# THE GREEN MOUNTAIN ANTICLINORIUM IN THE VICINITY OF WILMINGTON AND WOODFORD VERMONT

*By*

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VERMONT GEOLOGICAL SURVEY

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*Published by*

VERMONT DEVELOPMENT DEPARTMENT

MONTPELIER, VERMONT

BULLETIN NO. 17 1961


Looking northwest from central Whitingham, from a point near "G" in WHITINGHAM (Plate I). Looking across Sadawga Pond Dome to Haystack Mountain-Searsburg Ridge in the background; Stratton and Glastenbury Mountains in the far distance. Davidson Cemetery in center foreground on Route 8 serves as point of reference.

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# THE GREEN MOUNTAIN ANTICLINORIUM IN THE VICINITY OF WILMINGTON AND WOODFORD VERMONT

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## ABSTRACT

The Wilmington-Woodford area of southernmost Vermont (Fig. 1) comprises 385 square miles of metamorphosed pre-Cambrian and Lower Paleozoic strata. The area lies across the Green Mountain Anticlinorium although the greater part of it is east of the axis of this major fold. The Green Mountains represent the eroded remnant of the anticlinorium which is overturned and locally overthrust to the west.

The anticlinorium as a whole consists of a central area of pre-Cambrian rocks 90 miles long and is convex toward the east. The eastern flank of the Green Mountains is an easterly-dipping homoclinal sequence of regionally metamorphosed stratified rocks complicated locally by domal structures (Fig. 2). That of the western flank is composed of steeply-dipping or overturned and thrust-faulted Cambrian and Ordovician arenaceous, argillaceous, and carbonate beds.

The stratigraphic sequence of the Green Mountain core comprises pre-Cambrian gneisses, quartzite, conglomerate, and lime-silicate granulites. These are part of an earlier pre-Cambrian sequence which are overlain by a lithologically different succession, and are separated from this sequence by a profound angular unconformity. The latter are interpreted tentatively as being younger pre-Cambrian rocks and comprise conglomerate, albite schists, marbles, lime-silicate granulites and highly aluminous schists.

Early pre-Cambrian structures trend in an east-northeasterly direction; later pre-Cambrian structures trend north-northeasterly, essentially parallel to Paleozoic structures. Structures of the anticlinorium are related to compressional forces acting in an essentially horizontal direction.

Structures in the eastern third of the area are of two highly contrasting types, simple anticlinal and synclinal folds genetically related to the formation of the anticlinorium; and dome structures, related genetically to vertical forces.

Two major domes are present in the area, the Sadawga Pond and the Lake Rayponda domes which are 4 to 8 miles long. These have minor structural features which are the reverse of those in the simple anticlinal structures. The minor structures of anticlines indicate that the upper beds have moved toward the crest of the anticline relative to the beds below. Those of the domes indicate that the upper beds have cascaded toward the axis of the adjacent syncline relative to the lower beds. That these beds actually have reverse structures and are not merely the recumbent limb of an older nappe is suggested by the fact that these reverse structures are distributed symmetrically on all sides of the domes.

The rocks of the area are in the garnet zone of metamorphism with the exception of a locality in southwestern Whitingham and the extensive area of the Green Mountain core, which is in the diopside zone and is equivalent to the kyanite-staurolite zone. Also rocks of the chlorite and biotite zone are present in a few small, isolated districts. A study of the detailed variations in grade of metamorphism within the garnet zone was made and this zone subdivided into three subzones.

This subdivision was based on the ratio of the Fe: Fe+Mg in chlorite divided by that in biotite of garnetiferous rocks. The Fe: Fe+Mg ratio in turn was derived from the indices of refraction of these minerals. The high garnet subzone occupies much of the eastern half of the area and is associated with the domes. The low subzone occupies a small part of the southcentral part of the area. The zones of the western part of the area were established on the basis of macroscopic identification of minerals.

The East Dover ultramafic of the northeastern part of the area is probably the largest in Vermont. It has a dunite core and an outer rim of talc-carbonate. The large intermediate zone consists of serpentinized olivine in which the ratio of serpentine to olivine increases radially outward from the dunite core.

Major problems connected with the present study are the origin and tectonics of these mantled gneiss domes and the location of the roots of the Taconic Klippe.

Domes of the type found in the Wilmington-Woodford area seem to have resulted from vertical movement of the core gneisses relative to the overlying schists.

## INTRODUCTION

### *Location*

The Wilmington-Woodford area of southern Vermont (Fig. 1) is bounded on the north by latitude 43°00′; on the south by the Vermont-

Massachusetts state line; on the east by 72°45′ west longitude; and on the west by the Glastenbury-Shaftsbury and by the Woodford-Bennington township lines in the northern portion and in the southern portion by west longitude 73°10′. The present survey embraces a detailed study of the Wilmington quadrangle; the Vermont portion of the Heath, Rowe, North Adams and Williamstown, Vermont-Massachusetts quadrangles; and the southeastern part of the Stamford, Vermont, quadrangle. Reconnaissance mapping was done in the remainder of the Stamford, the entire Glastenbury, and in the eastern part of the Bennington, and Pownal quadrangles. In all, the area embraces one 15 minute, two 7½ minute quadrangles and parts of six others which together occupy about 385 square miles.

The area is almost evenly divided between Windham and Bennington Counties. It includes the whole or parts of the townships of Newfane, Wardsboro, Stratton, Dover, Somerset, Glastenbury, Woodford, Searsburg, Wilmington, Marlboro, Halifax, Whitingham, Readsboro, Stamford, and Pownal.

*Index to Geologic Mapping in Southern Vermont and Adjacent Areas*

1. Wilmington Area.
2. Castleton quadrangle. Phillip Fowler (1950).
3. Rutland quadrangle. William Brace (1953).
4. Woodstock quadrangle. Ping Hsi Chang (1950) Ph.D. thesis, Harvard University; Geologic Map incorporated in Geol. Surv. Bull. 6 (1953).
5. Hanover quadrangle. John B. Lyons, Dartmouth College. (1955).
6. Pawlet quadrangle. Eastern third by James B. Thompson, Jr. Work in progress. Western two-thirds by Robert Shumaker (1959).
7. Wallingford quadrangle. James B. Thompson, Jr. Work in progress.
8. Ludlow quadrangle. James B. Thompson, Jr. (1950) Ph.D. thesis, Massachusetts Institute of Technology.
9. Claremont quadrangle. C. A. Chapman (1942), and James B. Thompson, Jr. (1954, 1956).
10. Equinox quadrangle. Philip Hewitt. Work in progress.
11. Londonderry quadrangle. John Rosenfeld and Don Wilhelm. Work in progress.
12. Saxtons River quadrangle. John Rosenfeld (1954) Ph.D. thesis, Harvard University.
13. Bellows Falls quadrangle. F. Kruger (1946). Revisions by John Rosenfeld and Tom N. Clifford in progress.

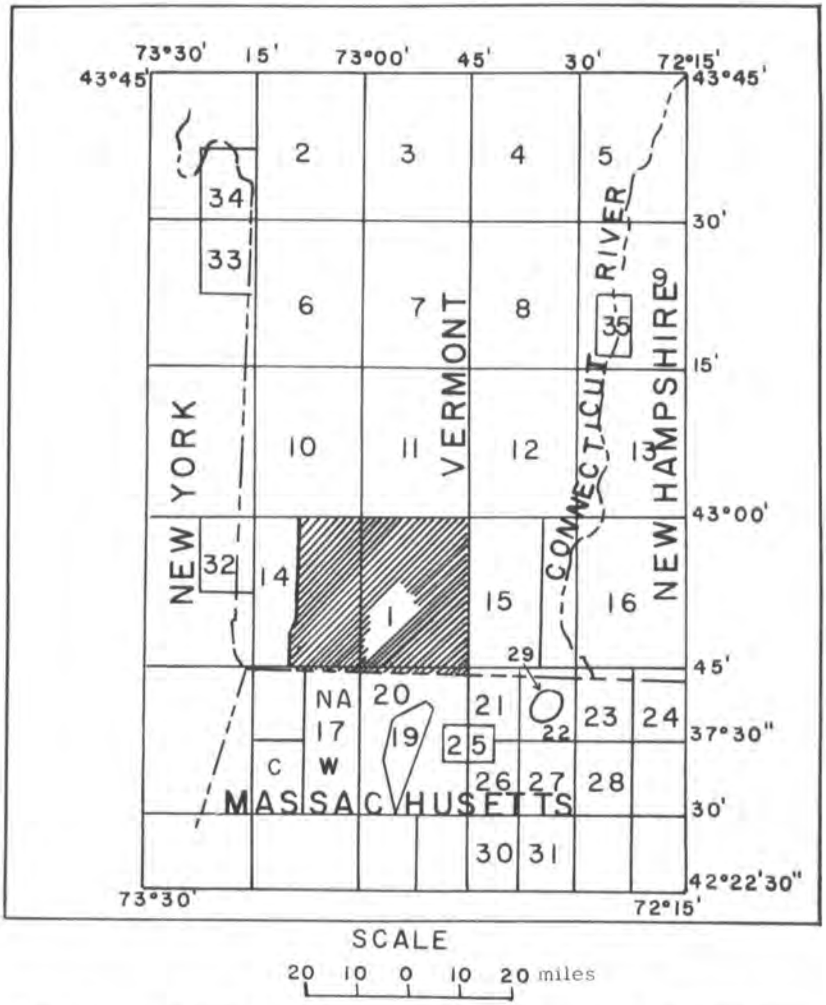



Figure 1. Index to Geologic Mapping in Southern Vermont and adjacent areas.

14. Bennington quadrangle. John A. MacFadyen, Jr. (1956).
15. Brattleboro quadrangle. Gordon J. F. MacDonald. Work in progress.
16. Keene-Brattleboro Area. George E. Moore (1950).
17. North Adams, Windsor and Cheshire quadrangles. Norman Herz. (1961; Windsor in progress; (1959).
18. Taconic quadrangle. Prindle and A. Knopf (1932).

19. Plainfield-Hawley Area. Alonzo Quinn and Donald C. Henderson. Open-file report, Boston Office, U.S. Geological Survey (1945).
20. Hawley quadrangle. Emerson, B. K. (1898a, 1898b, and 1917).
21. Colrain quadrangle. Kenneth Segerstrom (1956).
22. Bernardston quadrangle. Robert Balk (1956).
23. Northfield quadrangle. Robert Balk (1956).
24. Mount Grace quadrangle. Jarvis B. Hadley (1949).
25. Shelburne Falls Area. Robert Balk (1946).
26. Shelburne Falls quadrangle. Kenneth Segerstrom (1956).
27. Greenfield quadrangle. Max E. Willard (1952).
28. Millers Falls quadrangle. Robert Balk (1956).
29. Bernardston locality. Arthur J. Boucot, Gordon J. F. MacDonald, Charles Milton, and James B. Thompson, Jr. (1958).
30. Williamsburg quadrangle. Max E. Willard (1956).
31. Mount Toby quadrangle. Max E. Willard (1951).
32. Hoosick Falls quadrangle. Norman Potter. Work in progress.
33. Granville quadrangle. George Theokritoff (1959).
34. Thorn Hill quadrangle. George Theokritoff (1959).
35. Skitchewaug locality. Arthur J. Boucot, Gordon J. F. MacDonald, Charles Milton and James B. Thompson, Jr. (1958).

## *Regional Geologic Setting*

The Wilmington area is on the east flank of the Green Mountain anticlinorium (Fig. 2). The Green Mountains represent the eroded remnant of this fold which is overturned and locally overthrust toward the west. The anticlinorium consists of a central area of pre-Cambrian gneisses, schists, quartzites, and lime-silicate granulites mantled by a Lower Paleozoic sequence of dominantly argillaceous, arenaceous, and volcanic rocks.

The eastern flank of the Green Mountains and its southward continuation, Hoosac Mountain, is composed essentially of an easterly-dipping homoclinal sequence of regionally metamorphosed stratified rocks complicated locally by domes. The sequence of the western flank of the Green Mountain anticlinorium is composed of steeply dipping or overturned and thrust-faulted Cambrian and Ordovician beds. The lower part of the western sequence has been traced around the southern end of the anticlinorium just south of the place where the Vermont-Massachusetts line cuts across Clarksburg Mountain. From this locality, the writer has mapped this part of the sequence in a northeasterly direction to the western border of the Wilmington quadrangle. In general,

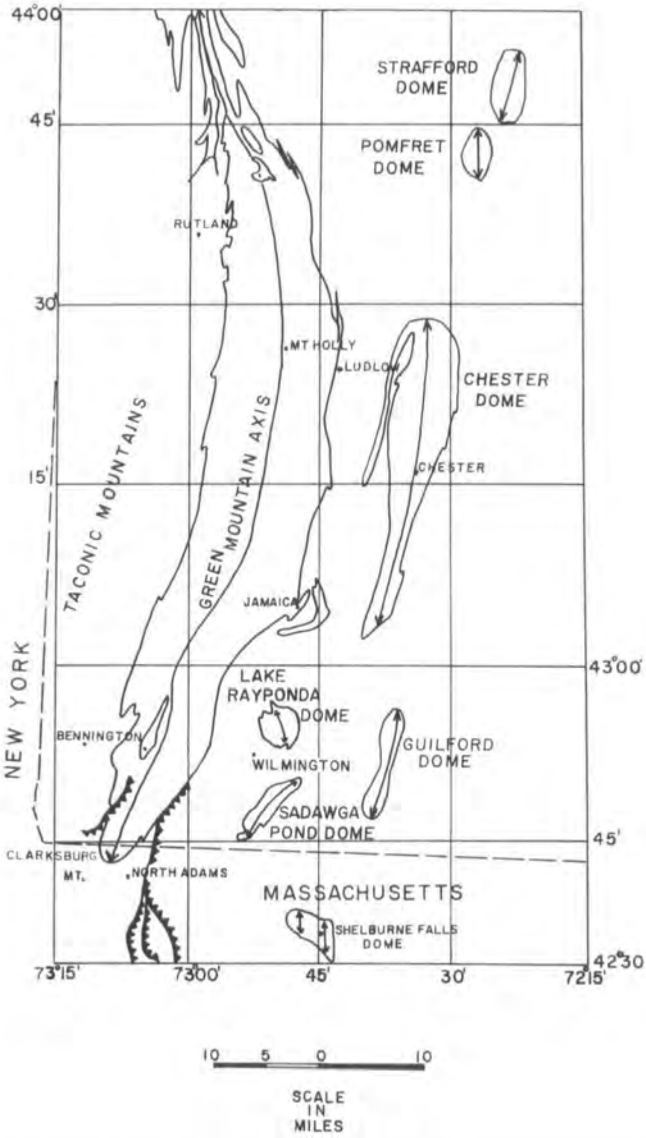



Figure 2. Map showing Major Structures of Southern Vermont and their relationship to the Wilmington-Woodford area.

the rocks west of the Green Mountains are less metamorphosed and more fossiliferous than those of the eastern sequence.

To the south and east of Clarksburg Mountain (Fig. 2) lies an area of pre-Cambrian, Cambrian, and Ordovician rocks, whose relationship to the homoclinal sequence of Hoosac Mountain seems to be such that the eastern sequence is here thrust over the western (Prindle and Knopf, 1932; Norman Herz, 1961).

Norman Herz's mapping shows that the major thrust fault is traced northward along the west flank of Hoosac Mountain in the North Adams quadrangle. The writer's mapping in the Wilmington-Woodford area confirms the existence of this Hoosac thrust (Prindle and Knopf, 1932) along the west flank of Hoosac Mountain in the Stamford Valley. This major thrust fault can be inferred with some degree of probability as far as Heartwellville. It is equally probable, however, that the line marked by the inferred Hoosac Thrust may be merely the western boundary of the more argillaceous facies consisting dominantly of albite schist and alternating green and black highly aluminous schist beds. The topographic low between Hoosac Mountain on the east and Clarksburg Mountain on the west may be the result of differential erosion along the line of facies change where a large-scale syncline has been developed.

## *Previous Geologic Work*

In the course of a reconnaissance survey of Vermont, prior to 1861, E. Hitchcock and fellow geologists made two traverses across the full width of the Wilmington-Woodford area and a third partial traverse. Besides recognizing the anticlinal structure of the Green Mountains, they contributed substantially to an understanding of the local structural geology of the area, and located many of the marble deposits and all of the ultramafic bodies of appreciable size. The geology of the Hawley, Massachusetts-Vermont quadrangle was mapped and described by Emerson (1898a, 1898b, 1917). The boundary interpreted as the Algonkian-Cambrian unconformity was described by Dale (1916), who also located and described many of the marble deposits of the area during a survey of the marble belt of eastern Vermont (1915). About that time also, students of Oberlin College (Hubbard, 1924) studied the southern third of the Wilmington area. The East Dover serpentine received considerable attention from G. W. Bain (1936) during his study of the ultramafic rocks of Vermont.

Gordon (1915) in the first of a series of papers discusses the geology of the western part of the area. In 1932, Prindle and Knopf published their well-known paper on the Taconic quadrangle in which they maintained that the argillaceous sequence of Hoosac Mountain was part of a thrust sheet rather than part of the normal stratigraphic sequence as it had been previously considered.

Pumpelly, Wolff, and Dale (1894) provided valuable information on the structures and stratigraphy of the Green Mountain anticlinorium and the relation of the arenaceous and carbonate sequence of its eastern side to the dominantly argillaceous succession of Hoosac Mountain complicated by anticlinal folds with known pre-Cambrian cores.

## The Problem

The writer was introduced to the complex problems of the regional geology of New England by Marland P. Billings and, more specifically, to those of the Green Mountains by John L. Rosenfeld and James B. Thompson, Jr. The field investigation of the Wilmington-Woodford area proposed to aid in the co-operative effort of unravelling the complex history of the Green Mountains in particular, and of the geology of New England in general.

At the time that the study of the Wilmington-Woodford area was begun in 1949, several problems of regional significance were considered and the Wilmington area was first chosen for special study. At that time, James B. Thompson, Jr. and John L. Rosenfeld were mapping the eastern flank of the Green Mountains in the Ludlow and Saxtons River quadrangles. They were continuing in southern Vermont the work of tracing the stratigraphic units southward from parts of northern and central Vermont to which they had been traced from Quebec. Geologic mapping in western Massachusetts had been completed by Emerson (1898a, 1898b, and 1917), at least on a reconnaissance basis. Therefore, the only gap remaining before the Vermont and Massachusetts sequences could be correlated was the Wilmington-Woodford area and the eastern and southeastern portion of the Londonderry quadrangle.

The Wilmington-Woodford area is strategically located since in a part of it is located the area where the dominantly arenaceous and carbonate sequence of western Vermont and western Massachusetts comes into contact with the dominantly argillaceous and arenaceous rocks of the eastern Vermont sequence. Various explanations of the relationship between these two sequences have been offered.

One of the earliest theories held that the eastern sequence of the Green Mountains should be correlated with the western sequence, but represents different facies deposited more or less contemporaneously.

A second theory held that the eastern sequence had been brought into contact with the western sequence by a major thrust fault (Prindle and Knopf, 1932). A third theory was that the lower arenaceous and carbonate part of the western sequence is essentially unrepresented on the eastern flank of the Green Mountains. This may be because of non-deposition here; or because of erosion; or because the arenaceous and argillaceous sequence overlapped them subsequent to their deposition; or it may possibly be the result of a combination of these reasons.

The writer's opinion is that either the first or second theory is acceptable in the light of known data. As this study has progressed the writer is more inclined to consider the lithologic differences between the eastern and western sequences as due to facies changes rather than to large-scale faulting. Within the Cheshire quartzite there are recognized facies changes and the same is true within the various units of albite and aluminous schist.

A further interesting problem was suggested by the fact that Hitchcock (1861) had described a major anticlinal structure occupying the southern part of the Wilmington quadrangle. The existence of this structure was indicated by scattered observations of the strike and dip along Hitchcock's line of traverse. It was also indicated by the map-pattern of the Rowe formation and Chester amphibolite in the northern part of the Hawley quadrangle (Emerson 1898a, 1898b, 1917), as well as reconnaissance traverses by Eleanora B. Knopf (written communication, 1948).

The present writer was interested to discover whether this were a domical structure such as the Chester Dome (Fig. 2) to the northeast (Thompson, J. B. and Rosenfeld, J. L., 1951) and the Shelburne Falls Dome (Balk 1946) to the southeast. Interest in this anticline was further intensified when James B. Thompson, Jr., at the time engaged in a study of the Chester Dome, brought to the writer's attention the importance of domes in New England geology.

A series of domes 15–25 miles wide extends for about 280 miles from the vicinity of Berlin, New Hampshire to the north shore of Long Island Sound. A second series of domes about the same width lies west of the Connecticut River and extends from east-central Vermont (Fig. 2) southward to Long Island Sound. The majority of those east of the Connecticut River have gneiss cores; those west of the River have either gneiss, schist, or limestone cores. The domes of New Hampshire and Vermont have thus far received greater attention than those farther south. Those in western New Hampshire lie along the crest of the

Bronson Hill anticline (Billings, 1937) and have gneiss cores. The domes of eastern Vermont are of two types, those having gneiss cores and others with schist or limestone cores.

It was thought that a further understanding of these domes would be of great help in unravelling the complex geology of the region. Moreover, recent studies of domes in Finland, Africa, and in the Appalachians indicate that the interest of structural geologists in the problems of domes is increasing the world over (Eskola, 1949; Hall and Molengraaf, 1925; Broedel, 1937; Cloos, 1937; Hadley, 1949; Thompson and Rosenfeld, op. cit.).

### *Present Investigation*

The writer spent about fourteen months in the field during the summers of 1949, 1950, 1951, 1958, 1959, and the fall of 1950. The most recent topographic maps of the area are the Wilmington quadrangle (1954 edition), on scale of 1:62,500; the Vermont portion of the Rowe, Heath, North Adams, and Williamstown, Massachusetts-Vermont quadrangles, all on a scale of 1:24,000; and the Stamford, Pownal, Glastenbury, and Bennington quadrangles on scale of 1:24,000 also.

The only suitable topographic maps available during the 1949–1951 field seasons were the Vermont portions of the Rowe, Heath, North Adams, and Williamstown quadrangles on scale of 1:31,680. During these early years of the study, aerial photographs supplemented, and previous to the publication of the Wilmington quadrangle in 1954 and of the other quadrangles on scale 1:24,000 later, largely supplanted the use of the Wilmington and Bennington quadrangle maps of the 1899 and 1898 editions, respectively.

Concurrent geologic mapping was carried on in the Saxtons River quadrangle to the northeast (Fig. 1) by John L. Rosenfeld who has extended his field work to portions of the adjoining Londonderry quadrangle. The Brattleboro quadrangle to the east has been partially mapped by Gordon J. F. MacDonald. The North Adams, Windsor, and Cheshire, Massachusetts quadrangles to the south of the Wilmington-Woodford area have been studied by Norman A. Herz of the United States Geological Survey. Prindle and Knopf (1932) mapped the North Adams, Massachusetts and the Bennington, Massachusetts-Vermont quadrangles as part of the larger Taconic quadrangle. John A. MacFadyen (1956) of Williams College has mapped in detail the Bennington and Pownal quadrangles and portions of the Stamford and Woodford quadrangles at least on a reconnaissance basis. The Hawley quadrangle to the south has been mapped by B. K. Emerson (op. cit.). The location of these and other nearby areas in which fieldwork is either in progress or completed is shown on the accompanying Index to Geologic Mapping in southern Vermont and northwestern Massachusetts (Fig. 1).


## *Acknowledgments*

The writer wishes to express his gratitude to His Excellency, the Most Reverend John J. McEleney, S.J., Rev. W. Edmund FitzGerald, S.J., and Very Reverend James E. Coleran, S.J., former and present Provincials of the Society of Jesus of New England, without whose continued encouragement and co-operation the field and laboratory investigations would have been impossible. He is most deeply indebted to Professors Marland P. Billings and James B. Thompson, Jr. of Harvard University for their continued interest and invaluable advice, both in the field and in the writing of the report.

The writer is greatly indebted to Misses Mary A. Quinlan, Mary F. Chevalier; Ellen M. Lynch, Lydia M. McCarthy; to Francis J. Accetta and John B. Kelley and to many others too numerous to mention for assistance in the preparation of the manuscript.

He thanks all those who assisted him so capably and enthusiastically during each of the field seasons. Specifically, Professor Arden L. Albee for assistance in the summer of 1949; Harry B. Groom and Edgar L. Weinberg in the summer of 1950; Dr. James Palacas, in the summer of 1951; William Luddy and Richard Sheehan during the summers of 1958 and 1959, respectively; and Michael Kelley, during the summers of 1958 and 1959. The writer wishes to thank also Professors John L. Rosenfeld, William Brace, Gordon J. F. MacDonald, Charles G. Doll, Philip H. Osberg; Drs. Wallace A. Cady, Lincoln R. Page, A. H. Chidester, E-an Zen, and other fellow students of Green Mountain geology for stimulating field and written consultations bearing on mutual problems.

To the Department of Mineralogy and Petrography of Harvard University, the writer expresses his deep gratitude for generous grants from the Holden fund for chemical analysis of minerals and for thin sections of rocks; to the Department of Geology and Geography for grants from the Wolff and Shaler funds to defray the expenses of his fellow students who assisted him in the field during the first three field seasons; and to the Vermont Geological Survey for financing the work of the 1958 and 1959 field seasons.


## *Topography*

The total topographic relief of the Wilmington-Woodford area is 2863 feet. The local relief averages about 500 to 1500 feet. The highest mountains and hills as well as those with the steepest slopes are in the western two thirds of the area. The eastern quarter consists of relatively steep slopes and rugged hills. The section east of Haystack-Mt. Pisgah Ridge and west of the eastern quarter has a mean elevation of about 1900 feet and consists dominantly of gently rounded hills. The highest points in the area and the most prominent topographic eminences of the district are Glastenbury Mountain, 3748 feet above sea level; Mt. Pisgah (now called Mt. Snow), 3556; Haystack Mountain, 3420; Hogback Mountain, 2410; Rice Hill (Dover), 2935; Searsburg Mountain (northeast of Heartwellville), 3119; Houghton Mountain (Stamford), 3095; and Hoosac Peak, 3005 feet.

The lowest points in the area are in Halifax, Newfane, and Stamford townships. The lowest point is in the southeast corner of the area along the East Branch of the North River in Halifax, with an elevation of approximately 885 feet. The next lowest is along the eastern boundary of the area on Rock River in Newfane with an elevation of about 895 feet. The lowest point in the western portion of the area is on the Vermont-Massachusetts boundary along the North Branch of the Hoosic River in Stamford at an elevation of approximately 1069 feet.

### *Rock Exposure*

In general, the rocks of the area are everywhere moderately well exposed with a few notable exceptions. They crop out most abundantly and consistently in the eastern quarter of the area, and in the Mt. Snow (Mt. Pisgah)—Haystack Mountain—Searsburg—Dutch Hill Ridges in the central part of the area. In the western half of the area outcrops are relatively abundant but not consistently well exposed. This is especially true in key localities such as in the Stamford-Heartwellville-Searsburg Valley, and the east side of the valley extending from Searsburg Reservoir to Somerset Reservoir. The older pre-Cambrian rocks of Woodford, Glastenbury, western Searsburg, and Somerset are well exposed on the higher parts of the hills but are poorly exposed in the lowlands and on the lower slopes.

### *Culture and Accessibility*

The area is crossed in an east-west direction by Route 9 and from north to south by Routes 8 and 8A, all of which are paved. The eastern third of the area with the exception of central Dover township is traversed by a close network of dirt roads and consequently all parts are readily accessible. The Mt. Snow—Dutch Hill Ridge is crossed by only

Figure 3. Haystack Mountain as seen from South Road, north of Bond Brook. Searsburg unconformity crops out in the intervening valley.

two roads (Fig. 3). The topography of this ridge is extremely rugged consisting of steep rock cliffs at the higher elevations, treacherous brush-covered talus piles and blocks and is overgrown by a dense cover of underbrush which renders mapping in this district most difficult and extremely laborious. The eastern part of the township of Glastenbury is traversed by two logging roads with much of it accessible only on foot. The greater part of Woodford, Stamford, and large parts of Somerset and Searsburg are accessible only on foot. (Fig. 4A).

Small towns and villages are scattered throughout the area, but the total resident population is probably less than 6000. Glastenbury has no permanent residents, and the total in Somerset is three persons. Wilmington is the largest community with a population of about 1800. Only the relatively level ground of the central third of the area is devoted to farming. This terrain is underlain largely by microcline gneiss and by albite schist with the lowland being developed largely on the former type. The rugged nature of the rest of the area renders farming unprofitable if not impossible. All except the central third is overgrown with woods which slowed down the mapping considerably.

looks more like Harriman reservoir.

Figure 4A. View looking from crest of Mount Snow (Pisgah) toward Stratton Mt.; hill having an elevation of 2682′ in foreground. Location of Searsburg unconformity shown by dashed line. 2862′ — see Geol map!

Figure 4B. View of Haystack Mountain—Mount Snow Ridge in distance and lowlands in vicinity of Harriman Reservoir, looking NNW from a point ½ mile northeast of Harriman Dam.

# STRATIGRAPHY AND LITHOLOGY

## *General Statement*

A general summary of stratigraphic and lithologic relationships is presented in the columnar section (Fig. 5). The pre-Cambrian rocks of the Wilmington-Woodford area, according to the interpretation preferred by the writer, consist of two principal divisions separated from each other by a profound angular unconformity. This remarkably well-developed angularity and the lithologic contrast between these two sequences suggests that the succession of the Green Mountain core (Pl. I & II) belongs to the early pre-Cambrian and that of the Mt. Pisgah (Snow)—Haystack Mountain Ridge is of later pre-Cambrian age.

In addition, there is some evidence for a two-fold division of the older sequence as well. The older rocks, called Mount Holly complex, form the larger part of the central portion of the Green Mountains of the western map area (Pl. I, Fig. 2) in the vicinity of the axis of the anticlinorium. This complex is comprised of two sets of rocks both of which have a strong northeasterly trend. They are, however, in part at least, composed of highly contrasting lithologic types.

Quartz-conglomerates, conglomeratic gneisses with associated quartzites and lime-silicate granulites are present in the eastern part of the district underlain by the Mount Holly complex, whereas reconnaissance studies failed to disclose these closely associated rock suites in the western portion of the area. Because of the prominence of these conglomerates, the writer infers a probable two-fold division of the Mount Holly complex separated by an erosion surface.

The rocks of probable later pre-Cambrian age of the east-central part of the map area consist of conglomerate, albite schist, and gneiss, grading laterally into calcite and dolomite marble and associated lime-silicate granulites. These are overlain by chloritic and graphitic, highly aluminous green and black schists.

This younger pre-Cambrian sequence in Somerset, Searsburg, and northern Stamford has a notably angular relationship to the older sequence but in Readsboro and in much of Stamford these rocks may be thrust over the Cambrian and Ordovician sequence of the east slope of Clarksburg Mountain and the Stamford Valley (Fig. 2 and Pl. I & II).

The Lower Paleozoic sequence occupies much of the eastern third of the map area as a continuous homoclinal sequence. These rocks seem to occur also as outliers in the pre-Cambrian district (Pl. I). These rocks are part of a highly folded and metamorphosed sequence.

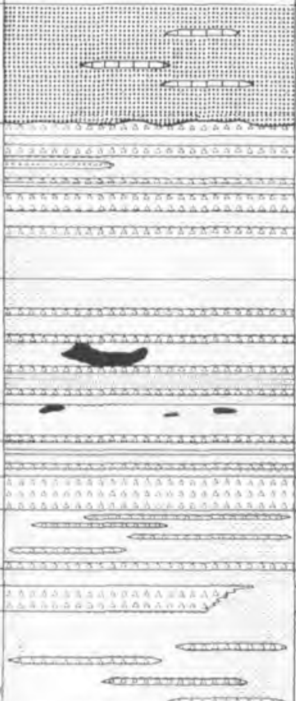
| AGE | SYMBOL | FORMATION | COLUMNAR SECTION | THICKNESS IN FEET | LITHOLOGY |
|---|---|---|---|---|---|
| SILURO-DEVONIAN | DSn | NORTHFIELD SLATE |  | 5000+ 7000 | Dark muscovite-garnet schists and thin limestones. |
| | | UNCONFORMITY | | | |
| ORDOVICIAN | Oc | CRAM HILL FORMATION | | 4000-6000 | Feldspathic-muscovite schists, massive gneiss; amphibolites abundant in all parts of the formation; black, fine-grained muscovite schists. Ocb-Barnard gneiss member, dominantly amphibolites. Ocs-schists with fewer amphibolites |
| ORDOVICIAN | Om | MORETOWN FORMATION | | 500-5000 | Banded amphibolites; light to dark biotite-muscovite-quartz schists; ultramafics, including that of East Dover, are zoned. All have talc-carbonate zone larger ones, a central serpentine zone; East Dover body, a dunite core. |
| ORDOVICIAN | Os<br>Osa | STOWE FORMATION<br>**AMPHIBOLITE MEMBER** | | 150-1000 | Chlorite-muscovite schists with thin, banded amphibolites. Smaller ultramafics. |
| CAMBRIAN and/or ORDOVICIAN | €o | OTTAUQUECHEE FORMATION | | 100-1200 | Dark muscovite-garnet schists with thin, dark quartzites, chlorite-muscovite schists and feldspathic-chlorite schist |
| CAMBRIAN and/or ORDOVICIAN | €c | CHESTER AMPHIBOLITE | | 300-600 | Banded amphibolites with thin schist beds. |
| CAMBRIAN and/or ORDOVICIAN | €ph | PINNEY HOLLOW FORMATION | | 300-2000 | Pale-green chlorite-muscovite schists with thin, banded amphibolites. |
| CAMBRIAN and/or ORDOVICIAN | €htm | HOOSAC FORMATION: TURKEY MOUNTAIN MEMBER | | 0-300 | Thick amygdaloidal amphibolite near the top of the formation. |
| CAMBRIAN and/or ORDOVICIAN | €h | HOOSAC FORMATION: HOOSAC SCHIST | | 700-2000 | Rusty, garnet-chlorite-biotite-albite augen-quartz schist, graphitic albite and non-albitic muscovite schist; thin amygdaloidal amphibolite €hg-green to dark schist. |
| CAMBRIAN and/or ORDOVICIAN | (€c) | HOOSAC FORMATION: (CHESHIRE QUARTZITE) | | (0-800) | (Massive, glossy white and buff quartzite) |
| CAMBRIAN and/or ORDOVICIAN | (€m) | HOOSAC FORMATION: (MENDON FORMATION) | | (400) | (Black, fine, banded chloritoid phyllite, dark quartzite) |
| CAMBRIAN and/or ORDOVICIAN | €t | TYSON FORMATION | | 0-200 | Quartzites, muscovite schists and conglomerate |
| | | UNCONFORMITY | | | |
| LATE PRE-CAMBRIAN | rp€h | HEARTWELLVILLE SCHIST | | 0-700 | Dark graphite-garnet-muscovite-quartz veined schist, light to dark, highly aluminous green, garnet-chlorite-chloritoid-quartz schists |
| LATE PRE-CAMBRIAN | rp€rs | READSBORO FORMATION: SHERMAN MARBLE MEMBER | | 0-200 | Coarse-grained calcite marbles, fine-grained dolomite marbles; actinolite and diopside granulites. |
| LATE PRE-CAMBRIAN | rp€r | READSBORO FORMATION: READSBORO SCHIST | | 700-3500 | Rusty to gray garnet-chlorite-biotite-albite augen-quartz schist and gneiss, graphitic near base, few amphibolites (p€ra). |
| LATE PRE-CAMBRIAN | rp€rsc | READSBORO FORMATION: SEARSBURG CONGLOMERATE | | 0-175 | Thin white to buff glossy quartzite, blue and white quartz conglomerate, albite and microcline conglomerate |
| | | UNCONFORMITY | | | |
| EARLY PRE-CAMBRIAN | ep€g | MOUNT HOLLY COMPLEX: PLAGIOCLASE GNEISS | | UNKNOWN | Blue and white quartz-plagioclase gneiss of Searsburg and Somerset, more closely associated with ep€l, ep€q, than ep€pg |
| EARLY PRE-CAMBRIAN | ep€l | MOUNT HOLLY COMPLEX: LIME SILICATES & WHITE GNEISS | | 0-150 | Fine to coarse lime-silicate, blue and white quartz-rich white gneiss |
| EARLY PRE-CAMBRIAN | ep€q | MOUNT HOLLY COMPLEX: QUARTZITE, CONGLOMERATE & CONGLOMERATIC GNEISS | | 0-500 | Massive buff to blue glossy quartzite, blue quartz conglomerate, "pebbly" rounded to angular microcline conglomeratic gneiss. |
| EARLY PRE-CAMBRIAN | rep€wg<br>ep€sgg | MOUNT HOLLY COMPLEX: WILMINGTON GNEISS<br>STAMFORD GRANITE GNEISS | | 0-6000 | Coarse gray, buff and pink microcline-augen and porphyritic-microcline in quartz-albite-microcline-biotite-epidote groundmass |
| | | UNCONFORMITY(?) | | | |
| EARLY PRE-CAMBRIAN | ep€pg | MOUNT HOLLY COMPLEX: PLAGIOCLASE GNEISS (HARMON HILL GNEISS) | | UNKNOWN | Blue and white quartz plagioclase gneiss of Green Mountain core |
| EARLY PRE-CAMBRIAN | ep€mg | MOUNT HOLLY COMPLEX: MICROCLINE GNEISS | | UNKNOWN | Banded microcline augen gneiss; porphyritic microcline gneiss, few amphibolites. |

Figure 5. Columnar section for the Wilmington-Woodford, Vermont, area shows sequence and character of sedimentary and volcanic rocks.

Symbols, names, and descriptions in parenthesis are those of western Vermont sequence.

The rocks of the area may be classified according to inferred origin as sedimentary, pyroclastic, extrusive, and intrusive rocks. Those of sedimentary and pyroclastic origin are far more abundant than either extrusive or concordant intrusive bodies. These concordant masses in turn are more numerous than ultramafic masses or mafic and felsic dikes.

The original sedimentary rocks were graywacke, arkose, shale, sandstone, carbonate and reworked volcanic rocks. Extrusive and concordant intrusives range from mafic to felsic tuffs and include also amygdaloidal basalt, mafic sill-like bodies and dunite masses. Discordant intrusive rocks include dunite as well as mafic and felsic dikes.

## *Stratigraphic Nomenclature*

In this paper, a number of stratigraphic terms which are commonly somewhat loosely defined, are used. The usage followed by the writer is that of the "Committee on Stratigraphic Nomenclature" (Ashley, 1933, pp. 423–431) which defines the formation as the fundamental unit in the local classification of rocks. Insofar as possible, formations are distinguished on the basis of lithologic unity and continuity of areal extent. These should be differentiated so that they may be readily traced in the field and represented on geologic maps. As defined by the "Committee on Stratigraphic Nomenclature" (Moore, 1947, pp. 513–518) the formation has no time connotation. In other words, the formation boundaries are not necessarily identical with hypothetical planes representing equivalent time.

In this paper, the definitions of formations and members are based on 1) practicability of field mapping, 2) geographic extent, and 3) near identity of chemical, if not mineralogical composition.

Formations are differentiated from members on the basis of practicability of mapping and geographic distribution. Rock units of at least 150 feet thickness can be mapped on scale of 1:62,500 no matter what the attitude of the beds. Formations should have a great geographic extent and be traceable under most circumstances for at least 50 miles. Members have a more limited distribution. Consequently, in this paper a formation is a rock unit at least 150 feet thick and which can be mapped over a geographic area of at least 50 miles. A member is a rock unit forming a distinctive part of a formation. Generally, it is less than 150 feet thick and may or may not have a geographic extent of 50 miles; or it may be over 150 feet thick but unable to be mapped over an extent of 50 miles.

Names of formations and members have been employed according to the rules set forth by the "Committee on Stratigraphic Nomenclature" (Ashley, 1933, pp. 433–434). The names are binomial, consisting of a geographic name followed by a descriptive lithologic name, but if no one lithologic term is suitable, the term formation is used.

