# GEOLOGY OF THE CAMELS HUMP QUADRANGLE VERMONT

*By*

ROBERT A. CHRISTMAN

AND

DONALD T. SECOR, JR.

VERMONT GEOLOGICAL SURVEY

Charles G. Doll, *State Geologist*

BULLETIN NO. 15 1961

# CONTENTS

## *List of Illustrations*

## *Figures*



## *Tables*

# GEOLOGY OF THE CAMELS HUMP QUADRANGLE VERMONT

*By*

ROBERT A. CHRISTMAN* AND DONALD T. SECOR, JR.**

## ABSTRACT

The rocks in the Camels Hump quadrangle were deposited in a Cambrian-Ordovician eugeosyncline principally as graywacke, siltstone and shale with minor amounts of quartzitic, volcanic and carbonaceous material. Units mapped are the Camels Hump group (Cambrian), consisting of the Pinnacle, Underhill, Hazens Notch and Pinney Hollow formations, the Ottauquechee formation (Cambrian) and the Stowe formation (Ordovician). The lower part of the section is characterized by coarser sediments, exemplified by the graywacke of the Pinnacle formation, and the upper part contains fine-grained and carbonaceous sediments exemplified by quartz-sericite schist and graphitic phyllite of the Hazens Notch and Ottauquechee formations.

Volcanic material, principally in the form of ash and debris, forms parts of the Underhill, Ottauquechee and Stowe formations. The only identified lava flows are found in the Underhill formation and a chemical analysis of one indicates that it is a spilite. A few small intrusions of ultrabasic rock and basic dikes occurred at a younger date.

The rocks have been complexly folded with the Green Mountain anticlinorium being the major structural feature. Large isoclinal folds also occur, particularly on the west side of the Green Mountains. Bedding schistosity, fracture cleavage and drag folds are the predominant small-scale features.

Accompanying the deformation the rocks were metamorphosed to the greenschist facies. The rocks along the anticlinorium are most highly metamorphosed, as indicated by garnet, and the rocks in the western part of the area are weakly metamorphosed, as indicated by original quartz grains and graded bedding which may still be recognized in

*The senior author is now Associate Professor and Chairman of the Geology Department at Western Washington State College.

**The junior author is now a graduate student at Stanford University.

places. The volcanic beds were recrystallized principally to calcareous greenstone and the lavas were recrystallized to amphibolitic greenstone. Chemical analysis and optical data suggest that the metamorphic amphibole from the spilitic lava is a calciferous hornblende, variety femaghastingsite. This is somewhat anomalous as the amphibole to be expected to occur with albite, epidote and chlorite in the greenschist facies is actinolite.

# INTRODUCTION

## General Statement

The present study of the Camels Hump quadrangle was undertaken to obtain an understanding of the geology of the area. Despite the easy accessibility and prominence of the various peaks in this part of the Green Mountains little geologic work had been done. With the completion of the geologic studies in adjoining quadrangles, it became desirable to complete the picture with this study. The complex folding and unsolved problems in stratigraphy and petrology made the area an interesting one for research.

The rocks occurring in the quadrangle are rather uniform schist, phyllite and metagraywacke of the greenschist facies which have resulted from the regional metamorphism of siltstone, shale and graywacke. The rocks were deposited in a eugeosyncline of Cambrian-Ordovician age. Associated with them are a few lava flows and a few small ultrabasic bodies and basic dikes which were intruded at a younger date.

## Location of Area

The Camels Hump quadrangle is located in northwestern Vermont (Fig. 1) between 44°15′ and 44°30′ north latitude and 72°45′ and 73°00′ west longitude. It includes parts of Addison, Chittenden, Lamoille and Washington counties, all of Bolton township and parts of Duxbury, Fayston, Huntington, Jericho, Moretown, Richmond, Starksboro, Stowe, Underhill and Waterbury townships.

## Regional Geologic Setting

The most prominent structural feature in the region is the Green Mountain anticlinorium, the axis of which extends roughly north-south through the center of the Camels Hump quadrangle. On the west side of the axis north of the quadrangle, the Cambridge syncline and the Enosburg Falls or Fletcher anticline have been identified (Christman

