from Deer and others, 1962.

paragonites may be estimated. All the γ indices of refraction for paragonite in the Windsor quadrangle (table 15) were between 1.602 and 1.608, indicating that the paragonite has little or no margarite in solid solution. The muscovite would have even less margarite in solid solution.

The relationship between the basal spacing d(002) of muscovite and paragonite and the amount of solid solution between them is not yet clear.

Zen and Albee (1964, p. 917) give a curve for the relationship between d(002) and mole fraction of paragonite for the muscovite-paragonite binary (figure 21). Their curve is linear with extrapolated end members having d(002) equal to 10.034Å and 9.607Å for muscovite and paragonite, respectively. It has not been established if the relationship between d(002) and mole fraction of paragonite is linear as Zen and Albee assume. It almost certainly is not (Fujii, 1966). Eugster (quoted in Zen and Albee, 1964, table 2, p. 916) gives end members with d-spacings for (002) of 10.042Å and 9.575Å. Evans and Guidotti (1966, p. 39) give a value of 10.006Å for pure muscovite.

Thus the mole fraction of paragonite component in the muscovite ranges from 0.1 to 0.2 (figure 21 and table 12). The mole fraction of paragonite in the paragonite ranges from 0.9 to 1.0.

TABLE 15

Index of Refraction of Paragonite

Hoosac Formation	γ
37e	1.602 ± 0.002
39	1.608 ± 0.002
326-2	1.605 ± 0.002
431	1.604 ± 0.002
2235	1.603 ± 0.002
Rowe Schist	
138	1.604 ± 0.002
Berkshire Schist	
518-a	1.605 ± 0.002

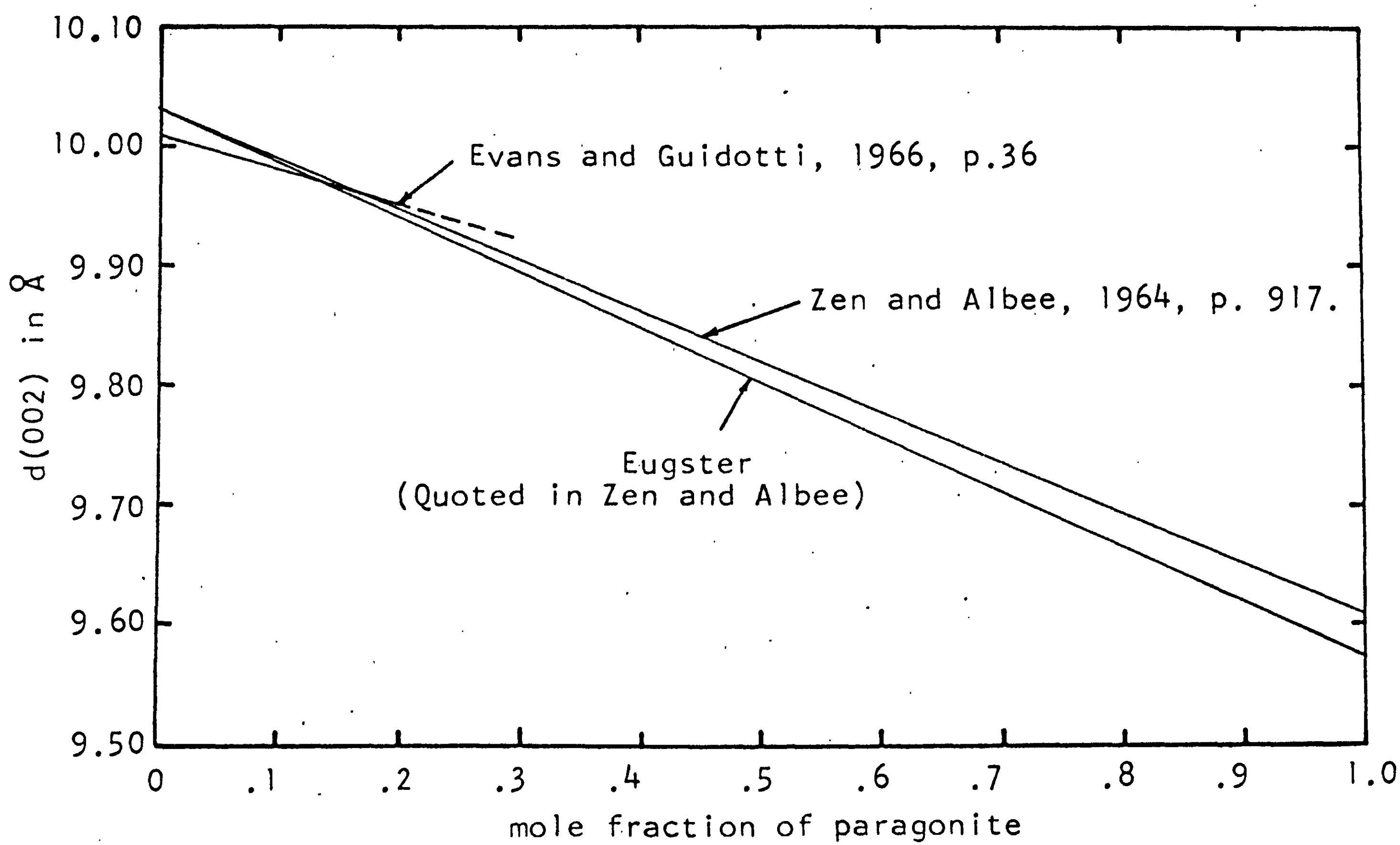


Figure 21. Relation between $d(002)_{mica}$ and mole fraction of paragonite.

Albite

The composition of plagioclase from the Hoosac Formation was approximately determined by the method of maximum extinction angle for albite twins. The albite content of the plagioclase was Ab_{95} or greater.

An analysis of albite from the eastern exposures of the Hoosac Formation in the Hoosac Tunnel in the North Adams quadrangle is reported in Pumpelly and others (1894, p. 60). The albite is $Ab_{98}An_0Or_2$.

CHAPTER V

METAMORPHISM

Paleozoic Metamorphism

General Statement

The entire Windsor quadrangle is within the garnet zone of metamorphism on the western flank of the westernmost Paleozoic metamorphic high affecting the northern Appalachian Mountains in New England.

Staurolite is present 5 miles to the east in the Goshen Formation of Devonian age (Osberg and others, in preparation). Three miles west of the quadrangle garnet it absent in the pelitic rocks of the Berkshire Schist and Greylock Schist. The garnet isograd lies approximately at the eastern base of Mount Greylock in the Cheshire quadrangle.

The metamorphic mineral assemblages of the pelitic rocks of the Berkshire Schist, Dalton Formation, Hoosac Formation, and the Rowe Schist suggest remarkably uniform metamorphic conditions for the entire area mapped.

Metamorphism of the Pelitic Rocks

The pelitic rocks of the Berkshire Schist, Dalton Formation, Hoosac Formation, and the Rowe Schist may be depicted and studied using either the A-F-M projection of Thompson or a portion of the Al_2O_3-K_2O-Na_2O projection or both.

A-F-M Projection

Thompson's projection (1957) is applicable only to rocks with quartz and muscovite (or orthoclase) and assumes an externally controlled chemical potential for $H_2O(\mu_{H_2O})$. All the pelitic rocks considered had assemblages consistent with these assumptions.

The muscovite generally contains appreciable paragonite in solid solution. Na^+ will not affect any phase already on the A-F-M face. Biotite is the only phase that will be affected because it lies within the A-K-F-M tetrahedron, will take up minor Na^+, and thus plot closer to the A-F-M face. This chemical variation is less significant than variations in the Al content of the biotite.

On figure 22 the positions of all phases are shown ideally with respect to Al_2O_3:FeO + MgO. This is a poor assumption since biotite and chlorite have variable amounts of aluminum, depending on the assemblage and the metamorphic grade. The $Fe^{+2}/(Fe^{+2} + Mg^{+2})$ ratios for garnet, biotite, and chlorite have been estimated from optical and X-ray measurements. Tie lines for coexisting phases are based on these measurements and are in agreement with the findings of other workers. The tendency for higher FeO:MgO ratios in the garnet zone of metamorphism is as follows: garnet > chloritoid > biotite > chlorite.