### Lithologic Nomenclature

In the nomenclature of metamorphic rocks, many terms are loosely defined and, therefore, ambiguous. The classification chosen by the writer is based on texture and mineral content and is therefore purely descriptive.

Texture refers to grain size, relative shapes of grains, and relative orientation of adjacent grains. In cases in which individual grains are notably larger than surrounding minerals, the large minerals are called "porphyroblasts", and smaller minerals are termed "groundmass". Mineral particles having a common size are called "grains". Minerals described by relative shapes are equidimensional and non-equidimensional minerals. The non-equidimensional minerals are "platy" or "needle-shaped".

The following textural names are used in this paper following the usage of Osberg (1952, pp. 18–19):

| | |
|---|---|
| Schist | a recrystallized rock consisting of minerals that can be resolved megascopically. At least 30 per cent of the non-equidimensional minerals have a common orientation. |
| Gneiss | a recrystallized rock consisting of minerals that are large enough to be recognized megascopically. Between 30 and 10 per cent of the nonequidimensional minerals have a common orientation. |
| Granulite | a recrystallized rock in which the minerals are recognized megascopically. Less than 10 per cent of the nonequidimensional minerals have a common orientation. |
| Marble | a granulite consisting dominantly of calcite or dolomite. |
| Quartzite | A granulite consisting dominantly of quartz. |
| Amphibolite | a gneiss or a schist consisting essentially of amphibole. |

The full rock name consists of a textural name preceded by a mineral name comprising the names of the most important minerals of the rock linked by hyphens. If the rock has the texture of a gneiss and if it is comprised of quartz, microcline, biotite, and epidote, the full rock name is quartz-microcline-biotite-epidote gneiss.

## *Pre-Cambrian Rocks*

### General Statement

Detailed geologic mapping has been carried out during the present investigation in the eastern two-thirds of the area. The western third has been mapped in reconnaissance fashion. On the basis of these two types of information, the writer concludes that the greater part of the southern Green Mountain core consists of a complex of plagioclase and microcline gneisses with only very small quantities of other lithologic types. The relative ages of the two older rock units (Fig. 5) is not known with certainty, but on the basis of available structural data (Pl. I) plagioclase gneisses seem to be younger than the microcline gneisses of Glastenbury Mountain and of the area around Houghton Mountain.

The eastern third of the central Green Mountain core contains a wide variety of distinctive and consistently associated lithologic types not seen elsewhere in the area. This sequence consists of "pebbly" arkosic microcline gneiss; massive buff, white and blue glassy quartzite with associated blue quartz conglomerates; and very coarse- to fine-grained tremolite-diopside, lime-silicate granulites. All of these rock types are widespread in Searsburg and Somerset townships as well as in one locality in Glastenbury and one in Woodford.

The Stamford granite gneiss of the southern end of the Green Mountain anticlinorium is a highly distinctive porphyritic gneiss and in many ways resembles the "pebbly" microcline gneiss of the Mount Holly complex.

Overlying this sequence of the Green Mountain core with profound angular unconformity is an intermediate series of arenaceous, argillaceous, and gneissic rocks with a small amount of dolomite and calcite marble and associated lime-silicate granulite. This sequence of Haystack Mountain—Dutch Hill Ridge, interpreted by the writer as being probably of later pre-Cambrian age, is separated from the overlying homoclinal sequence of the eastern part of the area by a slight unconformity which is regularly marked by conglomerate lenses.

This intermediate succession contains widespread rock types which are lithologically identical with several overlying arenaceous and argillaceous formations of Paleozoic age. These rocks are distinctive, however, in that they contain very few amphibolites in contrast to the Paleozoic sequence which abounds in this type rock. The late pre-Cambrian complex contains the Sherman marble, a distinctive dolomite and calcite marble and lime-silicate rock type in a distinctive stratigraphic succession not found in the homoclinal sequence. This difference in lithology, however, may be due to facies changes.

The fact that both the later pre-Cambrian and the lower part of the Paleozoic sequence contain widespread albite schist and gneiss beds as well as chlorite-muscovite and graphitic muscovite schists has presented a problem in differentiating them and consequently in interpreting the stratigraphy and structure. The relation of the Wilmington gneiss to these two sequences has added complexity to the above problems.

## Mount Holly Complex

*Name:* The term Mount Holly series was proposed by Whittle (1894) for the heterogeneous group of rocks exposed in the central core of the Green Mountain anticlinorium in the township of Mount Holly (Fig. 2). The term Mount Holly series has been emended by Osberg (1952, p. 21) to Mount Holly complex. This usage is adopted here since the term complex, as defined by the "Committee on Stratigraphic Nomenclature" (Ashley, 1933, p. 445) prescribes the following usage for the term complex: "where a large mass is composed of diverse rocks of any class, or classes, and is characterized by highly complicated structure, the term complex may be used". Moreover, the term complex does not have time connotations whereas the term series, in present stratigraphic usage, is restricted to rocks deposited during an epoch. The metamorphic rocks of the core of the Green Mountain anticlinorium, however, are not readily subdivided at present according to time-rock units representing such relatively short intervals.

The Mount Holly complex of the type locality in the central part of the southern Green Mountains consists of fine-grained biotite gneiss, microcline augen gneiss, garnet-quartz-muscovite schist, rusty muscovite schist, amphibolite, quartzite, and saccharoidal marble. In the Wilmington-Woodford area, the Mount Holly complex includes units lithologically similar to those of the Mount Holly series as defined by Whittle as well as associated units not exposed at the type locality.

*Distribution:* The Mount Holly complex occupies the central portion of the Green Mountain anticlinorium (Fig. 2 and Pl. I) from the base of the Mendon formation on the west flank of the anticlinorium to the prominent but complex boundary marked by the Hoosac Thrust, and in Searsburg and Somerset by the angular unconformity. Because several unconformities will be mentioned in the course of the paper, this angular unconformity will be referred to as Searsburg unconformity near the center of which village the angular relationships are excellently displayed.

Mount Holly microcline gneiss is well exposed in the vicinity of the Long Trail on the northern extension of Glastenbury Mountain on the line of the most direct route from an existing logging road to the Long Trail on the Woodford-Equinox quadrangle boundary. Here the gneiss consists of well-banded, biotite, blue-quartz, pink microcline augen gneiss. Some are thin-bedded and others thick-bedded. Blue quartz in some beds amounts to about 20 percent; biotite-rich beds are dark in color and this mineral makes up as much as 20 percent of some beds. The maximum dimension of the microcline augen observed at this locality is 1 cm. Except for the strikingly blue color of the quartz, this gneiss was noted as being identical in all other essential features with the Wilmington gneiss.

Besides the microcline gneiss in this locality, thin beds of coarse amphibolite-granulite (of gabbroic aspect) and feldspathic quartzite granulite are present.

Burt (1929, p. 68–70) distinguishes six types of gneiss of the pre-Cambrian core of the Green Mountains to which the kaolin deposits of Bennington are genetically related. Five of these he groups under the term Woodford gneiss. In the light of information gathered in the present survey, the term Woodford gneiss as defined does not seem helpful in establishing a satisfactory subdivision of the pre-Cambrian rocks such that they can be mapped as distinctive units over wide areas of the Green Mountain core.

The term Woodford gneiss, as originally proposed was not strictly intended as a stratigraphic term, but included the white gneiss, thought to be the correlative of the white gneiss of Berkshire County described by Pumpelly, Wolff, and Dale (1894, p. 80–86), as well as four other types of gneiss. In other words, Woodford gneiss seems to be more or less equivalent to the Mount Holly complex of this area. The writer prefers the latter term. The writer has mapped several lithologic types in the Mount Holly complex, but refrained from introducing new formational names at the present. More detailed mapping is in order before stratigraphic names may be appropriately given to these units.

Mount Holly microcline gneiss is well developed in the vicinity of Little Pond, in the northern part of Woodford township and 2.4 miles north of Big Pond along Route 9. Access is by private road. Microcline augen-gneiss, lithologically identical with that on Glastenbury Mountain, is excellently exposed here.

Besides being a prominent part of the sequence in the river bed at the outlet of Somerset Reservoir, this microcline gneiss is also found east of

Figure 6. Highly contorted plagioclase gneisses (ep€pg, Harmon Hill gneiss) at Dunville Hollow roadcut on Route 9, Molly Stark Trail, Woodford. Note highly deformed character as shown by folded light bands and shear zones.

the reservoir (Fig. 4A). This gneiss is also easily accessible and well exposed in the lower reaches of the northern branch of Clement Brook. The brook flows east to Searsburg Center, and is located one mile south of Route 9 (Pl. I).

This microcline augen gneiss has the same lithologic and textural features as the Stamford gneiss of the southern part of the central Green Mountain area. More specifically, this similarity consists in this that the rock is made up of coarse-grained microcline-augen, blue quartz, dark-green biotite and epidote.

Blue-quartz-microcline-augen gneiss is distinctive. It is a banded rock and contains lenticular microcline augen, ¼ to 8 inches long, lying in the plane of the foliation. Biotite is a minor but conspicuous mineral and forms the irregular laminae which enclose the feldspar augen. Blue quartz and epidote are scattered irregularly throughout the rock. The blue quartz, however, commonly occurs in the corners of the microcline augen, whereas epidote is more commonly associated with the biotite laminae.

A prominent rock in the plagioclase gneiss sequence is banded biotite gneiss. It is generally mottled rusty and yellow on weathered surfaces and gray where fresh. It consists of thin laminae of relatively coarse-grained brown biotite bands of medium-grained plagioclase and finer-grained quartz.

In the banded biotite gneiss, the ratio of biotite to muscovite is usually about 2:1. The biotite laths are generally oriented parallel to each other. The fine muscovite grains, on the other hand, are more randomly oriented. In those biotite gneisses, however, in which muscovite is coarser grained than the biotite the parallelism of the muscovite is more pronounced than that of the biotite. The plagioclase of these rocks is generally untwinned oligoclase. Epidote is usually present in minor amounts and is associated with plagioclase and muscovite. The individual granular layers are about 3 to 5 mm. thick. Most specimens contain widely-scattered, irregularly-shaped phenocrysts of buff or pink microcline. These grains average about 1 cm. in long diameter and are generally flattened into the plane of the foliation.

A conspicuous rock type associated with lime-silicate granulites is a white gneiss. It characteristically weathers rusty or yellow but is white on fresh surfaces. The banding is produced by alternation of blue-quartz bands 2 to 4 mm. thick with thin laminae of black biotite. In addition to blue quartz, the granular bands contain twinned albite and microcline, and the micaceous laminae contain muscovite as well as the more conspicuous biotite.

In thin section, the blue quartz is seen to be filled with minute rods of an unknown mineral. Some workers have attributed the blue color of the quartz to the presence of tiny rutile needles. Early investigators, such as Emerson (1917, p. 23), however, were more inclined to believe it

Figure 7. Plagioclase gneisses in roadcut on Route 9, Molly Stark Trail, Dunville Hollow, Woodford, cut by small-scale fault.

to be the result of strain developed during the deformation of the rock.

The blue-quartz gneiss is generally more highly deformed than the banded biotite gneiss. Crystals of graphite as coarse as 10 mm. in diameter are scattered throughout the gneiss. Thin, much contorted black

TABLE 1

MODES OF THE MOUNT HOLLY COMPLEX

| | 1 | 2 | 3 |
|---|---|---|---|
| Porphyroblasts | | | |
| Biotite | 3 | 10 | 5 |
| Oligoclase | 5 | 15 | 50 |
| Muscovite | .. | 5 | ... |
| Microcline | .. | 20 | ... |
| Groundmass or grains | | | |
| Muscovite | 10 | .. | 5 |
| Quartz | 8 | 50 | 31 |
| Zoisite | tr. | tr. | .. |
| Magnetite and Ilmenite | 1 | .. | tr. |
| Tourmaline | .. | .. | tr. |
| Allanite | .. | .. | 4 |
| Grain size in mm. | | | |
| Porphyroblasts | 0.4–2.0 | 0.5–2.0 | 0.5–1.0 |
| Groundmass or grains | 0.008–0.3 | 0.01–0.4 | 0.005–0.4 |
| Texture | G* | Gr** | Gr |

1. 96D/a*** Quartz-muscovite-biotite-feldspar gneiss (at Spillway, Somerset Res-servoir, Somerset).
2. 55D/f Quartz-muscovite-feldspar-biotite gneiss (0.2 mile northeast of junction of powerline and South Road, Searsburg).
3. 53D/c Feldspar-quartz-biotite-muscovite gneiss (.55 mile southeast of junction of Route 9 and South Rd., Searsburg).

*gneissose

**granulose

***specimen numbers refer to rock samples in the writer's collection and described in field notebooks under the appropriate number.

bands of biotite are abundant in this gneiss. The plagioclase grains are filled with finely disseminated sericite and allanite. Randomly scattered grains of allanite as large as 0.3 mm. are common. Modes are presented in Tables 1 and 2.

Harmon Hill gneiss is a term proposed by Gordon (1914, p. 349) for the plagioclase-rich gneiss cropping out along the Molly Stark Trail (Woodford Road). He describes the gneiss as consisting of coarser bands of quartz and feldspar alternating with finer micaceous bands. The quartz shows a pronounced wavy extinction, the plagioclase shows twinning lamellae, and zircon is present as an accessory mineral.

This plagioclase gneiss is well exposed in nearly vertical beds in a

TABLE 2

MODES OF THE MOUNT HOLLY COMPLEX

| | 4 | 5 |
|---|---|---|
| Porphyroblasts | | |
| Biotite | .. | 5 |
| Hornblende | 10 | |
| Diopside | 5 | |
| Microcline | 35 | .. |
| Oligoclase | .. | 60 |
| Graphite | 1 | 1 |
| Pyrrhotite | tr. | .. |
| Muscovite | .. | 2 |
| Groundmass or grains | | |
| Quartz | 49 | 30 |
| Magnetite and Ilmenite | tr. | .. |
| Allanite | .. | 2 |
| Grain size in mm. | | |
| Porphyroblasts | 0.5–2.0 | 0.5–20.0 |
| Groundmass or grains | 0.008–0.4 | 0.01–0.4 |
| Texture | G* | G |

4. 53D/b Quartz-feldspar-hornblende-diopside gneiss (0.5 mile E of junction of Bond Brook and Sleepy Hollow Road, Searsburg).
5. 490B/a Plagioclase-quartz-biotite-muscovite gneiss (.1 mile S 75° W. of junction of East Branch of the Deerfield and the Rake Branch of the Deerfield River, Searsburg.

*gneissose

roadcut excavated during this study on Molly Stark Trail (Route 9) at Dunville Hollow. This cut is just east of the outcrop described by Gordon. The writer proposes that if eventually this term proves to be a useful stratigraphic name, the locality be extended to include the rocks in the cut noted above because of the excellent exposures here (Fig. 6). Plagioclase gneisses seem to be confined to certain restricted areas and the writer would, therefore, group together all of the rocks of the southern Green Mountain core which are not rich in microcline and which are not at the same time associated with the distinctive lime-silicate quartzite, conglomerate, gneiss sequence of Searsburg and Somerset.

The present reconnaissance study indicates two northeasterly-striking zones of such gneisses, the first extending essentially across the full width of the anticlinorium in central Woodford. The second forms much of the lowland area east and northeast of Glastenbury Mountain within the confines of Glastenbury township.

The gneiss of Harmon Hill and Dunville Hollow is for the most part an excellently banded, light- to dark-gray biotite-rich, epidote-bearing plagioclase gneiss. Some thin bands are rich in relatively fine-grained microcline. A thin amphibolite, ranging from 2 to 20 feet thick is present in the sequence at the type locality, as described above. (Modes of the Mount Holly complex are presented in Tables 1 and 2).

### STAMFORD GRANITE GNEISS

*Name:* Hitchcock (1861, p. 561) describes the Stamford granite as a peculiar porphyritic granite in places well foliated and cropping out chiefly in Stamford and Pownal.

*Distribution:* The Stamford granite as it is called by Hitchcock, or the Stamford gneiss as it is called by Pumpelly, Wolff, and Dale (1894, p. 45), occupies the southern part of the Green Mountain anticlinorium in Massachusetts and extends northeasterly to the vicinity of Heartwellville.

Pumpelly, Wolff, and Dale describe Stamford gneiss as forming the core of Hoosac Mountain as well as the southern end of the Green Mountains. In addition, the present survey has shown that much of the microcline gneiss of the Green Mountain core has essentially the same lithologic character as the Stamford gneiss, but contains other types of rock which can be interpreted as of undoubted sedimentary origin, such as quartzites and schist beds. The Stamford gneiss contains a few quartzite beds which may represent chert or sand beds.

Certain rocks containing pebbles of Stamford gneiss and similar rock types crop out in the eastern part of the Green Mountain core, notably in Somerset and Searsburg. One such locality which is readily accessible is east of Castle Brook in Somerset at longitude 73°00′ (Fig. 8).

*Description:* In its most typical development, the Stamford granite gneiss is a coarse-grained porphyritic gneiss, with its most conspicuous constitutent being rectangular or rounded, perthitic microcline crystals. The finer-grained groundmass consists of blue quartz, albite, microcline, biotite, epidote, and magnetite.

In his excellent description of the Stamford gneiss from the type area and from Hoosac Mountain near Hoosac Tunnel, Wolff (Pumpelly, Wolff, and Dale, 1894, p. 45–48) presents those features characteristic of each locality.

In the area mapped as Stamford granite gneiss (ep€sgg, Pl. I) the rock on weathered surfaces displays numerous rounded elliptical masses or buff to white "rosettes" which stand out in strong contrast to the finer-grained, and darker groundmass. These rounded masses have the ap-

Figure 8. Coarse gneiss-pebble and microcline conglomerate at Longitude 73°00′, Castle Brook, Somerset, (ep€q).

pearance of pebbles and are composed of feldspar aggregates and patches of biotite. The feldspar is perthitic microcline with some albite. Wolff describes some of the gneiss west of Stamford Village as consisting of Carlsbad twins an inch or two across and having the appearance of pebbles.

Hitchcock (1861, p. 561) presents analyses of the feldspar of the gneiss and shows the range of the more significant oxides as follows: silica, 64–66.5 percent; alumina, 19–20.4 percent; lime, 0.63–1.8; potash, 9.8–10.5; and soda, 2.3–3.4.

The large microcline crystals are described by Wolff as having undergone great mechanical changes as seen by the fact that the crystals are in many cases faulted and the edges crushed, with small veins of secondary quartz mixed with tiny grains of albite traversing them along the fault lines. The microcline appears cloudy in thin section, in part the result of small epidote crystals, a mineral which is characteristic of the gneiss. In certain places, the feldspar contains garnets, flakes of biotite, muscovite and magnetite.

Characteristic of the older rocks of the Green Mountain core is blue quartz. In the Stamford gneiss the quartz is likewise characteristically blue, but Wolff has observed that when crushed by pressure in the rock it is white or sugary in appearance. In thin section he has observed that the original cores are still blue but are surrounded by masses of broken quartz, clearly derived from the parent blue quartz.

From the present reconnaissance study of the central Green Mountain region, the writer is of the opinion that microcline gneisses of the Mount Holly complex which make up the core of the anticlinorium are essentially of the same lithology as the Stamford gneiss, although in places showing pronounced textural differences from what is considered typical Stamford gneiss. The details of the stratigraphic relations between the Mount Holly microcline and the Stamford gneisses can be resolved only after they are mapped in detail.

*Origin of the Stamford granite gneiss:* From the fact that the feldspars are crushed, and the development of biotite, muscovite and quartz are parallel to planes of movement, the development of parallel structure is related to movements subsequent to the formation of the rock. These features are not incompatible with an igneous origin for the gneiss. The presence of interbedded quartzite and schist beds clearly excludes an intrusive origin. The gneiss may have originated as an arkose or as a rhyolitic tuff or agglomerate. The character of the rock as a whole is uniform. The "rosettes" of microcline filled with epidote and associated with albite, and surrounded by biotite give the rock a peculiar structural development. The chemical, mineral and structural character of this rock suggest that it originated as a scoriaceous rock or pumice of rhyolite composition, the gas holes having been filled with minerals appropriate to impart the appropriate bulk chemical composition.

## Wilmington Gneiss

*Name:* Wilmington gneiss is a term proposed by the writer for the microcline augen gneiss which crops out over a large part of the central lowlands of the eastern half of the Wilmington-Woodford area. This gneiss is widespread in this part of the area extending from the northern boundary almost to the southern. Excellent exposures of this distinctive rock are readily observed in the vicinity of Wilmington Village, in the Deerfield River Valley; and in ledges just west of the junction of Routes 8 and 9 one mile east of the village.

*Distribution:* Wilmington gneiss extends as a band of variable width from a point one-half mile south of the Stratton-Dover line southward to within 0.8 miles of the Vermont-Massachusetts line. It is narrowest in the northern and southern limits where it is inferred to plunge under younger strata. Its broadest expanse is in Wilmington-Marlboro-Readsboro where it crops out with only minor interruptions for about 8 to 9 miles. In northern Wilmington its breadth of outcrop is 4½ miles. Smaller bands of gneiss in all respects similar to the continuous band of Wilmington gneiss are found in several places in the area. On the northern boundary of the area in Stratton is a series of well-exposed outcrops which seem to form the southern limit of a southerly-plunging anticlinal structure which continues into the Londonderry quadrangle to the north.

Wilmington gneiss, in extremely interesting structural development, is well-exposed in the Sadawga Pond dome in Whitingham, Wilmington, Marlboro, and Halifax. Excellent exposures are easily accessible in roadcuts on Route 8, 0.7–1.0 miles northwest of Jacksonville (Figs. 9 and 10). Nearly vertical cliffs of this gneiss are present and easily accessible 0.3 miles northwest of Ryder Pond, east of Route 8, Whitingham, and in southwest Marlboro one mile northeast of Gates Pond (Whitingham).

*Lithology and Thickness:* The Wilmington gneiss consists essentially of medium- to very coarse-grained microcline gneiss. About 90 percent is well-banded, somewhat-foliated, microcline-augen gneiss. About 5 percent consists of schistose gneiss containing a small number of randomly distributed gray and pink microcline grains. The remaining 5 percent of the sequence comprises massive, medium-grained microcline gneiss.

The relationship of the Wilmington gneiss to the other formations is far from certain. The interpretation of the writer at present is that it is of the same age as the Stamford gneiss and the other microcline gneisses of the central Green Mountain core. It is lithologically identical with the Stamford except for the general lack of blue quartz in the former and its general presence in the latter, although it is lacking in some places. This absence of blue quartz in the Wilmington gneiss may be due to being in a different structural setting.

On the other hand it is highly probable, in the writer's opinion, that the Wilmington gneiss may represent a facies of microcline gneiss in the middle of the albite and green and dark schist sequence. The fact that the Wilmington gneiss is involved in extremely intricate structural complexities makes interpretation of its true stratigraphic position highly tentative.

TABLE 3

MODES OF WILMINGTON GNEISS

| | 1 | 2 | 3 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | .. | .. | 35 |
| Biotite | 3 | 10 | 10 |
| Microcline | 45 | 20 | 20 |
| Oligoclase | 35 | 27 | .. |
| Groundmass or grains | | | |
| Quartz | 10 | 35 | 30 |
| Muscovite | 5 | 5 | 1 |
| Magnetite | 1 | 1 | 2 |
| Epidote group | 1 | 2 | 2 |
| Zircon | .. | .. | tr |
| Sphene | .. | tr | tr |
| Tourmaline | .. | .. | tr |
| Pyrite | tr | tr | tr |
| Apatite | tr | tr | .. |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5 | 0.5–5.0 | 0.5–10.0 |
| Groundmass or Grains | 0.08–0.4 | 0.08–0.4 | 0.008–0.4 |
| Texture | G* | G | G |

1. 64D/a Microcline-oligoclase-plagioclase-quartz-biotite gneiss (.6 mile N 60°E of junction of Blue Brook and Handle Road, Wardsboro).
2. 79D/b Feldspar-muscovite-biotite-quartz-graphite gneiss (0.45 mile N 20° E of junction of Route 8 and 9, Wilmington).
3. 77D/e Microcline-quartz-albite-epidote gneiss (.7 mile N 45° E of junction of Binney Brook and Rose Brook, Wilmington).

*gneissose

The microcline of the microcline-augen gneiss is gray to pink and occurs as lenticular augen and flaser whose average long diameter is about 7 mm. Locally they are 8 inches long and are usually flattened into the plane of the foliation. In addition to the larger grains of microcline forming the augen, there are many smaller whose average size is about 0.3 mm. The microcline in thin section of the well-banded type characteristically shows undulatory extinction. Albite of finer grain than the microcline is present and many grains are twinned. Modes are presented in Tables 3, 4, and 5.

Biotite occurs as tiny shreds whose average diameter is about 0.15 mm. Light-green to colorless muscovite is present as a minor or accessory

Table 4

MODES OF WILMINGTON GNEISS

| | 4 | 5 | 6 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | 10 | 10 | 8 |
| Biotite | 10 | 8 | 5 |
| Microcline | 33 | 35 | 25 |
| Oligoclase | .. | tr | .. |
| Groundmass or grains | | | |
| Quartz | 35 | 30 | 45 |
| Muscovite | 5 | 2 | 12 |
| Magnetite | .. | tr | tr |
| Epidote group | 7 | 12 | 4 |
| Zircon | tr | .. | tr |
| Sphene | tr | 3 | 1 |
| Tourmaline | .. | .. | .. |
| Pyrite | .. | tr | tr |
| Apatite | tr | .. | tr |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–60.0 | 0.5–20.0 | 0.5–30.0 |
| Groundmass or Grains | 0.01–0.4 | 0.02–0.4 | 0.008–0.4 |
| Texture | G* | G | G |

4. L24/d Quartz-microcline-biotite-albite-epidote-muscovite gneiss (.8 mile N 20° E of junction of Hall Brook and Meadow Brook, Wilmington).
5. L28/d Quartz-microcline-plagioclase-epidote-sphene gneiss (.5 mile East of the junction of Route 8 and Route 9, Wilmington).
6. 14H/b Quartz-microcline-muscovite-biotite-epidote gneiss (on Handle Road, 1.6 miles South of junction of North Brook and Handle Road, Wilmington).

*granulose

constituent and is associated with biotite and epidote. Epidote is present in amounts up to 12 percent as very small grains. In general quartz is finer grained than albite in these rocks, although the coarser sizes of each have about the same relative diameters. Albite usually contains numerous inclusions which are generally absent from associated quartz.

Sphene is a conspicuous accessory or minor constituent of these rocks and forms grains as large as 0.4 mm. The average diameter, however, is about 0.04 mm. In rocks containing minor amounts of sphene, magnetite and ilmenite are generally absent. Tourmaline is conspicuously absent from the Wilmington gneiss. It is, however, abundant in the Heartwellville schist and in the albite schists of both the Readsboro and Hoosac

TABLE 5

MODES OF THE WILMINGTON GNEISS

| | 7 | 8 | 9 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | .. | 55 | .. |
| Biotite | 5 | 3 | 15 |
| Microcline | 47 | 15 | .. |
| Oligoclase | 15 | .. | .. |
| Groundmass or grains | | | |
| Quartz | 20 | 25 | 62 |
| Muscovite | 10 | 1 | tr |
| Magnetite and Ilmenite | .. | tr | 2 |
| Epidote group | 3 | 1 | 1 |
| Zircon | .. | tr | .. |
| Sphene | .. | .. | tr |
| Tourmaline | .. | .. | .. |
| Pyrite | .. | tr | .. |
| Apatite | tr | tr | tr |
| Chlorite | .. | .. | tr |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–20.0 | 0.5–10.0 | 0.5–6.0 |
| Groundmass or grains | 0.01–0.4 | 0.008–0.4 | 0.01–0.4 |
| Texture | G* | G | G |

4. L42/c Microcline-quartz-oligoclase-muscovite-biotite-epidote gneiss (0.1 mile east of Bench Mark 1798, southeast Wilmington).
5. 30W–d Plagioclase-quartz-microcline-biotite gneiss (0.65 mile north of Carley Cemetery, southwest Wilmington).
6. 16B/3 Quartz-feldspar-biotite gneiss (.95 mile southeast of junction of Route 8 and Route 112).

*gneissose

formations with which the Wilmington gneiss is closely associated in this area.

A second rock type within the Wilmington gneiss is a massive, microcline gneiss making up about 10 to 150 feet. It consists of medium- to coarse-grained microcline; medium- to fine-grained quartz; and minor or accessory amounts of biotite, muscovite, and epidote. Banding is either wholly lacking or crudely developed. Generally this gneiss forms thin beds randomly distributed through the sequence.

The Wilmington gneiss is exposed more widely in Wilmington and Whitingham townships than elsewhere. Whether or not the base of the

Figure 9. Recumbent Spruce-Tree or Cascade fold in coarse Wilmington gneiss in core of Sadawga Pond Dome looking northeasterly (0.65 mile NW of Jacksonville on Route 8). Note that the bed reverses direction near top of photograph. Shatter-zone in quartz vein indicated by trellis pattern on right.

formation is exposed is not known. In the northern part of the outcrop area in Dover, it is known on the basis of structural and sedimentary features in the Tyson conglomerate that these beds west of the Wilmington gneiss are overturned in a recumbent fold. Here also the gneiss plunges northerly under the Hoosac formation. The available data suggest to the writer that the Wilmington gneiss in the vicinity of its western boundary rests upon formations which are stratigraphically younger than the gneiss. Some of the conglomerates, albite schist and aluminous green and black schists under and west of the Wilmington gneiss may also represent Tyson and Hoosac formations. Since in most instances these rocks are associated closely with marble beds, they were classified with the Searsburg and Readsboro formations.

In addition to the typical and characteristic rock types of this formation, a small part of the sequence mapped as Wilmington gneiss consists of glassy white to black quartzite, dark feldspathic quartzite, schist and coarse white and gray plagioclase gneisses. Quartzite and schists are best

Figure 10. Recumbent folds in quartzite and schist beds on southeast limb of the Sadawga Pond dome, looking southwesterly (0.8 mile NW of Jacksonville on Route 8). Note that the man on the right has his hand parallel to the axial plane and that the folds cascade to SE.

displayed 0.8 miles northwest of Jacksonville; dark feldspathic quartzites, along Route 8, 1.3 miles southwest of Whitingham Village; and coarse plagioclase gneisses on the north side of Route 8 in the center of Whitingham Village.

*Age of the Wilmington gneiss:* The Wilmington gneiss is regarded by the writer as of probable pre-Cambrian age. From the stratigraphic relationships in the Whitingham anticline and in the Sadawga Pond Dome the writer infers that the central gneisses of the dome lie stratigraphically below the marble-lime-silicate beds and these in turn are overlain by black graphite-garnet-muscovite schists (Heartwellville formation). On the southern end of the Whitingham anticline, the field relations and detailed mapping suggest that the albite schists to the southeast (Hoosac schist) overlie the above sequence with an unconformable relationship.

There is a strong possibility, on the other hand, that the Wilmington gneiss may be a dominantly microcline facies of the albite schist and highly-aluminous green and dark schist sequence. If this is the case,

Figure 11. Spruce-tree folds in white quartzite and dark schist beds of the Wilmington gneiss sequence; 0.8 mile northwest of Jacksonville Village in roadcut looking southwest. At the same locality as Fig. 10.

Figure 12. Overturned fold in plagioclase gneiss beds of the Wilmington gneisses seen looking southwesterly. Note reverse drag folds near hammer (roadcut on Route 8, 0.65 mile N of Wilmington-Whitingham township line, near x1722 of map, Plate I).

the facies variations are very complex, and have been further complicated by the doming in the eastern part of the area and by the development of recumbent folds along much of its western border. The relationship of the Wilmington gneiss to the albite schist with conglomeratic beds and with interbedded amygdaloidal amphibolites of the eastern part of the area is such that the latter are all younger than the Wilmington gneiss. The writer is of the opinion, however, that a detailed study of facies variations in the Wilmington gneiss would be very illuminating. If the Wilmington gneiss is a facies of the albite and aluminous schist sequence, the writer would favor the interpretation that all the beds from the Searsburg conglomerate upward through the Hoosac formation are part of an essentially uninterrupted succession of Paleozoic age, but having intricate facies changes throughout.

What has been said regarding the origin of the Stamford gneiss is equally applicable to the origin of the Wilmington gneiss.

## Readsboro Formation

### *Searsburg conglomerate member*

*Name:* Searsburg conglomerate is a term presented by the writer for a series of schistose conglomerate beds exposed locally along the angular unconformity in Searsburg. It is used for a distinctive series of blue and white quartz-pebble conglomerates; thin glassy, white, gray, and buff quartzites; micaceous white quartzites; coarse feldspathic arkosic conglomerates; and coarse albite schists. The best exposure of this sequence is 0.1 miles south of the Somerset-Searsburg township line, and 0.3 miles east of the East Branch of the Deerfield River, from elevation 2000 to 2160. The most readily accessible exposures of this sequence are some 500 feet southeast of Searsburg Dam at the base of the slope, and along the slope itself. The rock here is a biotite-chlorite-rich quartz-pebble and microcline-pebble, carbonate-bearing conglomerate.

Thin, fine-grained, buff to gray micaceous quartzites are found in the biotite-muscovite schist. In these quartzites, quartz is present in amounts from 80 to 95 percent. They contain, in addition, 5 to 10 percent of untwinned, fine-grained albite. Muscovite is more abundant than green biotite and is also coarser grained. Muscovite and biotite, however, as well as magnetite and epidote are minor constituents. Pyrite, tourmaline, apatite, and zircon are accessory minerals.

Any individual specimen of quartzite contains layers which are nearly 90 to 97 percent quartz. Thin schistose or feldspathic layers or lenses are found, however, in nearly every hand specimen. In general

Figure 13. Searsburg Conglomerate with closely associated gneiss (0.35 mile N 60° E of intersection of Bennington-Windham County township line and Route 9).

these more schistose and feldspathic portions are richer in accessory and minor constituents than the quartzose parts. Modes of micaceous quartzite are presented in Table 6.

Stratigraphically above the Searsburg unconformity in many localities are a series of schists which are much like the schists associated with the conglomerate and quartzite beds. However, where the latter rocks are absent or not exposed the sequence is not sufficiently distinctive to be mapped separately. Consequently, only those parts of the Searsburg were mapped separately which contain these distinctive lithologic types.

The Searsburg conglomerate has a thickness ranging from 0 to approximately 175 feet.

*Description:* The Searsburg conglomerate is typically a quartz- and microcline-pebble conglomerate in a schistose matrix. Associated with it in this area are thin, glassy, white, gray, and buff quartzites; coarse-grained microcline and albite conglomerates; and coarse-grained albite schists with albite fragments up to 10 mm. Albitic and quartz-rich schists, some graphitic and others non-graphitic, are present near the base of the formation.

The larger pebbles consist of blue and white quartz elongated into

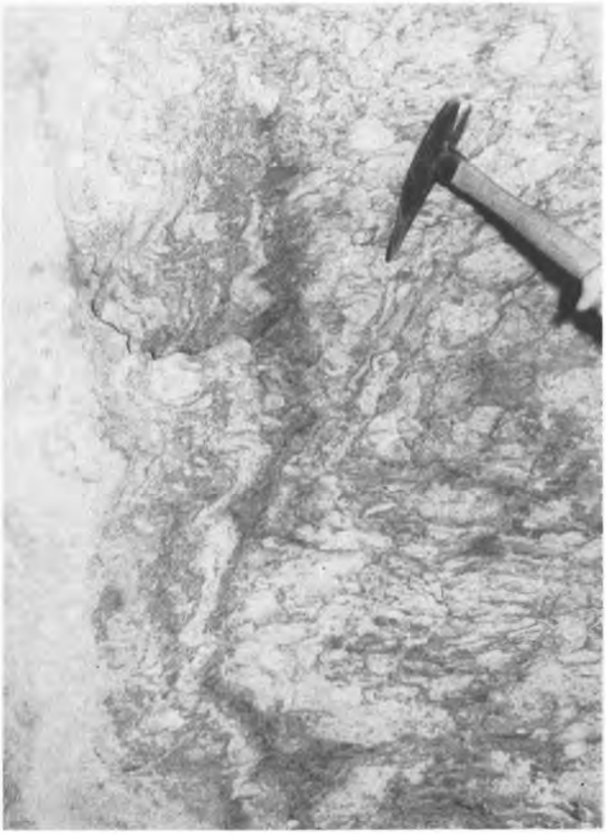


Figure 14. Closeup of Searsburg conglomerate. (same outcrop as preceding Figure)

cigar-shaped masses flattened into the plane of the foliation of the finer grained constituents in which the pebbles occur. Average dimensions of the pebbles differ from one locality to another. The largest are ¼ by ½ inch by 6 inches. The long axes of these pebbles trend east to northeast and plunge at a steep angle. The pebbles make up about one-third of the volume of the rock in which they are found. The remaining two-thirds of the rock is dark-gray where schistose and light-gray to light-blue where arenaceous.

The finer grained constituents of the conglomerate consist of quartz, microcline, albite, biotite, chlorite, muscovite, carbonate, magnetite, tourmaline, zircon, pyrite, apatite, and epidote. The microcline grains average about 0.3 mm. and quartz grains whose dimensions are as great

TABLE 6

MODES OF SEARSBURG CONGLOMERATE

| | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| Porphyroblasts | | | | | |
| Quartz | 35 | 43 | 81 | 89 | 85 |
| Albite | 10 | 15 | 10 | 6 | 5 |
| Microcline | 28 | .. | .. | 1 | 7 |
| Groundmass or grains | | | | | |
| Biotite | 4 | 7 | 2 | 1 | 2 |
| Muscovite | 13 | 25 | 5 | 3 | tr |
| Carbonate | .. | .. | .. | .. | tr |
| Zircon | .. | tr | tr | .. | .. |
| Tourmaline | .. | 1 | tr | .. | tr |
| Magnetite | tr | tr | 1 | tr | tr |
| Apatite | tr | .. | tr | .. | tr |
| Epidote group | 7 | .. | 1 | tr | 1 |
| Pyrite | tr | tr | tr | tr | tr |
| Garnet | .. | 10 | .. | .. | .. |
| Chlorite | .. | tr | .. | tr | .. |
| Grain size in mm. | | | | | |
| Porphyroblasts | 0.5–15.0 | .... | 0.5–2.0 | 0.5–2.0 | 0.5–3.0 |
| Groundmass or grains | 0.3–0.4 | 0.2–0.4 | 0.03–0.4 | 0.03–0.4 | 0.03–0.4 |
| Palimpsest grains | 2.0–150.0 | 2.0–20.0 | .... | .... | 1.0–150.0 |
| Texture | S* | S | G** | G | S |

1. R20/d Microcline-quartz-albite conglomerate (at the Pebble locality 500 feet Southeast of Searsburg Dam, Searsburg).
2. O13/e Quartz-muscovite-albite-garnet schist (1.25 miles N 30° E from intersection of East Branch and Deerfield River, Searsburg).
3. R20/da Quartz-albite-muscovite-biotite conglomerate (at Pebble locality 500 feet Southeast of Searsburg Dam, Searsburg).
4. R20/a Quartz-albite-muscovite-biotite conglomerate (at Pebble locality on East side of Searsburg Reservoir, Searsburg).
5. R20a/d Quartz-microcline-albite conglomerate (at Pebble locality 500 feet Southeast of Searsburg Dam, Searsburg).

*schistose
**granulose

as 0.60 mm. are scattered through the rock. The pebbles are coarse grained and deficient in minerals other than quartz. They consist of about 98 to 100 percent quartz as a general rule. Modes of the quartz-pebble conglomerate are given in Table 6.

Gray-weathering, coarse-grained biotite-albite augen schist is a distinctive rock and lies stratigraphically just above the conglomerate. The albite augen are usually 2 to 5 mm. in diameter, although not uncommonly they reach a length of 1 cm. In many cases the albite grains enclose garnets a millimeter or less in diameter. The minerals enclosing the albite augen consist in part of coarse-grained (0.5 mm.) muscovite which in thin section has a rather dark appearance. The dark-green biotite is extremely fine grained in comparison with the muscovite with which it is consistently associated. In most cases, biotite forms rims along the outer edges of the muscovite or forms actual overgrowths on the muscovite. Epidote is abundant and not uncommonly forms grains as large as 0.9 mm. Microcline grains 1 to 2 mm. in long diameter and filled with minute inclusions, are an important constituent of this rock. Carbonate and pyrite are conspicuous accessory minerals in the more schistose rocks. Crystals of green tourmaline forming overgrowths on blue tourmaline are 1 to 3 mm. in cross-section diameter. Tourmaline is a conspicuous minor constitutent of this schist. Modes of the biotite-albite augen schist are given in Table 6.

An important rock in the Searsburg conglomerate has a matrix of fine-grained, gray-weathering biotite, muscovite and albite. Angular pebbles of albite and microcline are set in this matrix.

### *Readsboro Schist*

*Name:* Hubbard (1924, p. 283) gave the name Readsboro schist to a series of feldspathic quartz-biotite-muscovite-quartz-chlorite schists that crop out widely over the townships of Readsboro and Whitingham. The term "Readsboro" is a useful term and quite appropriate to stratigraphic usage in the Wilmington-Woodford area. On the other hand, as mapped by Hubbard it is not quite suitable and for reasons that will be explained should be redefined as a stratigraphic unit that can be employed readily in field mapping.

As redefined by the writer, the Readsboro formation includes all of the muscovite-biotite-chlorite-garnet-albite schists, quartzites and marble beds lying between the Searsburg conglomerate and below the non-albitic Heartwellville green and black schists. As redefined, the term Readsboro formation includes the Sherman marble (Hubbard, p. 269) as a member of the Readsboro rather than a separate formation; and the Whitingham schist which is mapped separately by Hubbard (p. 276) is included in the Readsboro schist member as redefined by the writer for reasons which are given below.

Hubbard (Pl. XX, opposite p. 282) has mapped as part of the Readsboro, the microcline and plagioclase gneisses of the Sadawga Pond dome and of northwest Whitingham township. In addition, the rocks of Heartwellville lithology in the northwest flank of Sadawga Pond dome were mapped by Hubbard (p. 285) as thin beds of blue sericite schist (like the upper Heartwellville) stratigraphically part of the Readsboro.

Mapping by the writer in the Wilmington-Woodford area and reconnaissance inspection surveys on the Hoosac Range to the south strongly suggest that rocks of two different ages, but of largely the same lithology, are present in this region. The albite schist sequence (Readsboro schist) is lithologically the same as the younger sequence to the east, termed Hoosac formation in the report. These two sequences seem to be separated by an unconformity as indicated above under the discussion of the age of the Wilmington gneiss. The Hoosac formation as defined by Pumpelly, Wolff, and Dale (1894, p. 59) is an albite schist. In the type locality on the top of Hoosac Mountain, it conformably overlies conglomerate of the Vermont formation and as interpreted by the writer at this locality is the same sequence as the Tyson and Hoosac formations of the Wilmington-Woodford area. A part of the albite schists of the western slopes of Hoosac Mountain are interpreted by the writer as a southward continuation of the Readsboro schist and lie unconformably below the upper albite schists. For this reason, the writer uses the term Readsboro formation instead of the term Hoosac schist as proposed by Pumpelly, Wolff, and Dale. The writer wishes to retain the term "Hoosac schist" as being a useful stratigraphic name, but its usage in the light of new information should be somewhat restricted to include the Paleozoic albite schists containing amphibolites rather than the schists of probable pre-Cambrian age which do not contain that type of rock.

*Description:* The Readsboro schist is dominantly an albite-muscovite schist, the albite occurring as small to large augen. Abundant lenticular, untwinned albite augen up to 2 cm. in long diameter make up to 10 to 20 percent of the schist. The average grain diameter, however, is about 2–5 mm. These are set between the muscovite laminae in such a way that when the rock is viewed in sections normal to the foliation the highly contorted character of the schist is well displayed.

In thin section, the albite schist has a poikiloblastic texture as a result of the uniform distribution of abundant fine-grained magnetite and graphite in coarse-grained albite. The minerals occur in planes parallel to the foliation. In thin sections cut normal to fold axes the lines of magnetite and graphite record the amount and directions in which the rotational forces acted. The albite augen are the most conspicuous mineral in the schist. It is usually easy to distinguish this albite from quartz

TABLE 7

MODES OF THE READSBORO SCHIST MEMBER OF THE READSBORO FORMATION

| | 1 | 2 | 3 |
|---|---|---|---|
| Porphyroblasts | | | |
| Biotite | 5 | 8 | 15 |
| Albite | 15 | 20 | 20 |
| Chlorite | 2 | 1 | tr |
| Garnet | 1 | .. | .. |
| Tourmaline | tr | .. | .. |
| Hornblende | .. | 20 | .. |
| Groundmass or grains | | | |
| Muscovite | 20 | .. | 15 |
| Magnetite and Ilmenite | 2 | 4 | .. |
| Epidote | .. | 12 | 3 |
| Quartz | 55 | 35 | 47 |
| Rutile | tr | .. | .. |
| Sphene | .. | .. | tr |
| Zircon | tr | tr | .. |
| Carbonate | .. | tr | .. |
| Apatite | .. | tr | tr |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–4.0 | 0.5–1.5 | 0.5–7.0 |
| Groundmass or grains | 0.01–0.4 | 0.008–0.4 | 0.01–0.4 |
| Texture | S* | S | S |

1. 95D/b Quartz-muscovite-albite-biotite-garnet schist (1.2 miles northeast of crest of Mt. Pisgah).
2. L53/b Quartz-albite-hornblende-epidote-biotite schist (1 mile S 10° W of junction of Howe Pond Brook and West Branch of the Deerfield River, Readsboro).
3. M12/a Quartz-albite-muscovite-biotite schist (0.3 mile S 85° E of Readsboro Falls, Readsboro).

*schistose

because it contains many inclusions, whereas the quartz is relatively free from such impurities. Quartz is commonly present in the corners of the albite augen as well as in layers parallel to the foliation plane. The biotite in every case is a dark-brown variety and is associated with muscovite. A minor but conspicuous constituent is green tourmaline which is present in amounts up to 3 percent. The average grain diameter of the tourmaline is about 0.08 mm. Modes are presented in Tables 7 and 8.

Hubbard (op. cit., p. 276) distinguished the Whitingham schist from

Table 8

MODES OF THE READSBORO SCHIST MEMBER OF THE READSBORO FORMATION

| | 4 | 5 | 6 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | .. | 5 | 15 |
| Chlorite | 8 | 8 | .. |
| Garnet | 5 | .. | 5 |
| Tourmaline | 1 | tr | 1 |
| Groundmass or grains | | | |
| Quartz | 60 | 47 | 10 |
| Muscovite | 25 | 30 | 45 |
| Magnetite and Ilmenite | 1 | 1 | 1 |
| Epidote | tr | 7 | 2 |
| Graphite | .. | .. | 1 |
| Rutile | tr | .. | .. |
| Sphene | .. | .. | 1 |
| Apatite | .. | tr | tr |
| Pyrite | .. | 2 | tr |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–2.0 | 0.5–3.0 | 0.5–4.0 |
| Groundmass or grains | 0.01–0.4 | 0.008–0.4 | 0.01–0.4 |
| Texture | S* | S | S |

4. M3/a Quartz-muscovite-chlorite-garnet schist (1.35 miles S 80° E of Readsboro Falls, Readsboro).
5. L63/c Quartz-muscovite-chlorite-albite schist (on Route 8, 0.1 mile northwest of junction of Howe Pond Brook and West Branch of Deerfield River, Readsboro).
6. 95D/m Muscovite-albite-quartz-garnet schist (on Dover-Somerset boundary, 0.9 mile due south of intersection of Stratton-Somerset Boundary, Dover).

*schistose

the Readsboro and notes that it overlies the Sherman marble in every occurrence in which he had observed the latter, and he points out that they seem "to be special phases of the Readsboro".

This rock differs from typical Readsboro albite schist in that it usually has a high content of black biotite in which are set conspicuous white albite, quartz, and calcite grains. By reason of this color contrast between white and black minerals this rock has been referred to by Dale (1915, p. 51) as a "diorite schist". This rock type and associated Sherman marble are well displayed at the locality which he describes, and this is located along the slopes of the hills 0.4 miles south of the Deerfield River along the Wilmington-Searsburg township line.

A second development that seems typical of the albite schist in the vicinity of the marble where biotite is lacking, is that the rock becomes a light green muscovite-albite schist. Both of the above described albite schists characteristically have brown weathering pitted surfaces representing iron weathering from pyrite, ankerite, and possibly to some extent from biotite.

The writer has found this schist a useful marker since it is present quite commonly at localities which are stratigraphically at the usual position of the Sherman marble whether or not marble is developed or exposed there. It was not mapped separately, however, since in those localities where it is well developed, the marble is immediately below and is a much better marker. Where the marble is not present, this rock type is so thin that it cannot be mapped on scale 1:62,500.

*Sherman marble member of the Readsboro formation*

*Name:* Sherman marble is a term proposed by Hubbard (1924, p. 269) for a sequence of calcite and dolomite marbles and lime-silicate granulites. The type locality is adjacent to the Hoosac Tunnel and Wilmington Railroad, 2900 feet north of the Vermont-Massachusetts state line. It is the site of a lime quarry which formerly provided rock for a calcium carbide operation. Hubbard noted that the Sherman marble grades upward into the albite schist of the Whitingham schist formation. The writer's study in the Wilmington-Woodford area confirms this in a number of localities, but it is not everywhere true. In the Readsboro, Sherman, and Medburyville anticlines the marble forms the core of these folds. In the Whitingham, Haystack, Mount Snow anticlines and Sadawga Pond dome, rocks stratigraphically below the marble are exposed.

Data from the area as a whole indicate that the calcite and dolomite marbles mapped as Sherman marble are a facies of the albite schist sequence, and in every known instance but one are also underlain by albite schist. In the Sherman, Readsboro, Medburyville, Haystack Mountain, and Mount Snow anticlines the marble grades upward into albite schists which would be, in part at least, the equivalent of the Whitingham schist of Hubbard. In the Whitingham anticline and Sadawga Pond dome, on the other hand, the marble is directly overlain by dark schist of the Heartwellville formation. In the eastern part of the Mount Snow anticline, 1½ miles northeast of Mount Pisgah (Snow) in the North Branch of the Deerfield River, the marble seems to be stratigraphically enclosed by the Heartwellville schist.

The regional stratigraphic relationships therefore suggest that the Sherman marble be considered as a member of the Readsboro formation which is dominantly an albite schist sequence. The term Sherman marble, in the writer's opinion, should be redefined to include the calcite and dolomite marbles and lime-silicates lying stratigraphically in the dominantly albitic Readsboro schist sequence and below the dark schists of the Heartwellville schist formation. The marble of the eastern Mount Snow anticline which is enclosed in Heartwellville schist is indicated as marble in the Heartwellville rather than Sherman marble. It is, however, lithologically similar to the pink and gray calcite marble and buff dolomite of other localities of the Wilmington-Woodford area, and of Pike Hollow and other localities to the north in the Londonderry and Saxtons River quadrangles.

*Distribution:* Sherman marble crops out in eastern Readsboro, Whitingham, western Wilmington, southeastern Searsburg, western Dover and eastern Somerset townships. Marble is found in the eastern third of the area in only one outcrop in southeastern Wilmington but it is quite possible that this outcrop should not be considered part of the Sherman marble. The larger marble deposits are brought up in major anticlinal and domal folds. Smaller lens-shaped bodies of marble are present outside of these structures and form good marker beds for mapping.

*Description:* The Sherman marble consists of calcite and dolomite marbles and lime-silicate granulites. Lime-silicate minerals are particularly noteworthy and abundant west of the village of Whitingham and in the Sherman, Readsboro, and Whitingham anticlines. Most occurrences of this rock consist largely of coarse-grained, white or pink, siliceous calcite marble and fine-grained, siliceous dolomite marble. The dolomite marbles are present in the lower half of the member and the calcite marbles in the upper half.

A crude foliation is generally developed in the coarser grained marbles and is attributed to the presence of small amounts of phlogopite and a light-green mica, and in certain cases also to crystalline graphite scattered throughout the rock.

The grains of twinned calcite are 0.5 to 3 mm. in diameter and average 1 mm. In places crystals up to one inch have been observed. Dolomite grains are generally smaller and average about 0.5 mm. Most of the marbles contain 5 to 15 per cent of quartz, the grains of which average 0.3 mm. in diameter. Phlogopite is a minor constituent of these marbles

TABLE 9

MODES OF THE SHERMAN MARBLE MEMBER OF THE READSBORO FORMATION

| | 1 | 2 | 3 |
|---|---|---|---|
| Porphyroblasts | | | |
| Actinolite | .. | .. | 37 |
| Phlogopite | 2 | .. | tr |
| Graphite | tr | tr | tr |
| Pyrite | tr | tr | tr |
| Groundmass or grains | | | |
| Carbonate | 97 | 85 | 60 |
| Quartz | 1 | .. | .. |
| Talc | .. | 15 | tr |
| Epidote group | tr | .. | .. |
| Sphene | tr | .. | .. |
| Anthophyllite | .. | tr | 3 |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–3.0 | 0.5–2.0 | 0.5–3.0 |
| Ground mass or grains | 0.05–0.3 | 0.03–0.4 | 0.02–0.4 |
| Texture | G* | G | G |

1. 24S/g Calcite marble (0.6 mile N 75° E of Readsboro Center, Readsboro, Vermont).
2. 12W/f Lime-silicate marble (1.25 miles S 20° E of Harriman Dam in No. 9 Brook, Whitingham).
3. 50W/f Sherman marble (2 miles S 75° E. of Harriman Dam in No. 9 Brook, Whitingham).

*granulose

and forms grains as long as 12 mm. Graphite is a conspicuous accessory mineral. It is found as crystals up to 4 mm., some of which carry triangular markings on the base. These markings are the result of gliding along an undetermined second-order pyramid (Hurlbut, 1959, p. 245). All modes are presented in Tables 9 and 10.

Lime-silicate granulites comprise carbonate-phlogopite-actinolite, carbonate-phlogopite-tremolite, and biotite-diopside granulites. These form thin beds generally 10 to 20 feet thick and their mineralogy is related not only to bulk chemical composition of the rocks, but particularly to grade of metamorphism. The granulites are green and are fine- to medium-grained rocks.

The actinolite of the carbonate-phlogopite-actinolite granulites contains abundant inclusions of irregularly-shaped carbonate globules.

TABLE 10

MODES OF THE SHERMAN MARBLE MEMBER OF THE READSBORO FORMATION

| | 4 | 5 |
|---|---|---|
| Porphyroblasts | | |
| Actinolite | tr | 15 |
| Diopside | 35 | .. |
| Muscovite | .. | tr |
| Phlogopite | .. | 1 |
| Graphite | tr | 1 |
| Groundmass or grains | | |
| Quartz | 10 | 52 |
| Microcline | 5 | .. |
| Albite | .. | 5 |
| Oligoclase | 20 | .. |
| Epidote group | 15 | 25 |
| Magnetite | .. | tr |
| Sphene | .. | tr |
| Zircon | tr | tr |
| Apatite | tr | tr |
| Biotite | 15 | .. |
| Chlorite | .. | 2 |
| Grain size in mm. | | |
| Porphyroblasts | 0.5–6.0 | 0.5–5.0 |
| Groundmass or grains | 0.03–0.4 | 0.03–0.4 |
| Texture | G* | G |

4. 28 W/c. Lime silicate Marble (2.9 miles S 75° E of Harriman Dam, Whitingham).
5. B 9/c. 0.5 mile west-southwest of Clara Lake, Whitingham.

*granulose

Some of the coarser actinolite grains reach a length of 5 to 10 mm. Twinned carbonate grains range from 0.2 to 1.3 mm. and average about 0.5 mm. Minor constituents are phlogopite, talc, anthophyllite, and chlorite. Apatite, magnetite, pyrite, sphene, and zircon are accessory minerals. Modes are given in Tables 9 and 10.

Biotite-diopside granulites contain diopside grains 0.3 to 1.5 mm. in length and average about 0.7 mm. Epidote is relatively abundant and has a maximum grain diameter of 0.4 mm. The granulites contain small amounts of microcline and larger amounts of oligoclase and quartz. Accessory minerals are pyrite, apatite, and zircon. A mode of this rock type is given in Table 10.

The marble ranges in thickness from 0 to 300 feet. Its reduced thickness or absence may be attributable to tectonic thinning rather than to extreme variations in thickness at the time of sedimentation.

### Heartwellville Schist

*Name:* The term Heartwellville schist was proposed by Hubbard (1924, p. 278) for the green and dark sequence of muscovite-chlorite-quartz lensed schists above the carbonate-bearing albite schists of his Whitingham schist. This term serves as a useful stratigraphic term in the entire Wilmington-Woodford area. Hubbard considered that the Heartwellville was overlain by the Readsboro schist but the present investigation indicates that the Readsboro is stratigraphically under the Heartwellville in the entire area. The most complex portion of the area for working out the stratigraphy in the late pre-Cambrian sequence is in the Readsboro and Whitingham areas.

*Distribution:* The Heartwellville schist is widespread throughout the central part of the area. This rock is extremely resistant and caps the higher portions of the mountain ridges from the vicinity of Hoosac Peak (east of Stamford Village), northward through Readsboro, Searsburg, Somerset, Stratton, Dover, and Wilmington. Complexly-folded Heartwellville schist crops out in western Whitingham, and eastern Readsboro. Stratigraphic relationships seem more clear and consistent in the northern portions of the area and in the vicinity of Sadawga Pond dome. In spite of the stratigraphic simplicity suggested by the simple anticlinal structure of the Readsboro and Sherman anticlines, the stratigraphic succession is not completely clear here. The Heartwellville schist beds seem to be duplicated in these structures. This duplication may indicate that two sequences of Heartwellville lithology were deposited in the southern part of the area; or it may indicate that the Heartwellville schist has been deformed into recumbent folds and subsequently arched up into anticlinal folds. The writer's opinion, however, is that facies variations are responsible and that the alternation of albite schist with the green and black schist reflects conditions of deposition rather than tectonic duplication. Individual outcrops in many cases show this alternation on a small scale.

*Description:* The Heartwellville schist comprises two distinctive types of rock. The lower part of the unit, particularly in the Stamford, Dutch Hill, and northern districts of the area is a light- to dark- green chlorite-muscovite- (garnet)- quartz lensed schist (Fig. 15), and the upper part is a coaly, dark gray to black graphite-muscovite- (chlorite)- garnet-quartz lensed schist (Fig. 16).

The light- to dark-green schist contains abundant chlorite and chlori-

Figure 15. Light-green, chlorite-garnet-muscovite-quartz schist of lower part of Heartwellville schist. In Bond Brook, 0.1 mile west of Route 9, Searsburg.

toid. Modes are presented in Tables 11, 12 and 13. Garnets 5 to 15 mm. in diameter are common to each of the several rocks making up the unit. Green tourmaline is usually present as an accessory mineral. Locally, however, veins of fine-grained tourmaline are found which reach a length of two feet. These veins are easily seen along the Hoosac Tunnel and Wilmington Railroad track at Sherman in southwestern Whitingham (Pl. I).

The dark-gray to black muscovite schists are generally rusty and yellow due to weathering. Although this rock has several variations in mineral composition, it is characterized by an abundance of muscovite and absence of albite. One variety of muscovite schist contains 10 to 15

TABLE 11

MODES OF THE HEARTWELLVILLE SCHIST

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Porphyroblasts | | | | |
| Albite | 3 | 4 | .. | 15 |
| Garnet | 10 | 15 | .. | 3 |
| Biotite | 3 | .. | .. | 5 |
| Chloritoid | tr | .. | 20 | .. |
| Tourmaline | .. | tr | .. | 1 |
| Groundmass or Grains | | | | |
| Quartz | 25 | 15 | 15 | 40 |
| Chlorite | .. | 2 | 10 | .. |
| Muscovite | 55 | 60 | 51 | 34 |
| Magnetite and Ilmenite | 1 | 2 | 2 | 2 |
| Graphite | 1 | 1 | 2 | tr |
| Epidote group | tr | tr | .. | .. |
| Zircon | .. | tr | .. | tr |
| Sphene | .. | .. | .. | tr |
| Pyrite | .. | .. | tr | .. |
| Apatite | .. | .. | .. | .. |
| Rutile | .. | .. | .. | .. |
| Grain size in mm. | | | | |
| Porphyroblasts | 0.5–8.0 | 0.5–7.0 | 0.5–5.0 | 0.5–2.0 |
| Groundmass or Grains | 0.01–0.4 | 0.02–0.4 | 0.02–0.4 | 0.03–0.4 |
| Texture | S* | S | S | S |

1. 55 D/a. Muscovite-quartz-garnet-biotite schist (on Sleepy Hollow Road, 0.65 mile northeast of junction of Sleepy Hollow Road and South Road, Searsburg).
2. 9 H/g. Muscovite-quartz-garnet-albite-graphite schist (0.55 mile N 75° E of junction of Sleepy Hollow and South Road, Searsburg).
3. 9 H/d. Muscovite-chloritoid-quartz-chlorite schist (1.55 miles due East of intersection of Sleepy Hollow Road and South Road, Searsburg).
4. 55 D/b. Muscovite-albite-biotite-tourmaline schist (5.5 miles due South of junction of Sleepy Hollow Road and Routh 9, Searsburg).

*schistose

percent of garnet that ranges in size from 0.9 to 8 mm. in diameter. In thin section this dark rock has a poikiloblastic texture. This is due to the presence of 1 to 2 percent of magnetite and nearly equal amounts of finely disseminated graphite in garnet and muscovite. These minerals form layers parallel to the foliation planes and are traced continuously from the groundmass of the schist through the porphyroblasts of garnet and smaller grains of albite where this is present. Chlorite, present in

Figure 16. Characteristic appearance of dark graphite-muscovite-chlorite-garnet lensed schist at type locality (0.15 mile N of junction of Lamb Brook with West Branch of Deerfield River on Route 8, Heartwellville). Note highly contorted condition of schist as shown by ink lines near hammer.

minor amounts, is associated with garnet. In certain cases it forms partial or complete rims about the garnet. Some of these rims are several times larger than the garnet with which it is associated and from which it has formed by retrograde metamorphism. Accessory minerals are epidote, zircon and green tourmaline. Modes are given in Tables 11, 12 and 13.

Some beds of this muscovite schist contain chloritoid in amounts up to 15 percent. Chloritoid is associated with muscovite with which it is interlaminated. The muscovite associated with chloritoid is generally

TABLE 12

MODES OF THE HEARTWELLVILLE SCHIST

| | 5 | 6 | 7 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | 15 | .. | 10 |
| Garnet | .. | 7 | 5 |
| Biotite | 5 | 7 | .. |
| Chloritoid | .. | .. | 30 |
| Tourmaline | 3 | tr | 3 |
| Groundmass or Grains | | | |
| Quartz | 30 | 20 | 60 |
| Chlorite | tr | .. | 10 |
| Muscovite | 43 | 40 | 10 |
| Magnetite and Ilmenite | 2 | 3 | 1 |
| Graphite | tr | .. | .. |
| Epidote group | .. | .. | 1 |
| Zircon | tr | tr | tr |
| Sphene | tr | .. | .. |
| Rutile | .. | .. | .. |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–6.0 | 0.5–8.0 | 0.5–5.0 |
| Groundmass or Grains | 0.02–0.4 | 0.03–0.4 | 0.02–0.4 |
| Texture | S* | S | S |

5. L 49/c. Albite-biotite-tourmaline schist (on Stratton-Somerset township line 0.25 mile South of northern limits of the area).
6. R 18/e. Muscovite-quartz-chloritoid-garnet schist (0.7 mile east-southeast of Searsburg Reservoir Dam).
7. M 29/a. Quartz-muscovite-chloritoid-garnet schist (0.3 mile South of Massachusetts-Vermont state line on Sherman Reservoir).

*schistose

finer-grained, (average about 0.2 to 0.4 mm.), although grains 1 to 1.5 mm. are not uncommon. No garnet was observed in the thin sections associated with chloritoid although elsewhere in this formation they are found together. Rounded or irregularly-shaped masses of chlorite, however, which may be pseudomorphous after garnet are scattered throughout the rock.

The Heartwellville schist in the highly deformed Readsboro and Whitingham districts is very thin and in places is missing altogether, presumably as a result of tectonic thinning. Generally, however, this schist seems to have a thickness of some 700 feet. In mapping this formation

Figure 17. Looking southwesterly across Molly Stark Trail toward the Freezing Knolls (elevation 2612 and 2645 feet) in distance, and toward 2025 hill in center ground. Former are capped by resistant Heartwellville schist; latter is formed of Searsburg (or Tyson) conglomerate and of Wilmington gneiss.

one receives a first impression that this schist is extremely thick, since it is widespread over vast areas and shows highly contorted folds which are extremely difficult to follow for any distance. Good marker beds within the formation are absent and thin graphitic quartzites are not abundant nor consistently present. In the northern part of the area and in the Stamford-Dutch Hill district the Heartwellville schists cap the upper slopes of the ridges. This regional study leads the writer to conclude, that this formation is essentially flat-lying in spite of the highly deformed character of the schist and the much-brecciated appearance of the rock (see Figs. 16 and 17).

## *Paleozoic Rocks*

### General Statement

The Paleozoic stratified rocks of the area comprise an easterly dipping homoclinal sequence of arenaceous, argillaceous, and volcanic rocks.

TABLE 13

MODES OF THE HEARTWELLVILLE SCHIST

| | 8 | 9 |
|---|---|---|
| Porphyroblasts | | |
| Oligoclase | .. | 0.5 |
| Garnet | 5 | 5 |
| Biotite | .. | 0.5 |
| Pyrite | .. | tr |
| Tourmaline | 1 | .. |
| Groundmass or grains | | |
| Quartz | 60 | 54 |
| Chlorite | 8 | 10 |
| Muscovite | 25 | 20 |
| Magnetite and Ilmenite | 1 | 2 |
| Epidote group | tr | 7.5 |
| Zircon | ... | tr |
| Grain size in mm. | | |
| Porphyroblasts | 0.5–1.0 | 0.5–7.0 |
| Groundmass or grains | 0.03–0.4 | 0.008–0.4 |
| Texture | S* | S |

8. L 63/c. Muscovite-chloritoid-quartz-garnet schist (0.1 mile northwest of junction of Howe Pond Brook, West Branch of Deerfield River, Readsboro).
9. M20/b. Quartz-muscovite-chlorite-epidote-garnet schist (0.9 mile S 18° W. of the junction of Bennington-Windham County and Vermont-Massachusetts state lines).

*schistose

These crop out largely in the eastern third of the area and form a fairly well-defined sequence. The base of the Paleozoic succession has been difficult to establish because of the strong similarity in lithology between the rocks of probable late pre-Cambrian age and the lower part of the overlying Paleozoic sequence. The angular unconformity in Searsburg (Plate I) separating the late from the early pre-Cambrian rocks is most striking and has long been recognized as a major unconformity by earlier workers (Dale, 1914–16). By reason of the strong similarity between the probable late pre-Cambrian and the Paleozoic rocks they were considered as one and mapped as Paleozoic. The Searsburg unconformity, therefore, was considered to be the erosion surface between the pre-Cambrian sequence to the west and the Paleozoic to the east. It is quite possible that this earlier interpretation is correct.

TABLE 14

MODES OF THE TYSON FORMATION

| | 1 | 2 | 3 |
|---|---|---|---|
| Porphyroblasts | | | |
| Biotite | 10 | 10 | 7 |
| Microcline | 30 | 45 | 45 |
| Quartz | 42 | 35 | 34 |
| Albite | 10 | 5 | 5 |
| Groundmass or Grains | | | |
| Muscovite | 7 | 2 | 5 |
| Zircon | tr | .. | .. |
| Sphene | 0.5 | 2 | 2 |
| Epidote | tr | 1 | 2 |
| Pyrite | tr | tr | tr |
| Apatite | tr | tr | tr |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–6.0 | .. | .. |
| Palimpsest grains | 0.5–6.0 | 0.5–5.0 | 0.5–5.0 |
| Groundmass or grains | 0.03–0.4 | 0.03–0.4 | 0.03–0.4 |
| Texture | S* | S | S |

1. 16 w/b. Quartz-microcline-biotite-albite-muscovite gneiss (0.3 mile S35°E of junction of North Branch of Deerfield River and Route 8, Wilmington).
2. L 25/c. Microcline-quartz-biotite-albite-muscovite conglomerate (0.35 mile S45°W of Wilmington-Dover-Marlboro township junction).
3. L 25/c. Same

*schistose

The writer, however, considers the present division as equally valid until more decisive information is forthcoming.

The present survey has demonstrated the presence of a conglomeratic formation (Tyson conglomerate) in the eastern third of the area, associated with albite schists and amygdaloidal amphibolites of the Hoosac formation. In the eastern part of the area these schistose conglomerate lenses, albite schists and amphibolites rest on coarse microcline gneiss. In the southern part of the area they seem to rest on non-amphibolite bearing albite schists and highly aluminous schists. In the southern part of the Whitingham anticline the Hoosac schist seems to overlie the microcline gneiss and the Heartwellville schist. An alternate possibility is that the Hoosac formation is the youngest part of the albite and aluminous schist sequence with the Wilmington gneiss as an intermediate member which thins beyond recognition within the limits of the map area as a result of facies variations.

The sequence mapped tentatively as of later pre-Cambrian age is present also in the Wallingford quadrangle and in the Chester Dome (Thompson, 1960, oral communication). In the latter district, highly aluminous schists similar to the Heartwellville schist (Gassetts schist) occur in the central portion of the dome. Their structural features suggest the possibility that they may be unconformably below the Paleozoic sequence.

One of the most pronounced differences between the Lower Paleozoic sequence as mapped and the late pre-Cambrian rocks, is the presence of Sherman marble in the latter and no such marbles in the former. The writer has mapped large parts of the late pre-Cambrian district in great detail. As greater detail emerged from this study, the differences between these sequences became more pronounced.

An alternate interpretation that cannot be excluded in the light of available information, is that the Readsboro formation as mapped is a stratigraphically continuous sequence with the Hoosac formation but lies below it and that the Wilmington gneiss is a facies development between these formations. A third possibility is that the Readsboro and Hoosac formations are facies variations representing contemporaneous deposition on either side of an active positive area now represented by Wilmington gneiss outcrop.

## *Cambrian And/Or Ordovician Rocks*

### Tyson Formation

*Name:* The term, Tyson formation, was proposed by James B. Thompson, Jr. (1950) for the basal sequence of Lower Paleozoic rocks at the type locality in Tyson Village, in the northwestern part of the Ludlow quadrangle (Fig. 1). In the Wilmington area, as in the type locality, the formation includes the conglomerate lying between older gneisses and the overlying albite schists of the Hoosac formation (Grahamville formation of the Ludlow quadrangle).

*Distribution:* The Tyson conglomerate occurs as lenticular masses of blue and white quartz conglomerate; and of microcline and coarse-grained albite-pebble conglomerate. Localities at which the conglomerate is best displayed are in northwestern Dover, northeastern Wilmington township, one mile northeast of Wilmington Village, and in several localities on the northern and southeastern flanks of the Sadwaga Pond dome. In addition, two localities present west of Harriman Reservoir and adjacent to the Readsboro-Whitingham, Searsburg-Wilmington township boundaries, are considered to be part of the Tyson formation.

The Tyson formation is considered to be the equivalent of the Vermont formation as described by Pumpelly, Wolff, and Dale (1894, pp. 48–59).

*Description:* The Tyson formation consists chiefly of fine- to coarse-grained, white- to blue-quartz pebbles, fine- to coarse-grained, gray, buff, and pink microline pebbles, and coarse-grained albite fragments. At the locality northeast of Wilmington Village, boulder-like masses of carbonate up to 1½ feet long are present with quartz pebbles. The microcline of this formation has dimensions equal to those of the underlying Wilmington gneiss, but is not generally flattened into lenticular masses. Individual microcline grains are commonly as large as 1.5 cm. They have smooth and rounded, to rough and irregularly shaped boundaries (Fig. 18). Muscovite is the most abundant mineral adjacent to larger microcline grains, although biotite, quartz, fine-grained microcline, and pyrite are commonly found in contact with them also. Biotite is light to dark green. Sphene is present in amounts up to 2 percent. Twinned albite grains are more common in this formation than in the Wilmington gneiss. Muscovite is present in relatively small amounts (up to 7 percent) and is rather dirty in appearance in thin-section. Modes are presented in Table 14. The Tyson formation within the area ranges in thickness from 0 to 200 feet.

### Hoosac Formation

*Name:* The term Hoosac schist was proposed by Pumpelly, Wolff, and Dale (1894, p. 59) for a sequence of albite schists stratigraphically overlying the conglomerate of the Vermont formation. The Hoosac schist, as originally defined, included also schist on the west slope of Hoosac Mountain which is considered by the writer to be the same as the Readsboro and Heartwellville schists of probable pre-Cambrian age. Therefore, the writer wishes to redefine Hoosac formation in the present paper to include the albite schist sequence with interbedded amphibolites lying stratigraphically between the Tyson conglomerate (Vermont formation) and the green schists and amphibolites of the Pinney Hollow formation (or Rowe formation of Emerson (1898a)). The base of the formation is placed at the first break in lithology below the lowest amygdaloidal amphibolite.

*Distribution:* The Hoosac formation is found in Stratton, central Dover, Wilmington, Marlboro, Halifax, and Whitingham townships. In this district (Pl. I) it forms a continuous band of strata. In addition, rocks mapped as Hoosac formation by reason of the presence of amygdaloidal amphibolites are present in western Whitingham in Atherton

Figure 18. Closeup view of Tyson conglomerate outcrop, south of hill marked by elevation 2025′, western Wilmington, showing rounded to angular microcline and albite pebbles.

Meadow, as well as in much of Readsboro and Stamford. Isolated patches of Hoosac schist may be present locally in Searsburg but their similarity to schist of the Readsboro formation combined with extremely complex structure made differentiation impossible in the time given to this study.

*Age and Correlation of the Hoosac formation:* The Hoosac formation is thought to be present on the eastern flank of the southern Green Mountains between Clarksburg Mountain and Heartwellville (Pl. I., Fig. 2). Here the Cheshire quartzite, a gray to buff, massive glassy quartzite of the western sequence wraps around the southerly plunging nose of the Green Mountain anticlinorium and strikes northeasterly. The Cheshire quartzite which has been observed by the writer to be interbedded in and along strike within a horizontal distance of only 50 to 100 feet, seems to form a facies variation with dark biotite-albite schists. This interfingering is well displayed from the vicinity of Sumner Brook north of Stamford Village, northeasterly to the West Branch of the Deerfield River at Heartwellville.

TABLE 15

MODES OF THE HOOSAC SCHIST MEMBER OF THE HOOSAC FORMATION

| | 1 | 2 | 3 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | 15 | 30 | .. |
| Chlorite | tr | tr | tr |
| Biotite | 10 | 15 | 12 |
| Oligoclase | .. | .. | 20 |
| Groundmass or grains | | | |
| Quartz | 64 | 52 | 40 |
| Muscovite | 7 | 3 | 15 |
| Epidote group | tr | tr | .. |
| Apatite | 3 | tr | tr |
| Magnetite and Ilmenite | 0.5 | tr | 3 |
| Sphene | tr | .. | .. |
| Zircon | tr | .. | tr |
| Carbonate | .. | .. | tr |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–3.0 | 0.5–4.0 | 0.5–5.0 |
| Groundmass or grains | 0.008–0.4 | 0.01–0.4 | 0.007–0.4 |
| Texture | S* | S | S |

1. J37/a Quartz-albite-biotite-muscovite schist (.5 mile due N of junction of Ellis Brook and Dover, West Dover Road, Dover).
2. L37/a Quartz-albite-biotite-muscovite gneiss (1.05 miles N 40° E of junction of Route 9 and Route 8, Wilmington).
3. L4/b Quartz-oligoclase-muscovite-biotite schist (on road, 0.85 mile NW of bench Mark 1907, southeastern Whitingham township).

*schistose

This sequence of interfingering beds of the Cheshire-Hoosac formation is overlain by fine-grained chlorite-muscovite-garnet-quartz schist which in all respects is lithologically similar to the Heartwellville green schists. Estimates of minerals in hand specimens suggest that chlorite is present in amounts of 5 percent; biotite, 5 percent; and garnet 5 to 7 percent, with muscovite and quartz making up the rest. In the field, it was noted that this schist, cropping out excellently on the slopes one-half mile north of Sumner Brook, is quite similar to the green schists of Allen Hill and the district to the east of that locality.

The strike of the green schist at this locality is parallel to the strike of the underlying Cheshire-Hoosac strata. If this green schist is not to be

Table 16

MODES OF THE HOOSAC SCHIST MEMBER OF THE HOOSAC FORMATION

| | 4 | 5 | 6 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | 10 | 10 | .. |
| Biotite | 19 | 20 | 15 |
| Chlorite | tr | .. | .. |
| Microcline | .. | 5 | .. |
| Oligoclase | .. | .. | 12 |
| Epidote | 7 | 7 | 7 |
| Groundmass or grains | | | |
| Quartz | 60 | 53 | 55 |
| Muscovite | 1 | 4 | 10 |
| Magnetite and Ilmenite | 2 | tr | 1 |
| Apatite | tr | tr | tr |
| Pyrite | tr | 1 | .. |
| Grain size in mm. | | | |
| Porphyroblasts | 0.3–1.5 | 0.4–3.0 | 0.5–2.0 |
| Groundmass or grains | 0.006–0.2 | 0.008–0.3 | 0.006–0.4 |
| Texture | G* | G | G |

4. L36/a Quartz-biotite-epidote-albite gneiss (1 mile N 40° W of Mt. Olga, Wilmington).
5. N5/a Quartz-biotite-albite-epidote-microcline gneiss (0.75 mile W of longitude 73°50′ and latitude 72°50′).
6. 188/B Quartz-biotite-oligoclase-muscovite gneiss (1.50 miles due W of crest of Stickney Hill, Whitingham).

*gneissose

correlated with the Heartwellville of Allen Hill, the alternative is to correlate it with the Pinney Hollow formation and to consider that it overlies the Cheshire-Hoosac beds conformably. The fact that these green schists form a syncline strongly overturned to the east, should be considered in any interpretation of their relationship to major structures.

Fossils from the Mendon formation of Clarksburg Mountain and Bennington have indicated a Lower Cambrian age (Walcott, 1888). If the overlying green schists of the eastern flank of Clarksburg Mountain are in thrust relationship to the Cheshire quartzite, and if the pre-Cambrian age of the Heartwellville schist is correct, then pre-Cambrian rocks have been moved over Lower Cambrian along the northern part of the Hoosac Thrust. If, however, this is a zone of facies changes wherein

TABLE 17

MODES OF THE HOOSAC SCHIST MEMBER
OF THE HOOSAC FORMATION

| | 7 | 8 | 9 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | .. | 5 | 10 |
| Oligoclase | 8 | 40 | .. |
| Biotite | 20 | 5 | 10 |
| Chlorite | 2 | 3 | 0.5 |
| Groundmass or grains | | | |
| Quartz | 62 | 46 | 56 |
| Muscovite | tr | tr | 15 |
| Epidote group | 7.5 | tr | 7 |
| Magnetite and Ilmenite | .. | tr | .. |
| Zircon | .. | .. | tr |
| Apatite | tr | .. | .. |
| Pyrite | tr | 1 | tr |
| Grain size in mm. | | | |
| Porphyroblasts | 0.2–2.0 | 0.3–3.0 | 0.4–4.0 |
| Groundmass or grains | 0.006–0.1 | 0.008–0.2 | 0.008–0.3 |
| Texture | G* | G | S** |

7. L27/c Quartz-biotite-feldspar-epidote gneiss (0.4 mile N 80° E of junction of Route 8 and Route 9, Wilmington).
8. L26/b Quartz-oligoclase-biotite-albite gneiss (.35 mile S 45° W of the junction of Wilmington, Dover, and Marlboro township line).
9. 46D/b Muscovite-biotite-epidote schist (1.76 miles N 85° E of intersection of Dover-Somerset township line and Stratton boundary, Stratton).