The assemblages found throughout the quadrangle are plotted on figure 22. No changes in three phase triangles are present in the quadrangle, and no 4-phase assemblages in the A-F-M projection have been seen although these might

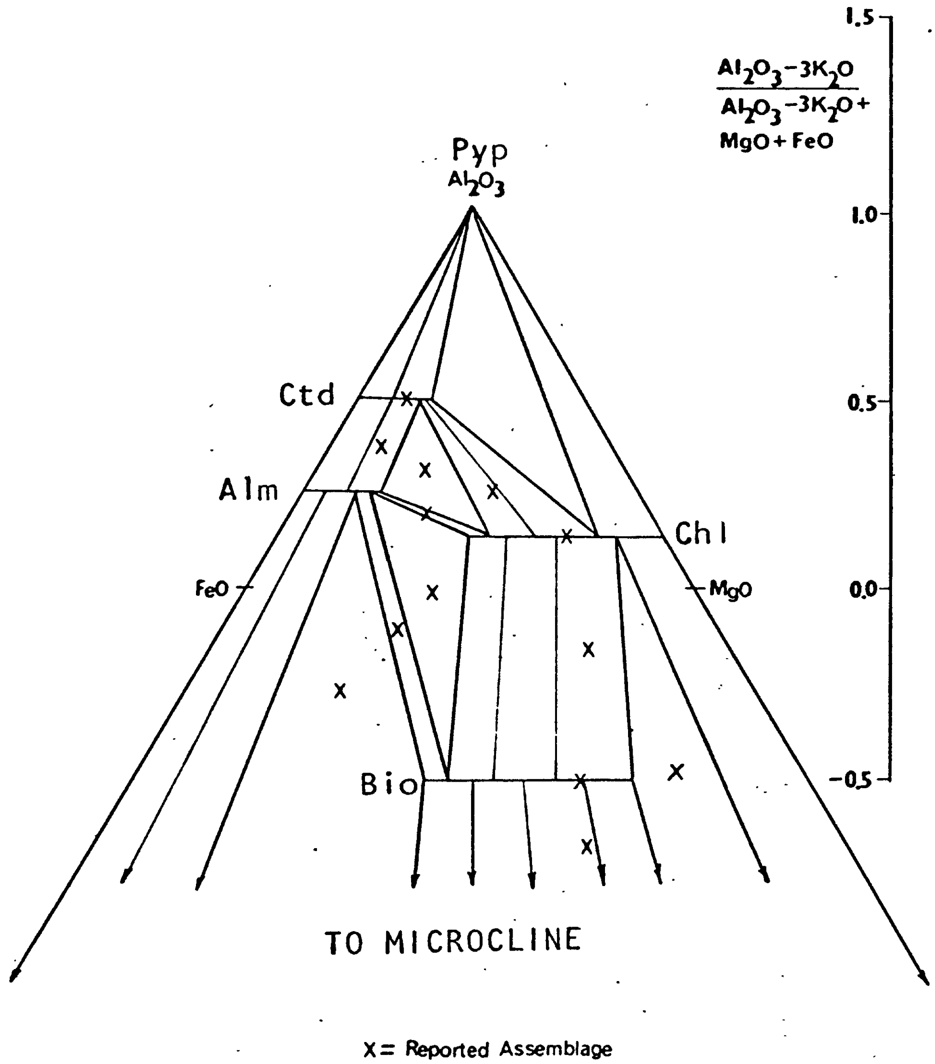


Figure 22. Assemblages in Thompson's A-F-M projection found in the Windsor quadrangle, Massachusetts. After Thompson, 1957.

be expected with a large amount of solid solution of other components in the iron-magnesium aluminum silicates. The garnets probably are the least suitable for plotting in the A-F-M. projections, containing appreciable Mn as spessartite $[Mn_3Al_2(SiO_4)_3]$. The absence of pyrophyllite or kyanite places a limit on the composition of the protolith. This will be discussed below.

Al_2O_3-$KAlO_2$-$NaAlO_2$ Projection

Eugster and Yoder (1955) studied the system $NaAlSi_3O_8$(Ab) -$KalSi_3O_8$(Or)-Al_2O_3 in equilibrium with an aqueous phase. They projected all hydrous phases in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-Al_2O_3-H_2O onto the Ab-Or-Al_2O_3 face of the tetrahedron. Their investigations of mica dehydrations are directly applicable only to quartz-free assemblages. However, their high temperature data on the degree of solid solution between muscovite and paragonite is independent of the presence or absence of quartz.

Thompson (1961) has considered a similar system but with an excess of SiO_2 (as quartz) and assumed an externally controlled μ_{H_2O}. The corners of the projection may then be written Al_2O_3, $KAlO_2$, and $NaAlO_2$ (henceforth A-K-Na) (figure 23).

This diagram is very applicable to metamorphic assemblages in the Windsor quadrangle because of the high $\frac{Na^+}{K^+}$ ratios found in certain of the pelitic rocks of the Berkshire Schist,

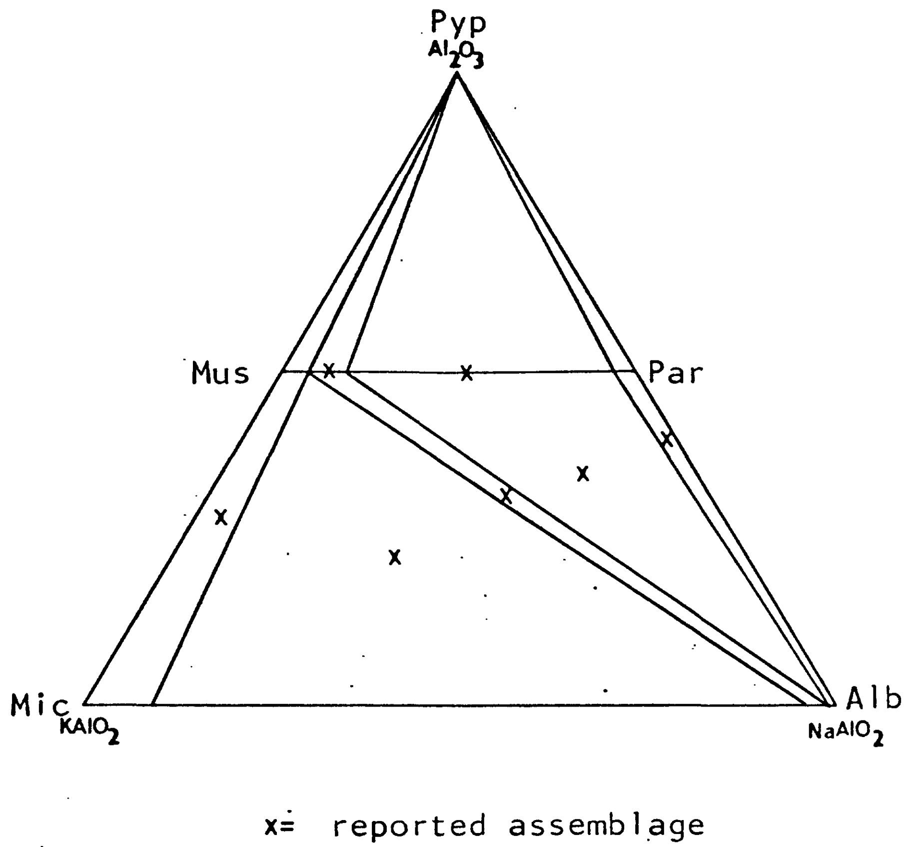


Figure 23. Assemblages found in the Windsor quadrangle, Massachusetts in the system Al_2O_3-$KAlO_2$-$NaAlO_2$-SiO_2-H_2O. After Thompson, 1961.

Dalton Formation, Rowe Schist, and the Hoosac Formation. Interestingly, the A-K-Na projection may be applied to the same rocks that are analyzed with the A-F-M projection.

The Hoosac Formation and the Rowe Schist have 3 phase assemblages in A-K-Na with a fourth Ca^{+2}-saturated phase (usually clinozoisite or epidote $\pm$ calcite) but with no addition of a fifth phase from the A-K-Na plane. The latter would correspond to a univeriant equilibrium.

Any attempt (see below) to use coexisting muscovite-paragonite pairs for geothermometry must assess the influence of a calcic component (and any other component for that matter) in solid solution on the equilibrium.

Metamorphism of Mafic and Ultramafic Rocks

The only pre-metamorphic mafic and ultramafic rocks in the quadrangle occur in the upper part of the Hoosac Formation, in the Rowe Schist, and in the serpentine body in the center of the quadrangle.

All associations are typical of the greenschist facies (table 16).

Retrograde Metamorphism

The only textural evidence for retrograde metamorphism in the Windsor quadrangle during the Paleozoic is the presence of garnet partially jacketed with chlorite and rarely chlorite is in a replacement texture with biotite.