*gneissose
**schistose

albite schists and green schists of the Readsboro and Heartwellville formations have been deposited in conformable sequence on the Cheshire quartzite, then the entire sequence from Cheshire through at least Ottauquechee formations may represent an essentially continuous sequence of Cambrian and possibly also of Ordovician age.

*Description:* The Hoosac formation consists dominantly of medium- to coarse-grained muscovite-chlorite-biotite-garnet-quartz schists and interbedded amphibolites.

The albite schist is a fine- to coarse-grained, light- to dark-gray albite augen schist. They are generally rusty weathering schists in contrast to the more gneissic and less highly weathered rocks of the Readsboro formation. These rocks have a relatively high biotite content while muscovite is generally present in amounts from a trace to less than 10 to 20 percent. This is notably true in the lower part of the formation in which the albite has a characteristic blue tint. The rusty weathering seems to be due to leaching of the garnet and biotite. The size of the albite grains in this formation ranges from 0.15 mm. to 1 cm. Usually albite is untwinned but scattered grains of twinned albite are found in accessory amounts, particularly in the lower portion of the formation. Characteristic of the albite is the presence in it of many small inclusions, particularly garnet. The quartz is relatively free of impurities and as a result the albite is differentiated from quartz merely by inspection in thin section. Average grain diameter of albite in very coarse grained rocks is 3 mm., but in medium-grained rocks it is about 1.5 mm. Modes are presented in Tables 15, 16, and 17.

Quartz occurs both as very fine grained and very coarse grained material. In most of the rocks of this formation, a small percentage of quartz is present in the form of grains as coarse as 2.5 mm. The bulk of the quartz, however, is very fine grained and, as a rule, is the finest of either the major or minor constituents. Biotite of the albite schists is light to dark brown. Locally, however, light-green biotite is found. Chlorite is present in accessory amounts in all of the rocks in which brown biotite is found. In addition, it is present in rocks in which the more iron-rich, green biotite occurs. The average length of biotite grain is 0.25 mm. and, except for albite and quartz, is consistently the coarsest grained mineral of the albite schist.

Muscovite is usually a minor or secondary constituent but the amount varies from bed to bed. Some muscovite- and garnet-bearing albite schists have up to 30 to 45 percent muscovite. In the bulk of the albite schist, however, muscovite and albite do not exceed 15 percent. Garnets are found in the more schistose parts of the formations. They range in size from 0.3 mm. to 1 cm. In many cases, tiny garnets are present not only in the groundmass but are conspicuous as inclusions in the albite porphyroblasts.

Apatite is a persistent and conspicuous constituent of these albite schists. A high percentage of the apatite is oriented in the schist so that sections cut normal to the foliation show a hexagonal outline. Epidote is associated with muscovite and biotite. The maximum grain length is about 1.25 mm.

A variation of the albite schist just described is a graphitic muscovite-albite schist. The albite occurs in this rock as small grains whose long diameter rarely exceeds 2 mm. and whose average diameter is about 2.6 mm. It is a fine grained rusty- and yellow-weathering schist, which on fresh surfaces is dark gray to black. It has a poikiloblastic texture both in thin section and in outcrop. Muscovite is present in fine to coarse grains in amounts up to 30 percent. Throughout the rock graphite is scattered as a very fine dust. Magnetite and pyrite occur as larger grains. Tourmaline is present in amounts up to 1 percent and forms grains which average about 1 mm. in length. It is a brown tourmaline, many grains of which have nuclei of aquamarine color. This dark schist is usually present most conspicuously in the lower part of the stratigraphic section.

Rocks of the Hoosac formation are well displayed on most of the slopes east of Route 8 north of West Dover. They are also well exposed but complexly folded east and northeast of Wilmington Village.

The Hoosac formation ranges in thickness from 700 to 2000 feet. It is thinnest in the vicinity of the Sadawga Pond dome and this is attributed to structural thinning.

### *Turkey Mountain member of the Hoosac formation*

*Name:* The term, Turkey Mountain amphibolite, was proposed by Rosenfeld (1954) for a distinctive and relatively thick amygdaloidal amphibolite in the upper part of the albite schist sequence. The type locality of this member is in Little Turkey Mountain Brook on the southern end of Turkey Mountain in the western part of the Saxtons River quadrangle.

*Distribution:* It is present as a continuous map unit in the northern third of the area and has been traced from the type locality by Rosenfeld, Thompson, and the writer to the northern part of the area. The Turkey Mountain member is present throughout the rest of the area as a discontinuous but distinctive unit. Other thin, discontinuous, amygdaloidal amphibolite beds are present in the Hoosac schist. These may be a part of the Turkey Mountain member which appear at the surface as a result of intricate folding and thinning in the central and southern portions of the area.

*Description:* The Turkey Mountain amygdaloidal amphibolite is a dense, dark-green epidote-amphibolite spotted gray to dark green by scattered mineral ovoids. The feature that distinguishes them and the other amphibolites of the Hoosac formation from those higher in the Paleozoic sequence is the presence of the so-called "amygdules". These are generally ellipsoidal ovoid aggregates 1 to 7 mm. long consisting of

TABLE 18

MODES OF TURKEY MOUNTAIN MEMBER
OF THE HOOSAC FORMATION

| | 1 | 2 | 3 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | 15 | 15 | 20 |
| Hornblende | 50 | 55 | 35 |
| Groundmass | | | |
| Quartz | 13 | 10 | 40 |
| Biotite | .. | .. | 1 |
| Chlorite | 2 | 5 | tr |
| Garnet | .. | .. | .. |
| Epidote | 15 | 15 | 2 |
| Magnetite and Ilmenite | 4 | 2 | 2 |
| Apatite | tr | .. | .. |
| Sphene | .. | tr | .. |
| Pyrite | 1 | .. | tr |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–5.0 | 0.5–5.0 | 0.5–4.0 |
| Groundmass | 0.02–0.4 | 0.05–0.4 | 0.02–0.4 |
| Texture | G* | G | S** |

1. 47 D/c. Actinolite-quartz-epidote amphibolite (4 miles N75° E from intersection of Blue Brook and Handle Road, Stratton).
2. 32 H/d. Quartz-hornblende-epidote amphibolite (0.8 mile NE of intersection of North Branch of Deerfield and Blue Brook, Dover).
3. 66 D/a. Actinolite-epidote-quartz-chlorite amphibolite (0.9 mile SE of West Dover Center, Ellis Brook near the Dover-West Dover Road).

*granulose
**schistose

very fine-grained quartz and albite filled with a central nucleus of epidote and hornblende grains. Accessory amounts of magnetite and more sparsely scattered apatite are also present in the "amygdules". Modes of the amphibolite are presented in Tables 18 and 19.

The amphibole of the Turkey Mountain member and of the other amygdaloidal amphibolites is a dark green variety and in composition is actinolitic-hornblende. It ranges in long diameter from 0.08 to 5mm. Average grain sizes differ from locality to locality but generally average 0.1 to 0.7 mm. Epidote is a major constituent of these amphibolites. It occurs as tiny grains which average about 0.08 mm. in diameter. The largest grains of epidote observed in thin section were 0.05 mm. On the

TABLE 19

MODES OF TURKEY MOUNTAIN MEMBER
OF THE HOOSAC FORMATION

| | 4 | 5 | 6 |
|---|---|---|---|
| Porphyroblasts | | | |
| Albite | 15 | 17 | 20 |
| Hornblende | 45 | 48 | 35 |
| Groundmass | | | |
| Quartz | 15 | 17 | 27 |
| Chlorite | 1 | 3 | .. |
| Garnet | 10 | .. | .. |
| Epidote | 10 | 10 | 10 |
| Magnetite and Ilmenite | 4 | 5 | 7 |
| Apatite | tr | tr | tr |
| Pyrite | tr | .. | 1 |
| Grain size in mm. | | | |
| Porphyroblasts | 0.5–3.0 | 0.5–3.0 | 0.5–4.0 |
| Groundmass or grains | 0.02–0.4 | 0.03–0.4 | 0.02–0.4 |
| Texture | G* | G | G |

4. 472 B. a. Actinolite-quartz-epidote amphibolite (0.6 of a mile due N of Route 9 on Marlboro-Wilmington township boundary).
5. M 34/c. Hornblende-quartz-epidote amphibolite (0.3 mile S 55° W from intersection No. 9 Brook and Route 8, Whitingham).
6. L 28/b. Hornblende-quartz-epidote amphibolite (0.6 mile N 85° E of intersection of Route 9 and Route 8, Wilmington).

*granulose

other hand, epidote grains up to 1 mm. are present in the coarse epidote knots and stringers which are present locally in the northern part of the area.

Quartz in these amphibolites is variable in amount but is uniformly fine grained. It ranges in amount from 10 to 50 percent. The average grain diameter is less than 0.15 mm. Chlorite is present in minor or accessory amounts and generally forms rounded or irregular masses.

Magnetite and ilmenite constitute about 2 to 7 percent of the rocks. They are scattered more or less uniformly throughout as fine grains or aggregates which average 0.02 to 0.4 mm. in diameter. Pyrite is present in accessory amounts.

In certain of the amygdaloidal amphibolites of the Hoosac formation minor variations in the mineral composition of the "amygdules" from those of the Turkey Mountain member are found.

Certain amphibolites, as for example in Ellis Brook (Pl. I & Table 18, specimen 3), contain "amgydules" showing well-formed, garnet-shaped, outlines. The "amgydules" average 3 mm. in diameter. In thin section they are seen to consist of a rim of quartz and albite with a large central nucleus of isolated grains of garnet as much as 2 mm. in diameter. Generally, however, the nuclei are a fraction of a mm. in diameter. Between the outer rim of albite and quartz and the central nucleus of garnet is a thin zone filled with chlorite, epidote, magnetite, and ilmenite. In thin section, the contact of the garnet-shaped "amgydules" with the surrounding amphibolite is sharp. In certain other amgydaloidal amphibolites the percentage of amphibole is greater and the amount of albite less than in the Turkey Mountain member. Biotite is not present in the Turkey Mountain member ordinarily, but does occur to the extent of 8 percent in some of the amgydaloidal amphibolites mapped as part of the Readsboro formation. Modes are presented in Tables 18 and 19.

The Turkey Mountain member of the Hoosac formation in the northern portion of the area is about 300 feet thick. South of Higley Hill, this member is generally much thinner and in places does not crop out at the surface. In the vicinity of Lake Rayponda, it attains a thickness of approximately 300 feet which may be the result of tectonic thickening.

### Mendon Formation and Cheshire Quartzite

*Name:* The Mendon formation was named by Whittle (1894) and the Cheshire by Emerson (1898a).

*Distribution:* The Mendon formation crops out in a synclinal fold in the central part of the Green Mountains in Woodford Village. This synclinal fold is approximately 5 miles in length. The Mendon formation crops out on the west flank of the Green Mountain core and the base of this forms the westernmost boundary on the geologic map (Pl. I). This boundary on the west flank of the Green Mountains has been mapped by Professor John A. MacFadyen, Jr. (1956).

The Cheshire quartzite crops out on the east flank of the Green Mountain core in Stamford and extends from the Massachusetts boundary as far northeast as Heartwellville. Beds of the Cheshire are present in the Woodford syncline. The Cheshire quartzite crops out as a continuous band and as a conspicuous ridge former in the extreme western portion of the map area lying west of the boundary mapped by MacFadyen (Pl. I).

*Description:* The Mendon formation is represented in the Woodford syncline by the Moosalamoo member consisting of a black, sericite-biotite-chloritoid phyllite with thin interbeds of gray quartzite. It is sandy phyllite having thin layers of quartz. Thin beds of Cheshire quartzite seem to be present in the upper portions of the syncline, although these quartzites may represent quartzites interbedded in the Moosalamoo member of the Mendon formation.

*Age of the Mendon formation:* Whittle (1894), Keith (1932), and Fowler (1950) have considered the Mendon to be of late pre-Cambrian age. MacFadyen (1956) considers it to be of lower Cambrian age, for the reason that Walcott (1888) found fragments of *Olenellus* about 100 feet above the pre-Cambrian contact near North Adams, Massachusetts. The rock in which it was found is a quartzitic graywacke and is stratigraphically beneath a band of black phyllite considered to be the equivalent of the Moosalamoo.

### Pinney Hollow Formation

*Location:* The Pinney Hollow formation is found in southwestern Wardsboro, central Dover, Marlboro, southeastern Wilmington, northwestern Halifax, and southeastern Whitingham. It crops out as a band about one-quarter of a mile broad south of Dover. In Dover and Wardsboro, however, it is about a mile broad.

In the western half of the area, vast tracts of rugged topography are underlain by rocks of the Heartwellville schist which are lithologically similar to the green and black schists of the Pinney Hollow and Ottauquechee formations, respectively. In the Mt. Pisgah, Haystack Mountain, Readsboro Falls, and eastern Stamford districts, it is clear that in present sequence these green and black schists are stratigraphically above the lower albite schist and gneiss of the Readsboro formation proper. There are, however, certain dissimilarities between the Haystack Mountain schists and the Pinney Hollow-Ottauquechee sequence that make correlation between them very improbable at present. However, such a possibility of their correlation cannot be definitively excluded although their possible equivalence should be investigated further.

The Pinney Hollow formation contains a number of amphibolite beds; the prominent Chester amphibolite and the abundant amphibolites in the Ottauquechee formation of the eastern part of the area are not present in the green and black schist sequence of the Heartwellville schists. For these reasons and more especially because of structural relationships, the Heartwellville is regarded as having a much lower stratigraphic position than the Pinney Hollow-Ottauquechee sequence.

TABLE 20

MODES OF THE PINNEY HOLLOW FORMATION

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Porphyroblasts | | | | |
| Chlorite. . . . . . . . . . | 7 | 10 | 5 | 8 |
| Garnet . . . . . . . . . . | 8 | 10 | 10 | 6 |
| Magnetite and Ilmenite. . | 2 | 2 | 3 | .. |
| Groundmass or grains | | | | |
| Quartz . . . . . . . . . . | 61 | 40 | 50 | 45 |
| Albite . . . . . . . . . . | 2 | 3 | 3 | 3 |
| Muscovite. . . . . . . . . | 20 | 35 | 30 | 20 |
| Apatite. . . . . . . . . . | tr | tr | tr | tr |
| Tourmaline . . . . . . . . | tr | .. | tr | tr |
| Sphene . . . . . . . . . . | .. | tr | .. | tr |
| Zircon . . . . . . . . . . | .. | tr | tr | tr |
| Oligoclase. . . . . . . . | .. | .. | .. | 15 |
| Grain size in mm. | | | | |
| Porphyroblasts . . . . . | 0.5–13.0 | 0.5–2.0 | 0.5–2.0 | 0.5–1.0 |
| Groundmass or grains. . . | 0.03–0.4 | 0.02–0.4 | 0.01–0.4 | 0.01–0.4 |
| Texture. . . . . . . . . . | S* | S | S | S |

*Pinney Hollow formation*

1. 482B/a. Muscovite-chlorite-garnet-quartz schist (.75 mile west of South Wardsboro on West Wardsboro Road, southern border of Londonderry quadrangle).
2. 50D/c. Muscovite-chlorite-garnet-quartz schist (6.5 miles southeast of Stratton-Wardsboro-Dover township junction at el. 2550).
3. 47D/k. Muscovite-chlorite-garnet-quartz schist (0.95 mile eastnortheast of 2887 hill on Stratton-Wardsboro line).
4. 93D/a. Muscovite-garnet-chlorite-oligoclase schist (0.5 mile northeast of Snow Cemetery, on west slope of 2398 hill).

*schistose

*Name:* The name, Pinney Hollow formation, was proposed by Perry (1928, p. 24) for the sequence of pale-green, chlorite schists with interbedded amphibolites which lie stratigraphically between the upper muscovite schists of the Hoosac formation and the black schists of the Ottauquechee formation.

*Lithology and Thickness:* The Pinney Hollow formation consists of pale-green, well-foliated chlorite schists with thin epidote amphibolite and amygdaloidal epidote-amphibolite beds. The chlorite schists make up about 90 percent of the sequence and the amphibolites comprise the remaining 10 percent.

TABLE 21

MODES OF THE PINNEY HOLLOW FORMATION
AND THE CHESTER AMPHIBOLITE

| | *Pinney Hollow formation* | | | *Chester amphibolite* |
|---|---|---|---|---|
| | 5 | 6 | 7 | 1 |
| Porphyroblasts | | | | |
| Chlorite | 3 | 1 | 5 | 5 |
| Garnet | 80 | 40 | 5 | 3 |
| Magnetite and Ilmenite | .. | 11 | tr | .. |
| Hornblende | .. | 2 | .. | 70 |
| Groundmass or grains | | | | |
| Quartz | 7 | 34 | 51 | 15 |
| Albite | .. | .. | 5 | 5 |
| Muscovite | 10 | 20 | 15 | .. |
| Biotite | tr | 2 | 12 | tr |
| Epidote | tr | tr | .. | 5 |
| Apatite | tr | tr | tr | tr |
| Tourmaline | .. | .. | .. | tr |
| Sphene | .. | .. | .. | 2 |
| Zircon | tr | tr | tr | .. |
| Andesine | tr | .. | .. | .. |
| Rutile | .. | tr | .. | .. |
| Pyrite | .. | .. | .. | tr |
| Microcline | .. | .. | 7 | .. |
| Grain size in mm. | | | | |
| Porphyroblasts | 0.5–2.0 | 0.5–3.0 | 0.5–4.0 | 0.5–0.2 |
| Groundmass or grains | 0.01–0.4 | 0.02–0.4 | 0.02–0.4 | 0.02–0.4 |
| Texture | S* | S | S | S |

*Pinney Hollow formation*

5. 2H/b. Coarse garnet-muscovite-chlorite-quartz schist (0.5 mile south southwest of Mt. Olga, Wilmington).
6. K54/d Garnet-chlorite-biotite-muscovite-quartz schist (1.75 miles southeast of Adams School on Green River, southernmost Marlboro township. Chemical analysis of chlorite presented in Table 30).
7. 043/b Chlorite-garnet-muscovite-biotite schist. (6.5 miles South of Jacksonville).

*Chester amphibolite*

1. 474B/b Actinolite-chlorite-epidote-garnet-quartz amphibolite (Sky View, Route 9, Hogback Mt., Marlboro).

*schistose

The chlorite schists are quite different in appearance from any part of the pre-Cambrian or Lower Paleozoic sequence below, with the exception of the lower part of a relatively thin but widespread band in the upper part of the pre-Cambrian sequence referred to as the Heartwellville schist. These rocks are, however, identical in appearance with some of the younger schists; notable are those of the Stowe and a part of the Moretown formation. Modes of the Pinney Hollow formation are given in Tables 20 and 21.

Chlorite usually constitutes only about 10 to 15 percent of the schist, but this is enough to give a distinctive, pale-green color to the rock. The chlorite is generally interlaminated with muscovite and thus imparts to the schist a uniform, light-green color. Some beds within the formation, on the other hand, contain chlorite in knots or blebs that give the rock a spotted appearance. Such schists, however, make up only a small percentage of the whole.

Coarse-grained muscovite is present in amounts from 40 to 80 percent. Garnet constitutes about 5 to 10 percent of the schist. Nearly every hand specimen is made up of layers that contain abundant garnets as well as those in which the garnets are large and widely spaced. The larger ones average about 0.05 to 0.1 cm. Quartz is the most abundant constituent of the schist and is concentrated in layers between the foliation planes.

Magnetite is an abundant and conspicuous minor constituent. Most hand specimens contain many euhedral magnetite crystals 1 to 2 cm. in diameter. In this section magnetite is abundantly present in the muscovite, chlorite, and garnet layers and forms elongate grains parallel to the foliation.

Epidote and biotite are present only locally and in minor or accessory amounts. Albite is generally not present and where it does occur it is a minor constituent. Accessory minerals are apatite, tourmaline, sphene, zircon, and rutile.

The banded epidote-amphibolites of the Pinney Hollow formation contain hornblende needles and epidote grains that generally form matted laminae separated by thin, granular quartz, and feldspar layers. The relative thickness of the bands ranges from a fraction of a millimeter to 2 to 3 cm. The color contrast between the white and dark bands emphasizes the pronounced deformational structures produced in these rocks. The detailed lithologic description of the amphibolites is given under the Chester amphibolite inasmuch as they are lithologically identical.

Figure 19. Chester amphibolite in northerly sloping surface, Hosley Hill, Jacksonville; looking southwesterly parallel to direction of plunge of fold axes. Shear-sense and the relation of these to the major folds of Stickney Hill is shown on Plate 1.

The thickness of the Pinney Hollow formation is between 1000 and 8000 feet in the less disturbed and attenuated sections. In the more highly folded localities, the thickness is between 300 and 800 feet.

## Chester Amphibolite

*Location:* The Chester amphibolite lies directly east of the Pinney Hollow formation. Consequently, it is found in Wardsboro, central Dover, Marlboro, Halifax, and southeastern Whitingham. It crops out as a band less than half a mile broad in the northern third of the area and is about a quarter mile broad in the southern two-thirds.

Figure 20. Northeasterly plunging fold in the Chester amphibolite near the axis of Hogback Mountain syncline, Sky View, Route 9, Hogback Mountain in Marlboro township. Hammer head points in direction of plunge of fold axis.

*Name:* The Chester amphibolite received its name from the type locality in Chester, Massachusetts (Emerson, 1898b, p. 78). Here the amphibolite is about 3200 feet broad and is a banded epidote amphibolite as it is in the Wilmington quadrangle.

*Lithology and Thickness:* The Chester amphibolite is the first thick epidote amphibolite above the chlorite-muscovite schists of the Pinney Hollow formation. It lies just below the dark muscovite schists of the Ottauquechee formation. The Chester amphibolite is lithologically the same as the thin epidote amphibolites of the Pinney Hollow formation. It is distinguishable from them only by its greater thickness and by its position between the chlorite schists below and the dark muscovite schists above.

The Chester amphibolite is a dark-green, well-foliated, fine-grained epidote amphibolite. The amphibole is a dark-green actinolitic hornblende. The average diameter of the amphibole is about 0.1 mm. Larger grains average about 0.25 mm., and a few needles 2 to 3 mm. in diameter and 1 cm. in length are scattered randomly throughout the rock.

Figure 21. Well-banded epidote-ankerite-actinolite amphibolite of Chester amphibolite, Route 9, Hogback Mountain. These folds are near the axis of the Hogback Mountain syncline. Watch in center gives scale.

Hornblende makes up between 50 and 80 percent of the amphibolite and in thin section is seen to form laminae separated by thin layers of quartz and albite. Epidote is concentrated in the hornblende layers. Chlorite in small amounts forms rounded masses and flakes that are randomly distributed throughout the rock. In the Cooper Hill district, Dover, ankerite and chlorite are present in greater amounts than elsewhere in the area. Here also hornblende decreases correspondingly in amount. Pyrite, apatite, tourmaline, sphene and rutile are accessory minerals.

In addition to amphibolite very thin beds of dark-gray to black muscovite schist and pale-green chlorite-muscovite schist are found locally in the formation. These beds are lithologically the same as those of the Ottauquechee and Pinney Hollow formations, respectively.

South of Dover the Chester amphibolite is between 300 and 600 feet thick. In Dover and Wardsboro it is about 600 feet thick.

## Ottauquechee Formation

*Location:* The Ottauquechee formation lies just east of the Chester amphibolite. It is found in Wardsboro, Dover, Marlboro, Halifax and

Figure 22. Closeup of bedding of highly contorted Chester amphibolite, Sky View, Hogback Mountain, Route 9, Marlboro, showing folds and associated slip cleavage.

southeastern Whitingham townships. In the southern two-thirds of the area the breadth of outcrop is between 400 and 800 feet. In Dover and Wardsboro, however, it attains a breadth of 1 to 2 miles.

*Name:* The rusty-weathering, dark, muscovite schist above the Chester amphibolite and below the chlorite-muscovite schists of the Stowe formation was first called Ottauquechee formation by Perry (1928, p. 27). It is named from exposures in the Ottauquechee River valley in central Vermont.

*Lithology and Thickness:* About 80 percent of the Ottauquechee formation is a rusty-weathering, monotonously uniform sequence of muscovite schists. It is dark-gray to coaly-black on fresh surfaces. Lithologically it is similar to the upper portion of the Heartwellville schist. It consists dominantly of muscovite and quartz with smaller amounts of garnet and fine-grained graphite. Albite is either lacking or is present in minor amounts.

Well-banded feldspathic-chlorite schist makes up about 15 percent of the formation. It consists of bands of fine-grained, granular quartz and albite alternating with laminae of biotite, muscovite and chlorite. The gray granulose bands are 1 to 6 mm. thick, whereas the micaceous lami-

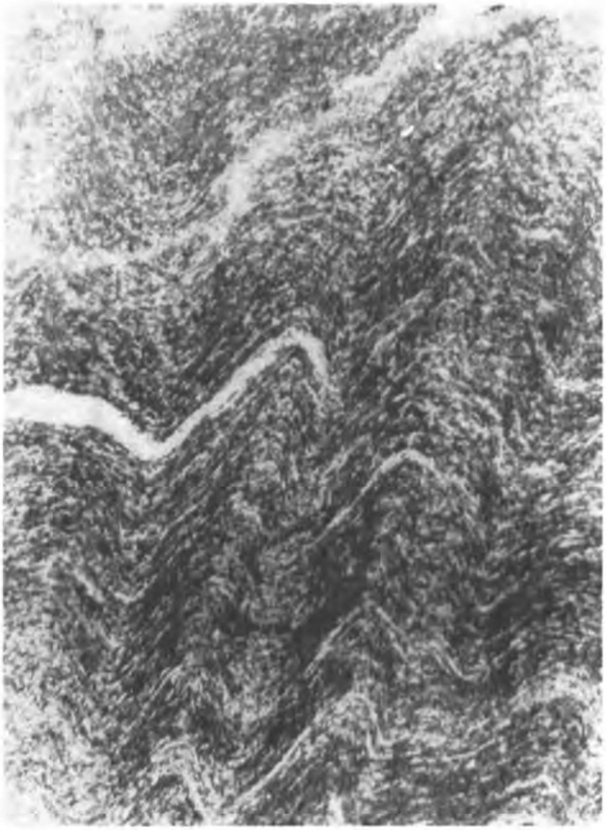


Figure 23. Photomicrograph of Chester amphibolite, Sky View, Hogback Mountain, Route 9, Marlboro. Same specimen as in Fig. 22. Dark bands are actinolite, chlorite, and magnetite and light layers are quartz. X4½.

nae have a thickness of a fraction of a millimeter up to 2 mm.

In this schist, brown biotite in crystals up to 2 mm. in diameter are commonly oriented at an angle to the foliation. Chlorite occurs as rounded balls and as rims around the garnet. Zircon with pleochroic halos is scattered through the chlorite. Tourmaline grains average about 0.08 mm. and are present throughout the schist. Pyrite is a conspicuous accessory mineral.

Thin, gray to black quartzites, thin chlorite-muscovite schists and banded epidote amphibolites constitute about 5 percent of the formation.

Figure 24. Highly contorted chlorite-garnet schist and amphibolite of the Stowe formation, Route 9, Hogback Mountain, Marlboro. Near axis of Hogback Mountain syncline with hammer handle pointing parallel to fold axis.

The Ottauquechee formation in Wardsboro and Dover is between 800 and 1200 feet thick. In the rest of the area it is reduced to a thickness of between 100 and 500 feet, partially as a result of structural thinning and partially also as a result of original conditions of sedimentation.

## *Ordovician Rocks*

### Stowe Formation

*Location:* The Stowe formation lies just east of the Ottauquechee formation throughout the area. It crops out in eastern Wardsboro and Dover, western Marlboro and Halifax and in the southeastern part of Whitingham. In Wardsboro and Dover the maximum breadth of outcrop is 1¾ miles. South of Dover it ranges in breadth from 200 to 1600 feet.

*Name:* Cady (1956, p. 1811) applies the name Stowe formation to a sequence of green chlorite schists and thin amphibolites exposed in the town of Stowe, in northern Vermont. This sequence lies stratigraphically above the black muscovite schists and feldspathic-biotite schists of the

Figure 25. Amphibolite at top of the Stowe formation. Small fault shows that upper beds moved northeasterly relative to beds below. This is the same movement sense as in associated spruce-tree folds. Roadcut on Molly Stark Trail, Marlboro.

Ottauquechee formation and below the banded biotite-quartz schists and amphibolites of the Moretown formation.

*Lithology and Thickness:* The lithologic character of the Stowe formation is identical with that of the Pinney Hollow formation described above. Chlorite, muscovite, garnet and quartz are present in about the same quantities as in the chlorite-muscovite schists of the Pinney Hollow formation. Chlorite is usually associated with biotite, muscovite and garnet. It occurs as rounded balls or as grains smaller in size than the muscovite of the same specimen. In some rocks chlorite forms crystals elongated in a direction transverse to the foliation. In such cases the orientation of trails or lines of magnetite and other minerals included in the chlorite is parallel to that of the same minerals outside the chlorite in the surrounding schist. Scattered through the chlorite are small zircon crystals surrounded by pleochroic halos. Magnetite makes up 2 to 3 percent of the rock. Garnets have an average diameter of 0.08 to 0.6 mm. Apatite and tourmaline are accessory constituents of nearly every specimen. Modes are presented in Table 22.

TABLE 22

MODES OF STOWE FORMATION

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Porphyroblasts | | | | |
| Albite | 10 | tr | .. | .. |
| Chlorite | 10 | 5 | 20 | 5 |
| Biotite | 1 | 10 | .. | .. |
| Hornblende | .. | .. | .. | 60 |
| Groundmass or grains | | | | |
| Quartz | 34 | 50 | 42 | 15 |
| Muscovite | 40 | 30 | 20 | .. |
| Garnet | 3 | 3 | 15 | .. |
| Epidote | .. | tr | .. | 10 |
| Magnetite and Ilmenite | 2 | 2 | 3 | tr |
| Apatite | tr | tr | .. | tr |
| Tourmaline | tr | tr | .. | .. |
| Sphene | .. | .. | .. | tr |
| Zircon | tr | tr | .. | .. |
| Oligoclase | .. | .. | .. | 10 |
| Rutile | .. | tr | .. | .. |
| Pyrite | .. | tr | .. | .. |
| Grain size in mm. | | | | |
| Porphyroblasts | 0.5–1.5 | 0.5–35.0 | 0.5–1.0 | 0.5–10.0 |
| Groundmass or grains | 0.02–0.4 | 0.03–0.4 | 0.01–0.4 | 0.01–0.4 |
| Texture | S* | S | S | S |

1. 463 B/m Muscovite-feldspar-chlorite-garnet schist (1.1 miles N 15° W of crest of Whites Hill, Dover).
2. 2F Quartz-muscovite-biotite-chlorite schist (0.4 mile due West of Adams School, Marlboro).
3. 478 B/e Quartz-muscovite-chlorite-garnet schist (1.1 miles due North of Goose City, Dover).
4. 21 D/b Hornblende-quartz-plagioclase-chlorite schist (in Bellows Brook, 1.3 miles N 30° E of Adams School, Marlboro).

*schistose

In the Cooper Hill district in Dover, the epidote amphibolite at the base of the formation is well enough developed to be mapped separately. South of Dover, however, it is very thin and irregularly developed and therefore is not mapped separately. Lithologically it is the same as the Chester amphibolite except that the ratio of the amount of chlorite to amphibole is generally higher in this rock than in the Chester amphibolite.

The Stowe formation in Wardsboro is between 500 and 1000 feet thick. South of Dover the formation is between 150 and 600 feet thick. A photograph of structure in a bed of the Stowe is shown in Fig. 25.

*Age of the Stowe formation:* Osberg (1956, p. 1820) has indicated that the rocks beneath the Stanbridge formation in the west flank of the Green Mountain-Sutton Mountain anticlinorium have been traced from southern Quebec into the Cambrian section of west-central Vermont. The rocks east of the Green Mountain-Sutton Mountain anticlinorium in Quebec have been traced southward into Vermont. The Beauceville of Ordovician age had been traced into the Moretown as mapped in north-central Vermont. The upper part of the Mansonville is in part the equivalent of the Stowe and therefore may be of Ordovician age. Stratigraphic relations beneath the Mansonville, the lower part of which has been traced into the Ottauquechee, are still obscure.

## Moretown Formation

*Location:* The Moretown formation makes up a broad band of rocks in the eastern third of the area. It is found in eastern Wardsboro and Dover, western Newfane, Marlboro, Halifax and southeastern Whitingham. Its breadth of outcrop in the northern half of the area is between 6000 and 11,000 feet. In the southern half, it is reduced to a breadth of 2900 to 8000 feet. It has a greater outcrop area than any other formation except the Hoosac formation and the pre-Cambrian formations.

*Name:* Moretown formation is the name proposed by Cady (1956, p. 1812) for a sequence of mica-quartzites, quartzose mica schists and amphibolites in the type locality on Mad River, Moretown, in northern Vermont. This sequence is stratigraphically above the chlorite-muscovite schists and thin amphibolites of the Stowe formation and below the black slates and metamorphosed volcanics of the Cram Hill formation.

*Lithology and Thickness:* The Moretown formation comprises a distinctive sequence of gray, well-banded, mica-quartz schists and banded epidote amphibolites. The mica-quartz schists make up about 80 per cent and the amphibolites about 20 percent of the formation.

Grouped under the general heading of mica-quartz schists are several varieties. The most distinctive of these is a well-banded, biotite-muscovite-quartz schist. The bands consist of quartz which alternates with thinner laminae of biotite and muscovite. The quartz bands range in thickness from 0.02 to 5.0 mm. The quartz is generally very fine grained, more so than that of the older rocks of the area. Biotite is usually brown but is locally green and is the coarsest grained mineral in the rock.

Figure 26. Outcrop area of Moretown formation looking NNE from Molly Stark Trail (Route 9), one mile west of Wilmington-Brattleboro quadrangle. Central mountain is on the right in middle distance, Stratton Hill on extreme left, Newfane Hill in center, and Saxtons River quadrangle area in far distance.

It has an average length of 0.1 mm., but in some rocks is 3 mm. long. In those micaceous quartz schists in which the banding is most pronounced, biotite is more abundant than muscovite. In those mica-quartz schists that are less perfectly banded, muscovite is more abundant than biotite. Some of these schists are further distinguished from simple mica-quartz schists by the presence of chlorite fasciculites up to 2 inches long. These fasciculitic clusters of chlorite lie in the plane of the foliation and are associated with the micaceous layers.

Other mica-quartz schists are dark-gray to black because of fine-grained graphite disseminated through the rock. These dark schists generally have more muscovite and biotite than the lighter colored schists. Garnets are usually less than 2 mm. in diameter and are only a minor constituent of the rock. Modes are given in Tables 23, 24 and 25.

The bulk of the amphibolites of the Moretown formation are banded epidote amphibolites of the same type found in many of the older formations. The mineralogical composition and appearance in outcrop is the same as that of the amphibolites given above under Chester amphibolite. One point, however, in which some of these amphibolites differ

TABLE 23

MODES OF MORETOWN FORMATION

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Porphyroblasts | | | | |
| Biotite | 5 | 5 | 6 | 3 |
| Chlorite | 2 | tr | tr | 2 |
| Garnet | 2 | 2 | 3 | 2 |
| Groundmass | | | | |
| Quartz | 85 | 89 | 89 | 50 |
| Muscovite | 4 | 3 | 1 | 41 |
| Epidote group | 1 | tr | tr | .. |
| Magnetite and Ilmenite | 1 | tr | 1 | 1 |
| Apatite | tr | tr | tr | tr |
| Tourmaline | .. | .. | tr | tr |
| Sphene | tr | tr | tr | .. |
| Zircon | tr | tr | tr | tr |
| Average grain size in mm. | | | | |
| Porphyroblasts | 0.4 | 0.1–3.0 | 0.1–3.0 | 1.0–3.0 |
| Groundmass or grains | 0.1–0.2 | 0.1–0.2 | 0.1–0.2 | 0.1–0.4 |
| Texture | S* | S | S | S |

1. 485B/a Micaceous quartz schist (North of Blodgett Brook and 0.25 mile southeast of Dover Village).
2. 4F/a Micaceous quartz schist (0.2 mile northwest of junction of Rock River and Hunter Brook along road, Newfane).
3. 4F/b Same specimen; section normal to 2.
4. 23D/c Muscovite-garnet-quartz-chlorite schist (on hill 0.75 mile northeast of Adams School and southwest of Bellows Brooks, Marlboro).

*schistose

from the older is that these contain sphene in amounts up to 2 percent and magnetite is absent. Thin, coarse-grained garnet amphibolites are found in every part of the formation but they constitute only a few percent of the amphibolite sequence. Garnets 1½ cm. in diameter are not uncommon.

The Moretown formation ranges in thickness from 500 to 5000 feet. It is thinnest in northern Halifax and thickest in southeastern Whitingham. North of Halifax the full thickness of the formation is not found in the Wilmington area, but the upper part occupies the western part of the adjacent Brattleboro quadrangle.

TABLE 24

MODES OF THE MORETOWN FORMATION

| | 5 | 6 | 7 | 8 |
|---|---|---|---|---|
| Porphyroblasts | | | | |
| Biotite | 7 | 5 | 5 | 10 |
| Chlorite | tr | tr | 3 | 1 |
| Garnet | 3 | 1 | 2 | 3 |
| Groundmass | | | | |
| Quartz | 79 | 88 | 50 | 66 |
| Muscovite | 10 | 5 | 40 | 20 |
| Epidote group | .. | tr | .. | .. |
| Magnetite and Ilmenite | 1 | 1 | tr | tr |
| Apatite | tr | tr | tr | tr |
| Tourmaline | tr | tr | tr | tr |
| Sphene | tr | tr | .. | tr |
| Zircon | tr | tr | .. | tr |
| Graphite | .. | .. | .. | tr |
| Average grain size in mm. | | | | |
| Porphyroblasts | 2.0–30.0 | 2.0–7.0 | 0.5–2.0 | 1.5 |
| Groundmass or grains | 0.1–0.4 | 0.01–0.2 | 0.1–0.3 | 0.1–0.4 |
| Texture | S* | S | S | S |

5. K68/d Garnet-chlorite-quartz schist (0.5 mile north of Valley School and Route 112, Halifax).
6. 31W/h Microcline-biotite-quartz schist (0.07 mile south of intersection of Brook River and Hunter Brook, Newfane).
7. 84D/f Muscovite-biotite-quartz-chlorite schist (0.9 mile southeast of three corners of Newfane, Dover and Marlboro, on west slope of Stratton Hill, Marlboro).
8. 31W/f Muscovite-quartz-garnet-chlorite schist (1.01 miles southwest of intersection of Brook River and Hunter Brook, Newfane).

*schistose

## CRAM HILL FORMATION

*Location:* The Cram Hill formation is a northeasterly trending band in central Halifax and the extreme southeastern part of Whitingham. It ranges in breadth of outcrop from one to nearly two miles.

*Name:* Cram Hill formation is the name applied by Currier and Jahns (1941, p. 1493) to a sequence of black micaceous slates, amphibolites and gneisses exposed at Cram Hill in central Vermont. These rocks are stratigraphically above the upper micaceous quartz schist bed of the More-

TABLE 25

MODES OF THE MORETOWN FORMATION

| | 9 | 10 | 11 | 12 |
|---|---|---|---|---|
| Porphyroblasts | | | | |
| Biotite | 15 | .. | . | .. |
| Chlorite | tr | 3 | .. | 10 |
| Garnet | 5 | .. | .. | 2 |
| Hornblende | .. | 65 | 45 | .. |
| Groundmass | | | | |
| Quartz | 45 | 15 | 35 | 56 |
| Muscovite | 35 | tr | .. | 30 |
| Epidote group | .. | 15 | 10 | .. |
| Magnetite and Ilmenite | tr | .. | .. | 2 |
| Apatite | tr | tr | tr | tr |
| Tourmaline | tr | .. | .. | tr |
| Sphene | tr | 2 | tr | .. |
| Zircon | tr | .. | .. | .. |
| Graphite | tr | .. | .. | .. |
| Albite | .. | .. | 10 | .. |
| Average grain size in mm. | | | | |
| Porphyroblasts | 3.0 | 2.0–3.0 | .. | 0.5–2.0 |
| Groundmass or grains | 0.1–0.4 | 0.1–0.5 | 0.1–1.0 | 0.1–0.4 |
| Texture | S* | S | G** | S |

9. K15a Muscovite-quartz-biotite-chlorite schist (100 feet west of junction of Brook River and Hunter Brook, Newfane).
10. 2D/k Schistose actinolite-epidote-chlorite-quartz amphibolite (0.15 mile north of junction with Rock River, in Adams Brook, Newfane).
11. 2D/e Massive amphibolite (in Adams Brook, 1.45 miles due N of East Dover, Newfane, Vt.).
12. 35D/a Muscovite-chlorite-biotite-garnet-quartz schist (0.4 mile west-northwest of East Dover).

*schistose
**gneissose

town formation and below the Shaw Mountain unconformity. In the Wilmington-Woodford area the Cram Hill formation is overlain directly by Northfield slate because the Shaw Mountain formation is absent here.

*Lithology and Thickness:* The Cram Hill formation is a varied complex of black slates, banded greenstones, massive amphibolites, gneisses and feldspathic-muscovite schists.

The black slates (Ocs) constitute about 40 percent of the formation

TABLE 26

MODES OF THE CRAM HILL FORMATION

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Porphyroblasts | | | | |
| Albite | .. | .. | 15 | 15 |
| Biotite | .. | tr | .. | .. |
| Chlorite | .. | 25 | 2 | 2 |
| Muscovite | 35 | .. | .. | .. |
| Pyrite | 1 | .. | .. | tr |
| Groundmass or Grains | | | | |
| Quartz | 54 | 66 | 27 | 33 |
| Garnet | .. | 3 | .. | tr |
| Hornblende | .. | 5 | 50 | 40 |
| Epidote group | tr | 1 | 3 | 5 |
| Magnetite and Ilmenite | .. | 10 | .. | .. |
| Sphene | .. | .. | 3 | 5 |
| Graphite | 2 | .. | .. | .. |
| Unidentifiable | 8 | .. | .. | .. |
| Grain size in mm. | | | | |
| Porphyroblasts | 1.0–5.0 | 0.5–1.0 | 1.0–6.0 | 1.0–5.0 |
| Groundmass or grains | 0.05–0.3 | 0.05–0.4 | 0.1–0.5 | 0.05–0.3 |
| Texture | S1* | S** | S | G*** |

1. N 43/C Quartz-muscovite-graphite slate (1 mile S 85° E of West Halifax Center, Halifax).
2. 33 D/b Quartz-chlorite-actinolite-magnetite amphibolite (1.5 miles due S. of intersection of North River and Hager Brook, Halifax).
3. 30 D/i Hornblende-feldspar-chlorite schist (.4 miles N 20° W from junction of Fowler Road and Panel Hill Road, Halifax).
4. 13 D/m Quartz-actinolite-feldspar-epidote amphibolite (1.1 miles S 85° E of Blue Mt., Halifax, in Brattleboro Quadrangle).

*slaty
**schistose
***granulose

and are stratigraphically above the volcanics (Ocb) mapped as the Barnard gneiss. They are extremely fine-grained, rusty-weathering, fissile rocks which, on fresh exposures, are jet black. The only mineral large enough to be recognized in hand specimens is pyrite. In thin section the slate is seen to consist of layers of very fine-grained quartz between finer grained muscovite. The average diameter of the quartz grains is 0.015 mm. Only a very few muscovite grains are coarser than the quartz and these reach a length of 0.15 mm. Pyrite cubes as large as 0.5 mm.

TABLE 27

MODES OF THE CRAM HILL FORMATION

| | 5 | 6 | 7 | 8 |
|---|---|---|---|---|
| Porphyroblasts | | | | |
| Albite | .. | .. | .. | 57 |
| Chlorite | 1 | 1 | 3 | 7 |
| Muscovite | .. | .. | .. | 1 |
| Oligoclase | .. | 19 | 15 | .. |
| Pyrite | tr | tr | .. | .. |
| Groundmass or grains | | | | |
| Quartz | 10 | 8 | 34 | 35 |
| Hornblende | 65 | 60 | 60 | .. |
| Epidote | tr | 10 | .. | tr |
| Magnetite and Ilmenite | tr | .. | 3 | tr |
| Sphene | 5 | 6 | tr | .. |
| Grain size in mm. | | | | |
| Porphyroblasts | 0.5–1.0 | 0.5–1.0 | 0.5–2.5 | 1.0–5.0 |
| Groundmass or grains | 0.05–0.3 | 0.05–0.3 | 0.05–0.3 | 0.03–0.4 |
| Texture | G* | G | G | G |

5. 51B Hornblende-sphene-chlorite amphibolite (on Fowler Brook Road, 4 miles SE of junction of Pease Brook and Fowler Brook Road, Halifax).
6. 42B Hornblende-epidote-plagioclase-sphene-quartz amphibolite (.4 mile S 65° W of junction of Brook River and Sperry Brook, Halifax).
7. 13D Hornblende-quartz-chlorite amphibolite (.25 mile due west of crest of Blue Mountain, Halifax).
8. 16D/b Feldspar-quartz-chlorite-muscovite gneiss (1.2 miles N 45° E of West Halifax Center).

*granulose

are not uncommon but the average diameter is about 0.3 mm. The slate is jet black because of abundant graphite. Modes are given in Tables 26 and 27.

Massive epidote-amphibolite makes up about 25 percent of the formation. In outcrop as in thin section they are most distinctive in appearance. These rocks comprise two types, the massive porphyritic and non-porphyritic amphibolites.

Massive porphyritic amphibolites are dark-green rocks that contain many buff to gray feldspar porphyroblasts. These are albite and have a large central nucleus consisting of numerous fine-grained zoisite grains. Fine-grained sphene is present in amounts from 3 to 5 percent and is associated with amphibole. Magnetite is associated with pyrite but is

Figure 27. Boudinage structure in amphibolite bed of the Cram Hill formation. Boudin-line nearly parallel to strike of bed and hammer head (0.4 mile due N of Valley School, southern Halifax).

absent from those rocks containing sphene. Quartz is very fine grained and fills the interstices between amphibole grains.

Massive non-porphyritic amphibolites are dark gray to dark green. They consist dominantly of hornblende and albite. They, as well as the porphyritic amphibolites, occur as thin beds or small lenticular bodies in the black slates described above. The green hornblende grains are rarely less than 1 mm. long and average 1.5 mm. They are randomly oriented and the interstices are filled with plagioclase and quartz. Associated with the hornblende is abundant sphene. Modes are presented in Tables 26, 27 and 28.

Light green amygdaloidal amphibolites are prominent in the amphibolite sequence of this formation. They consist largely of hornblende with lesser amounts of interstitial quartz. The "amygdules" are lenticular masses of coarse-grained chlorite that average about 0.7 mm. in long diameter. Modes are given in Tables 26, 27 and 28.

Thin beds of coarse-grained light-green gneiss constitute about 10 per cent of the Cram Hill formation. The gneiss consists dominantly of coarse-grained, twinned albite grains that average about 0.7 mm. in diameter. The light green color is imparted to the rock by minor amounts of fine-grained chlorite. It forms a lacework about each feldspar grain and in addition occurs as grains scattered randomly among the quartz grains. Quartz is relatively fine grained and occupies the interstices between the much larger plagioclase grains.

Other important rocks in the Cram Hill formation are light- to dark-gray, banded, feldspathic-muscovite schists. These rocks contain in addition small amounts of chlorite and brown biotite, and locally also graphite and pyrite. The micaceous minerals are concentrated as thin folia separated by thicker layers of granular feldspar and subordinate quartz. In some of these rocks muscovite forms large grains about 0.7 mm. in diameter. The average diameter of biotite is about 0.10 mm. Quartz and feldspar are extremely fine grained and average about 0.03 mm. in diameter. Zoisite has about the same grain size as quartz and is present in amounts up to 20 percent.

Of the feldspathic muscovite schist a pebble-bearing rock is particularly noteworthy. The schist is light-gray and is composed dominantly of muscovite and fine-grained feldspar with small amounts of biotite and chlorite. Rather uniformly distributed throughout the schist are lenticular blue-quartz pebbles as long as 1 cm. The Cram Hill formation is of the order of 4000 to 6000 feet in thickness.

## *Siluro-Devonian Rocks*

### NORTHFIELD SLATE

*Location:* The lower part of the Northfield slate occupies the extreme southeastern part of the area in Halifax. Here it has a maximum breadth of outcrop of 9000 feet.

TABLE 28

MODES OF CRAM HILL FORMATION AND NORTHFIELD SLATE

| | *Cram Hill formation* | | | *Northfield slate* |
|---|---|---|---|---|
| | 9 | 10 | 11 | 1 |
| Porphyroblasts | | | | |
| Albite | 25 | 43 | 20 | .. |
| Biotite | 1 | 7 | .. | tr |
| Chlorite | 3 | .. | 10 | tr |
| Garnet | .. | .. | .. | 7 |
| Muscovite | .. | 10 | 15 | 35 |
| Pyrite | .. | tr | .. | .. |
| Quartz pebbles | .. | .. | .. | 5 |
| Groundmass or grains | | | | |
| Quartz | 67 | 20 | 50 | 58 |
| Hornblende | 3 | .. | .. | .. |
| Epidote group | tr | 20 | tr | .. |
| Magnetite and Ilmenite | 1 | .. | tr | tr |
| Graphite | .. | .. | .. | tr |
| Grain size in mm. | | | | |
| Porphyroblasts | 0.5–2.0 | 0.5–1.5 | 0.5–8.0 | 1.0–4.0 |
| Groundmass or grains | 0.05–0.4 | 0.04–0.4 | 0.04–0.4 | 0.05–0.3 |
| Texture | G* | S** | S | P*** |

9. 13D/k Quartz-feldspar-hornblende-chlorite gneiss (1.8 miles N 25° E of West Halifax center on slope of Blue Mt., Halifax).
10. 30D/j Feldspar-quartz-epidote-muscovite-biotite gneiss (1.4 miles N 20° W from junction of Fowler Road and Panel Hill Road, Halifax).
11. 15D/d Quartz-feldspar-chlorite-muscovite schist. (0.8 mile S 35° W. of Blue Mt., Halifax, in Brattleboro Quadrangle).

Northfield slate

1. N43/a. At Bench Mark 947, 0.15 mile northwest of junction of East Branch of the North River, Halifax.

*granulose
**schistose
***phyllitic

*Name*: The Northfield slate comprises a group of slates as redefined by Currier and Jahns (1941, p. 1501) which lie between the older Shaw Mountain formation and the younger calcareous rocks of the Waits River limestone. The Northfield slate grades upward into the Waits River limestone, the difference between the two being in the predominance of slate in the former and of impure brown limestones in the latter.

These slates are correlated with the Goshen formation of Massachusetts named by Emerson (1898b) for the sequence of dark muscovite phyllites exposed in the Goshen dome in Goshen, Massachusetts.

The Northfield slate of the Wilmington-Woodford area lies unconformably on the rocks of the Cram Hill formation and grades upward into the limestones of the Waits River formation. However, the absence here of the Shaw Mountain formation and its presence to the northeast, and the presence of conglomerate and cummingtonite rocks to the southwest at approximately this horizon, suggests that the Wilmington-Woodford area was emergent as the Shaw Mountain formation was deposited to the southwest and northeast. The geological map of the Hawley quadrangle (Emerson, 1898a, 1898b) suggests that the Goshen (Northfield slate) was deposited on an erosion surface bevelling the beds of the Hawley formation.

*Lithology and Thickness:* The Northfield slate consists of a relatively uniform sequence of dark-gray to black fissile muscovite phyllite. These rocks consist dominantly of fine-grained quartz which averages 0.05 mm. in diameter. Much finer grained than the quartz is muscovite which imparts an excellent foliation to the rock. Garnets that average 1 mm. in diameter are rather uniformly distributed through the phyllite. They bulge out the muscovite laminae and give the rock the spotted appearance that is so characteristic of the rocks of the Northfield formation. A mode is given in Table 28. A few beds of dense, brown-weathering limestone up to 15 inches thick are found in these rocks but comprise only about one percent of the lower half of the formation.

The thickness of that part of the formation within the area is estimated at between 5000 and 7000 feet.

*Age of the Northfield slate:* The age of the Northfield slate is established as Silurian or Devonian on the basis of stratigraphic correlation of this formation with the fossiliferous Bernardston formation (Boucot, MacDonald, Milton, and Thompson, 1958). The Shaw Mountain formation is correlated with the Clough formation at Skitchewaug Mountain, near Springfield, Vermont, where fossil tetracorals of Silurian or Devonian age occur. The above-named authors conclude that the Bernardston formation as mapped by Emerson (1898a, p. 262–271) is probably equivalent to all the Silurian and Devonian units of adjacent New Hampshire which include the Clough, Fitch, and Littleton formations. The Northfield slate would probably be equivalent to the Fitch formation of New Hampshire.

## *Unstratified Rocks*

### Ultramafic Rocks

#### *General Statement*

Ultramafic rocks are confined to the eastern third of the area and are present in Newfane, Dover, Marlboro, and Halifax Townships. They form part of the ultramafic belt traced southward from Canada to Massachusetts, which, in turn, is part of the belt that extends the length of the Appalachian Highlands.

The ultramafic bodies of the Wilmington–Woodford area are extremely varied both in size and in mineralogical composition. The East Dover serpentine body in the northeastern corner of the area is one of the largest of such masses in Vermont. Other ultramafics of small size within the area range from serpentine to talc-carbonate in composition. The country rocks are amphibolite or chlorite schist.

#### *East Dover Ultramafic*

*Location:* The East Dover ultramafic is 4¼ miles long and slightly less than a mile wide. It lies partially in Dover and partially in Newfane and is more or less bisected parallel to its long axis by the township boundary. For the most part it is concordant with the enclosing metamorphic rocks but seems to crosscut them locally at the northeastern and southern ends.

*Description:* The East Dover ultramafic is unusual in the Vermont ultramafic belt. It is not only an extraordinarily large mass but differs from any other such bodies in that it is only partially altered to serpentine. It differs conspicuously from the other ultramafics of Vermont in its weathering. It is dull orange to dark red on weathered surfaces in all but its northern end. Here the body is a white-weathering, verde antique ultramafic which is the more common type throughout Vermont. This body as a whole has a partial zonal arrangement and in each of the zones the relative amounts of various minerals are characteristically different.

The East Dover ultramafic is excellently exposed inasmuch as it is cut diagonally by Rock River and by three tributaries. These streams flow in different directions across the body to converge east of the serpentine. They have cut deep gorges, and as a result of this its entire length and breadth are excellently exposed. Moreover the serpentine generally forms an extremely varied and rough topography, thus affording opportunity for detailed study of the rock.

The East Dover ultramafic is, in general, a massive, very fine-grained rock with a more or less horizontal or somewhat inclined banding. It consists essentially of four zones. The central core (Zone 1, Pl. I) which is about half a mile long, is over 90 per cent olivine. Surrounding the olivine-rich nucleus is Zone 2 which consists of dunite having more than 60 percent olivine. This zone has a maximum length of two miles. Zone 3 does not seem to surround Zone 2 entirely but is found only to the north of it. Likewise, Zone 4, which consists of a mixture of serpentine and talc-carbonate is found only north of Zone 3. Zones 3 and 4 are each about 3500 feet long.

The central dunite core consists of olivine grains which average 0.7 mm. in long diameter. The olivine grains are transected by several sets of parallel serpentine veins that average about 0.025 mm. in width. Minor amounts of magnetite and smaller amounts of chromite are scattered throughout the ultramafics. In general, the less altered rocks contain less magnetite and chromite than the more serpentinized rocks. Modes are presented in Table 29.

Zones 2 and 3 consist of ultramafics in which the olivine grains have been progressively more and more highly serpentinized. In these zones the olivine is reduced to smaller and more isolated nuclei. For example, in serpentine with about 10 per cent olivine, the olivine crystals are reduced to an average length of about 0.12 mm. Even these grains are cut through by tiny veins of serpentine. The average distance between such tiny veins is rarely more than 0.05 mm. Radially outward from Zone 1, olivine is reduced to smaller and more isolated nuclei.

The northern part of the East Dover ultramafic, Zone 4, consists of serpentine cut by numerous talc-carbonate veins. The northern periphery of this zone, where it contacts the country rock, is completely talc-carbonate.

By optical methods and using Winchell's determinative tables, certain of the following identifications were made. Specimen 38D/a (Table 29) contains a lemon-yellow anthophyllite which, in crossed nicols, is a reddish-yellow, and a blue-gray in z-sections. This specimen has $+2V=$ approximately 50–60°, and the mineral shows parallel extinction. For this reason, it is tentatively inferred to be bowlingite. Specimen 491B/b (Table 29) contains olivine which was calculated to contain 13% FeO. The olivine crystals are much altered to antigorite except for the central portion which remains unchanged and except for serpentine stringers transecting the olivine. Magnetite occurs as a dust-like mineral more or less uniformly distributed throughout the specimen; chromite occurs as

Table 29

MODES OF THE ULTRAMAFIC ROCKS

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| Olivine | 92 | 80 | 88 | 80 | .. | 19 | .. |
| Serpentine | 7 | 10 | 9 | 16 | 78 | 72 | .. |
| Anthophyllite | .. | 5 | .. | .. | tr | 3 | .. |
| Chromite | tr | 1 | 1 | 1 | tr | 1 | 2 |
| Magnetite | 1 | 4 | 2 | 3 | tr | 5 | .. |
| Magnesite, Ankerite & Calcite | .. | .. | .. | .. | 22 | .. | .. |
| Talc | .. | .. | .. | .. | .. | .. | 97 |
| Pyroxene | .. | .. | tr | .. | .. | .. | .. |
| Hematite | .. | .. | .. | .. | .. | .. | 1 |
| Grain size in mm. | | | | | | | |
| Olivine | .7 | <.7 | <.7 | <.7 | — | .12 | — |
| Serpentine vein width | .025 | — | — | — | — | — | — |
| Texture | Gr* | Gr | Gr | Gr | Gr | Gr | Gr |

1. 2D/a In Adams Brook, 0.1 mile north of junction with Bemis Brook.
2. 38D/a 0.4 mile southwest of East Dover.
3. 491B/b 0.35 mile east of East Dover on road.
4. 491B/a 0.3 mile east of East Dover on road.
5. 2D/c In Adams Brook, 1.45 miles due north of East Dover, near contact with Moretown formation.
6. 2D/f At waterfall in Adams Brook, 0.5 mile west of Dover-Newfane township boundary in dike or fracture zone.
7. 2D/e At second waterfall in Adams Brook, 0.1 mile west of Dover-Newfane township boundary in dike or fracture zone.

*granulose

subhedral grains and as fairly large globules throughout. Specimens 2D/a and 2D/f (Table 29) show preferred orientation features. The olivine of Specimen 2D/a shows a strong alignment, while the antigorite forms small stringers cutting through the rock. The antigorite of Specimen 2D/f is characteristically uniformly oriented where the alteration is complete. A random orientation is observed where alteration of olivine is only partially complete. The anthophyllite forms in tiny veins and is optically positive with $c \frown z = 43°$.

Specimen 2D/e is salmon-pink in color and is semi-friable to chalky in texture, and shows a prominent layering. It forms a dike-like mass in the central portion of the main ultramafic body. The writer interprets this as an alteration along a prominent fracture or movement zone.

### *Other Ultramafic Rocks*

*General Statement*: Other ultramafic rocks of the area are either small serpentine bodies surrounded in whole or in part by an aureole of talc-carbonate or are entirely talc-carbonate bodies. The former are 200 to 300 feet long, about 50 feet wide and are conformable with the surrounding amphibolites. Those which consist of talc-carbonate alone are generally less than 30 feet in long diameter.

*Location*: Five ultramafic plutons besides the East Dover body have been mapped by the writer. All but one of these are in Marlboro township, and the fifth is in Halifax. Three of these are serpentine masses with talc-carbonate rims and average 300 feet in length and 100 feet in breadth. One of these masses is located 0.8 mile south-southeast of the junction of Dover, Newfane and Marlboro township boundaries and is in the base of the Moretown formation. One is 0.7 mile northwest of the Adams school and about 0.1 mile west of Route 9, and is near the top of the Stowe formation. Another is 2 miles north of the Halifax-Marlboro township boundary at the contact of the Stowe and Moretown formations.

Two talc-carbonate bodies too small to be shown on the map scale, are indicated on the map by an X. One is 1.35 miles north-northeast of Adams School in Bellows Brook along the road and is in the lower part of the Moretown formation. The other one is 1.3 miles west-northwest of Blue Mountain in Halifax.

## Pegmatites and Veins

Pegmatites and quartz veins are common in the rocks of the area. They are found in gneisses, schists and amphibolites. Pegmatites and quartz veins are found together in many localities and grade into one another even in the same outcrop. In general, however, pegmatites are more common in gneisses, whereas quartz veins are more abundant in schists and amphibolites.

Pegmatites of the Wilmington area seem to be unrelated to intrusive igneous bodies, at least directly, in contrast with pegmatites of some other areas, such as in central New Hampshire. Unlike the pegmatites of such areas, these are generally small and comparatively fine grained.

In the pre-Cambrian rocks of the western part of the area, there are small pegmatite bodies that range from a few inches to fifteen feet long. They consist dominantly of pink or white, coarse-grained microcline, small amounts of quartz, and minor amounts of biotite and metallic minerals. Generally, the pegmatites of the pre-Cambrian are highly sheared and fractured.

In the rocks east of the Green Mountain core, pegmatites consist for the most part of pink and white microcline, with minor amounts of quartz, biotite, chlorite, ilmenite and pyrite. These pegmatites range from two inches to seven feet in length. Many of them tend to be concordant with the banding of the country rock. The Wilmington gneiss is outstanding in this respect. The veins are usually lenses or bands lying in the plane of the foliation. They are very abundant and are highly folded. The strong color contrast between the white quartz of the veins and the dark schist emphasizes the extremely contorted condition of the rocks.

### Mafic and Felsic Dikes

Only two small dikes have been observed within the area by the writer. These dikes are both unmetamorphosed basalt. One is exposed in a roadcut along Route 9, about one mile west of the eastern border of the area. The other is in the roadcut on the same route in western Woodford, in the new roadcut at Dunville Hollow. These dikes are like the Triassic diabase of the Connecticut Valley in Massachusetts. Since they are unmetamorphosed, they were probably intruded after the Taconic Disturbance and probably also subsequent to the Acadian and Appalachian orogenies. They are regarded by the writer as of probable Triassic Age. The former cuts amphibolite of the Stowe formation (Fig. 25); the latter forms a sill-like mass of variable thickness which, in places, is discordant, with small stringers reaching out into the enclosing early pre-Cambrian plagioclase gneiss.

Although felsic dikes are more common than mafic types, they are well exposed in relatively few places. They are light-colored granular rocks of granitic composition but are finer grained than typical granite. These dikes are small masses that cut across the enclosing schists. Relatively numerous occurrences of this kind are found in southern Wilmington township.

## STRUCTURAL GEOLOGY

### *Major Structures*

*General Statement:* The rocks of the Green Mountain anticlinorium comprise deformed pre-Cambrian and Lower Paleozoic clastic sediments

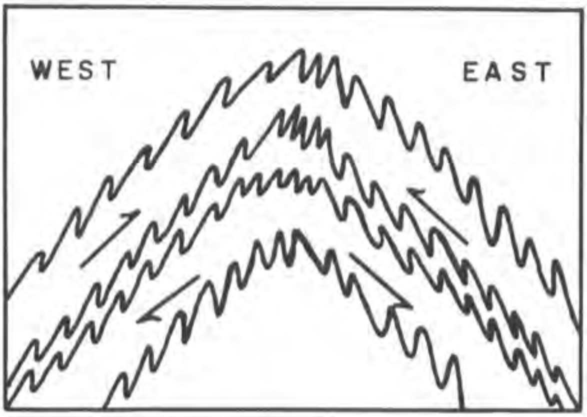



Figure 28. Diagrammatic cross-section of Readsboro anticline showing minor folds and inferred directions of movement.

and volcanics. This major fold is overturned and in part also overthrust toward the west. In plan, (Fig. 2) the 90 mile-long central pre-Cambrian core is convex toward the east. The Wilmington-Woodford area lies athwart the anticlinorium and comprises the lower portion of the western sequence and a much more complete section of the eastern sequence.

The eastern sequence is complicated by the development of two major dome structures superimposed on pre-existing folds. The latter are genetically related to the development of the Green Mountain anticlinorium. The Hoosac thrust crops out in the western half of the area on the west flank of Hoosac Mountain. The early pre-Cambrian sequence is separated from a later pre-Cambrian succession by a profound angular unconformity extending from the vicinity of Heartwellville (Pl. I) through Searsburg and Somerset, known as the Searsburg unconformity. A second unconformity, less conspicuous than the Searsburg unconformity, seems to separate the later pre-Cambrian sequence from the Lower Paleozoic.

The major structures are shown on the Geologic Map (Pl. I), in the cross sections (Pl. III) and in Figure 2. These folds comprise two contrasting types. One consists of simple anticlinal and synclinal folds, the other type is known as the Spruce-Tree or Cascade folds, but is more commonly referred to simply by the somewhat misleading name, "dome". The term, spruce-tree, refers to the cross-sectional appearance of the fold (Fig. 29). Simple anticlinal and synclinal folds are found dominantly in the western two-thirds of the area and are of relatively small

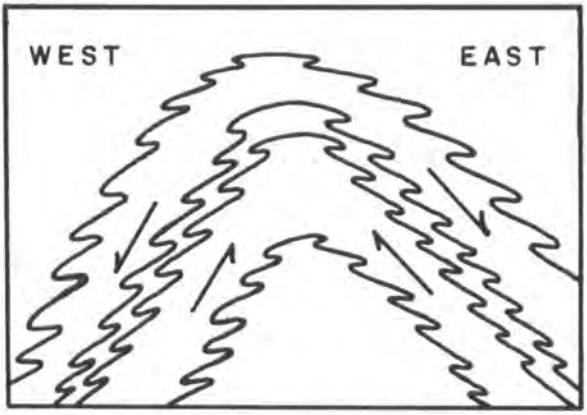


Figure 29. Diagrammatic cross-section of Sadawga Pond dome showing Spruce-Tree folds and inferred directions of movement.

size. The spruce-tree domes are represented by two folds in the eastern half of the area and are larger scale features than the simple anticlines.

These simple anticlines and synclines are part of the Green Mountain anticlinorium and seem to be the product of compressional forces. The domes are interpreted as being the product of vertical movement of the central gneisses relative to the overlying mantle of schists.

## Simple Anticlines and Synclines

### *General Statement*

Within the Wilmington-Woodford area numerous anticlines and synclines are present. In the south-central part, there are three prominent, small, doubly-plunging anticlines that are well developed and easily accessible. The structure here is seemingly rather simple and these folds display all of the minor structures usually associated with anticlines formed by compression. These are nearly symmetrical anticlines.

### *Readsboro Anticline*

The axis of the Readsboro anticline lies about half a mile east of Readsboro Village (Fig. 28). This anticline has a central core of coarsely-crystalline marble of the Sherman member of the Readsboro formation. The central core of the marble is inferred to be about 1600 feet long in a northwesterly-southeasterly direction. The beds of both the eastern and western limbs dip about 50°E and W, respectively. The axial plane of the

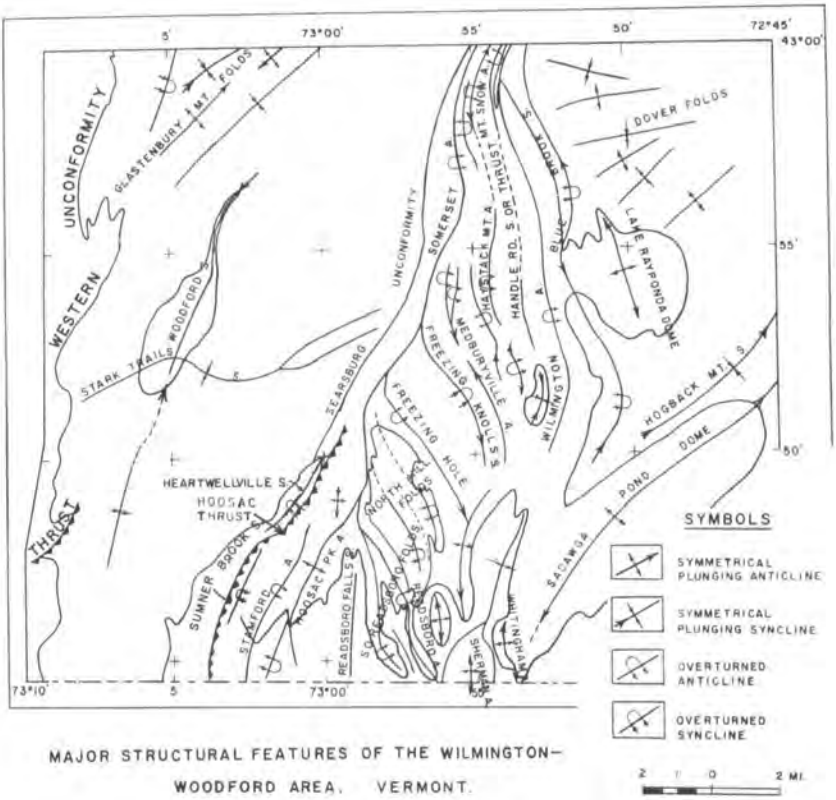



Figure 30. Major structural features of the Wilmington-Woodford area. Base of the Hoosac formation indicated by lines.

anticline as a whole strikes 15°W and dips steeply to the E, being essentially vertical. Axial planes of the minor folds on the western limb dip 40° to the west; those of the eastern limb, about 75°E. The fold axes on the northern half of the eastern flank plunge about 15°N. or NE; on the western flank they plunge about 20° N. to NW. In the southern half of the east flank, the folds plunge 30°SE; and on the western limb, 20°S to SW.

### *Sherman Anticline*

The Sherman Anticline occupies the extreme southwest corner of Whitingham (Fig. 30). Its axis trends a few degrees W of N and its central marble core is a little over half a mile long. Both this one and the Readsboro anticline are incised by the Deerfield River. The beds of the eastern limb of the Sherman anticline dip 45°E; those of the western limb dip 35°W. The axial plane of the major fold dips about 85°W. The fold axes in the northern part of the structure plunge about 20°N; those in the southern half plunge 60°SE.

### *Whitingham Anticline*

The Whitingham anticline is located one-half mile southwest of Sadawga Pond in Whitingham. Like the anticlines described above, its central core is about a mile long and trends northerly. Unlike them, however, the central part of the fold exposes rocks lower stratigraphically than those of the Sherman and Readsboro anticlines. In the central part of this fold are exposed albite schists and plagioclase gneisses of the Readsboro formation or possibly of older rocks.

The beds of the eastern flank dip about 45°E and those of the western limb at about the same angle to the west. The axial plane as a whole is essentially vertical. The axial planes of the minor folds of the eastern flank strike about north and dip 70°E; those of the western limb strike essentially north and dip 50°W.

### *Folds of the Central Green Mountains*

Reconnaissance studies in the central core of the Green Mountains indicate that the plagioclase gneisses (ep€pg, Harmon Hill gneiss) are preserved in synclinal folds. A series of northeasterly trending folds seem to be present in Glastenbury. The Moosalamoo member of the Mendon formation and the associated Cheshire quartzite are present in a synclinal fold indicated in Figure 30 as the Woodford syncline. This syncline plunges southerly in its northern portion and seems to plunge northerly in the southern part. This fold lies just west of the axis of the Green Mountain anticlinorium. Smaller deposits of Cheshire quartzite are preserved in synclinal folds in Stamford and Glastenbury townships. The pre-Cambrian-Cambrian boundary of the western margin of the area is folded into a series of southwesterly plunging synclinal and anticlinal folds (Fig. 30). Plagioclase gneisses in the vicinity of Woodford are preserved in an ENE-trending fold referred to in Figure 30 as Stark Trail syncline. In its eastern portion, lime-silicate granulites of the upper part of the early pre-Cambrian sequences are preserved.

### *Heartwellville Syncline*

The Heartwellville syncline is located southwest of the village and is overturned to the southeast. The rocks here are glassy quartzites of the Cheshire formation containing also some albitic schist. A highly deformed and mylonitized hematite-rich rock is exposed along what is interpreted as the thrust zone some 300 feet west of Route 8. The glassy quartzites, the most northerly of these which are typical of the Cheshire formation east of the Green Mountains, are exposed in the West Branch of the Deerfield River.

### *Sumner Brook Syncline*

North of Stamford Village a series of quartzites and albite schists of the Cheshire-Hoosac formations overlie the Stamford granite gneiss of the central Green Mountain core. East of these formations a series of garnetiferous-chlorite schists interpreted by the writer as those of the Heartwellville schist which is widespread in the area immediately east of the Stamford Valley, crops out. The boundary between these green schists and the underlying Cheshire-Hoosac beds is regarded by the writer as marking the location of the Hoosac thrust. These green schists are folded into a synclinal fold overturned toward the east. This structure is not inconsistent with an interpretation that the allochthonous rocks to the east have moved to their present position either as a recumbent nappe-like structure or as a more rigid thrust plate. Regional relationships, and particularly those structures associated with the Searsburg unconformity and/or Hoosac thrust (?) and Somerset anticline, suggest the possibility that the older rocks of the area may have moved westward as a nappe.

### *Haystack Mountain Anticline*

The Haystack Mountain anticline is located in northwestern Wilmington township and its center is southeast of both Haystack Mountain and Haystack Pond (Pl. I). Its axis trends north. In its central core, pre-Cambrian gneisses are overlain by albite schists of the Readsboro formation, by the Sherman marble, and by the aluminous schists of the Heartwellville formation. The beds of the eastern limb strike essentially north and dip 30°E. That part of the western limb in the vicinity of southern Haystack Mountain is overturned. Here, the beds strike northwesterly to northeasterly and dip to the east. The axial plane of the anticline is somewhat folded, but in general strikes northerly and dips 40°E. Fold axes in the northern half of this fold plunge 10°NE, and those of the southern half plunge 20°SE. In general, the axial plane seems to be essentially convex to the west.

### *Mount Snow Anticline*

The Mount Snow anticline is located in the north central part of the area and occupies small parts of Somerset, Dover, and Stratton townships (Pl. I). It is a well-delineated fold, the central core of which is north of Mount Snow (Mount Pisgah). The axis of the fold trends northeasterly in its northern part and southeasterly in its southern portion and lies about 0.2 miles east of the Somerset-Dover line. The central part of the anticline brings up coarse-grained, biotite-muscovite-albite augen-gneisses of the Readsboro formation. Pink calcite and buff dolomite marbles of the Sherman formation crop out as a continuous band in the southern part of the fold. Marbles of the same lithology, but enclosed in black schist, are well developed on the east flank of the structure in Binney Brook. These beds seem to be the same kind of marble as that found in Pike Hollow in the southern part of the Londonderry quadrangle immediately north of this district. These marbles, however, were not traceable as a continuous unit from this district to Pike Hollow.

The Mount Snow anticline does not close completely on the north within the confines of the Wilmington-Woodford area. The rocks of the eastern limb of the anticline strike northwesterly and dip about 40°E. Those of the western flank strike in about the same directions but are overturned and consequently dip 50°E. The axial plane of the anticline seems to be convex to the west in harmony with the convex pattern of the Paleozoic formations in the Dover district. The fold axes in the northern part of the anticline plunge 30°N or NE. Those on the southern part show greater variation. Most of the minor structures plunge on an average 20°SE; some, however, plunge 15°NE.

### *Somerset Anticline*

The Somerset anticline is inferred to be present just east of the Searsburg unconformity and/or the Hoosac thrust (?), from Heartwellville northeasterly to the northern limits of the area. In the lower strata of the Searsburg conglomerate and Readsboro schist fold, the shear sense of minor folds indicates an anticline directly east of the Searsburg unconformity. This relationship has been observed in many places from northern Somerset to Heartwellville. The Stamford anticline is likewise an inferred, overturned structure lying between the known Sumner Brook syncline and Hoosac Peak syncline.

### *Wilmington Anticline*

The Wilmington anticline is a narrow, elongate structure extending from just south of the Stratton township line to about 3 miles south of Wilmington Village. It consists of beds in essentially horizontal to gentle easterly dipping attitude. Coarse microcline gneisses of the Wilmington gneiss are brought up in the core of this structure. It is inferred that the beds on the west flank of this structure are overturned inasmuch as schists of the Readsboro and Heartwellville formation dip easterly under the microcline gneiss. Likewise, conglomerate beds, which seem to be those of the Tyson formation, are present west of the Wilmington gneiss and they dip easterly under the latter. Fold axes throughout all but the southern portion plunge consistently to the southeast as do the quartz-rodding and mineral streaming structures.

### *Hogback Mountain Syncline*

Hogback Mountain Syncline is located in southwestern Marlboro and southeastern Wilmington townships. Its synclinal axis is traced 5 miles to the eastern boundary of the area. Its axial trace trends N 50°E and its axis plunges 25°NE (see Fig. 30). The axial plane of the syncline as a whole strikes NE and dips about 65°NW. The axial planes of the minor folds on the southern limb strike NE and dip 30°NW. Those on the northern limb generally strike northeasterly and dip from 65°NW to 65°SE.

### *Other Anticlinal and Synclinal Folds*

A number of other recognizable anticlinal and synclinal folds are present in the area, and their relationship to previously mentioned structures and domes is rather interesting. An inspection of Figure 30 shows that the axes of those folds have a pronounced northwesterly trend. Hoosac Peak syncline, South Readsboro folds, and the North Hill folds, Freezing Hole anticline, Freezing Knolls syncline, and Medburyville anticline have strong northwesterly trends, but in the vicinity of the projection of Somerset anticline they swing into parallelism with this northeasterly trending structure.

## Domes of the Wilmington-Woodford Area

### *Sadawga Pond Dome*

*General Statement*: Sadawga Pond is a northeasterly trending dome, 7½ miles long by 1 to 2½ miles wide. It lies almost wholly in Whitingham, but projects into western Marlboro, Halifax, and southern Wilmington townships. It is named for Sadawga Pond southeast of the village of Whitingham.

The central gneisses of the Sadawga Pond dome are nearly horizontal. The beds of the east limb strike N 40°E and dip 30-75°SE. Those of the western limb strike essentially in the same direction and dip 20-70°NW. The axial plane of the fold as a whole is somewhat overturned to the northwest and dips 85°SE.

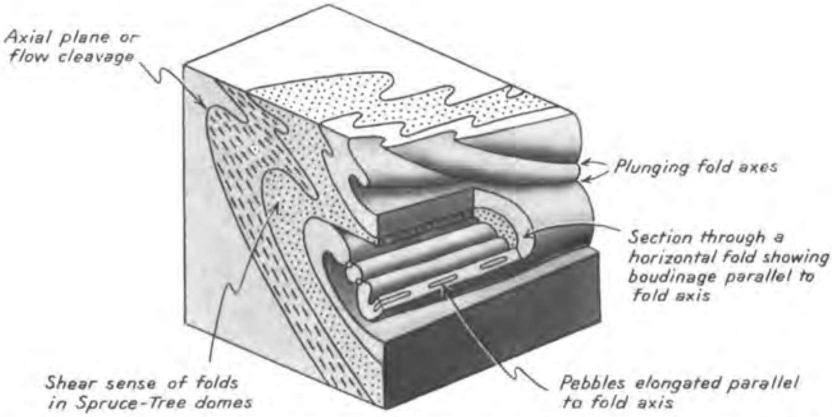



Figure 31. Block diagram of minor structural features of Sadawga Pond Dome associated with horizontal or gently-plunging Spruce-Tree folds.