Table 16

Assemblages from Mafic and Ultramafic Rocks

	Greenschist in the Rowe Schist		Amphibole Gneiss in the Rowe Schist	Ultramafic Rocks* in the Hinsdale Gneiss and Rowe Schist	
Quartz	X	X	X		
Plagioclase	X	tr	tr		
Chlorite	X	X	X	X	
Biotite	tr				
Epidote group	X	X	X		
Hornblende		tr	X		
Magnetite	tr	tr		X	X
Apatite	tr				
Calcite + dolomite	X	X			X
Sphene	tr	tr	tr		
Talc				X	X
Serpentine				X	X

* Paleozoic age

Garnet, chlorite, biotite, and chloritoid associations and compositions suggest that the Windsor quadrangle was subjected to a rather uniform grade of metamorphism.

The amount of solid solution between coexisting muscovite and paragonite is a function mainly of temperature. The change in the d(002) values is greater for muscovite than for paragonite. However, the data in table 13 and figure 19 show muscovite with a narrower spread of values for d(002) than paragonite. Furthermore, the peaks for (006) for paragonite on diffractometer charts are commonly very broad and multiple. This suggests either retrograde effects or the lack of complete equilibration during the maximum temperature of metamorphism. Retrograde metamorphism is also suggested by the large number of paragonite values which fall in the range 9.57 to 9.59Å, the former being the value for the pure paragonite end-member.

One specimen from the Hinsdale Gneiss offered an opportunity to check the hypothesis that most of the paragonite not present represents extensive retrograde metamorphism. The assemblage consisted of white mica, quartz, scapolite, garnet, biotite, and secondary chlorite. The white mica appeared to be retrograde after the scapolite. X-ray determinations showed it to be composed of muscovite and paragonite; the d(002) values for these minerals are plotted on figure 19 and they fall in the maximum for the Paleozoic sediments.

Apparently very few of the paragonite d(002) values represent the conditions reached during the peak of metamorphism.

Precambrian Metamorphism

Rocks in Vermont probably equivalent to the Hinsdale Gneiss have been radiometrically dated and yield an age corresponding to the Grenville Orogeny (0.9 to 1.1 billion years) (Faul and others, 1963, p. 3).

Evidence from central Vermont (e.g. Thompson, 1950, p. 21) and southern Massachusetts (e.g. Emerson, 1899, p. 110; Emerson, 1917, p. 23; Eskola, 1922, p. 293,294) suggests that the Mount Holly Gneiss and Hinsdale Gneiss and equivalent rocks were subjected to a Precambrian metamorphism that was more intense than the Paleozoic metamorphism and reached the sillimanite zone. In the Windsor quadrangle there are few relic features that verify this former grade of metamorphism.

A metamorphic discontinuity between the Precambrian and Paleozoic rocks is indicated by textural and mineralogical evidence.

<u>Textural Evidence</u>:

1. Retrograde effects are more pronounced in the Hinsdale Gneiss than in the Paleozoic rocks. Both garnet and biotite have commonly retrograded to chlorite whereas the garnet in the Paleozoic pelitic rocks is usually only jacketed with chlorite

or is fresh.

2. Muscovite in the Precambrian rocks is typically minor in amount and very fine-grained, commonly having a shredded appearance and a sieve-like texture with numerous inclusions -- atypical for the muscovite in the overlying sediments.
3. The scapolite-bearing assemblages (table 18) have a very fine-grained matrix of muscovite and paragonite that appears retrograde after the scapolite. The muscovite and paragonite have d(002) spacings of 9.966 $\pm$ 0.004Å and 9.583 $\pm$ 0.002Å respectively. The muscovite value suggests a temperature of about 520° C. This paragonite-muscovite assemblage is saturated with Ca^{+2} and thus the temperature would be a minimum.
4. The contrast in the size of porphyroblasts of microcline across the Precambrian-Cambrian unconformity is great. South of MacDonald Brook the microcline in the Hinsdale Gneiss is as much as 25 mm. in diameter, whereas less than 500 feet away the Dalton Formation contains only 1 to 2 mm. porphyroblasts of microcline.

Mineralogical Evidence:

1. Diopside, scapolite, and tremolite are common in the calc-silicate rocks in the Hinsdale Gneiss, whereas they are not

present in the Paleozoic metasediments.

2. The mafic rocks in the Hinsdale Gneiss have a predominance of hornblende over chlorite. The latter is not retrograde. The mafic rocks in the Paleozoic rocks are more hydrous.
3. Graphite in the Hinsdale Gneiss is highly crystalline, occurring in places as much as 5 mm. in diameter. "Graphite" in the Paleozoic rocks is finely divided dust.

There are very few rocks in the Precambrian terrane that have the proper composition to indicate metamorphic conditions. The bulk of the rocks are quartz-feldspar-biotite gneisses with minor amounts of other phases.

Calc-silicate and mafic rocks compose less than 5 percent of the Hinsdale Gneiss but are useful in indicating a metamorphic discontinuity between the Precambrian and Paleozoic rocks.

Assemblages from mafic rocks and calcium-rich rocks are listed in tables 17 and 18 respectively. The assemblages in the mafic rocks are not diagnostic but the association diopside-quartz indicates metamorphic conditions higher than those found in the Paleozoic rocks that have the proper composition for the development of diopside. Scapolite also indicates higher grade metamorphism than garnet zone.

TABLE 17

Assemblages from Mafic Rocks from the Hinsdale Gneiss

	1	2	3	4	5	6	7	8	9
Quartz	tr	X	X		X	X	X	X	X
Plagioclase	tr	tr	tr	X	tr	X	X		X
Microcline						X	X		
Chlorite			tr[2]		X				tr
Garnet				X					X
Biotite	tr	X	X	X	X	X	X		X
Epidote group		X	X	X	X	tr	X		
Hornblende	X	X	X	X	X	X	X	X	X
Magnetite								tr	
Ilmenite				tr					
Hematite	tr							tr	
Apatite		tr			tr		tr		tr
Calcite + dolomite							tr		
Zircon				tr					
Sphene		tr	tr	tr	tr				
Pyrrhotite									tr

[2] = retrograde

TABLE 18

Assemblages from calcium-rich rocks from the Hinsdale Gneiss

	1	2	3	4	5	6	7	8	9	10	11	12	13
Quartz	X	X	X	X	X	X	X	X	X	X	X		X
Muscovite			X^{2}										
Paragonite			X^{2}										
Plagioclase	X	X						X			X		
Microcline					X			X		X	X		
Chlorite			X^{2}						X				
Garnet			X		X		X			X			
Biotite	X		X^{2}					X	X	X	X		X
Epidote group	X	X		X	X	X		X					X
Tremolite		X		X		X		X	X	X	X^{2}	X	X
Diopside	X			X	X	X	X				X	X	
Scapolite			X	X			X						
Calcite	X			X	X		X						
Sphene				X	X								
Graphite	X										X		
Pyrrhotite					X								
Magnetite									X				

2 = retrograde

Conditions of Metamorphism

Temperature

Various writers have postulated the use of solid solution between coexisting phases as a geothermometer in metamorphism. To accomplish this successfully, all of the following conditions must be met:

1. The equilibrium must be frozen in, recording the highest temperature of formation and not later less intense thermal events or cooling.
2. The solid solution, ideally, should contain only those components whose influence on the equilibrium is known.
3. The amount of solid solution between the two phases should be independent of the pressure unless an independent estimate of pressure can be made.
4. Changes in physical properties should be easily measurable and correlated with change in temperature of formation.

The writer selected the coexisting phases muscovite and paragonite to test the applicability of this pair as a geothermometer. The only condition of the above 4 that is met is the third. Fujii (1966) has demonstrated that for dehydroxylated muscovite and paragonite, the amount of solid solution is relatively insensitive to total pressure. The amount of paragonite component in the muscovite is decreased by the presence of margarite component in the paragonite. The result is an increase in the basal spacing of the muscovite and any attempt to estimate temperatures from the basal spacing would yield anomalously low values (Zen and Albee, 1964).

The paragonite side of the muscovite-paragonite (Eugster and Yoder, 1955, p. 127) solvus is very steep and a wide variation in temperature does not appreciably affect the amount of muscovite in solid solution in the paragonite. Significant variations in temperature might not be reflected by measurable variations in d(002).

Figure 24 is a schematic isothermal section at 600° C in the ternary margarite-paragonite-muscovite. Six hundred degrees represents a temperature slightly higher than the maximum estimate for metamorphism in the Windsor quadrangle.