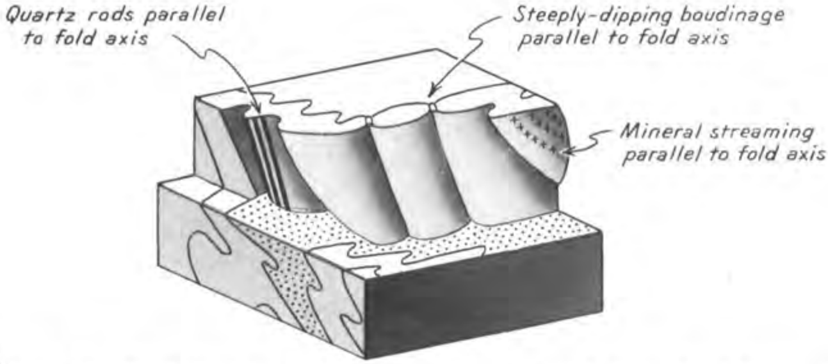



Figure 32. Block diagram of minor structural features of Sadawga Pond Dome associated with steeply-plunging Spruce-Tree folds.

The axial planes of minor folds in the Sadawga Pond dome dip at a smaller angle than the bedding and in general dip in the same direction as the bedding (Figs. 29 and 31). The fold axes of the northeastern part of the dome plunge 35°NE. In the southern part of the dome on the southwestern limb, on the other hand, the fold axes plunge 20°N to NW and NE.

### *Lake Rayponda Dome*

The Lake Rayponda dome is located in the northeastern corner of Wilmington, with extensions of the central core gneiss in the adjacent

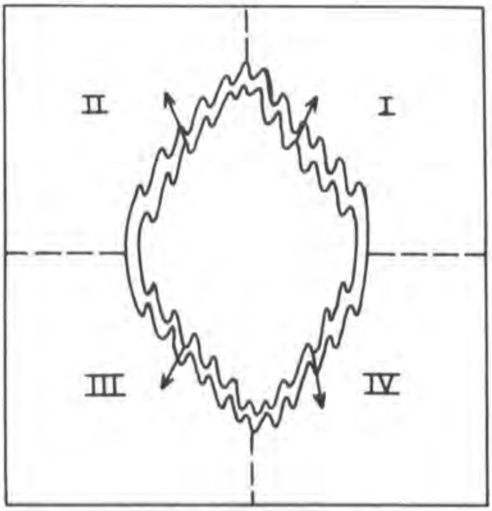



Figure 33. Doubly plunging anticline in plan view.

townships of Marlboro and Dover. The central district consists of a circular area of outwardly-dipping, banded, microcline-augen gneisses of the Wilmington gneiss. The central gneiss area is mantled by conglomerates of the Tyson and schists of the Hoosac formations. The gneisses of the central part of the dome form an area 4 miles in diameter.

This dome is essentially circular but seems to trend about N 30°W. The axial plane is essentially vertical. In the northern part of the dome, the fold axes plunge 20°N to NE. In the southwestern part of the fold, they plunge 15°NW. In the northwest section, they plunge 10 to 15° both to the NW and SE.

### *Minor Structures and Their Relation to Major Structures*

*General Statement*: Minor structures studied in the course of the present survey are minor folds (their map pattern, axial plane orientation and shear sense); lineation of fold axes, quartz rods, pebbles, mineral streaming and boudinage; cleavage; joints; and quartz veins.

### *Structural Features Associated with Folds*

A structural feature of importance in the study of deformed rocks is the map pattern of folds. There are two patterns, one right-handed and the other left-handed. Right-handed folds are those which in plan view are offset to the right (Fig. 33, quadrants II and IV). Left-

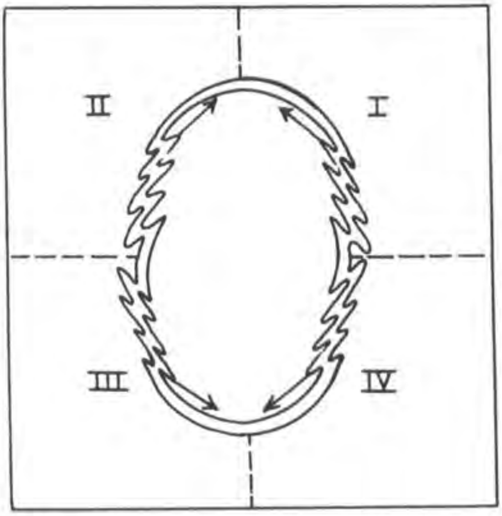



Figure 34. Spruce-Tree dome in plan view.

handed folds, on the contrary, are those which in plan view are offset to the left (Fig. 33, quadrants I and III).

Every doubly-plunging anticline and dome of the Wilmington area has both right- and left-handed folds associated with it. Left-handed folds are found in the northeast and southwest portions of the simple anticlines, and right-handed folds are found in the northwest and southeast parts. This is illustrated by the pattern of folds in the various quadrants. The map pattern of the rocks around the central portion of the Whitingham anticline (Pl. I) exemplifies the right- and left-handed patterns in the various quadrants.

In these domes, on the other hand, the distribution of right-handed and left-handed folds is just the reverse of that in the simple anticlines. In the Sadawga Pond dome, for example, right-handed folds are observed in quadrants I and III and left-handed folds are observed in quadrants II and IV (Fig. 34). The map pattern of folds around Sadawga Pond dome on the geologic map (Pl. I) shows this distribution of map patterns. Thus the observation and recording of map patterns is of importance if correlated with other structural information such as the relation to the direction of plunge of folds and the direction of dip of the beds (see Pl. II).

In certain cases it is impossible to observe the folded strata carefully in plan view. If in such cases one can see the fold in cross section and can at the same time observe the direction in which the fold plunges, its map

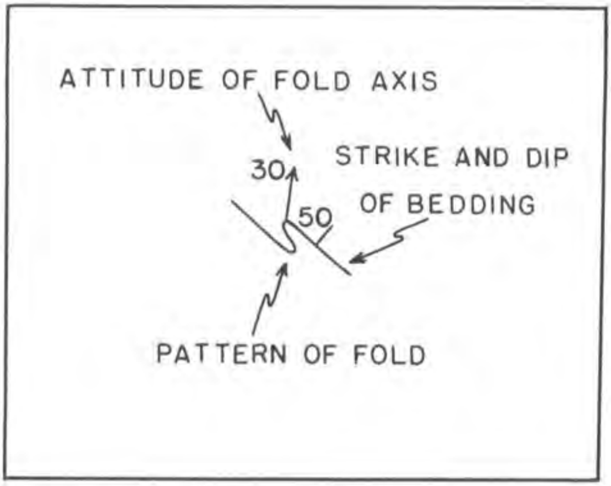



Figure 35. Structural symbol for combined strike and dip of bedding, pattern of fold, and attitude of fold axis.

pattern can be deduced. Thus, if one views a fold in cross section and is looking in the direction of plunge (Fig. 28), the map pattern is the same as the cross-section pattern.

Figure 28 is a diagrammatic east-west section through the Readsboro anticline. If the observer faces north and if the fold plunges to the north, the map pattern of the eastern half of the fold is left-handed. This is the actual case one meets in the field in the northern half of the Readsboro anticline.

The same principles are applicable in deducing the map pattern of folds in the Sadawga Pond dome from cross-section views of the folded beds. Figure 29 is an idealized cross section of folds in the Sadawga Pond dome.

To bring out the relationship between map pattern of folds, direction of plunge of the folds and the dip of the folded beds, the writer wishes to describe a useful symbol. It represents the combined map pattern of the fold, attitude of the bedding and the attitude of the fold axis (Fig. 35). Various parts of this symbol have been used for the individual structural features, but this composite symbol synthesizes several related pieces of information. Moreover, the directions shown by the symbol correspond in orientation to the directions of the structural elements as they exist in the rock.

Figure 36 is a composite symbol which shows the relation of the atti-

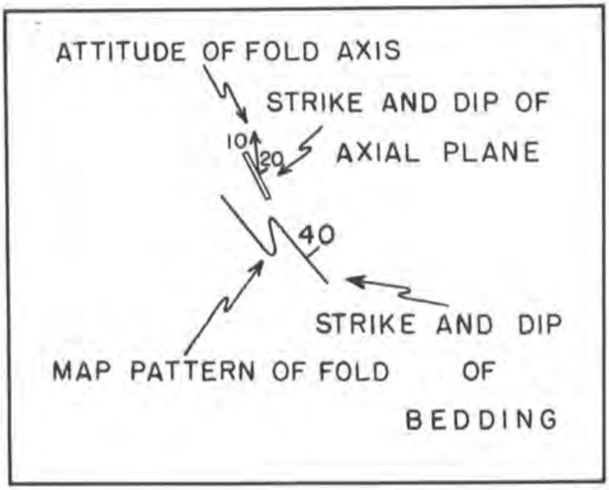



Figure 36. Structural symbol showing combined strike and dip of bedding, pattern of fold, and attitude of axial plane and fold axis.

tude of the axial plane of folds to that of the associated beds. It is a combination of part of the symbol in Figure 35 and the standard symbol for the axial plane and fold axis. Such a symbol gives extremely valuable structural information.

The axial planes of the minor folds of the major structures have also been studied. The Readsboro, Sherman, and Whitingham anticlines are essentially small anticlinoria. According to Billings' definition (1950, p. 51), these would be classed as abnormal anticlinoria because the axial planes of their folds converge upward. In general, in the above-mentioned anticlines, the dips of the axial planes of the minor folds are steeper than the average dip of the bedding. Moreover, they dip in the same general direction as the beds (Fig. 28).

In the Medburyville, Haystack Mountain, and Mount Pisgah anticlines, the axial planes of the minor folds on the eastern limb generally dip at an angle greater than that of the bedding (Fig. 37). On the overturned western limbs of these folds, however, the dips of the axial planes are gentler than those of the beds.

In the Sadawga Pond and Rayponda domes, the relationship between the angle of dip of the axial plane and that of the beds is quite the reverse of that in the simple anticlines (Figs. 29 and 28). In general, the dips of the axial planes of folds in the domes are gentler than those of the beds and usually dip in the same general direction except in the center of the

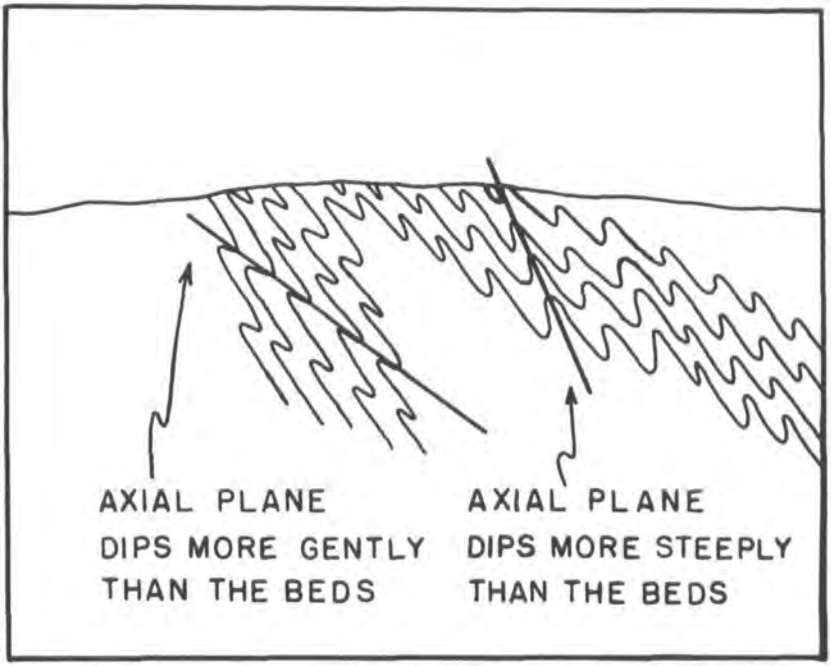



Figure 37. Diagrammatic cross-section through an overturned anticline shows relation of dip of axial plane to that of beds.

dome. In some cases, however, the axial plane dips in the opposite direction from that of the beds. In such cases, however, the angle of the dip of the axial plane is usually very low.

In the deformed incompetent beds of the simple anticlines of the area, there is a characteristic relationship between the orientation of the axial planes of the folds and the main bedding planes described above (Fig. 28). In these simple anticlines, the acute angle between the axial plane of the drag fold and the bedding points in the direction of shearing. In the Readsboro anticline (Figs. 28 and 38) the orientation of axial planes of the minor or drag folds is such that they indicate that the upper beds moved toward the crest of the anticline relative to those below. The directions of relative shearing are shown by the arrows in Figure 28.

In the case of the Sadawga Pond dome, on the other hand (Figs. 29 and 31), the orientation of the axial planes of the drag folds relative to the bedding is such that it indicates that the upper beds have cascaded away from the crest of the dome and toward the axis of the adjacent syn-

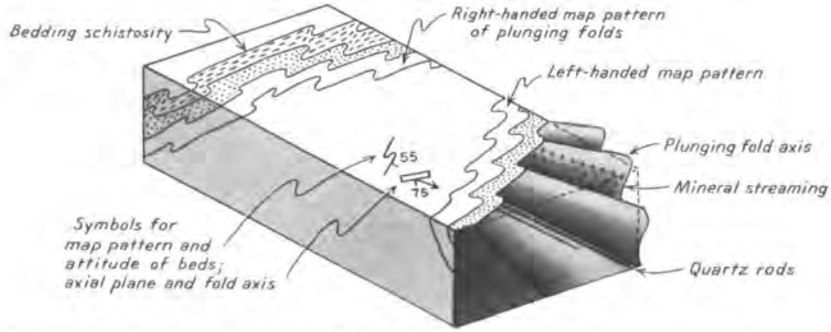



Figure 38. Block diagram of half of doubly-plunging anticline, showing relation of minor structures to major fold.

cline relative to the beds below. For this reason they may be referred to as Cascade folds. The directions of relative shearing are indicated by the arrows in Figure 29.

The attitudes of fold axes have been recorded in the field and a representative selection of those observed are presented in Plate II. In the region just east of the Searsburg unconformity in the central part of the area, most of the axes of the minor folds plunge to the northeast (Pl. I; Fig. 41). In the northern part of the Readsboro, Whitingham, and Sherman anticlines the minor folds plunge to the north on both the eastern and western limbs. Likewise, in their southern parts, the folds plunge to the south on both the eastern and western flanks, respectively. A diagrammatic representation of the Readsboro anticline in plan view is given as Figure 33. It shows the directions of plunge of the fold axes and their relationships to the map pattern of the folds.

In the Haystack Mountain and Mount Pisgah anticlines (Fig. 41), the dominant direction of plunge of the fold axes is northeast in the northern half and southeast in the southern half (Fig. 41). These anticlines are overturned to the west. Thus, northwesterly and southwesterly plunging folds are the exception.

In the overturned Somerset, Medburyville, and Freezing Knolls synclines the minor folds plunge dominantly toward the southeast. In the probable southward extension of the Medburyville and Wilmington synclinal axis, north of the Sherman anticline, the folds plunge dominantly to the north and northwest.

The attitude of fold axes around the Sadawga Pond and Lake Rayponda domes follows a rather striking pattern (Fig. 41). On the north-

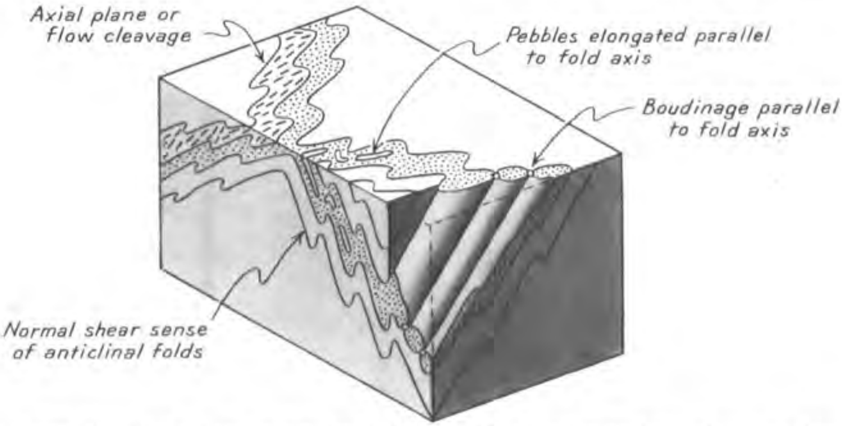



Figure 39. Block diagram of other half of doubly-plunging anticline showing other typical minor structures.

west part of the northeastern nose of the Sadawga Pond dome, the folds plunge systematically to the northeast as do those of the Mount Olga syncline which is directly north of that part of the dome. On the southeast part of the northeast nose of the dome (Pl. II), the fold axes, in part, plunge northeast or north and, in part, plunge to southeast (Fig. 41). On the southwestern portion of the western limb, some of the fold axes plunge to the northwest and some to the southeast.

In the Lake Rayponda dome, the fold axes of the northern half plunge dominantly to the northeast and northwest on the eastern and western limbs, respectively. Likewise, in the southern half they plunge predominantly to the southeast and southwest on the eastern and western limbs, respectively.

In the northern part of the Blue Brook Syncline, fold axes of the western flank plunge southeast. North of the Haystack Mountain anticline and northwest of Lake Rayponda dome, they plunge to the northeast and northwest on the western and eastern limbs, respectively. Southeast of the Haystack Mountain anticline and southwest of Lake Rayponda dome, they plunge to the southeast on the eastern limbs of the fold and to the northeast and south on the western flank.

In the northeastern part of the area, fold axes plunge consistently to the east. Their attitudes are parallel to the axes of larger scale flexures (Pl. I; Fig. 41).

## Lineation

*General Statement:* The present discussion of lineation treats of those features of the deformed rocks of the Wilmington area that are the prod-

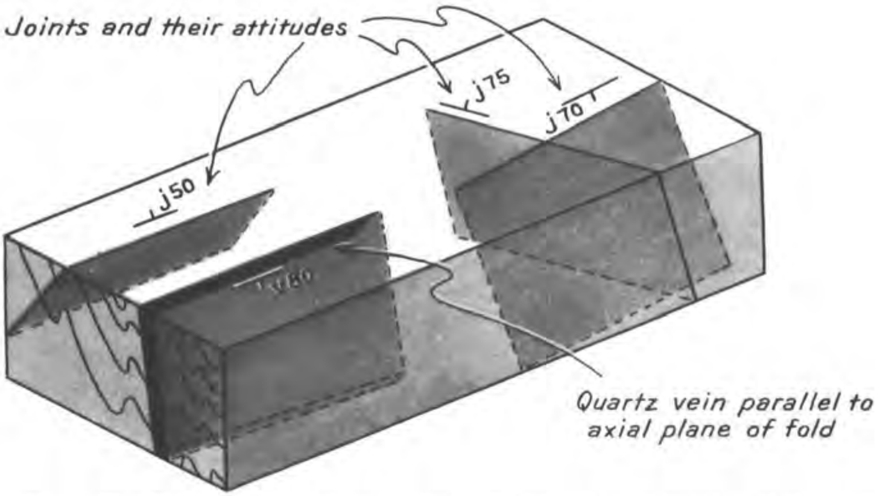



Figure 40. Block diagram showing attitude of joints and quartz veins relative to associated folds.

ucts of "parallelism of some directional property in the rocks" (Billings, 1954, p. 287). Under lineation are included quartz rods, linear streaking of platy minerals, elongated pebbles in conglomerate, and boudinage.

*Orientation and Lineation of Quartz Rods*

Quartz rods are a conspicuous structural element in the area. They are best developed in the schists of the Readsboro and Hoosac formations, and in the Wilmington gneiss. The rods consist of milky- to clear-white quartz which, in general, are elongate, rod-shaped masses of quartz oriented parallel to associated fold axes. In many cases where the rods are most abundant and the rocks are best exposed, some at least are observed to consist of quartz concentrated along the axes of the folds. In many cases the relationships of the quartz rods to folds is not clear.

Many of these concentrations of quartz along the axes of folds are actually veins of quartz parallel to the foliation of the schist or banding of the gneiss in which they occur. In certain cases, such as in the Deerfield River in the village of Readsboro, some rods are connected to others along the strike by a thin layer of quartz which becomes thicker in the crests of the folds.

The quartz rods are broken normal to their length by parallel fractures which are present at regular intervals of a fraction of an inch to two inches apart. In every case the fractures are either vertical or dip in the

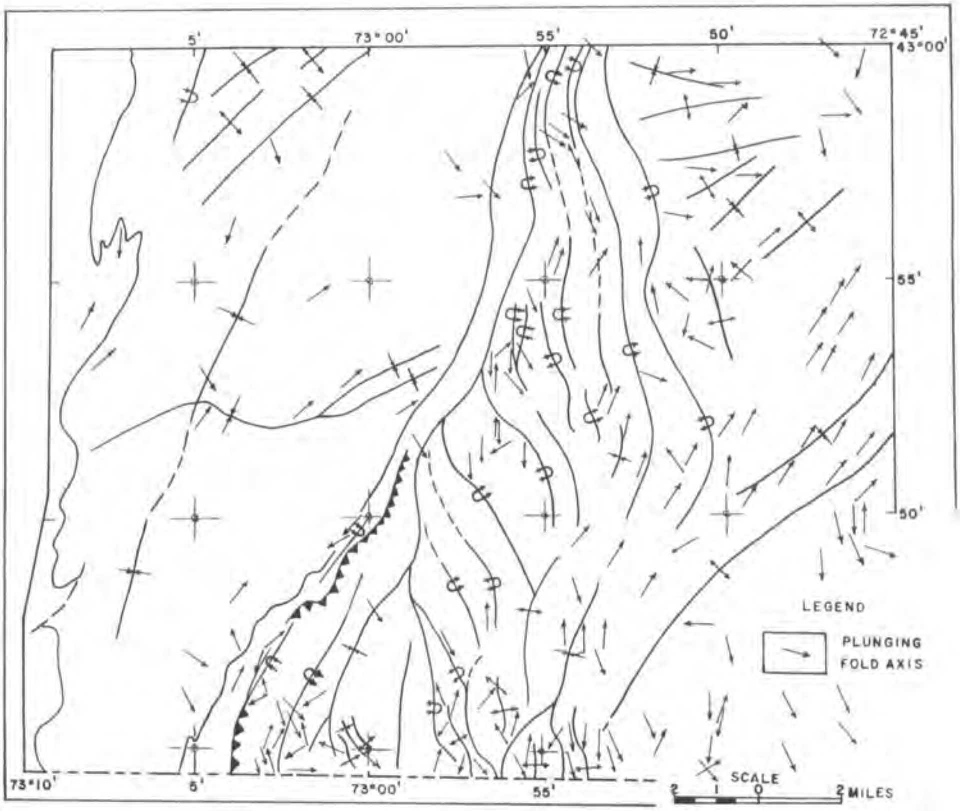



Figure 41. Relation of Lineation of fold axes to Major Structures.

direction of plunge of the rods. These fractures have a steeper angle of dip, however, than the plunge of the rods. They generally do not cut through the rods in the surrounding schist or gneiss in which the rods are contained. Commonly, the surface of the rods has fluted grooves parallel to their length. These grooves resemble very small-scale mullion structure. Parallel to the steeply-dipping fractures of the quartz rods and dipping in the same direction and angle, are quartz veins usually one-half to one inch in width and 10 to 20 feet long. These veins cut across the foliation and banding of the schists and gneisses in which they are found. They are most conspicuous in those rocks in which the rods are most abundant.

The Hoosac formation contains abundant quartz rods. In the closely associated strata of the Tyson formation there are many blue-quartz pebbles of small to medium size, and gneissic and carbonate masses up to 1½ feet long. Although many of the quartz rods are without doubt quartz veins, certain of them may be deformed pebbles.

Plate II shows the average directions of plunge of the quartz rods in relation to the axial traces of the major structures. For the most part, the attitude of the rods is the same as that of the fold axes of associated minor folds.

### *Pebble Orientation and Lineation*

As early as 1861 Hitchcock recognized that pebble elongation was of significance in the understanding of the structural history of the Green Mountains when he wrote:

> "Most of the pebbles are somewhat elongated in the direction of the strike on a horizontal surface, so as to give them an egg-shaped appearance. But where joints or other fissures have exposed the edges of the strata at right angles to the strike, the elongation, flattening and bending of the pebbles are much more striking . . . (p. 34, Vol. I).
>
> "they (the pebbles) show, we think, that the elongating and flattening forces in Vermont must have operated most energetically in the direction of the dip, whereas in Rhode Island it was most powerful in the direction of the strike. In the latter case it was as if two men had taken hold of the ends of a plastic mass and pulled it out horizontally: but in Vermont it is as if one had stood at the top of a steep hill and the other at the bottom. This is evident from the fact that when we look at the edges of a rock laid bare along the line of dip, we see little more than flattened edges of the pebbles in the form of folia; but if laid bare along the line of strike, we see flattened and even lenticular ends of the pebbles as shown on Fig. 17 (Vol. I, p. 136) already given. The fact, however, that the pebbles are lenticular on the basset edges (i.e. edges normal to the dip) the strata show that the whole force was not exerted in the direction of the dip. They were a good deal flattened horizontally but more so vertically" (Vol. I, p. 38).

Hitchcock's description cited above is appropriately descriptive of the conglomerates of the area, particularly those of Searsburg.

In the east-central part of the area, known pebbles are confined to the Tyson formation. They are tiny blue-quartz pebbles having a maximum diameter of about one-half to three-quarters of an inch. Their long axes are parallel to fold axes as they are in the beds of the Searsburg conglomerate of the western part of the area. The horizontal projections of pebbles in the east-central part of the area, however, are about parallel to the strike of the beds in the periphery of the domes. Those of the western part of the area, on the other hand, are at a large angle to the strike of the Searsburg unconformity.

### *Mineral Streaming*

The term, mineral streaming or streaking, refers to lineation best developed by the alignment of biotite flakes along the foliation planes. It is a common linear feature both in the anticlinal structures and in the domes. It is especially well developed in the Wilmington gneiss and in the arenaceous rocks of the Moretown formation. The mineral streaks consist dominantly of biotite smeared out on the foliation planes. These smears are generally 1 to 3 feet long, one-half to one inch broad and less than one millimeter thick. Throughout the area, the attitude of mineral streaming is consistently parallel to that of fold and quartz-rod lineation (Fig. 32; Pl. II).

### *Boudinage*

Boudinage or "sausage structure" is a conspicuous minor structure of the area, but is better developed in certain localities than in others. Boudinage refers to the cross-section appearance of the deformed competent bed between relatively incompetent strata. Figure 42 shows boudinage as it appears both in sections normal to the bedding and in sections parallel to the bedding. The linear character of the structure is best displayed in those parallel to the bedding. The lineation results from the depression formed along the line of greatest thinning of the beds and is in many places marked by a line of fracture. The line along which this attenuation occurs is referred to as the boudin-line. The attitude of the boudinage is defined in terms of the attitude of the boudin-line. For example, in Figure 42 the boudin-line trends S 3°E and plunges 7°.

Concentrated along the zone of greatest thinning in the competent beds, are small pegmatites (Fig. 42). They are referred to as boudinage fillings and average about 6 inches to 2 feet in diameter. They consist of white and pink microcline, quartz, and small amounts of chlorite, biotite, ilmenite, pyrite and magnetite (Figs. 42 and 43).

In the area, boudinage is best developed in thin-bedded amphibolites between schist and slate beds. In the present study, the attitudes of

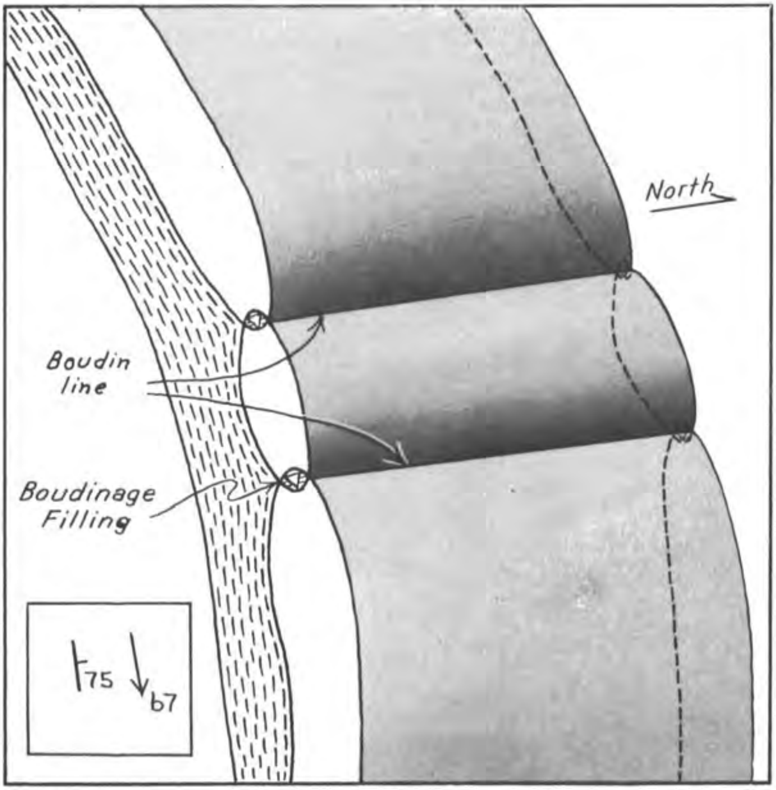



Figure 42. Boudinage structure showing the relation of the incompetent bed to the competent; boudin line and boudinage filling are indicated; inset shows symbols which are used for recording the structural data.

boudinage have been observed in approximately two dozen localities. Of these, 8 are on the southeast flank of Sadawga Pond dome. On the northwestern limb, 4 were observed, and 1, on the northeastern nose of the dome. On the western limb of Whitingham anticline, 2 were observed; 1 in the Medburyville anticline, 1 west of Haystack Mountain anticline; and 2 near the Searsburg unconformity in Readsboro and Searsburg. One boudinage occurrence was recorded near the southern end of the East Dover ultramafic.

Of the 8 boudinage localities southeast of Sadawga Pond dome, 5 trend parallel to the axial trace of the dome and 3 trend essentially normal to it and parallel to associated fold axes. On the northeast nose of

Figure 43. Boudinage in amphibolite of Stowe formation, Molly Stark Trail, Marlboro, 1.1 miles west of eastern border of area. Shows quartz boudin-filling.

Sadawga Pond dome, boudinage is normal to the axial trace of the dome and parallel to the strike of the beds. Of the 4 localities investigated northwest of Sadawga Pond dome, 3 trend parallel to the axial trace and 1 normal thereto. Boudinage lineation of the Deerfield River anticline trends essentially parallel to associated fold axes. In the Medburyville anticline it trends at a large angle to that of the axial trace of the major structure, but essentially parallel to the general trend of the beds that outline the gneisses of the central part of the fold (Pl. II). The trend of boudinage northwest of the Haystack Mountain anticline is parallel to the strike of the beds. Boudinage in the south end of the East Dover ultramafic body is essentially parallel to associated fold axes. Its horizontal projection is oriented at a large angle to the strike of the beds in which it is found, and normal to that of the beds of the nearest portions of the Lake Rayponda and Chester domes (Pl. I). The Chester dome is located in the Ludlow and Saxtons River quadrangles (Figs. 1 and 2).

*Cleavage*

Cleavage may be defined as the property of rocks whereby they break along parallel surfaces of secondary origin. Two kinds of cleavage are


Figure 44. Specimen of schistose and gneissic beds of Wilmington gneiss, on Route 8, 1.75 miles SSE of Wilmington at el. 1799. Shows schistosity cross-cutting noses of folds and parallel to beds on flanks of folds.

well developed in the Wilmington–Woodford area, namely schistosity and fracture cleavage. By schistosity is meant a cleavage resulting from mineral parallelism. It is sometimes referred to as flow cleavage and slip cleavage according to its particular character.

Schistosity is produced both by the parallel orientation of micaceous minerals or by dimensional orientation of non-micaceous minerals such as quartz, feldspar or amphibole, and also by parallel orientation of fractures with accompanying orientation of minerals parallel to the fractures. It is developed in all of the argillaceous rocks, and in most of the arenaceous rocks and many of the amphibolites of the area.

Throughout much of the area there is a well-developed schistosity essentially parallel to the bedding. On the noses of the folds, however, it cuts across the bedding and is parallel to the axial plane of the fold.

Figure 45. Albite schist beds of the Hoosac formation showing the relation between cleavage and bedding. True bedding is parallel to hammer handle and wavy ink lines; cleavage, parallel to longer ink lines (2 miles ESE of Wilmington Center and Northeast of x1631).

Figure 44 shows a fold in which this transition is indicated. It shows the schistosity passing from parallelism with the bedding on the limbs of the fold to parallelism with the axial planes on the noses of the folds (Figs. 44 and 45).

Excellent exposures in which this structural transition from bedding schistosity to axial plane schistosity is shown, are easily accessible along the Jacksonville-Wilmington Road (Route 8), 0.7 mile north of the Wilmington-Whitingham line (Pl. I, Fig. 44).

In the Wilmington area, two types of cleavage are well exemplified, namely, fracture and slip cleavage. Fracture cleavage is essentially micro-scale jointing and refers to essentially parallel fracture surfaces a fraction of an inch apart (Figs. 22 and 23). Fracture cleavage is essentially independent of parallel arrangement of mineral constituents. In this respect it differs from axial plane and from slip cleavage. It is best developed in the more gneissic rocks and amphibolites (Pls. I and II).

Slip cleavage refers to secondary surfaces of splitting parallel to which micaceous minerals are oriented. These slip surfaces are essentially small faults along which there has been a very small amount of differential movement. These slip surfaces are oriented at an angle to the bedding schistosity. Slip cleavage is well developed throughout the area, but especially in the more argillaceous rocks. Slip cleavage differs from axial plane cleavage in that it is characterized by orientation of dimensional minerals parallel to the fractures. In axial plane cleavage the mineral parallelism is not accompanied by fractures. Fracture cleavage, on the other hand, consists of parallel fractures without the orientation of dimensional minerals that are characteristic of slip cleavage.

### *Joints and Quartz Veins*

Although the study of joints and quartz-filled veins has received relatively little detailed attention during the present study, a certain amount of pertinent data has been presented in Plate II and Figure 40.

Joints are numerous in the rocks of all formations, but especially in the more gneissic beds. The Heartwellville schist and the Wilmington gneiss above all offer excellent opportunity for the study of jointing and the orientation of quartz veins in metamorphic rocks.

From the observations made, certain broad patterns in their orientation have been noted. A prominent zone along which large quartz veins are consistently present, is in the rocks of the Somerset anticline and syncline (Pl. II and Fig. 30). There are two prominent sets of veins, one of which strikes parallel to the strike of the beds and dips in the same direction as the beds, but more steeply. The average dip of these veins is 70 to 90°E. The second set is oriented normal to the axial trace of the syncline and dips steeply at 60 to 90°S.

The veins range from 2 inches to 2 feet broad in this locality. Those parallel to the strike of the schist are cut by cross fractures orientated parallel to the second set of quartz veins described above.

In the Medburyville anticline, three sets of joints have been observed. One set strikes parallel to the axial trace of the fold and dips 45° SW. A second set strikes normal to the axial trace and strike of the beds and dips at angles of 60° to 90°, both north and south. These joints are normal to associated plunging fold axes. A third set strikes essentially parallel to the strike of the beds and the axial trace, and dips steeply to the east at an angle which is larger than the dip of the beds. Many of these occurrences are easily accessible near Route 9 in western Wilmington (Pl. I).

In the central part of the area there are numerous quartz veins that range from a fraction of an inch to 5 feet in width and are as much as 30 to 50 feet long. They are consistently and spectacularly developed in the Wilmington gneiss in central Dover and Wilmington. Their strike is parallel to the axial traces of the Wilmington anticline. The veins dip steeply to the east at angles greater than the dip of the associated gneisses. They generally dip at angles of 60 to 90°E., whereas the gneiss dips 20 to 30°E. The contact of the vein quartz with the gneiss is sharp. Many of these veins are highly contorted.

East of Lake Rayponda dome, the quartz veins strike at a high angle to the strike of the beds and normal to the axial trace of the dome. One set dips southeast and the other northwest at large angles. Here the quartz veins are 4 to 5 feet broad.

In the Readsboro Falls anticline and South Readsboro syncline the quartz veins strike essentially parallel to the axial traces of the folds. They dip steeply to the east at angles of 60 to 90°. The veins range in breadth from 6 inches to 30 feet.

The East Dover ultramafic is highly fractured by two sets of joints. One of these strikes northwest and has a dip that is essentially vertical. The other set strikes north to northeast and dips 25 to 90°W.

### Summary of Relationships of Minor to Major Structures

The simple anticlines of the area, described in the foregoing discussion of major and minor structures, have certain features in which they differ notably from the domes. Although both have right- and left-handed map patterns of folds, the simple anticlines have the left-handed map patterns of folds, the simple anticlines have the left-handed pattern in quadrants I and III and right-handed patterns in II and IV (Figs. 33 and 38). On the other hand, the domes have the same patterns, but in just the opposite quadrants from those of the simple anticlines. Thus, the right-handed pattern is present in quadrants II and IV, and the left-handed patterns in I and III in the domes (Figs. 31 and 34).

More striking than the differences in map pattern between the anticlines and domes is the cross-section appearance of these folds and the shear sense of the minor folds. In the anticlines the attitude of the minor folds or drag folds indicates that the upper beds have moved toward the crest of the fold relative to the underlying beds (Fig. 38). In the domes, on the other hand, they show that the upper beds have moved toward the axis of the adjacent syncline relative to the underlying beds (Figs. 29 and 31). The difference in shear sense in these two types is probably the most striking difference between the two.

Closely connected with the difference in shear sense is the contrast in the attitude of the axial plane of minor folds relative to that of the bedding in these two kinds of folds. In the anticlines, the axial planes generally dip in the same direction as the bedding, but more steeply. In the domes, however, their angle of dip is gentler than that of the bedding.

Fracture and slip cleavage in the rocks of both the anticlines and domes is parallel to the axial planes. Thus, in a cross-section view of the domes the cleavage forms an arch convex upward (Fig. 29). Commonly such domes are referred to as cleavage arches. A corresponding architectural term for the inverted V-attitude of the steeply-dipping cleavage of the anticlines, that describes them is "cleavage gables", for this is what they resemble in cross section. In the anticlines, the cleavage dips steeply and the acute angle formed by the cleavage points upward (Fig. 38).

Lineation resulting from quartz rods and mineral streaming is parallel to fold axes in both types of structures (Figs. 32 & 38; and Pl. II). Pebbles of the anticlinorium are elongate parallel to fold axes. Pebbles of the domes are of relatively small size, but they are oriented with their long axes essentially parallel to associated fold axes. Those mineral aggregates of larger size than the pebbles and which are inferred to be deformed pebbles, likewise have their long axes parallel to fold axes.

Boudinage is better developed in the domes of the eastern part of the area than in the simple anticlines of the central portion. This may be due, in part, to the difference in the relative competency of the rocks, and, in part, to variations in the mechanics of deformation in each structural type. In the Sadawga Pond dome about 60 percent of the observed boudin-lines trend parallel to the axial trace of the dome and to the strike of the beds in which they occur. About 40 percent trend parallel to the dominant trend of fold axes, quartz rods, and mineral streaming.

Joints and quartz veins of the anticlines have three orientations. The two most prominent sets strike essentially parallel to the axial traces of the folds. One of these dips in the same direction as the beds, but at a steeper angle. The second dips in the opposite direction from the dip of the beds. The third set strikes normal to the trend in many cases and also to the plunge of the fold axes (Pl. II and Fig. 40).

Topographically the domes differ from the complex of anticlines and synclines. The frontispiece is a panoramic view taken from the axis of the Sadawga Pond dome in central Whitingham. The gently rolling farmland in the foreground stands in contrast to the rugged mountainous topography of the Green Mountain anticlinorium proper, reflecting both differences in structure and lithology.

ORIGIN OF THE SIMPLE ANTICLINES AND DOMES

From the foregoing, it is evident that there are two contrasting structural types in the area. The major structures of the western half are structurally a part of the anticlinorium. All the minor structures indicate that the upper beds have moved toward the crest of the anticlinorium relative to the beds below. These have been produced by lateral compression.