Conclusions from examination of figures 19, 20, and 24 are:

1. Many paragonites have d(002) values higher than would be predicted by solvus data.
2. Margarite component in solid solution in paragonite coexisting with muscovite will cause a reduction of $d(002)_{paragonite}$ and an increase in $d(002)_{muscovite}$. No paragonite, however, coexisting with muscovite, should have a d(002) less than 9.60Å if there is no solubility gap between paragonite and margarite. This is a good assumption.

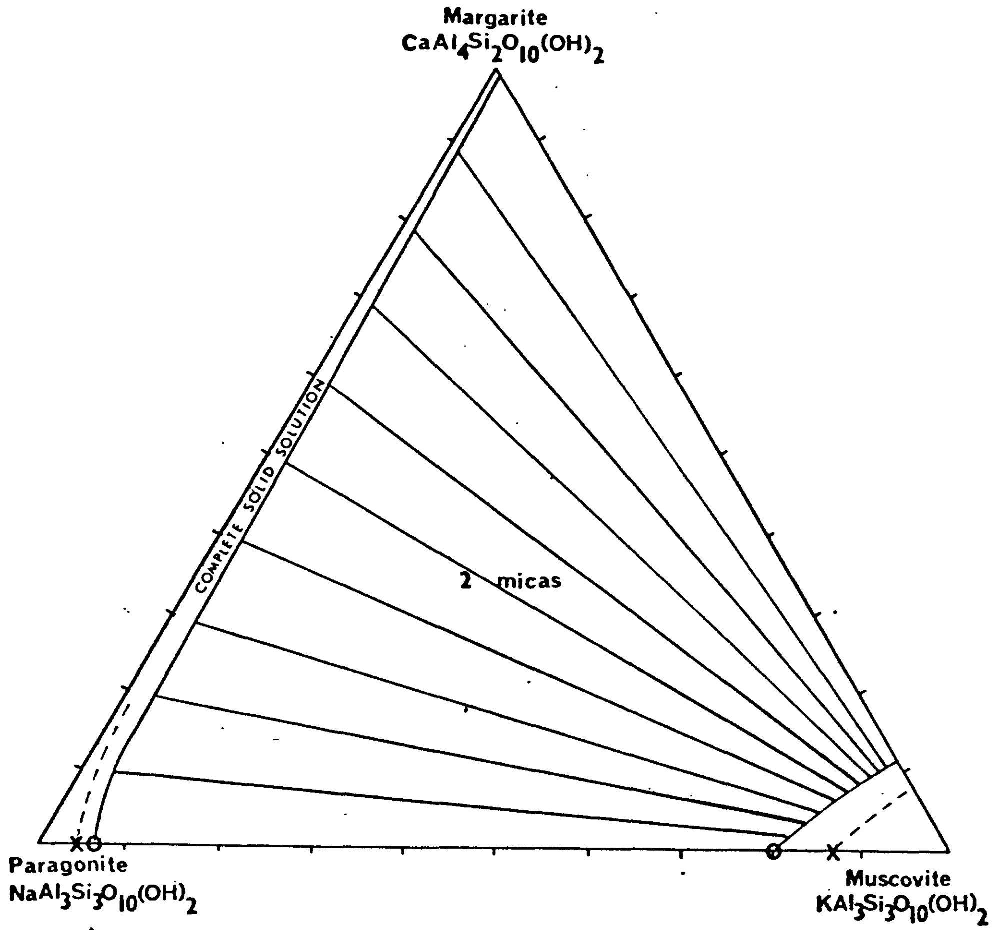


Figure 24. Isothermal section through the ternary margarite-muscovite-paragonite. Section is schematic at 600°C. Circled points are from data of Fujii, 1966 and Eugster and Yoder, 1955. Solvus points for 500°C are indicated by X and are also from Eugster and Yoder, 1955.

3. There are numerous paragonites with d(002) less than 9.60Å that are coexisting with muscovite. This may be the product of retrograde metamorphism.
4. Muscovites should have d-spacings of approximately 9.92 to 9.95Å for a temperature range of 600° to 500° C. Most of the d(002) values are larger than 9.95Å.

There is a maximum around 9.95 to 9.96Å for $d(002)_{muscovite}$. This would correspond to a temperature range (figure 25) between 540 and 560° C. A maximum temperature of about 600° C is represented by a spacing of 9.935Å. A minimum temperature of 490° C is yielded by the value 9.977Å.

The d(002) values for muscovite coexisting with paragonite are as high as values found in the staurolite zone by Zen and Albee (1964), (figure 26). Albee's (1965) temperature for the kyanite zone of Lincoln Mountain, Vermont is estimated at 550° C which is comparable to the estimate for the rocks in the Windsor quadrangle. Albee also has the tie lines chloritoid-chlorite and garnet-chlorite. The lack of kyanite and/or staurolite in the Windsor quadrangle may be caused by the compositional restrictions of the rock formations.

One other method may serve to indicate the thermal conditions during metamorphism. Within the quadrangle there are several ultramafic bodies. The largest of these is located

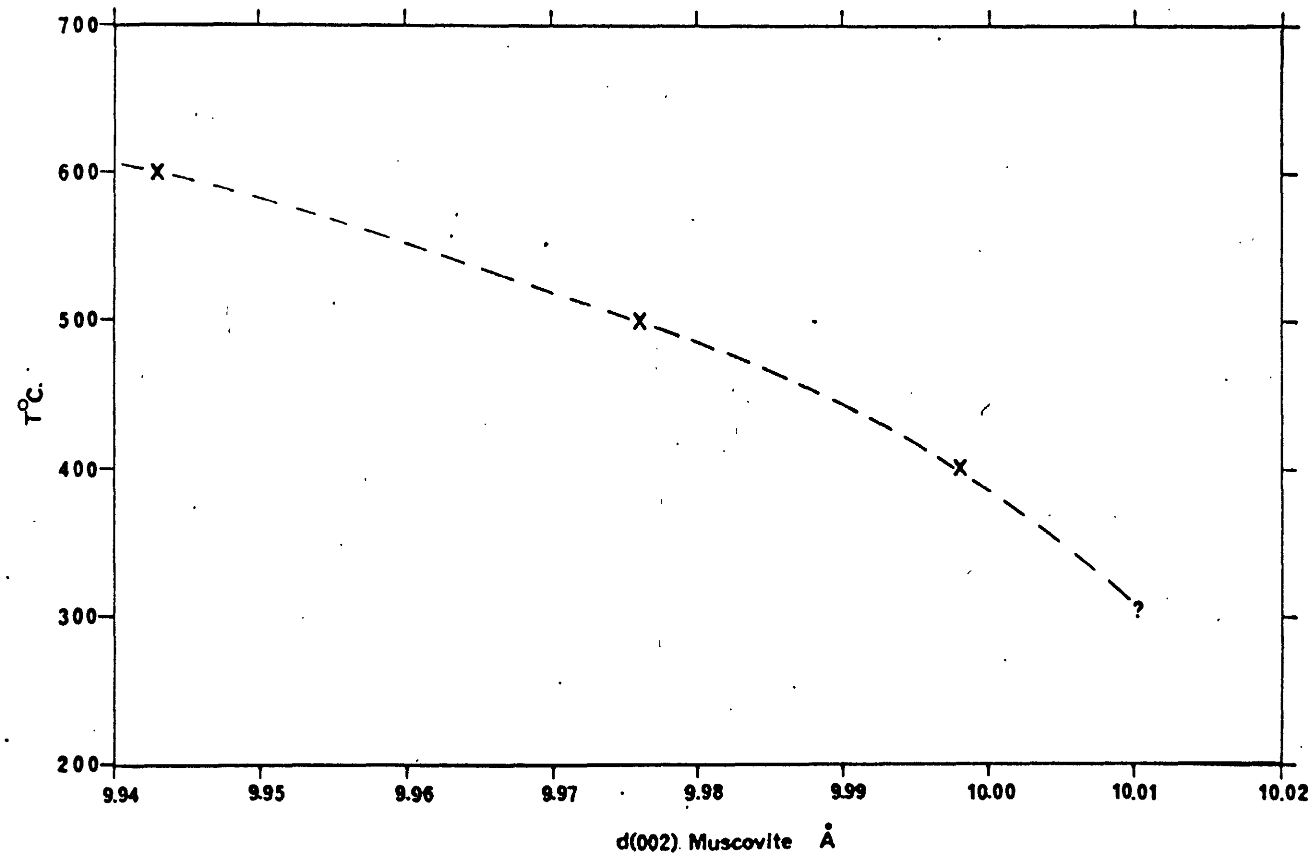


Figure 25. Relation between temperature and d(002) for muscovite coexisting with paragonite. After data of Eugster, quoted in Albee, 1965.

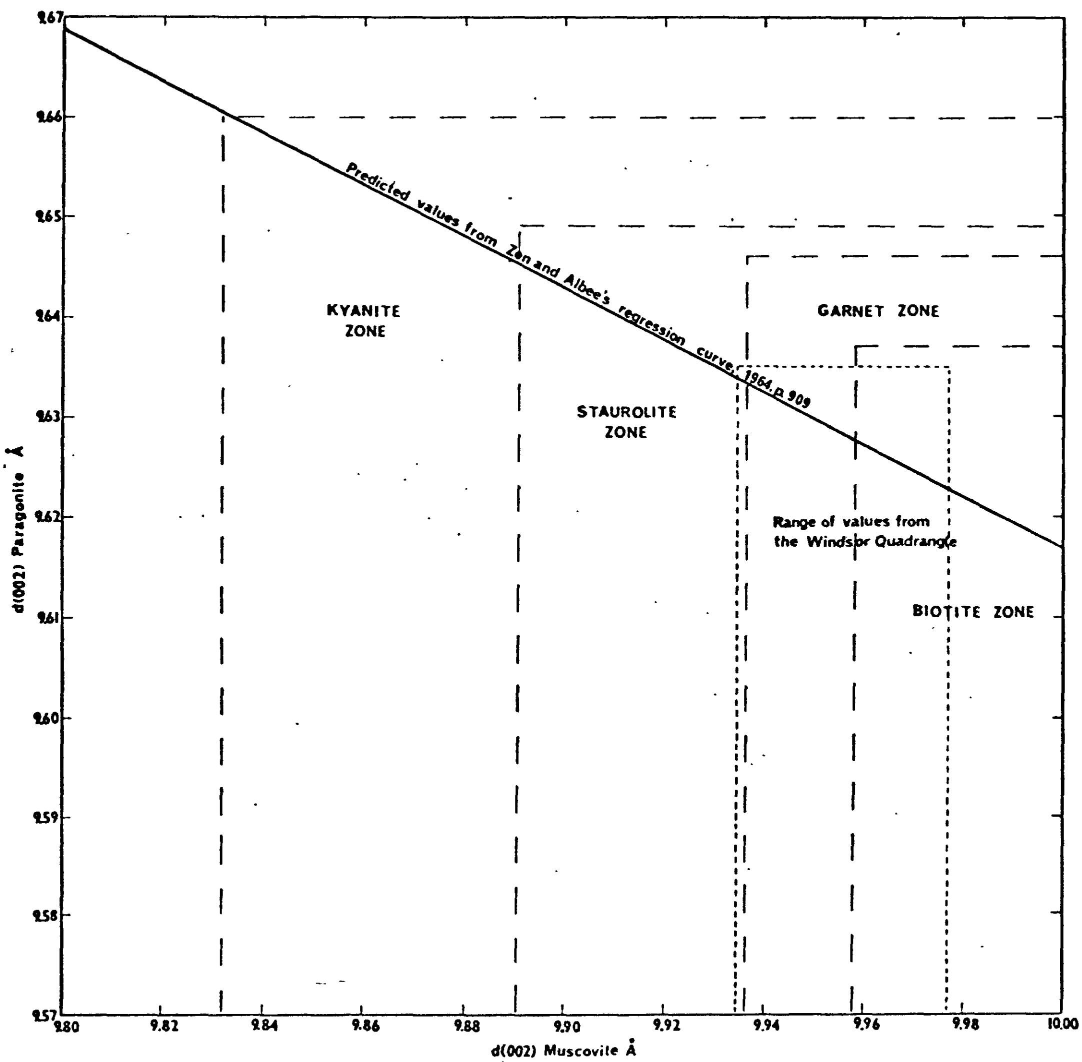


Figure 26. Relation between d(002)$_{muscovite}$ and d(002)$_{paragonite}$ with limits of metamorphic zones.
Modified from Zen and Albee, 1964.

in the Hinsdale Gneiss and is zoned with a serpentine core and a talc-carbonate rim. The zoning implies that the body was emplaced and then steatitized.

The upper stability limit for serpentine is 500° C, probably lower if the humidity (Thompson, 1957) is less than 100 percent. The steatitization of this serpentine body may have been caused by incomplete dehydration of the serpentine body during the regional metamorphism in the Devonian. Temperatures would be near 500° C. There are several smaller ultramafic bodies in the Rowe Schist that are talc and talc-carbonate rock. Because rates of diffusion would partially control the extent of steatitization, these smaller bodies would have a better chance for complete dehydration.

The widespread occurrence of paragonite and chloritoid in the quadrangle suggests a maximum possible temperature for the metamorphism of 660° C. because the decomposition temperatures for these phases are 660° C (Eugster and Yoder, 1955, p. 127) and 700° C (Halferdahl, 1961, p. 99) respectively. Halferdahl states that the assemblage almandite-chloritoid (pure Fe)-quartz is stable between 485° and 600° C at water pressures ranging from 1,010 to 2,000 bars. Addition of magnesium to the assemblage would raise the upper stability limit.

Pressure

Both compositional restrictions (figures 22 and 23) and metamorphic grade prevented the appearance of an aluminum silicate (e.g. kyanite) in the Windsor quadrangle. Thus there is no pressure-sensitive phase to estimate the total pressure.

In central and southern Vermont and western Massachusetts the sequence of appearance of aluminum silicates with increasing grade of metamorphism is kyanite followed by sillimanite. Kyanite is present in Paleozoic rocks in central Vermont (Lincoln Mountain; Albee, 1965:Townshend; Rosenfeld, 1954) and just south of Chester (Emerson, 1917), and at Williamsburg, Massachusetts. It seems reasonable to assume that the Paleozoic rocks in the Windsor quadrangle were probably buried at depths at which the pressure exceeded the pressure of the triple point for the Al_2SiO_5 polymorphs.

An aluminum silicate would not appear until either the stability field of chloritoid was exceeded or the tie line chloritoid-chlorite was broken.

CHAPTER VI

ECONOMIC GEOLOGY

Talc

There are two ultramafic bodies of mappable size in the quadrangle. The first of these was mined and is about 1,500 feet north of the intersection of Bush Road and Bush Cemetery Road (6). No outcrops of the body were observed on the surface that is now heavily overgrown pasture. Talc was mined in the early part of the century from a shaft, and tunnels were driven along strike to the north and south. Talc was taken by team to the town of Charlemont for processing. Judging by the size of the dumps (200 feet long and 30 feet wide) the workings must have been fairly extensive when the mine shut down. Specimens taken from the dump are either pure talc rock or talc-carbonate rock; no actinolite or relics of serpentine or olivine were seen.

In the center of the quadrangle (5), in Hinsdale Gneiss, there is a large ultramafic body 800 feet long (north-south) and at least 300 feet wide in places (plate 1). The body has been prospected in the last ten years by surface exploration and core drilling by Al Chamberlin of Cheshire. The body strikes north and the foliation of the talc dips 60 to 90° east and is concordant with the country rock. Drilling indicates that the talc body plunges steeply to the northeast. The bulk of the body is a serpentinized dunite with relic olivine[1] in some specimens. Near the foot wall of the body, the northwest corner of the body, open pit exploration has been undertaken in the last ten years. Solid talc has been exposed in a band ranging from 10 to 30 feet wide for a distance of 150 feet along the strike. Core-drilling exploration down the plunge of the talc band revealed a zone of talc-carbonate at a depth of 60 feet. This is the only part of the ultramafic body that has been explored for talc. The talc is of good quality, being free of actinolite, serpentine, and carbonate within the 10 to 30 foot band. The serpentine is highly sheared but contains no fibrous minerals. It would not be of value as either ornamental stone or as a source of asbestos. Opaques in the serpentine and talc (presumably chrome-bearing magnetite) are rare. The body is in well drained, open country with good access to roads. If the market for talc warrants, this body might be worth further surface and at depth investigation.

Ornamental Stone

The albite schist of the Hoosac Formation was quarried for an ornamental stone or a construction stone in a small pit 800 feet northeast of Lewis Hill (3). The pit is completely overgrown. The rock is a calcareous quartz-albite-muscovite-chlorite-clinozoisite schist, and is unusually homogeneous with good splitting properties. Calcite composes

[1] Reported by the owner, not seen by the writer.

as much as 2 percent of the rock.

Lime

Although carbonate rocks are exposed in the quadrangle, both dolomitic and calcitic, no prospects or workings are present. The areas of outcrop of these rocks are now heavily populated and the carbonate rocks present probably will never be developed. The quarries to the northwest of Adams (outside the quadrangle) in the Shelburne Marble will accommodate future needs in the area.