The domes likewise have a characteristic shear sense. Their minor structures are satisfactorily explained only as being the product of upward movement of the gneisses of the central part of the dome relative to the overlying mantle of schists and volcanic rocks (Figs. 29 and 31).

From the structural data of the Wilmington-Woodford area alone it is impossible to state more precisely the exact nature of the causes of domes. The relation of minor to major structures in the area indicates only the facts stated above, that the central gneisses have moved vertically upward relative to the overlying mantle of rocks. Other data on the mechanism of dome formation are, however, provided by studies in other parts of North America and elsewhere. More information will undoubtedly be forthcoming from detailed geologic studies of these structures in various parts of the world. In addition, geophysical investigations utilizing detailed gravity, and magnetic coverage will undoubtedly shed much light on the subject. These aspects of the subject will be discussed in succeeding pages under *Major Problems*.

## METAMORPHISM

### General Statement

The stratigraphic sequence of the Wilmington-Woodford area consists of metamorphosed pre-Cambrian and Lower Paleozoic rocks. The early pre-Cambrian sequence consists dominantly of microcline and plagioclase gneiss; and of lime-silicate gneiss, quartzite, conglomerate, and conglomeratic gneiss. The later pre-Cambrian sequence consists dominantly of arenaceous and argillaceous rocks in the form of albite schists and gneisses; of highly aluminous muscovite-chlorite schists with some schistose conglomerate. The Lower Paleozoic sequence consists largely of the same types of arenaceous and argillaceous rocks as

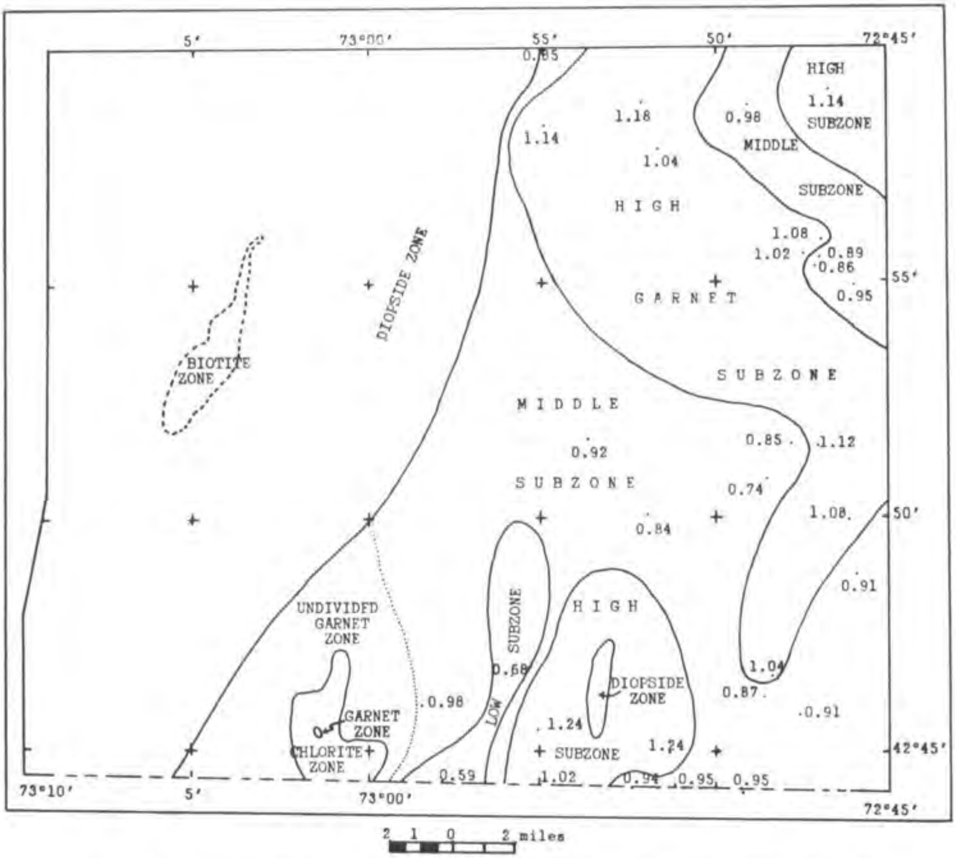



Figure 46. Isograd map of Wilmington-Woodford area. The eastern half is based on $\frac{Fe}{Fe + Mg}$ ratio of biotite divided by that of chlorite in garnetiferous rocks. Garnet zone divided into three subzones. The western half is based on megascopic observation.

the late pre-Cambrian types with abundant amphibolites and amphibolitic schists.

## *Metamorphic Zones*

Most of the rocks of the Wilmington-Woodford area are in the garnet zone of metamorphism, with local districts showing indications of higher and others of lower grades. No staurolite or kyanite has been observed in any of the schists. In the southwestern part of the Sadawga Pond dome, however, the Sherman marble includes rocks of as high grade as diopside granulites. Elsewhere in southern Vermont, there is evidence that diopside in rocks of equivalent chemical composition comes in at approximately the staurolite-kyanite grade of metamorphism (Thompson, 1950). Therefore, it seems safe to conclude that in the Sadawga Pond dome and the central core of the Green Mountains, and specifically in the older pre-Cambrian rocks of Searsburg, Somerset, eastern Glastenbury, and eastern Woodford where diopside granulites are present, the staurolite-kyanite zone of metamorphism is reached. The minerals of the lime-silicate localities of the Mount Holly complex are described above.

The diopside granulite localities of the eastern part of the area are about one mile southwest of Sadawga Pond as described under the Sherman marble member of the Readsboro formation (Pl. I and Fig. 5). The bulk of the rocks of the Green Mountain core and of the Sadawga Pond and Lake Rayponda domes are microcline and plagioclase gneisses and are therefore relatively insensitive to changes in grade of metamorphism.

Two districts of metamorphic grade lower than the garnet zone have been mapped and these are in the central Green Mountain district and on the Hoosac Range in eastern Stamford and western Readsboro townships (Pl. I, Fig. 46). The first district of the central Green Mountains is that which is underlain by the Cheshire quartzite and the dark chloritoid phyllite of the Moosalamoo member of the Mendon formation lying between 42°50′ and 42°55′ north latitude and near 73°05′ west longitude. These Paleozoic rocks are of a considerably lower grade of metamorphism than the pre-Cambrian rocks upon which they rest unconformably. Chloritoid was the only mineral observed in these phyllitic rocks by which the grade of metamorphism could be fixed. By itself, chloritoid is not diagnostic but in relatively nearby areas to the northwest on the west flank of the Green Mountains, chloritoid-bearing rocks of similar composition are closely associated with biotite-bearing strata (Thompson, oral communication). From this it is concluded that these strata are in the biotite zone of metamorphism.

In the eastern part of Stamford and western Readsboro, the highly aluminous chlorite-muscovite-quartz schists of the Heartwellville schist are found (Pl. I). Ordinarily these rocks would have garnet developed in the garnet zone since their bulk chemical composition is suitable for this mineral development. In the area indicated by the chlorite zone of Figure 46 (near latitude 42°45′ and longitude 73°00′) no garnet was

Table 30

CHEMICAL ANALYSIS OF CHLORITE FROM THE PINNEY HOLLOW FORMATION*

| | Weight % | Molecular Props. |
|---|---|---|
| $SiO_2$ | 25.19 | .420 |
| $Al_2O_3$ | 23.34 | .228 |
| $TiO_2$ | 0.13 | .0016 |
| $Fe_2O_3$ | 2.43 | .015 |
| $FeO$ | 22.63 | .314 |
| $MnO$ | 0.03 | .0004 |
| $MgO$ | 14.90 | .370 |
| $H_2O^+$ | 11.57 | .643 |
| Total | 100.22 | 1.9920 |

Chemical Formula of Chlorite:—

$$(Fe,Mg)_{4.2}Al_{0.8}(Fe,Ti,Al_{0.8})\ (Al_{1.4}Si_{2.6})\ O_{10}(OH)_8$$

Indices of Refractions:— $\alpha = 1.625$

$\beta = 1.627 \pm .002$

$\gamma = 1.628$

Specific Gravity:— 2.895

Chlorite from Specimen 258B, coarse garnet-chlorite-muscovite-quartz schist (in Green River, Marlboro, at latitude 42°49′53″ and longitude 72°46′06″, 1.65 miles N 20° W of Blue Mountain).

*analysis by F. A. Gonyer

observed except in the small district so indicated (Fig. 46). For these reasons, the rocks of this district are mapped in the chlorite zone of metamorphism.

The writer wishes specifically to point out that the green schists of the Paleozoic sequence of the eastern slope of the Green Mountains between Stamford and Heartwellville contain garnet and, therefore, belong to this zone of metamorphism.

## *Method of Study*

The present investigation is concerned with a study of the distribution of various mineral assemblages and the variations in chemical composition of some of the mineral components. The distribution of assemblages has been correlated with the major structural features of the area. In the eastern portion of the area detailed studies of the grade of metamorphism were made in the manner outlined below and in the western part the determination of grade of metamorphism was made on

TABLE 31

OPTICAL AND CHEMICAL DATA ON GARNET-BEARING ROCKS OF THE WILMINGTON-WOODFORD VERMONT AREA

Locations are shown on Fig. 46

| | Biotite B index | $\frac{Fe}{Fe + Mg}$ ratio | Chlorite B index | $\frac{Fe}{Fe + Mg}$ ratio | $\frac{Fe}{Fe + Mg}$ ratio $\frac{Bi}{Chl.}$ |
|---|---|---|---|---|---|
| North-central Marlboro | 1.625 | 0.38 | 1.618 | 0.42 | 0.91 |
| Hogback Mt., Marlboro | 1.635 | 0.47 | 1.618 | 0.42 | 1.12 |
| Northwestern Marlboro | 1.633 | 0.45 | 1.620 | 0.44 | 1.02 |
| Southeastern Wilmington | 1.627 | 0.40 | 1.629 | 0.54 | 0.74 |
| Southern Halifax | 1.636 | 0.48 | 1.627 | 0.53 | 0.91 |
| South-central Whitingham | 1.628 | 0.41 | 1.619 | 0.43 | 0.95 |
| Central Dover | 1.633 | 0.45 | 1.615 | 0.38 | 1.18 |
| Southern Whitingham | 1.635 | 0.47 | 1.615 | 0.38 | 1.24 |
| West-central Halifax | 1.635 | 0.47 | 1.627 | 0.53 | 0.87 |
| Northeastern Dover | 1.628 | 0.41 | 1.615 | 0.38 | 1.08 |
| Northwestern Dover | 1.675 | 0.57 | 1.625 | 0.50 | 1.14 |
| Hogback Mt., Marlboro | 1.621 | 0.34 | 1.617 | 0.40 | 0.85 |
| Southern Marlboro | 1.645 | 0.55 | 1.626 | 0.51 | 1.08 |
| South-central Whitingham | 1.615 | 0.30 | 1.609 | 0.32 | 0.94 |
| Northeastern Somerset | 1.635 | 0.47 | 1.630 | 0.55 | 0.85 |
| South-central Readsboro | 1.637 | 0.49 | 1.625 | 0.50 | 0.98 |
| Southwestern Whitingham | 1.630 | 0.43 | 1.620 | 0.44 | 1.02 |
| Southwestern Whitingham | 1.635 | 0.47 | 1.615 | 0.38 | 1.24 |
| Northeastern Dover | 1.632 | 0.44 | 1.622 | 0.45 | 0.98 |
| Northeastern Dover | 1.638 | 0.50 | 1.620 | 0.44 | 1.14 |
| Readsboro Village | 1.623 | 0.36 | 1.627 | 0.53 | 0.68 |
| Northeastern Whitingham | 1.643 | 0.54 | 1.638 | 0.64 | 0.84 |
| Eastern Whitingham | 1.635 | 0.47 | 1.621 | 0.45 | 1.04 |
| Eastern Whitingham | 1.633 | 0.45 | 1.628 | 0.53 | 0.85 |
| Southeastern Dover | 1.635 | 0.47 | 1.628 | 0.53 | 0.89 |
| North-central Marlboro | 1.640 | 0.52 | 1.630 | 0.55 | 0.95 |
| Central Dover | 1.645 | 0.55 | 1.628 | 0.53 | 1.04 |
| Southeastern Dover | 1.635 | 0.47 | 1.630 | 0.55 | 0.86 |
| Central Wilmington | 1.649 | 0.60 | 1.640 | 0.65 | 0.92 |

megascopic identification of mineral constituents. Inasmuch as the rocks belong to the garnet zone for the most part, it was necessary to undertake detailed investigations of variations in mineral composition within the garnet zone.

Many rocks in the area contain one or two ferro-magnesian components. Only those rocks, however, which contain the assemblage, biotite-chlorite-garnet were considered in drawing the subisograds. In rocks which contain only two ferromagnesian components, such as chlorite and biotite, or chlorite and garnet, or biotite and garnet, the relative compositions of these phases is determined not only by grade of metamorphism but also by the bulk composition of the rocks themselves. In rocks containing three or more ferro-magnesian phases, such as biotite, chlorite and garnet, the minerals have a fixed composition for any given temperature-pressure condition.

The method employed in the present investigation assumes that in a rock containing the ferromagnesian phases, chlorite, biotite and garnet, the garnet forms at the expense of the iron component of the other two minerals, and that biotite, in certain cases at least, forms at the expense of the iron component of the chlorite. If these are valid assumptions, the indices of refraction of chlorite steadily decrease with increase in metamorphic intensity. One might expect also that in the lower part of the garnet zone the indices of refraction of biotite would increase as both biotite and garnet are formed at the expense of chlorite.

Rocks containing multiple ferromagnesian phases such as chlorite-amphibole-ankerite and chlorite-amphibole-garnet were not considered in the present investigation because of the uncertainties that are involved in attempting to deduce the Fe:Fe+Mg ratios from the indices of refraction of the amphiboles.

The Fe:Fe+Mg ratios in this investigation were obtained from the refractive indices of chlorite and biotite. The indices were obtained by oil immersion. The $\beta$ index of biotite and chlorite was measured and the corresponding Fe:Fe+Mg ratio was obtained from graphs given by Winchell (1951, pp. 374, 385). The kind of biotite assumed is the average for analyzed biotites and is Annite-Phlogopite 70%-Siderophyllite-Eastonite 30%. The composition of chlorite is the average composition of chlorite from the garnet grade of metamorphism and is Daphnite-Amesite 70%-Ferroantigorite-Antigorite of the Winchell diagram (p. 385).

The chemical formula for the average chlorite is very close to that of chlorite frcm arenaceous rocks of the Wilmington area. A chemical analysis of chlorite was made from the muscovite-chlorite schist of the Pinney Hollow formation of southern Marlboro. The chemical analysis of the chlorite, its formula, optical and physical data are presented in Table 30. The chemical analysis was performed by Forest A. Gonyer in 1951.

Mineral assemblages were studied optically and indices of refraction of iron-and magnesium-bearing minerals were determined. Thus among other data, variations throughout the area in the Fe:Mg molar ratios in biotite and chlorite were investigated.

From the β indices of biotite and chlorite the Fe:Fe+Mg ratios were calculated from the diagrams presented by Winchell (op. cit.). The molar ratio of biotite was then divided by the ratio of chlorite. Where the molar ratio of biotite equals that of chlorite the resultant is unity. Where the ratio is greater or less than that of chlorite the resultant is greater and less than unity, respectively. On the basis of variations in these ratios, a triple subdivision of the garnet zone has been mapped.

The variations observed during the study ranged from 0.89 to 1.24 (Table 31). In general the regions of highest biotite-chlorite ratios were adjacent to the area in southwestern Whitingham where the rocks are in the diopside zone (Fig. 46). In a general way the zone whose biotite-chlorite ratios were greater than unity is associated with the Sadawga Pond and Lake Rayponda domes. This zone occupies a much greater area than the central gneiss cores of the domes.

Since the rocks of the garnet zone show such a wide range in the biotite-chlorite ratio, it was decided to sub-divide it into three subzones. The upper garnet subzone includes those rocks having a ratio greater than unity. Those of the middle subzone include rocks having a ratio less than unity and greater than 0.7; and the low subzone, those whose ratio is 0.7 or less.

The high garnet subzone (Fig. 46) occupies three parts of the area. An irregularly oval-shaped area of this subzone is located in southwestern Whitingham and surrounds the small locality of the diopside zone. This high garnet subzone also occupies a larger, irregularly shaped, elongate region in the central and eastern part of the area. It extends through northeastern Whitingham, northwestern Halifax, western Marlboro, northern Wilmington, Dover, and the portions of Stratton and Wardsboro within the Wilmington area. A third locality occupies the extreme northeastern corner of the area.

The low garnet subzone occupies a small part of eastern Readsboro township. The middle garnet subzone is found throughout the rest of the area adjacent to the regions of the high subzone.

A comparison of the geologic map (Pl. 1) and the isograd map (Fig. 46) reveals that the high subzone in Whitingham and that in Marlboro and Halifax coincide with the two ends of the Sadawga Pond dome. In the northern part of the Wilmington area the high subzone is broadest in the vicinity of the Lake Rayponda dome. It is to be noted that this subzone reappears in the northeastern part of the area adjacent to the southwestern portion of the Chester dome of the Brattleboro and Saxtons River quadrangles.

The highest biotite-chlorite ratios were in southern Whitingham where they reached 1.24 at distances of 1½ to 2 miles from the diopside zone. Values of comparable magnitude were found for the Hogback Mountain and Dover regions where the ratios were as high as 1.12 to 1.18. In the northeastern corner of the area the ratio is 1.14. The lowest value was from the Massachusetts-Vermont line in southern Readsboro where the ratio was 0.59.

Other methods for studying the metamorphic intensity within the garnet zone in the area have yielded only partial success. Data on the Fe:Fe+Mg ratio of biotite or chlorite alone in garnetiferous rocks, was collected. In rocks containing only one of these minerals, however, the bulk chemical composition in addition to the metamorphic intensity of the rocks is an important factor in determining this ratio. It is therefore a rather unreliable method. In rocks of approximately the same bulk composition, however, the method could be used as indicative of at least general trends in intensity.

The average sizes of garnets in rocks of approximately the same bulk composition have been recorded. This method is subject to great limitations because the rate of reaction is an important factor in controlling the size of garnets. For this reason, the data are not presented in this report. The rate of reaction may vary from place to place and therefore introduce considerable error into the results. In the Wilmington-Woodford area, however, garnets 1 to 3 cm. in diameter have been observed only in those localities where the biotite-chlorite ratio is greater than unity, that is, in the high subzone of Figure 46.

### *Metamorphism of Argillaceous Rocks*

There are several different kinds of argillaceous rocks in the area. These include biotite-muscovite-albite schist, biotite-muscovite schist, muscovite schist and chlorite-muscovite schist. Throughout the area argillaceous rocks consistently contain garnets. They usually contain in addition small amounts of either chlorite or biotite, or both. A few rocks contain chloritoid with or without associated chlorite. In the garnetiferous argillaceous rocks there is a fairly consistent relationship between the amount of chlorite, the amount and size of the garnets and the Fe:Fe+Mg ratio of the chlorite, and to a lesser extent of the biotite also.

The megascopic appearance of argillaceous rocks of the upper garnet subzone differs somewhat from those of the middle subzone. Rocks of the middle subzone contain garnets ranging in size from those too small to be seen by the naked eye to those as large as 1 to 2 mm. These rocks also contain a conspicuous amount of ankerite. Argillaceous rocks of the upper garnet subzone usually contain garnets several millimeters to 3 centimeters in diameter. Ankerite is either absent or at least is not a conspicuous constituent of these rocks. Retrograde metamorphic effects have been observed in areas of the middle subzone but not in those of the upper garnet subzone.

### *Metamorphism of Carbonate Rocks*

Carbonate rocks of the area show a greater variety of metamorphic minerals than any other rock type. These rocks are found in eight localities in the area. Five of these localities are located in the western half of the area in the Mount Pisgah, Haystack Mountain, Freezing Knolls, Readsboro, Sherman and Whitingham anticlines. In these localities the rock is coarse-grained siliceous calcite marble and fine-grained dolomite marble. It contains also small amounts of microcline, phlogopite, tremolite, actinolite, talc, anthophyllite, epidote and graphite and is associated with garnetiferous muscovite schists above and muscovite-biotite-albite augen schist below.

Diopside-lime-silicate granulite is found one mile southwest of Sadawga Pond. This rock contains biotite, oligoclase, microcline and epidote in addition to quartz and diopside. North of the diopside-lime-silicate granulite locality and in the brook west of Sadawga Pond, carbonate and lime-silicate rocks are associated. In the brook northwest of the village of Whitingham, actinolite-granulite without associated carbonate is overlain directly by dark, garnetiferous muscovite schists. These rocks are thought to belong to the upper garnet subzone. This conclusion is consistent with data on the indices of the ferromagnesian constituents and on the relative sizes of garnets in argillaceous rocks associated with these rocks.

Only one thin limestone lens has been found in the eastern third of the area. This is found in southeastern Wilmington. It is a siliceous phlogopite-calcite marble. (Pl. I).

## *Metamorphism of Mafic Rocks*

With increase of intensity of metamorphism in the mafic rocks of the area, the indices of refraction of amphibole increase. This increase is accompanied by a decrease in the indices of chlorite and biotite in the same rock. The exact nature of these changes are, however, not accurately known. Therefore in the present investigation, although a certain amount of data was collected on this phase of the subject, the results are inconclusive. Therefore without a great number of chemical analyses of the amphiboles they would not form a reliable basis for an isograd map of the subzones of metamorphic intensity.

Most of the mafic rocks of the area contain two or more iron-and magnesium-bearing minerals. These minerals are commonly amphibole, chlorite, biotite, garnet and ankerite. The mafic rocks of the Wilmington area are essentially of uniform composition. They occur in the upper part of the Hoosac formation and in the younger formations, but are generally not present in the older rocks of the area. The megascopic appearance of mafic rocks of all the garnet subzones is essentially the same. Generally, however, the rocks of the upper subzone contain greater amounts of, and coarser grained amphibole than those of the middle subzone. Mafic rocks of the middle subzone contain a higher percentage of chlorite and the amphibole is rather fine grained.

## *Metamorphism of Ultramafic Rocks*

Ultramafic rocks of the area are of three types. The smaller bodies consist of talc-carbonate rock. The larger masses have a rather large central core of serpentine surrounded by a thin outer rim of talc-carbonate rock.

The East Dover ultramafic, the largest in the area and possibly the largest in Vermont, has a central dunite core (Zone 1, Pl. I). Surrounding this nucleus are broad zones (Zones 2 and 3) in which the ratio of olivine to serpentine decreases radially outward from the center. Partial rims of serpentine-talc-carbonate and talc-carbonate are found in the northern part of the body (Zone 4, Pl. I).

The mineralogical constitution and zonal arrangement in the various ultramafics follows a rather consistent pattern. The outer rim consists of magnesite and talc. In the smaller bodies the transition zone from talc-carbonate rock to serpentine is thin; in the East Dover ultramafic it is rather broad (Zone 3). The serpentine zone consists of serpentine, magnetite and chromite with varying amounts of olivine. The central core of the East Dover body contains over 90 percent of olivine and small amounts of serpentine, magnetite and chromite.

Although the mineral assemblages in these bodies differ from zone to zone, their bulk chemical composition is the same except for certain constituents. The central dunite core is essentially anhydrous. The serpentine-olivine zones, on the other hand, contain the hydrous mineral, serpentine, in amounts which become greater radially outward from the center. The serpentine zones also have a slightly higher content of silica. The outward zones contain not only hydrous but also carbonate minerals. Evidently carbon-dioxide is a constituent which has been added from a source outside the ultramafic.

With regard to the method by which the original ultramafics were metamorphosed, there are several possibilities. One theory has been proposed by Hess (1933, p. 640). He concluded that their metamorphism was an autometamorphic reaction. His interpretation was made for the following reasons:

> "1) serpentine distribution rarely shows a relation to the borders of the ultramafic mass in contrast to steatitization, 2) field evidence indicates that it occurs during the same cycle of igneous activity as the intrusion of the ultramafic itself, and 3) serpentinization always precedes the steatitization. Moreover the even distribution of serpentinization in many ultramafics indicates that not only was it autometamorphic but largely deuteric in the most restricted sense. Cores of olivine uniformly distributed and nearly equally corroded are commonly found in partially serpentinized ultramafics."

Hess accepts these reasons as evidence that the olivine grains were attacked by a residual liquid present in the interstices between the grains rather than as representing attack by solutions concentrated in the more widely spaced fissures. In referring to factors that control the degree of serpentinization attained by an ultramafic mass he writes as follows:

> "It is undoubtedly true that ultramafics intruded at shallow depths into water-bearing sediments suffer much greater serpentinization than do ultramafics intruded at greater depths into relatively anhydrous gneisses." (Hess, 1938, p. 331).

From his study of the East Dover ultramafic, Bain (1936) concludes that the serpentine developed during two stages. One alteration is thought to be autometamorphic and the other a late alteration. He notes that the antigorite is limited to fracture systems along which some slippage has taken place. He, therefore, relates its development to tectonic movements. The columnar serpentine is developed along fissures related to the horizontal sheeting of the East Dover body as a whole. Bain interprets this early serpentine as a product of autometamorphic serpentinization resulting from attack on the olivine grains by a residual liquid present in the interstices between the grains.

Phillips and Hess (1936, p. 348), from their studies of ultramafic bodies of Vermont and elsewhere, have concluded that these alterations are the result of the action of slowly migrating or nearly stagnant solutions during changes of temperature that were regional in extent. Their studies reveal that deposits subjected to the higher temperature type of metamorphic differentiation are associated with country rocks which have undergone a correspondingly high grade of metamorphism, but which in addition show some younger temperature effects as well. Likewise, outside of the area in which the metamorphic differentiation at serpentinite contacts is of the lower temperature type, the country rocks show only the effects of low-temperature regional metamorphism.

The writer concludes that in the Wilmington-Woodford area the only autometamorphic effect in the ultramafics is that mentioned by Bain in which serpentine is developed in fractures parallel to the horizontal sheeting and which are younger than other fissures. It is possible, however, that serpentinization and steatitization may be entirely related to later tectonic movements. In any case, a large part of the serpentinization and all of the steatitization is certainly related to later tectonic movements.

The evidence known to the writer at present suggests that serpentinization occurred sometime after intrusion. It seems quite certain that a large part of the serpentinization and all of the steatitization did not occur as an autometamorphic process for the following reasons. The original ultramafic seems to have been of dunite composition. The only component of common occurrence in the serpentine zone not found in the dunite core is water. Carbon-dioxide is the only constituent in the talc-carbonate band not found in the serpentine zone.

Apparently water was more mobile than carbon-dioxide, at least under the conditions of metamorphism that prevailed during the serpentinization and steatitization. In the East Dover ultramafic, water seems to have penetrated through all except the central core. Carbon-dioxide, and probably also iron seems to be less easily moved than water under the conditions prevailing at the time of steatitization. This is shown by the fact that the carbonate band occurs only around the periphery of the ultramafic bodies within a few feet of the contact with the country rocks. In the northern end of the East Dover ultramafic, carbonate has migrated as much as a quarter of a mile from the country rock along small fissures; and in the central part of Zone 1 a dike-like body of talc is present along a fracture zone. Water is believed to have migrated greater distances than carbonate because the hydrous mineral, serpentine, is found in all but a tiny central core which is unserpentinized dunite. The serpentine increases in amount outward from the central dunite core. Iron seems to have moved short distances from the epidote amphibolites that everywhere surround the ultramafics.

The material which is added, is probably derived from the surrounding country rocks because hydrous and carbonate minerals are common constituents of these rocks. With rise in intensity of metamorphism in the country rocks, the carbon-dioxide and water of greenstone, amphibolite and limestone would tend to be driven off and would settle in these anhydrous and non-carbonate-bearing rocks.

Thin sections from the serpentine-olivine zones reveal that the olivine grains are cut by several series of sub-parallel, crosscutting serpentine veins. These veins seem to represent planes of fracture, or at least, planes of weakness. They seem best explained as fractures along which serpentinizing solutions moved.

If the production of serpentine and steatite were the result of autometamorphism, one would not expect the zonal arrangement that is everywhere present. Especially difficult to explain by autometamorphism is the presence of carbonate in the talc-carbonate zone. Hess (1933) recognized this as a difficulty in his attempt to reconcile the talc-carbonate rim with his theory of autometamorphism. Therefore he proposed that the formation of steatite was subsequent to that of serpentine.

In partially serpentinized ultramafics examined by Hess, cores of olivine were uniformly distributed throughout the body and these he discovered were nearly equally corroded. He accepted this as evidence that the olivine grains were attacked by a residual liquid present in the interstices between the grains rather than attack by solutions concentrated in the more widely spaced fissures.

Thus the production of serpentine and steatite in the Wilmington area seems to have been accomplished by solutions derived from outside sources. The solutions were probably directly related to the regional metamorphism for reasons given above. Reasons for maintaining that the source of solutions by which the metamorphism was effected, was outside the ultramafics and not in the dunite magma from which the ultramafics crystallized, are summarized as follows.

at this general time or at least associated with this period of mountain building, that earlier rocks described above were metamorphosed and subjected to a long period of erosion.

Following this interval of earlier pre-Cambrian orogeny, the region was presumably depressed beneath sea level while the central Green Mountain area probably stood near or somewhat above sea level. At this time began the deposition of shaly gravel and sand deposits, some of which are represented by the Searsburg conglomerate. Thin beach sands and shales were deposited along the shore and in shallow water.

The source area of the Readsboro schists was probably the plagioclase gneisses of the central Green Mountains. On the other hand the albite schists may represent analcime shales or possibly the source of the soda may have been interbedded salt beds in an otherwise shaly and sandy succession. The beds of green and dark highly aluminous schist of Heartwellville lithology probably represent shales and sandy shales. The dark shales contain graphite which suggests an organic origin. The marble deposits of the Sherman member are for the most part filled with irregular structures suggestive of the possibility that they represent reefs rather than even-bedded limestones. The discontinuous nature of the marble suggests the same origin. The presence of graphite crystals in the marble may indicate the former presence of life in the reefs.

If the later pre-Cambrian age is valid for the Searsburg through Heartwellville formations there was a period of folding in which horizontal compressional forces acted in a direction approximately N75°W. This was followed by a period of erosion which exposed the older formations down to the older pre-Cambrian rocks. On this erosion surface were deposited the Tyson gravels and sandy beds. More widespread were the analcime shales or whatever was the sediment from which the albite schists were derived. Near the beginning of Hoosac time basaltic lava flows were poured out intermittently as is recorded in the amygdaloidal amphibolites and pillow lavas.

During Lower Cambrian time the crest of the Green Mountains was submerged as is evidenced by the presence of black shales and sandstones of the Mendon and Cheshire formations. The presence of marble and lime-silicate rocks associated with shales of Heartwellville lithology in the northern part of the area high on the Green Mountains suggests that this part of the Green Mountains may have been submerged at the time when these deposits were laid down. From this we may conclude that the higher parts of the Green Mountains were submerged at least twice during later pre-Cambrian and/or Lower Paleozoic time.

From Hoosac time until the deposition of the Ottauquechee sands and black shales, volcanic activity is recorded in the presence of amphibolites and greenschists. The amphibolites seem to represent in part reworked volcanic rocks, and in part lava flows and tuffs. The greenschists probably represent a mixture of reworked volcanic material in arenaceous shales. These alternating conditions of deposition continued through Stowe time.

The beds of Moretown lithology are in part similar to those described above but there is a higher percentage of recognizable pyroclastic material in the Moretown. Many outcrops throughout the area show clearly the pyroclastic character of these deposits in the presence of angular to rounded volcanic fragments in a finer grained matrix. Amphibolites representing flows or sills are likewise present. Some of the volcanic deposition undoubtedly took place contemporaneously with the accumulation of sediments in the form of arenaceous and argillaceous deposits, giving rise to a bedded deposit partaking of the character of both types of rock.

The lower part of the Cram Hill deposits consist of mafic and felsic volcanics. These are in part exceptionally well-bedded volcanic ash or reworked and water-laid volcanics, and in part are pyroclastics, flows, and sills. During the latter half of Cram Hill time the dominant activity was the deposition of dark shales and arenaceous shales, lithologically like those of the Northfield phyllite. These Cram Hill slates mark a new sedimentary environment such as was present previously only during Ottauquechee time. Subsequent to the deposition of these Cram Hill shales and prior to metamorphism, mafic sills and some dikes were intruded into the sequence. These were intruded prior to the deposition of the Northfield shales and presumably, therefore, prior to the development of the Shaw Mountain unconformity.

The main episode of compression by which the Green Mountains were folded took place following the deposition of the Cram Hill formation. The land surface was uplifted and eroded to a limited extent as is recorded in the Shaw Mountain unconformity. It is a widespread erosion surface but the beds on either side are essentially parallel in this region. No deposits of Shaw Mountain gravels or associated rocks are recognized in the Wilmington-Woodford area. On this erosion surface were deposited the black shales of the Northfield slate. On the basis of fossil evidence these are known to have been deposited in marine waters during Siluro-Devonian time. Interbedded in the Northfield shales are brown-weathering limestones which gradually increase in abundance eastward from the Wilmington-Woodford area where they are mapped as Waits River limestone (Conway formation of Massachusetts). Conditions of deposition favorable to the formation of calcium carbonate prevailed at this time.

The East Dover ultramafic and the other ultramafics of the area may represent intrusions of igneous rock of dunite composition into rocks of the Moretown and Stowe formations. Precisely why, in this area, the ultramafics are concentrated in the Stowe and Moretown formations is unknown.

The deposition of this sequence of Paleozoic strata was followed by a period of orogeny during which the Sadawga Pond dome and the Lake Rayponda dome were developed. These domes show structures which indicate that they were formed by the upward movement of the central gneisses relative to the overlying mantle of schists.

The Paleozoic strata and probably also the later pre-Cambrian rocks were metamorphosed at the same general time in which the domes were formed. The writer infers that this is true for the reason that the rocks east of the Searsburg unconformity are uniformly metamorphosed, and that the doming is attributed to deep-seated intrusions of igneous rocks.

## ECONOMIC GEOLOGY

### Iron Mines

Abandoned iron mines are found in the western part of the area. Two mine-shafts occur one-quarter mile west of Handle Road and south of the abandoned road from West Dover to Somerset Reservoir. The ore is magnetite in a dark schist matrix and was mined for munitions at the time of the Battle of Bennington. Another mine was operated for some years in Stamford, 300 feet west of East Street between Goodrich Brook and Harris Brook. The ore at this abandoned mine was limonite and hematite, and, as far as the writer was able to determine, was mined from the bedrock.

### Other Mines

A number of other abandoned prospects have been indicated on the geological map. These are all located in the eastern half of the area. Two of these are located in southwest Readsboro, one in Searsburg ⅜ mile west of Searsburg Village, and one on the Newfane-Wardsboro boundary on the east side of Baker Brook. The geological setting in which these prospects were made, suggests that they were highly unsuccessful for the shareholders.

### *Marble Quarries*

Marble quarries have been developed exclusively in the Sherman marble member of the Hoosac formation and in the marble in the Heartwellville schist on the North Branch of the Deerfield River. All of these quarries have been abandoned for many years. The largest operation was at Sherman on the Hoosac Tunnel and Wilmington Railroad where lime was quarried in a calcium carbide operation. Marble for agricultural lime has been quarried in the Readsboro anticline and in several places in the Sherman marble of the Whitingham anticline, and in No. 9 Brook in Whitingham. Marble has also been quarried along the Searsburg-Wilmington township line, although not extensively.

Some marble has been quarried for agricultural lime southwest of Binney Brook on the property of Dr. Beebe. Small amounts of marble have been worked for the same purpose in the small deposits along the North Branch of the Deerfield River. Marble deposits of the same types as those described above are found high up on the north slope of Mount Snow (Pisgah) and south of Haystack Pond. To the writer's knowledge, the latter have never been operated.

### *Ultramafic Rocks*

As far as the writer is aware, no commercial use has ever been made of the East Dover ultramafic. Certain sections of this body are reported to contain nickel minerals notably in the vicinity of the junction of Adams and Bemis Brooks. Soapstone has been quarried from the ultramafic west of Stratton Hill and from the ultramafic southeast of Hogback Mountain.

### *Sand and Gravel*

Sand and gravel deposits within the area are extremely small. Those which are found in the northern half of the area are concentrated along the North Branch of the Deerfield River in Dover and Wilmington. One small gravel operation is worked north of the Deerfield River in eastern Searsburg. Small-scale gravel operations utilizing river gravels can be worked in most of the streams of the area, but particularly in Searsburg and Wilmington and in the East Branch of the North River.

Two sand and gravel pits found adjacent to Beaver Brook in eastern

Wilmington and additional gravels in small quantity, could undoubtedly be worked in the terraces along Beaver Brook.

### *Crushed Rock and Concrete Aggregate*

Within the area, crushed stone and concrete aggregate are worked at only two localities. One of these utilizes feldspathic quartzite and gneiss talus blocks ½ mile SW of Searsburg Reservoir. Crushed rock is obtained from the coarse glacial gravels north of the Deerfield River in eastern Searsburg.

The writer is of the opinion that crushed rock and concrete aggregate could be had from some of the microcline gneisses, feldspathic quartzites, and coarser grained, biotite-poor plagioclase gneisses, if it were economically feasible to quarry these rocks for construction purposes. Probably these gneisses would have a Los Angeles rattler test rating of between 25 and 40, with the average being in the vicinity of 33. The gneisses of the Sadawga Pond dome would likewise be suitable with the same qualifications. Some of the core gneisses of Sadawga Pond dome would be of higher quality than others.

## MAJOR PROBLEMS

### *Origin of Domes*

Although many pertinent facts have emerged from recent studies of domes, not only in the United States but also in Scandinavia and other parts of the world, the genesis, causes of, and relationship of domes to orogenesis is, as yet, but imperfectly understood. The problems related to mantled gneiss domes show promise of becoming one of the many fascinating and enlightening problems of structural geology, petrogenesis and orogenesis.

Foremost among the students of domes are Eskola in Finland; Cloos and Broedel in Maryland; Billings in New Hampshire; and Thompson and Rosenfeld in Vermont.

Eskola (1949) has summarized his views on the origin of mantled gneiss domes and concludes that they represent granite intrusions of an earlier orogenic period. These have been eroded, levelled and subsequently subjected to a period of sedimentation. He believes that during a subsequent orogenic cycle the pluton was remobilized and palingenetic granitic magma injected into the pluton contemporaneously with its deformation into gneiss.

Recent investigations in southern Vermont have shown that the domes there consist essentially of a core of gneiss of granodiorite to granite composition, the foliation of which is essentially parallel to the bedding and foliation of the overlying mantle of Lower Paleozoic schists. The attitude of the gneiss is generally concordant with that of the schists. The Chester dome, investigated by Thompson and Rosenfeld (1951), has in its northern part a very small stock of late-tectonic Paleozoic granite intruding the pre-Cambrian gneisses.

The central parts of the domes of the Wilmington area have no granite stocks exposed. In the adjacent Brattleboro quadrangle, a stock of muscovite-biotite granite is found in the northern part of the Guilford dome. Broedel (1937, p. 161) describes domes in Maryland that have granite stocks intruding gneisses of the central part of the domes.

The presence of these granites crosscutting the central gneisses in many domes lends considerable plausibility to Eskola's theory of the causes and mode of development of the domes. With regard, however, to Eskola's belief that the gneisses of the central part of the domes represent earlier granites, certain observations should be made on those of the Wilmington-Woodford area that are relevant to the general problem.

Undoubtedly, what Eskola says is true in areas near the central part of orogenic belts. In the Wilmington area, however, there seems to be ample evidence that the central gneisses of both the Sadawga Pond and Lake Rayponda domes are a bedded sequence. They contain thin schist and quartzite beds throughout. In this area, at least, there seems to be no reason for believing that these rocks were anything but metamorphosed clastic or volcanic rocks. Thus the writer prefers to extend Eskola's concept of core-gneisses of domes to include not only those of igneous but also of sedimentary origin.