Glass Sand

Glass sand has been quarried in the past in the Windsor quadrangle in two places. Both sites are in the Cheshire Quartzite and represent disaggregated quartzite (see discussion in stratigraphy section). Both deposits are limited in size. The most accessible locality is on Main Road at the height of land between the towns of Savoy and Adams. The road passes through the outcrop. The other locality, at an elevation of 1,500 feet, lies 100 feet east of Dry Brook. Pits of this type are plentiful in the Cheshire and Windsor area and have been described in full by Chute (1943). None of these pits are currently being operated as a source of glass sand.

Crushed Quartzite

The Cheshire Quartzite could conceivably serve as a large reserve of crushed quartz of high purity, suitable for glass-making or as an ornamental stone aggregate in pre-stressed concrete. The Cheshire Quartzite typically is well exposed in most places, has a rather gentle dip, and forms dip slopes. This type of occurrence would be amenable to open pit methods of development. The areas of best outcrop are readily accessible by road and sparsely populated. They include:

Burlingame Hill (1) - north-south continuous exposure of 2,000 feet, east-west continuous exposure 200+ feet; probable thickness in excess of 50 feet.

Dry Brook (4) - exposures of 100 x 100 feet and a thickness exceeding 25 feet.

MacDonald Brook area (4) - on the western slope at the break in slope at the western edge of the quadrangle; continuous exposure of the Cheshire Formation for greater than 4,000 feet along strike; the formation is 100 feet thick here and could be quarried for at least 200 feet along the dip slope.

Graphite

Abundant graphite was observed in several outcrops of Hinsdale Gneiss. An old "lead mine" on the southwest slope of Hill 2,145 feet (2) at an elevation of about 2,030 feet consists of a tunnel driven to the north parallel to the strike of the banded gneisses. Judging by the dumps, the tunnel was not driven very far and subsequently has collapsed. The graphite occurs as disseminated flakes as much as 5 mm. in diameter in calc-silicate gneisses derived from carbonaceous shale and silty limestone.

Another prospect, 1,200 feet northeast of Hill 2,043 feet on the southwest slope of the knob (5), has graphite in disseminated flakes as much as 5 mm. in diameter in calc-silicate gneisses similar to those reported above, but containing diopside knots as much as 6 inches in diameter and disseminated small grains of pyrrhotite.

Sand and Gravel

In the northwest corner of the quadrangle there are large sand and gravel deposits in a kame terrace. They are about 2,000 feet long, 500 feet wide and 200 feet thick. These deposits have been worked extensively in the past but are now nearly inactive. Although the reserves are large, the culture is encroaching on the deposits and making further development impossible.

CHAPTER VII

GEOLOGIC HISTORY

The oldest rocks in the Windsor quadrangle are the Hinsdale Gneiss and Stamford Granite Gneiss of probably Grenville Age (1.1 billion years). They represent an eugeosynclinal suite of rocks that were metamorphosed to the sillimanite (?) zone and folded into east-west to northwest-southeast structures. Sillimanite has been reported from Otis, Massachusetts in the Washington Gneiss (Emerson, 1899, p. 110).

The area was then subjected to uplift and erosion for some 400 to 500 million years.

By Late Precambrian time the area had been reduced to a highly saprolitized terrane. At this time or in the Early Cambrian there was a transgression of the seas from east to west with deposition of a basal clastic unit.

The south-central part of the area remained nearly a positive area early in the Cambrian while subsidence of sedimentary troughs began to the east and west.

Although deposition was probably not continuous in the western sequence (Cady, 1945), there are no great erosional breaks in the record. The eastern sequence also has no erosional breaks up to the Middle Ordovician.

During the Middle Ordovidian the central and western part of the area underwent epeirogenic uplift. The result was two-fold:

1. Erosion of the western sequence of clastic and carbonate rocks, in places removing the entire section.
2. Gravity sliding or thrusting of sediments to the west from the present axial trace of the Hoosac nappe.

The protolith of the Berkshire Schist was deposited unconformably on all other Cambro-Ordovician sediments and on the Hinsdale Gneiss.

The interval of time between Upper Ordovician and Middle Devonian is not represented by any rocks in the quadrangle.

During the Middle to Upper Devonian the whole area underwent a severe phase of simultaneous deformation and garnet-grade metamorphism. The deformation resulted from compressive stresses with an east-west orientation with the development of large scale recumbent folding (the Hoosac nappe). Minor serpentinite bodies were emplaced at this time or slightly prior to the main phase of metamorphism. Low angle thrust faults were developed on the inverted limb of the Hoosac nappe because of either higher strain rates or less plastic conditions of folding.

The deep burial required for metamorphism and folding was probably accomplished by loading of the area with Devonian rocks in nappes rooting to the east.

Later in the Paleozoic the area was gently arched and folded by epeirogenic uplift along the axis of the Hoosac nappe. Reverse drag folds developed to the east and west of the axis of uplift. This deformation also produced normal faults with the displacement being east side down to the east of the axis and west side down to the west. These faults may, on the other hand, be related to Triassic faulting.

Since the end of the Paleozoic the area has been subjected to several epeirogenic uplifts and peneplanations during the Tertiary.

REFERENCES

Albee, A. L., 1957a, Geology of the Hyde Park quadrangle, Vermont: Unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, 137 p.

_______, 1957b, Bedrock Geology of the Hyde Park quadrangle, Vermont: U. S. Geol. Survey Geol. Quad. Map GQ 102.

_______, 1965, Phase equilibria in three assemblages of kyanite zone pelitic schists, Lincoln Mountain quadrangle central Vermont: Jour. Petrology, v. 6, no. 2, p. 246-301.

Bean, R. J., 1953, Relations of gravity anomalies to the geology of central Vermont and New Hampshire: Geol. Soc. America Bull., v. 64, p. 509-538.

Berry, W. B. N., 1962, On the Magog, Quebec, graptolites: Am. Jour. Sci., v. 260, p. 142-148.

Billings, M. P., Rogers, J., and Thompson, J. B., Jr., 1952, Geology of the Appalachian highlands of east-central New York, southern Vermont, and southern New Hampshire - Field trip No., 1, in Geol. Soc. America, Guidebook for field trips in New England, p. 1-71.

Billings, M. P., 1954, Structural Geology: Englewood Cliffs, New Jersey, Prentice-Hall, Inc., 514 p.

Bowen, N. L., and Tuttle, O. F., 1949, The system $MgO-SiO_2-H_2O$: Geol. Soc. America Bull., v. 60 p. 439-460.

Brace, W. F., 1953, The geology of the Rutland area, Vermont: Vermont Geol. Survey Bull. 6, 120 p.

Bucher, W. H., 1951, Infolded Mid-Ordovician limestone on pre-Cambrian north of Peekskill, New York and its bearing on the region's orogeny: Geol. Soc. America Bull., v. 62, p. 1426-1427.

_______, 1957, Taconic Klippe: A stratigraphic structure problem: Geol. Soc. America Bull., v. 68, Part 2, p. 657-674.

Cady, W. M., 1945, Stratigraphy and structure of west central Vermont: Geol. Soc. America Bull., v. 56, p. 515-588.

_______, 1956a, Bedrock geology of the Montpelier quadrangle, Vermont: U. S. Geol. Survey Geol. Quad. Map GQ 79.

Cady, W. M., 1956b, Stratigraphic relationships in northern Vermont and southern Quebec (abs.): Geol. Soc. America Bull., v. 67, p. 1811-1812.

_______, 1960, Stratigraphic and geotectonic relationships in northern Vermont and southern Quebec: Geol. Soc. America Bull., v. 71, p. 531-576.

Chang, P. H., and others, 1965, Bedrock geology of the Woodstock quadrangle, Vermont: Vermont Geol. Survey Bull. 29, 65 p.

Chidester, A. H., 1962, Petrology and geochemistry of selected talc-bearing ultramafic rocks and adjacent country rocks in north-central Vermont: U. S. Geol. Survey Prof. Paper 345, 207 p.

Chidester, A. H., and others, in press, Geology of the Rowe quadrangle, Massachusetts-Vermont: U. S. Geol. Survey Geol. Quad. Map GQ.

Christensen, M. N., 1963, Structural analysis ~~of~~ the Hoosac nappe in northwestern Massachusetts: Am. Jour. Sci., v. 261, p. 97-107.

Chute, Newton, 1943, Cheshire glass-sand deposits: Unpublished report, U. S. Geol. Survey Office, Boston, Massachusetts, 12 p.

Cloos, E., 1946, Lineation, a critical review and annotated bibliography: Geol. Soc. America, Mem. no. 18, 122 p.

Cohee, G. V., and West, W. S., 1966, Changes in stratigraphic nomenclature by the U. S. Geological Survey, 1965, U. S. Geol. Survey Bull. 1244-A.

Craddock, J. C., 1957, Stratigraphy and structure of the Kinderhook quadrangle, New York, and the Taconic Klippe: Geol. Soc. America Bull., v. 68, Part II, p. 675-724.

Currier, L. W., and Jahns, R., 1941, Ordovician stratigraphy of central Vermont: Geol. Soc. America Bull., v. 52, p. 1487-1512.

Dale, T. N., 1891, The Greylock synclinorium: Am. Geologist, v. 8, p. 1-7.

_______, 1896, Structural details in the Green Mountain region and in eastern New York: U. S. Geol. Survey, 16th Ann. Rept., Part 2, p. 543-570.

Dale, T. N., 1898, The slate belt of eastern New York and western Vermont: U. S. Geol. Survey, 19th Ann. Rept., Part 3, p. 153-307.

_______, 1910, Cambrian conglomerate of Hyston, Vermont: Am. Jour. Sci., v. 30, p. 267-270.

_______, 1914-1916, Field notes on the Algonkian-Cambrian boundary east of the Green Mountain axis in Vermont: U. S. Geol. Survey open-file report, Boston, Massachusetts.

_______, 1915, The calcite marble and dolomite of eastern Vermont: U. S. Geol. Survey Bull. 589, 67 p.

_______, 1916, The Algonkian-Cambrian boundary east of the Green Mountain axis in Vermont: Am. Jour. Sci., ser. 4, v. 42, p. 120-124.

_______, 1923, The lime belt of Massachusetts and parts of eastern New York and western Connecticut: U. S. Geol. Survey Bull. 744, 71 p.

Dana, J. D., 1884, On the decay of quartzite, and the formation of sand, kaolin, and crystallized quartz: Am. Jour. Sci. 3rd, v. 28, p. 448-452.

_______, 1885, Decay of quartzite: Pseudo-breccia: Am. Jour. Sci. 3rd, v. 29, p. 57-58.

Deer, W. A., and others, 1962, Rock-forming minerals: Vols. 1-4, New York, New York, John Wiley and Sons, Inc.

DeVore, G. W., 1955a, Crystal growth and the distribution of elements: Jour. Geology, v. 63, p. 471-494.

_______, 1955b, The role of adsorption in the fractionation and distribution of elements: Jour. Geology, v. 63, p. 159-190.

Diment, W., 1954, A regional gravity survey in Vermont, western Massachusetts and eastern New York: Unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University.

_______, 1956, Regional gravity survey in Vermont, western Massachusetts and eastern New York (abs.): Geol. Soc. America Bull., v. 67, p. 1688.

Doll, C. G., 1951, Geology of the Memphremagog quadrangle and southeastern portion of the Irasburg quadrangle, Vermont: Vermont Geol. Survey Bull. 3, 113 p.

Doll, C. G., and others, 1961, Centennial geologic map of Vermont: Vermont Geol. Survey.

Donnelly, T. W., 1963, Genesis of albite in early orogenic volcanic rocks: Am. Jour. Sci., v. 261, p. 957-972.

Eckstrand, O. R., 1963, Crystal chemistry of chlorite: Unpublished Ph. D. Thesis, Department of Geolgocial Sciences, Harvard University, 128 p.

Emerson, B. K., 1892, Outlines on geology of the Green Mountain region in Massachusetts: U. S. Geol. Survey, Geologic Atlas, Hawley Sheet.

_______, 1898a, Geology of Old Hampshire County, Massachusetts: U. S. Geol. Survey, Mon. 29, 790 p.

_______, 1898b, Holyoke quadrangle, U. S. Geol. Survey Geol. Atlas, Folio 50.

Emerson, B. K., 1899, The geology of Eastern Berkshire County, Massachusetts: U. S. Geol. Survey Bull. 159, 135 p.

_______, 1917, Geology of Massachusetts and Rhode Island: U. S. Geol. Survey Bull. 597, 289 p.

Emmons, E., 1844, The Taconic system: based on observations in New York, Massachusetts, Maine, Vermont, and Rhode Island: Albany, Carroll and Cook Printers.

Eskola, Pentti, 1922, On contact phenomena between gneiss and limestone in western Massachusetts: Jour. Geology, v. 30, p. 265-294.

Eugster, H. P., 1956, Muscovite-paragonite join and its use as a geologic thermometer (abs.): Geol. Soc. America Bull., v. 67, p. 1693.

Eugster, H. P., and Yoder, H. S., 1955, The join muscovite-paragonite: Carnegie Inst. Washington, Annual Rept. Dir. Geophysics Lab., 1954-1955, p. 124.

Evans, B. W., and Guidotti, C. V., 1966, The sillimanite-potash feldspar isograd in western Maine, U. S. A.: Contr. Mineralogy and Petrology, v. 12, p. 25-62.

Faul, H., and others, 1963, Ages of intrusion and metamorphism in the northern Appalachians: Am. Jour. Sci., v. 261, p. 1-19.

Foster, M. D., 1962, Interpretation of the composition and a classification of the chlorites: U. S. Geol. Survey Prof. Paper 414-A, 33 p.

Fowler, P., 1950, Stratigraphy and structure of the Castleton area, Vermont: Vermont Geol. Survey Bull. 2, 83 p.

French, B. M., and Eugster, H. P., 1965, Experimental control of oxygen fugacites by graphite-gas equilibrium: Jour. Geophysical Research, v. 70, p. 1529-1541.

Gates, R. M., and Bradley, W. C., 1952, The geology of the New Preston quadrangle: Connecticut Geol. Nat. History Survey Bull., Misc. Series no. 5, p. 1-45.

Goodwin, B. K., 1963, Geology of the Island Pond area, Vermont: Vermont Geol. Survey Bull. 20, 111 p.

Green, J. C., 1960, Geology of the Errol quadrangle, New Hampshire-Maine: Unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, 258 p.

Halferdahl, L. B., 1961, Chloritoid: Its composition, X-ray, and optical properties, stability, and occurrence: Jour. Petrology, v. 2, no. 1, p. 49-135.

Hatch, N., and Hartshorne, J., in preparation, Geology of the Heath quadrangle, Massachusetts-Vermont: U. S. Geol. Survey Geol. Quad. GQ.

Hawkes, H. W., 1941, Roots of the Taconic fault in west-central Vermont: Geol. Soc. America Bull., v. 52, p. 649-666.

Hemley, J. J., and Jones, W. R., 1964, Chemical aspects of hydrothermal alteration with emphasis on hydrogen metasomatism: Econ. Geology, v. 59, p. 538-569.

Herz, N., 1955, Alternative to the "Taconic Overthrust" hypothesis in western Massachusetts (abs.): Geol. Soc. America Bull., v. 66, p. 1714.

_______, 1958, Bedrock geology of the Cheshire quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ 108.

_______, 1961, Bedrock geology of the North Adams quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ 139.

Hess, H. H., 1933, The problem of serpentinization and the origin of certain chrysotile asbestos, talc and soapstone deposits: Econ. Geology, v. 28, p. 634-657.

Hitchcock, E., 1832, Report on the geology of Massachusetts: Am. Jour. Sci., v. 22, p. 1-70.

_______, 1833, Report on the geology, mineralogy, botany and zoology of Massachusetts: 692 p.

_______, 1841, Final report on the geology of Massachusetts: J. S. and L. Adams, Amherst; J. H. Butler, 831 p.

_______, and others, 1861, Report on geology of Vermont: Claremont, New Hampshire, v. 2.

Hollister, L., 1966, An electron probe investigation of metamorphic reactions and mineral growth histories, Kwoiek area, British Columbia: Unpublished Ph. D. Thesis, Cal. Inst. of Tech., 207 p.

Jopling, A. V., 1965, Hydraulic factors controlling the shape of laminae in laboratory deltas: Jour. Sed. Petrology, v. 35, p. 777-791.

Knopf, E. B., 1935, Recognition of overthrusts in metamorphic terranes: Am. Jour. Sci., 5th series, v. 30, p. 198-209.

Konig, R. H., 1961, Geology of the Plainfield quadrangle, Vermont: Vermont Geol. Survey Bull. 16, 86 p.

Kothavala, R. Z., 1964, Wall rock alteration at the United Verde Mine, Jerome, Arizona: Unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, 112 p.