In those domes which have stocks of granite intruding the central core gneisses, the granites are for the most part tectonic to post-tectonic. The folding of the gneisses and schists seems to be the result of essentially vertical movements of the central core-gneisses relative to the overlying mantle of schists. Consequently the shear sense in these rocks is such that it was produced by the movement of the upper beds toward the axis of the adjacent syncline relative to the lower beds. The increase in metamorphic intensity in the direction of the central parts of the domes may be ascribed either to heat from the granite of the intruding stocks or to the higher thermal conductivity of the granite and central gneisses over that of the superjacent mantle of schists, or to some other mechanism whereby the hotter zones are brought up in the central parts of the domes.

The distribution of domes and metamorphic zones west of the main mass of igneous intrusives in central New Hampshire suggests a decrease in the intensity of large-scale igneous activity away from central New Hampshire. For example, the map patterns and metamorphic zoning of the Bronson Hill anticline of western New Hampshire is more complex than that of the Vermont domes. Thus, it would not be at variance with known facts to attribute the vertical uplift of the central gneisses, the increase of metamorphism toward the center of the domes, and the minor structures characteristic of such folds, to the intrusion of granite bodies.

## *Roots of the Taconic Klippe*

The Taconic Klippe, as the term is commonly known at present, refers to a large mass of arenaceous and argillaceous sediments of Cambrian and Ordovician age that occupies several hundred square miles in western Massachusetts, western Vermont, and eastern New York. These rocks are underlain by Cambrian and Ordovician rocks of dominantly carbonate composition. The argillaceous and arenaceous sequence forms a mass elongate in a northerly direction for about 165 miles and having an average width of about 35 miles.

On the basis of fossil evidence, this sequence has been considered by some geologists to be entirely allochthonous. These workers believe that this mass of rocks has attained its present position above the carbonate sequence of the same age by being thrust there from the east.

One of the major problems of New England and New York geology is the location of the root zone or source of the Taconic Klippe and the mechanism by which it attained its present position. At present, the precise location of the root zone of the Klippe is unknown.

From his mapping in the north end of the Taconic Range in Vermont, Zen (1956, 1959) concludes that, if the slaty rocks are autochthonous, the structure must be some kind of mushroom fold. But if they are allochthonous, then it must be either a folded simple thrust or a gigantic recumbent fold refolded subsequent to emplacement. Zen (1959) prefers the latter theory that the Taconic rocks represent a refolded nappe further complicated by faulting and may have been derived from the crest of the rising Green Mountain anticlinorium.

Other geologists believe that small-scale faulting has occurred in areas where the arenaceous and argillaceous sequence is known to overlie the calcareous unit of the same age. The argillaceous rocks, they believe, are interbedded with carbonate rocks and where faulted on a small scale may give the appearance of large-scale movements. More detailed and diagnostic paleontologic data and extremely detailed studies of the structure must be undertaken before the problems related to the Taconic Klippe are resolved even in a general way.

Up to the present, some study of the paleontology and stratigraphy has been made in the Taconic Klippe, but structural and petrologic data are extremely meager. Before any controlled speculation can be made on the size and source of the Klippe, these studies must be carried to completion.

Geologic mapping in the Wilmington-Woodford area, in the North Adams and Cheshire quadrangles, and in certain other areas to the south, suggests that there is a thrust fault along the western base of Hoosac Mountain at least as far north as Heartwellville.

The Hoosac thrust, recognizable in the southern part of the Green Mountains where it cuts across younger pre-Cambrian, Cambrian and Ordovician rocks, may pass northward into the central portion of the Green Mountains. Here it may disappear partially because it is not recognized in the pre-Cambrian area or more probably because the rising crest of the Green Mountain anticlinorium involved in the thrusting, is now eroded to depths where the plane of the thrust is not recognizable as such. More detailed studies of the Green Mountain core may ultimately shed more light on this fascinating problem.

At the present time, however, there are certain data available that must be considered in any attempt to solve the problem of the existence of the Taconic Klippe or the location of its root zone. These are presented briefly as follows:

(1) The rocks of the Klippe are essentially an argillaceous and arenaceous sequence of Cambrian and Ordovician age.

(2) They are lithologically similar to certain more highly metamorphosed rocks east of the Green Mountains, namely parts of the Hoosac and Pinney Hollow formations and possibly others also.

(3) The Taconic sequence is somewhat less metamorphosed than the rocks east of the Green Mountains and yet the grade of metamorphism seems to be somewhat higher than generally considered by earlier writers. If these rocks are derived from a place east of the Green Mountains they must have been moved to their present position either before metamorphism was completed or, if afterwards, they must have their place of origin in an area not so greatly affected by metamorphism.

(4) The Green Mountain anticlinorium is somewhat overturned to the west. Domes of the eastern flank of the Green Mountains are either vertically uplifted or overturned somewhat to the west. Thus, if rocks of the

Klippe attained their place as part of a nappe originating east of the Green Mountains and if large parts of the limbs of the nappe were eroded away, one would expect that no root zone would be detectable. If, however, the Klippe were not part of a nappe but were thrust as a rigid block, fewer predictable restrictions would be placed on the character of the root zone. It should, in that event, however, be a prominent structural feature and should be detected by careful geologic mapping of the source area.

(5) The average distance between the western border of the pre-Cambrian core of the Green Mountains and the eastern flank of the Klippe in Vermont is less than five miles.

(6) The northern limit of the Klippe is about due west of the northern extent of the pre-Cambrian rocks of the Green Mountains.

(7) Prindle (Prindle and Knopf, 1932) mapped a fault at the western base of Hoosac Mountain, in the Taconic quadrangle (Fig. 2). He believed that this fault marks the contact of autochthonous carbonate rocks of the Berkshire Valley sequence and the overthrust sequence derived from the east. Prindle regarded this fault as the root zone of the Taconic Klippe. He mapped the fault as far north as the western border of the Wilmington area. If it were projected northeasterly into the Wilmington area, it would continue into the region of pre-Cambrian rocks of the central Green Mountain core or follow the prominent discontinuity shown in Searsburg and Somerset of Plate II. A great deal of careful mapping and correlation of structural and petrologic data must take place before the problems of the Taconic Klippe are either solved or brought into closer focus.

(8) In Eastern New York Bucher (1951, pp. 18–19) describes an unconformity between the Wappinger limestone-Hudson River slates sequence and older rocks. Thus, it is also possible that the Heartwellville schist, Sherman marble and most of the albite schists are equivalents of this sequence since they too seem to be resting unconformably on older rocks. It is possible that the Hoosac fault mapped by Prindle (op. cit.) and the writer at the base of Hoosac Mountain, in the Taconic quadrangle (Fig. 1), is developed essentially along an unconformity.

## SUGGESTED FUTURE INVESTIGATIONS IN THE WILMINGTON-WOODFORD AREA

Future investigation in the Wilmington area should emphasize among other methods that of chemical analyses of a more extensive nature than was possible during the present study. Analyses of ferromagnesian minerals and garnets of arenaceous, argillaceous and mafic rocks should be undertaken.

In mafic rocks, chemical analyses should be made on the above-mentioned minerals in both garnet- and non-garnet-bearing types. These analyses will provide greater control over the composition of various minerals and establish for this area their relation to the grade of metamorphism.

Detailed mapping in the central core of the Green Mountains should be undertaken. There is sufficient lithologic variety in the northeastern portion of the early pre-Cambrian region to suggest that greater variations may also be present in the area mapped as microcline and plagioclase gneisses. In many parts of the world, these early pre-Cambrian rocks have yielded economically valuable mineral deposits. The writer has found no evidence to suggest that such deposits are present here, but this possibility should not be overlooked.

The later pre-Cambrian rocks of Dover which contain magnetite bodies mentioned above may elsewhere contain small to large deposits of iron ore. For this reason, it is suggested that magnetic studies be undertaken in these older regions.

Magnetic studies of a reconnaissance nature have been initiated by the writer and undertaken by some of his students. Indications of extremely high magnetic values have been obtained in the area immediately west of the above-mentioned magnetite mines of Dover. In addition, high magnetic values have been recorded in reconnaissance traverses in the vicinity of the ultramafic bodies of the eastern part of the area. This is to be expected, but excessively high magnetic values may indicate economically valuable concentrations of magnetite and/or chromite.

The writer suggests that an extremely interesting and valuable investigation would be more detailed studies in the vicinity of the Hoosac thrust, the Searsburg unconformity, and the possible continuation of the Hoosac thrust in Woodford and Glastenbury. These studies would undoubtedly shed more light on the nature of the thrusting and the relationships in this area between the earlier and later pre-Cambrian rocks.

A detailed study of facies changes particularly in the sequence from the Searsburg conglomerate through the Wilmington gneiss and the Hoosac formation will yield valuable information on the sedimentary, volcanic and tectonic history of this region. Such a study would undoubtedly clarify the solution of most of the more pressing problems within the area.

# BIBLIOGRAPHY


ALBEE, ARDEN L. (1957) Geology of the Hyde Park Quadrangle, Vt., U. S. Geol. Surv., Geol. Quad. Map GQ 102.

ASHLEY, G. H. (1932) Stratigraphic nomenclature: Geol. Soc. Am. Bull., v. 43, pp. 469–476.

—— (1933) Classification and Nomenclature of rock units: Geol. Soc. Am. Bull., v. 44, pt. 2, pp. 423–459.

BAIN, G. W. (1926) Geological history of the Green Mountain front: Vt. Geol. Survey, State Geologist 15th Rpt. (1925–26), pp. 222–241.

—— (1931) Flowage Folding: Amer. Jour. Sci., 5th ser., vol. 22, pp. 503–530.

—— (1936) Serpentinization of Vermont Ultrabasics: Geol. Soc. Am. Bull., v. 47, pp. 1961–1980.

BALK, R. (1946) Gneiss Dome at Shelburne Falls, Mass.: Geol. Soc. Am. Bull., v. 57, pp. 125–159.

—— (1956) Bedrock Geology of the Massachusetts Portion of the Bernardston Quadrangle, Mass.-Vt.: U. S. Geol. Surv., Geol. Quad. Map GQ 90.

—— (1956) Bedrock Geology of the Massachusetts Portion of the Northfield Quadrangle, Mass.-Vt.: U. S. Geol. Surv., Geol. Quad. Map GQ 92.

—— (1956) Bedrock Geology of the Millers Falls Quadrangle, Mass.: U. S. Geol. Surv., Geol. Quad. Map GQ 93.

BEAN, R. J. (1953) Relations of gravity anomalies to the geology of central Vermont and New Hampshire: Geol. Soc. Amer., Bull., vol. 64, pp. 509–538.

BILLINGS, M. P. (1937) Regional metamorphism of the Littleton-Moosilauke area, New Hampshire: Geol. Soc. Amer. Bull., v. 48, pp. 463–566.

—— (1950) Structural Geology: Prentice-Hall, Inc., 473 pp.

—— (1954) Structural Geology: Prentice-Hall, Inc., 514 pp.

——, RODGERS, J., and THOMPSON, J. B. (1952) Geology of the Appalachian Highlands of East-Central New York, southern Vermont and southern New Hampshire: Geol. Soc. Amer., Guidebook for field trips in New England, 65th Ann. Meeting, 71 p.

BOUCOT, A. J., MACDONALD, G. J. F., MILTON, C., THOMPSON, JR., J. B. (1958) Metamorphosed middle Paleozoic fossils from central Mass., eastern Vermont, and western N. H.: Geol. Soc. Amer., Bull., v. 69, pp. 855–864.

BRACE, W. F. (1953) The Geology of the Rutland area, Vermont: Vt. Geol. Surv., Bull. 6, 120 p.

BROEDEL, C. H. (1937) The Structure of the Gneiss Domes near Baltimore, Maryland: Md. Geol. Surv., v. XIII, pp. 149–183.

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. Amer., Bull., v. 62, pp. 1426–1427.

BURT, F. A. (1929) The origin of the Bennington Kaolins: Vt. State Geol. Rpt. 16, pp. 65–84.

CADY, WALLACE M. (1956) Stratigraphic Relationships in Northern Vermont and Southern Quebec (abs): Geol. Soc. Am. Bull., v. 67, pp. 1811–1812.

CHANG, P. H. (1950) Geology of the Woodstock, Vermont quadrangle; Unpub. Ph.D. Thesis, Harvard University; data of Geol. Map incorporated in Vt. Geol. Surv. Bull. 6, 120 p.

CHAPMAN, C. A. (1942) Intrusive domes of the Claremont-Newport area, N. H.: Geol. Soc. Am. Bull., v. 53, pp. 889–915.

CHIDESTER, A. H., BILLINGS, M. P., and CADY, W. M. (1951) Talc Investigations in Vermont: Prelim. Rept., U. S. Geol. Surv., Circ. 95., 33 pp.

CLOOS, E. (1937) The Application of Recent Structural Methods in the Interpretation of the Crystalline Rocks of Maryland: Md. Geol. Surv., v. XIII, pp. 27–105.

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

—— (1947) Boudinage: Trans. Geop. U., v. 28, no. 4, pp. 626–632.

CURRIER, L. W. and JAHNS, R. (1941) Ordovician stratigraphy of Central Vermont: Geol. Soc. Am., Bull., v. 52, pp. 1487–1512.

DALE, T. N. (1896) Structural details in the Green Mountain region and in eastern New York: U. S. Geol. Surv., 16th Ann. Rept., pt. 2, pp. 543–570.

—— (1898) The Slate belt of eastern New York and western Vermont: U. S. Geol. Surv., 19th Ann. Rept., pt. 3, pp. 153–307.

—— (1914–1916) Field Notes on the Algonkian-Cambrian boundary east of the Green Mountain Axis in Vermont: Open file in U. S. Geol. Surv. Office, Boston, Mass.

—— (1915) The Calcite Marble and Dolomite of Eastern Vermont: U. S. Geol. Surv. Bull. 589, 67 p.

—— (1916) The Algonkian-Cambrian boundary East of the Green Mountain Axis in Vermont: Amer. Jour. Sci., v. XLII, no. 248, pp. 120–124.

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

DOLL, CHAS. G. (1943) A Paleozoic Revision in Vermont: Am. Jour. Sci., v. 241, pp. 57–64.

—— (1944) A Preliminary report of the Geology of the Strafford Quadrangle, Vermont: Vt. Geol. Survey, State Geologist 24th Rpt. (1943–44), pp. 14–28.

EMERSON, B. K. (1898a) Geology of Old Hampshire County, Mass.: U. S. Geol. Surv., Mon. 29, 790 p.

—— (1898b) Geology of the Hawley, Mass.-Vermont Quadrangle: Unpub. U. S. Geol. Surv., Folio O.

—— (1917) Geology of Massachusetts and Rhode Island: U. S. Geol. Surv., Bull. 597, 289 p.

ESKOLA, P. (1949) The Problem of Mantled Gneiss Domes: Quart. Jour. Geol. Soc. Lond., v. 104, pp. 461–476.

FOWLER, P. (1950) Stratigraphy and Structure of the Castleton Area, Vermont: Vt. Geol. Surv. Bull. 2, 83 pp.

FOYE, W. G. (1949) The Geology of Eastern Connecticut: State of Conn., St. Geol. and Nat. Hist. Surv. Bull. 74, 100 p.

GORDON, C. E. (1914) Notes on the geology in the vicinity of Bennington, Vermont: Vt. Geol. Survey, State Geologist 9th Rpt. (1914–15), pp. 337–371.

—— (1921) Studies on the geology of western Vermont: Vt. Geol. Survey, State Geologist 12th Rpt. (1920–21), pp. 114–279.

HADLEY, J. B. (1949) Bedrock geology of the Mt. Grace Quadrangle, Mass.: U. S. Geol. Surv., Quad. Map GQ 4.

HALL, A. L. and MOLENGRAAFF, G. F. (1925) The Vredefort Mountain Land: Shaler Mem. Ser., Amsterdam, 183 pp.

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

—— (1938) A primary peridotite magma: Am. Jour. Sci., 5th ser., v. 35, no. 209, pp. 321–344.

HERZ, N. (1958) Bedrock Geology of the Cheshire Quadrangle, Mass.: U. S. Geol. Surv., Geol. Quad. Map GQ 108.

—— (1961) Bedrock Geology of the North Adams Quadrangle, Mass.: U.S. Geol. Surv., Geol. Quad. Map GQ 139.

HITCHCOCK, E. et al (1861) Report on the Geology of Vermont: Vt. Geol. Surv., 2 vols., 982 p.

HUBBARD, G. D. (1924) Geology of a small tract in south-central Vermont: Vt. Geol. Survey, State Geologist 14th Rpt. (1923–24), pp. 260–344.

HURLBUT, C. S. (1959) Dana's Manual of Mineralogy: 17th ed., John Wiley and Sons, Inc., 609 p.

JACOBS, E. C. (1944) General Petrography in Strafford Township, Vermont: Vt. Geol. Survey, State Geologist 24th Rpt. (1943–44), pp. 29–37.

KEITH, A. (1914) A pre-Cambrian Unconformity in Vermont (abs.): Geol. Soc. Am. Bull., v. 25, p. 39.

KNOPF, E. B. (1935) Recognition of overthrusts in metamorphic terranes: Am. Jour. Sci., 5th ser., v. 30, pp. 198–209.

KRUGER, F. C. (1946) The structure and metamorphism of the Bellows Falls quadrangle of New Hampshire and Vermont: Geol. Soc. Am. Bull., v. 57, pp. 161–206.

LEITH, C. K. (1905) Rock Cleavage: U. S. Geol. Surv. Bull. 239, 216 p.

LYONS, J. B. (1955) Geology of the Hanover quadrangle, New Hampshire-Vermont: Geol. Soc. Am. Bull., v. 66, pp. 105–145.

MACFADYEN, JR., J. A. (1956) The Geology of the Bennington area, Vermont: Vt. Geol. Surv. Bull. 7, 72 p.

MAYNARD, J. E. (1954) The petrographic reexamination of quartz-bearing plutonites from Vermont: Jour. Geol., v. 42, pp. 146–162.

MOORE, R. C. (1947) Stratigraphic Commission; Note 1, Organization and objectives of the Stratigraphic Commission; Note 2, Nature and classes of stratigraphic units: Am. Assoc. Petrol. Geol., v. 31, pp. 513–528.

MOORE, G. W. (1950) Structure and metamorphism of the Keene-Brattleboro area, New Hampshire-Vermont: Geol. Soc. Am. Bull., v. 60, pp. 1613–1670.

OSBERG, P. H. (1950) The Green Mountain Anticlinorium in the vicinity of Rochester and East Middlebury, Vermont: Vt. Geol. Surv. Bull. 5, 127 p.

—— (1956) Stratigraphy of the Sutton Mts., Quebec: Key to stratigraphic correlation in Vermont: Geol. Soc. Am. Bull., v. 67, p. 1820.

—— (1959) The Stratigraphy and structure of the Coxe Mountain Area, Vt., Guidebook for the 51st Annual Meeting of the New England Intercollegiate Geological Conference, pp. 45–51.

PERRY, E. L. (1928) The Geology of Bridgewater and Plymouth Townships, Vermont: Vt. Geol. Survey, State Geologist 16th Rpt. (1927–28), pp. 1–64.

PHILLIPS, A. H. and HESS, H. H. (1936) Metamorphic Differentiation at contacts between Serpentinite and Siliceous Country Rocks: Amer. Min., v. 21, no. 6, pp. 333–362.

PRINDLE, L. M. and KNOPF, E. B. (1932) Geology of the Taconic Quadrangle: Am. Jour Sci., 5th ser., v. 24, pp. 257–302.

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

QUINN, A. and HENDERSON, D. C. (1945) Report on the Plainfield-Hawley area: On open file in U. S. Geol. Surv. Office, Boston, Mass.

ROSENFELD, J. R. (1954) Geology of the southern part of the Chester Dome, Vt.: Ph.D. thesis, Harvard U., 303 p.

SEGERSTROM, K. (1956) Bedrock Geology of the Colrain Quadrangle, Mass.: U. S. Geol. Surv., Geol. Quad. Map GQ 86.

—— (1956) Bedrock Geology of the Shelburne Falls Quadrangle, Mass.: U. S. Geol. Surv., Geol. Quad. Map GQ 87.

SHUMAKER, R. (1959) Pawlet Quadrangle: Guidebook for the 51st Annual Meeting of the New England Intercollegiate Geological Conference, pp. 59–60.

SKEHAN, J. W. (1952) Structural geology of the Wilmington, Vermont area (abs): Geol. Soc. Am. Bull. v. 63, pp. 1297–1298.

—— (1953) Anticlines and Domes of the Wilmington, Vermont area (abs.): Geol. Soc. Am. Bull., v. 64, p. 1565.

—— (1960) Major and Minor Structures of the Wilmington, Vermont Area, U. S.: Rept. of XXIst Sess. of Int. Geol. Cong., Norden. Vol. of Abs., p. 190.

—— (1961) Stratigraphy and Structure in the Southern Part of the Green Mountain Anticlinorium (abs.): Geol. Soc. Am. Bull., v. 71, no. 12, Pt. 2, p. 1975.

THOMPSON, J. B. (1950) Geology of the Ludlow, Vermont area: Unpub. Ph.D. thesis, Mass. Inst. Tech.

—— and ROSENFELD, J. R. (1951) Tectonics of a Mantled Gneiss Dome in southern Vermont (abs.): Geol. Soc. Am. Bull., v. 62, pp. 1484–1485.

—— (1954) Structural geology of the Skitchewaug Mt. area, Claremont Quadrangle, New Hampshire-Vermont: New England Intercollegiate Geol. Conf. Guidebook. 46th Conf., Dartmouth College, Hanover, N. H.

—— (1956) Skitchewaug Nappe, a major Recumbent Fold in the area near Claremont, New Hampshire (abs.): Geol. Soc. Am. Bull., v. 67, pp. 1826–27.

—— (1959) Stratigraphy and structure in the Vermont Valley and the Eastern Taconics between Clarendon and Dorset: Guidebook for the 51st Annual Meeting of the New England Intercollegiate Geological Conference, pp. 71–87.

WALCOTT, C. D. (1888) The Taconic System of Emmons: Am. Jour. Sci., 3rd ser., v. 35, pp. 307–327.

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

—— (1949) Cleavage in east-central Vermont: Trans. Am. Geop. U., v. 30, p. 587.

WHITTLE, C. H. (1894) The occurrence of Algonkian rocks in Vermont and evidence for their subdivision: Jour. Geol., v. 2, pp. 396–429.

WILLARD, M. E. (1951) Bedrock Geology of the Mount Toby Quadrangle, Mass.: U. S. Geol. Surv., Geol. Quad. Map GQ 9.

—— (1952) Bedrock Geology of the Greenfield Quadrangle, Massachusetts: U. S. Geol. Surv., Geol. Quad. Map GQ 20.

—— (1956) Bedrock Geology of the Williamsburg Quadrangle, Mass.: U. S. Geol. Surv., Geol. Quad. Map GQ 85.

WINCHELL, A. N. (1951) Elements of Optical Mineralogy: John Wiley and Sons, Inc., 551 p.

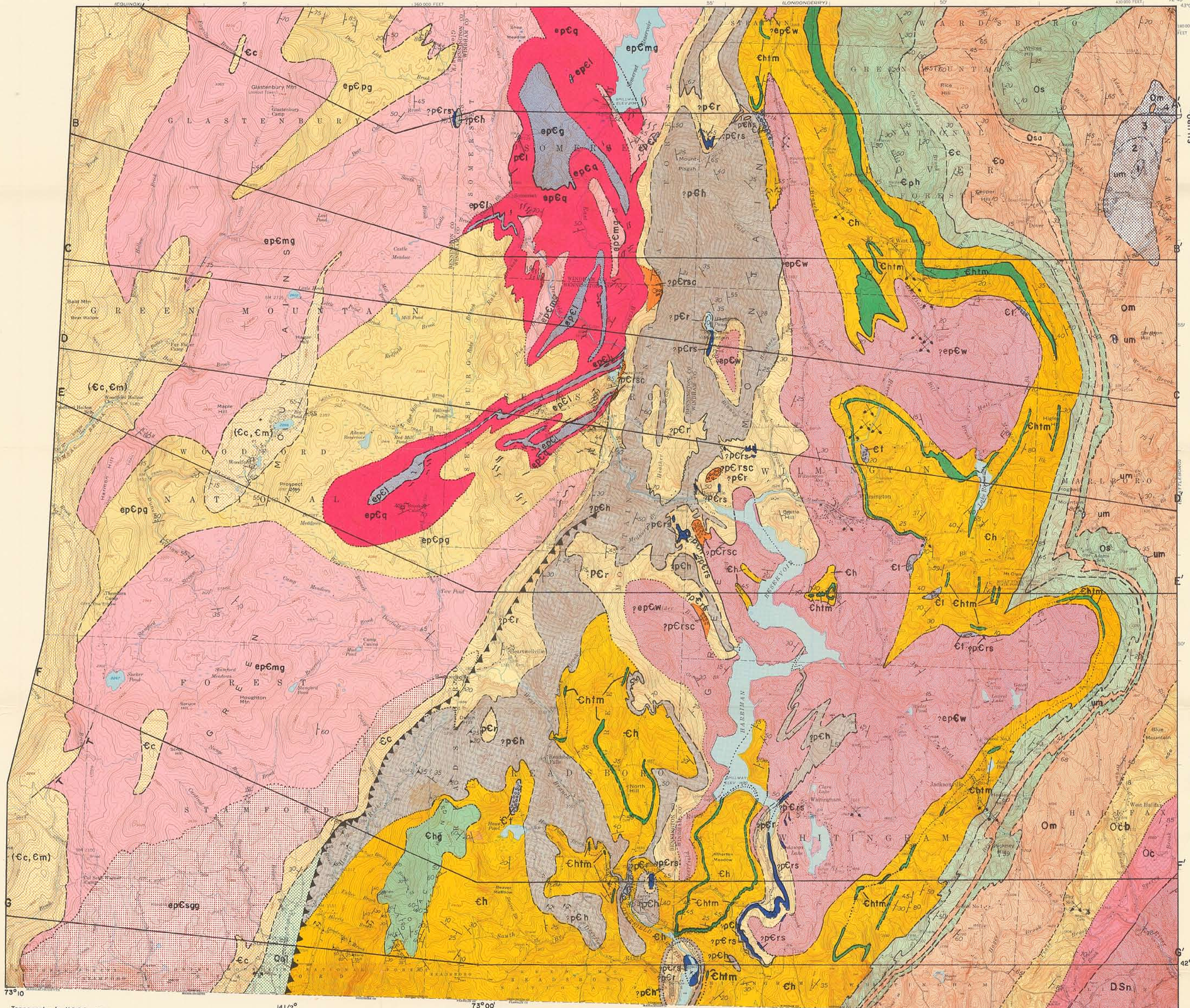
WOLFF, J. E. (1891) On the Lower Cambrian age of the Stockbridge limestone: Geol. Soc. Am. Bull., v. 2, pp. 331–337.
ZEN, E-AN (1956) Stratigraphy and structure of the North End of the Taconic Range, Vermont (abs.): Geol. Soc. Am. Bull., v. 67, pp. 1829–30.
—— (1959) Stratigraphy and structure at the North End of the Taconic Range and adjacent areas: Guidebook for the 51st Annual Meeting of the New England Intercollegiate Geological Conference, pp. 1–15.

Plate 1

BENNINGTON QUADRANGLE
VERMONT
15 MINUTE SERIES (TOPOGRAPHIC)

UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

WILMINGTON QUADRANGLE
VERMONT
15 MINUTE SERIES (TOPOGRAPHIC)




Topography by U.S.G.S. 1942 and 1951
Control by U.S.G.S. U.S.C.&G.S. and U.S.F.S.

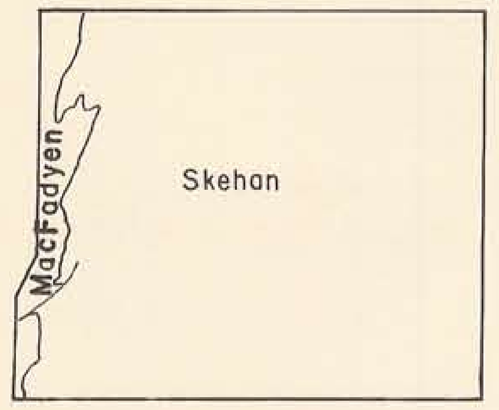
14 1/2°
True North
Magnetic North
APPROXIMATE MEAN DECLINATION 1954

Scale $\frac{1}{62500}$

1 1/2 0 1 2

Contour interval 20 feet
Datum is mean sea level

# GEOLOGIC MAP OF THE WILMINGTON - WOODFORD AREA, VT.




Geology by James W. Skehan, S.J., assisted by A. Albee, E. Weinberg, H. Groom J. Palacas, W. Luddy, R. Sheehan, and M. Kelley. Western boundary by John A. MacFadyen, Jr. Geology surveyed in 1949-51, 1958 and 1959.

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 17)

## LEGEND

Qal

Quaternary alluvium (Qal): stratified glacial sand and gravel deposits overlain locally by recent alluvial sand gravel deposits.

SILURO-DEVONIAN

DSn

Northfield slate (DSn): thin beds of brown weathering, dark gray, impure limestone; relatively uniform sequence of dark gray to black, fissile, muscovite-garnet slate.

----------------------Unconformity----------------------

ORDOVICIAN

um

Ultramafic rocks (um): talc-carbonate bodies of small size; zoned talc-carbonate-serpentines of larger size; zoned talc-carbonate-serpentine-dunite body of very large size; Zone 1 is 90% or more olivine; Zone 2 is 60-90%; and Zone 3 is 30-60% olivine; Zone 4 is 0-70% talc-carbonate and 30-100% serpentine.

Oc Ocb

Cram Hill formation (Oc): light to dark gray, banded, feldspathic muscovite schists including blue quartz pebble conglomerate; thin beds of coarse-grained, massive, light green gneiss; black, very fine-grained rusty weathering muscovite-pyrite slates. (Ocb): Barnard gneiss, dominantly amphibolites including typically massive, dark green, poorly banded, porphyritic and non-porphyritic amphibolites; and light green "amygdaloidal" amphibolites.

Om

Moretown formation (Om): thin beds of banded epidote amphibolites; thin coarse-grained garnet amphibolites; dark gray to black biotite-muscovite-quartz schists; gray to light green, well-banded biotite-muscovite-quartz schists.

Os Osa

Stowe formation (Os): pale green, chlorite-muscovite-magnetite schist with thin, banded, epidote amphibolites and greenstone beds. A prominent greenstone member (Osa) at the base of the formation is mapped in the Cooper Hill locality where the formation is unusually broad.

CAMBRIAN AND/OR ORDOVICIAN

Ꞓo

Ottauquechee formation (Ꞓo): rusty weathering, dark gray to black muscovite-chlorite schist; thin beds of gray to black quartzite; thin beds of well-foliated chlorite-muscovite-magnetite schist; well-banded, fine-grained, feldspathic-chlorite schist.

Ꞓc

Chester amphibolite (Ꞓc): banded, well-foliated, epidote-chlorite amphibolite; thin beds of dark gray to black schist and chlorite-garnet-muscovite schist.

Ꞓph

Pinney Hollow formation (Ꞓph): pale green, well-foliated chlorite-muscovite-magnetite schists; thin beds of well-banded epidote-chlorite amphibolite.

Ꞓh Ꞓhg Ꞓhtm (Ꞓc, Ꞓm) Ꞓh

Hoosac formation (Ꞓh): medium to coarse-grained, rusty weathering muscovite-chlorite-biotite-garnet-quartz schists; thin graphite-albite and non-albitic muscovite schist; thin interbedded amygdaloidal amphibolites (Ꞓhtm); muscovite-chlorite green schist mapped separately (Ꞓhg).

(Cheshire quartzite, Ꞓc): grades laterally into rocks of Hoosac lithology on east flank of Green Mountains in Stamford; massive, glassy white and buff quartzite.

(Mendon formation, Ꞓm): black, fine-grained, banded, sericite-biotite-chloritoid phyllite with thin interbeds of gray quartzite representing the Moosalamoo member.

Turkey Mountain member (Ꞓhtm): dense, dark green amphibolites spotted gray to green or dark brown by "amygdules" of fine-grained quartz and albite filled with epidote, hornblende and garnet.

Ꞓt

Tyson formation (Ꞓt): fine to coarse-grained, schistose, white to blue quartz conglomerate; fine to coarse-grained, gray, buff and pink microcline, and coarse-grained albite-pebble conglomerate; quartzites.

----------------------Unconformity (?)----------------------

LATE PRE-CAMBRIAN

?pꞒh ?pꞒhs

Heartwellville schist (?pꞒh): coaly black to dark gray graphite-muscovite-(chlorite)-garnet-quartz lensed schist; light to dark green chlorite-muscovite-(garnet)-quartz lensed schist. In western Dover marbles lithologically identical with those of Sherman member are mapped as ?pꞒhs.

?pꞒrs ?pꞒr ?pꞒrsc

Sherman marble member of Readsboro formation (?pꞒrs): coarse-grained, white or pink, siliceous calcite marble, and fine-grained, siliceous dolomite marble; actinolite-diopside-phlogopite lime-silicate granulites.

Readsboro schist (?pꞒr): rusty to gray garnet-chlorite-biotite-quartz-albite augen schist and gneiss; graphitic near base.

Searsburg conglomerate (?pꞒrsc): thin, white to buff glassy quartzite; blue and white quartz conglomerate; albite and microcline conglomerate; micaceous white quartzites; coarse feldspathic, arkosic conglomerates, and coarse albite schists.

---------------Profound Angular Unconformity---------------

EARLY PRE-CAMBRIAN

Mount Holly complex:

epꞒg

Plagioclase gneiss (epꞒg): blue and white quartz-plagioclase gneiss of Searsburg and Somerset, more closely associated with epꞒl, epꞒq, than epꞒpg.

epꞒl

Lime-silicates and white gneiss (epꞒl): fine to coarse lime-silicate; blue and white quartz-rich white gneiss.

epꞒq

Quartzite, conglomerate and conglomeratic gneiss (epꞒq): massive buff to blue glassy quartzite; blue quartz conglomerate; "pebbly" rounded to angular microcline conglomeratic gneiss.

?epꞒw epꞒsgg

Wilmington gneiss (?epꞒw): Stamford granite gneiss (epꞒsgg): coarse gray, buff and pink microcline-augen and porphyritic-microcline in quartz-albite-microcline-biotite-epidote groundmass.

----------------------Unconformity (?)----------------------

epꞒpg

Plagioclase gneiss (Harmon Hill gneiss) (epꞒpg): blue and white quartz-plagioclase gneiss of Green Mountain core.

epꞒmg

Microcline gneiss (epꞒmg): banded microcline augen gneiss; porphyritic microcline gneiss; few amphibolites.

Contact

Dashed where gradational or inferred in part; dotted where inferred or concealed.

35 inclined vertical

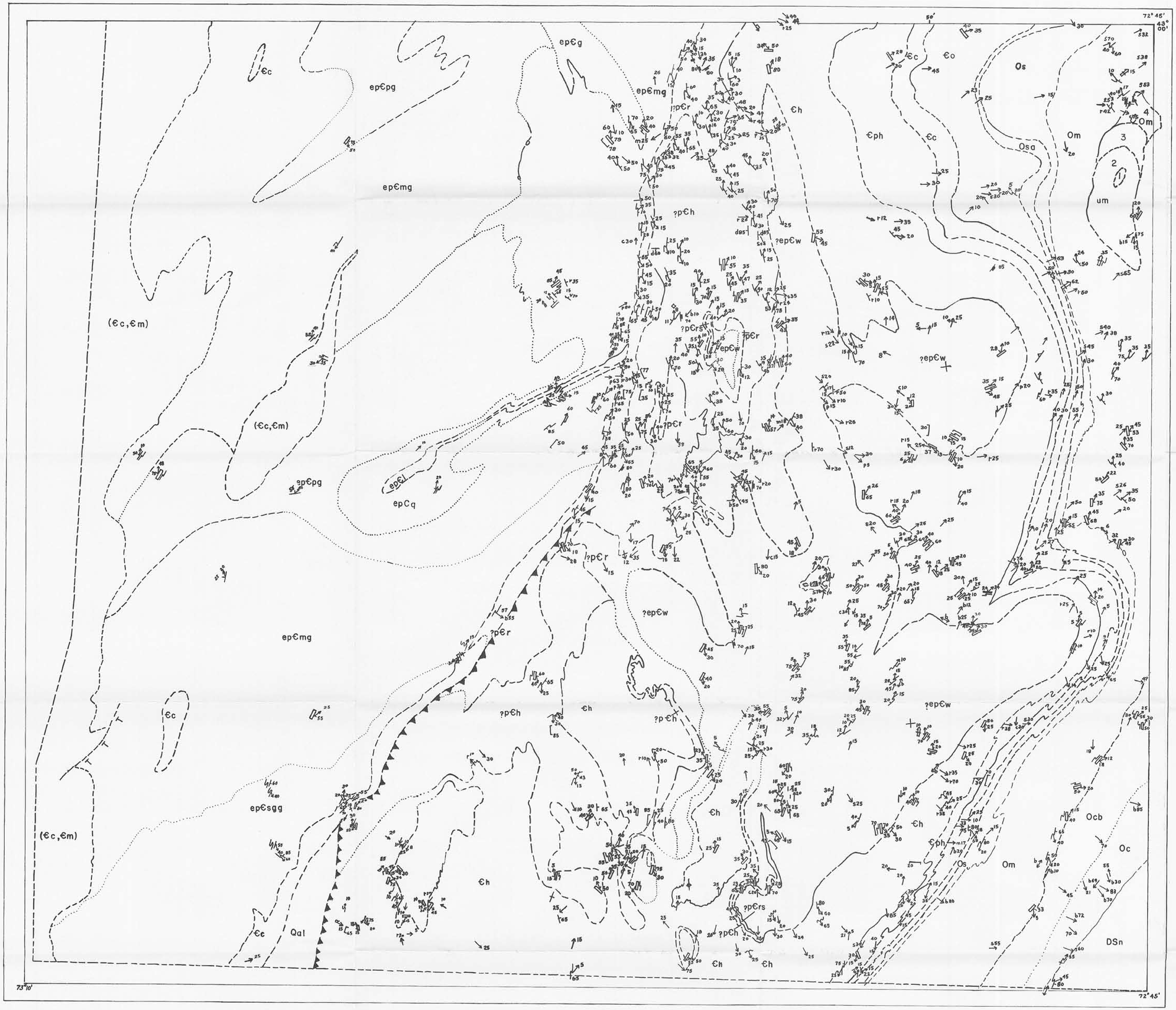
Strike and dip of bedding, including normal and inverted strata.

45

Strike and dip of schistosity.

Abandoned mines and prospects.

Sand and gravel pits.




## LEGEND

- Attitude of compositional banding or bedding.
- Vertical compositional banding or bedding.
- Horizontal compositional banding or bedding.
- Attitude of compositional banding or bedding showing pattern of folds in plan view.
- Attitude of compositional banding or bedding showing rapid variation from west to east; attitude of western part of outcrop indicated by upper part of symbol and that of eastern part by lower.
- Attitude of highly folded compositional banding or bedding in which pattern is unknown.
- Attitude of plunging lineation.
- Direction of trend of horizontal lineation
- Attitude of compositional banding or bedding showing pattern of folds and attitude of fold axis.
- Attitude of compositional banding or bedding showing pattern of folds and attitude of fold axis where either the pattern or fold axis changes direction in same or adjacent outcrops.
- Attitude of axial plane of fold.
- Vertical axial plane showing direction and value of plunge.
- Horizontal axial plane.
- Attitude of slip cleavage.
- Vertical slip cleavage.
- Horizontal slip cleavage.
- Attitude of flow cleavage.
- Vertical flow cleavage.
- Horizontal flow cleavage.
- Attitude of joints.
- Attitude of relatively large quartz veins.

Subscripts for various types of lineation:
no subscript for fold.

- g – Rotational axis of garnet or other mineral.
- b – Boudinage.
- s – Streaming.
- r – Quartz rodding.
- m – Long axis of mineral.
- p – Long axis of pebble.

# PL II. TECTONIC MAP OF THE WILMINGTON-WOODFORD, VERMONT AREA

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Plate 3


## STRUCTURE SECTIONS OF THE WILMINGTON-WOODFORD AREA, VT.

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