Lobeck, A. K., 1948, Physiographic Provinces of North America: The Geographical Press.

MacFayden, J. A., Jr., 1956, Geology of the Bennington area, Vermont: Vermont Geol. Survey Bull. 7, 72 p.

Martin, G. C., 1898, An occurrence of dunite in western Massachusetts: Am. Jour. Sci., v. 6, p. 244-248.

Mason, B., 1960, Principles of geochemistry: John Wiley and Sons, Inc., New York, second edition, 310 p.

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

_______, 1961, Evolution of metamorphic belts: Jour. Petrology, v. 2, p. 277-311.

Miyashiro, A., 1964, Oxidation and reduction in the Earth's crust with special reference to the role of graphite: Geochim. et Cosmochim. Acta, v. 28, p. 717-730.

Myer, G. H., 1966, New data on zoisite and epidote: Am. Jour. Sci., v. 264, p. 364-385.

Osberg, P. H., 1952, The Green Mountain anticlinorium in the vicinity of Rochester and East Middlebury, Vermont: Vermont Geol. Survey Bull. 5, 127 p.

_______, 1956, Stratigraphy of the Sutton Mountains, Quebec: Key to stratigraphic correlations in Vermont (abs.): Geol. Soc. America Bull., v. 67, p. 1820.

_______, 1959, The stratigraphy and structure of the Coxe Mountain area, Vermont, Guidebook for the 51st Annual Meeting of the New England Intercollegiate Geological Conference, p. 45-51.

_______, 1965, Structural geology of the Knowlton-Richmond area, Quebec: Geol. Soc. America Bull., v. 76, p. 223-250.

_______, and others, in preparation, Geology of the Plainfield quadrangle, Massachusetts: U. S. Geol. Survey Geol. Quad. Map GQ.

Patterson. N. S., and Weiss, L. E., 1966, Experimental deformation and folding in phyllites: Geol. Soc. America Bull., v. 77, p. 343-374.

Phillips, A. H., and Hess, H. H., 1936, Metamorphic differentiation at contacts between serpentinite and siliceous country rocks: Am. Mineralogist, v. 21, p. 333-362.

Prindle, L. M., and Knopf, E. B., 1932, Geology of the Taconic quadrangle: Am. Jour. Sci., 5th series, v. 24, p. 257-302.

Pumpelly, R., Wolff, J. E., and Dale, T. N., 1894, Geology of the Green Mountains in Massachusetts: U. S. Geol. Survey Mon. 23, 206 p.

Robinson, P., 1963, Gneiss domes of the Orange area, Massachusetts and New Hampshire: Unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, v. 1, 253 p.

Rodgers, J., 1937, Stratigraphy and structure in the Upper Champlain Valley: Geol. Soc. America Bull., v. 48, p. 1573-1588.

Rosenfeld, J. L., 1954, Geology of the southern part of the Chester Dome, Vermont: Unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, 303 p.

_______, 1960, Rotated garnets and the diastrophic-metamorphic sequence in southeastern Vermont (abs.): Geol. Soc. America Bull., v. 71, p. 160.

Schuchert, C., 1937, Cambrian and Ordovician of northwestern Vermont: Geol. Soc. America Bull., v. 48, p. 1001-1078.

Shaw, A. B., 1958, Stratigraphy and structure of the St. Albans area, northwestern Vermont: Geol. Soc. America Bull., v. 69, p. 519-568.

Shumaker, R., 1959, Pawlet quadrangle: Guidebook for the 51st Annual Meeting of the New England Intercollegiate Geological Conference, p. 59-60.

Skehan, J. W., 1961, The Green Mountain anticlinorium in the vicinity of Wilmington and Woodford, Vermont: Vermont Geol. Survey Bull. 17, 159 p.

Stanley, R. S., 1964, The bedrock geology of the Collinsville quadrangle: State Geological and Natural History Survey of Connecticut, Quadrangle Report 16, 99 p.

Sturt, B., 1962, The composition of garnets from pelitic schists in relation to the grade of metamorphism: Jour. Petrology, v. 3, p. 181-191.

Theokritoff, G., 1964, Taconic stratigraphy in northern Washington County, New York: Geol. Sco. America Bull., v. 75, p. 171-190.

Thompson, J. B., Jr., 1950, Geology of the Ludlow, Vermont, area: Unpublished Ph. D. Thesis, Department of Geology and Geophysics, Massachusetts Institute of Technology.

_______, 1954, Structural geology of the Skitchewaug Mt. area, Claremont quadrangle, New Hampshire-Vermont: New England Intercollegiate Geological Conference Guidebook, 46th Conference, Dartmouth College, Hanover, New Hampshire.

_______, 1955, The thermodynamic basis for the mineral facies concept: Am. Jour. Sci., v. 253, p. 65-103.

_______, 1956, Skitchewaug nappe, a major recumbent fold in the area near Claremont, New Hampshire (abs.): Geol. Soc. America Bull., v. 67, P. 1826-27.

Thompson, J. B., Jr., 1957, The graphical analysis of mineral assemblages in pelitic schists: Am. Mineralogist, v. 42, p. 842-858.

_______, 1959, Stratigraphy and structure in the Vemont Valley and the eastern Taconics between Clarendon and Dorset: Guidebook for the 51st Annual Meeting of the New England Intercollegiate Conference, p. 71-87.

_______, 1961, Mineral facies in pelitic schists: In Fiziko-Khimicheskie Problemii Formirovaniya Garnyikh Porod H Rud (Physio-chemical problems of the formation of rocks and ores), G. A. Sokolov, ed., (in Russian), Akad, Nauk S. S. R., Inst. Geo. Ore Deposits, Petrog., Mineral., Geochem., p. 313-324, English summary, p. 323-324.

Thompson, J. B., and Rosenfeld, J. R., 1951, Tectonics of a mantled gneiss dome in southern Vermont (abs.): Geol. Soc. America Bull., v. 62, p. 1484-1485.

Tobisch, O. T., 1966, Large scale basin-and-dome pattern resulting from the interference of major folds: Geol. Soc. America Bull., v. 77, p. 393-408.

Trask, N. J., Jr., 1964, Stratigraphy and structure in the Vernon-Chesterfield area, Massachusetts, New Hampshire, Vermont: Unpublished Ph. D. Thesis, Department of Geological Sciences, Harvard University, 99 p. plus appendix.

Walcott, C. D., 1888, The Taconic system of Emmons, and the use of the name Taconic in geologic nomenclature: Am. Jour. Sci., 3rd series, v. 35, p. 229-242.

Walton, M. S,, and DeWaard, D., 1963, Orogenic evolution of the Precambrian in the Adirondack highlands, a new synthesis: Koninkl. Nederlandse Akad. Wetenoch Proc. Amsterdam, B., v. 66, p. 98-106.

Weaver, J. D., 1957, Stratigraphy and structure of the Copke quadrangle, New York: Geol. Soc. America Bull., v. 68, p. 725-762.

White, W. S., and Jahns, R., 1950, Structure of central and east-central Vermont: Jour. Geology, v. 58, p. 179-220.

Whittle, C. H., 1894, The occurrence of Algonkian rocks in Vermont and evidence for their subdivisions: Jour. Geology, v. 2, p. 396-429.

Wones, D. R., 1963, Physical properties of synthetic biotites on the join phlogopite-annite: Am. Mineralogist, v. 48, p. 1300-1321.

Wones, D. R., and Eugster, H. P., 1965, Stability of biotite: Experiment, Theory, and Application: Am. Mineralogist, v. 50, p. 1228-1272.

Zen, E-An, 1956, Stratigraphy and structure of the north end of the Taconic Range, Vermont (abs.): Geol. Soc. America Bull., v. 67, p. 1829-1830.

_______, 1959, Stratigraphy and structure at the north end of the Taconic Range and adjacent areas: Guidebook for the 51st Annual Meeting of the New England Intercollegiate Geological Conference, p. 1-15.

_______, 1961, Stratigraphy and structure at the north end of the Taconic Range in west-central Vermont: Geol. Soc. America Bull., v. 72, p. 293-338.

_______, 1964, Taconic stratigraphic names: Definitions and synonymies: U. S. Geol. Survey Bull. 1174, 95 p.

_______, 1966, Geologic map of the Bashbish quadrangle, Massachusetts, Connecticut, and New York: U. S. Geol. Survey Geol. Quad. Map GQ 507.

Zen, E-An, and Albee, A. L., 1964, Coexistent muscovite and paragonite in pelitic schists: Am. Mineralogist, v. 49, p. 904- 925.