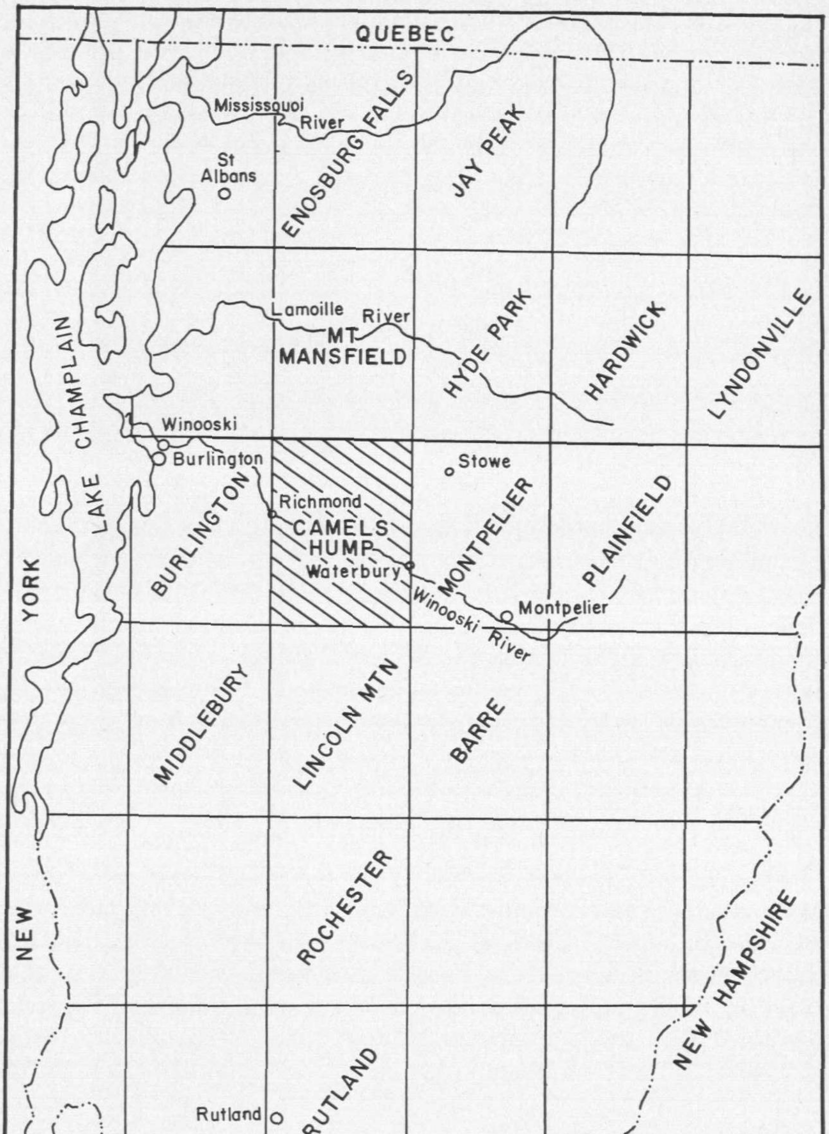



Figure 1. Index map of northwestern Vermont. Camels Hump quadrangle is shaded.

1959, Cady, 1960). As these structures are traced southward the Cambridge syncline strikes into two synclines separated by an anticline in the Camels Hump quadrangle and the Enosburg Falls-Fletcher anticline lies to the west of the quadrangle.

At one time the Green Mountains in northern Vermont were considered to be part of the Taconic allochthone in which the rocks reached their present position by thrusting from the east (Cady 1945). The traces of these thrust planes had been mapped three to five miles west of the quadrangle. However, this theory has been abandoned and the Green Mountains are considered to be autochthonous and the thrusts mapped to the west are considered to be smaller thrusts or faults (Cady 1960). As no faults or thrusts have been recognized in the Camels Hump quadrangle and the rocks show no evidence of having been part of a thrust sheet, the authors are happy to discard the allochthone theory for the Green Mountains.

The sediments in the Camels Hump quadrangle were deposited in a eugeosyncline and the oldest coarse-grained material, graywacke, was probably derived from the Adirondack Mountains to the west. Slightly later, a miogeosynclinal zone developed in the site of the Champlain Valley and a carbonate-quartzite assemblage was deposited. With the miogeosyncline present between the craton and eugeosyncline, it is believed that most of the sediments deposited in the subsiding eugeosyncline were derived from the eugeosyncline itself (Cady, 1960).

## *Topography and Drainage*

The Green Mountains which extend through the center of the quadrangle are the most prominent topographic feature. The higher peaks along this range, north to south, are Dewey Mountain, Mount Mayo, Bolton Mountain, Woodward Mountain, Camels Hump, Mt. Ethan Allen, Mt. Ira Allen and Burnt Rock Mountain with the highest elevation being Camels Hump at 4083 feet. As the elevation of the Winooski River, less than four miles to the north is 340 feet, the relief is as much as 3743 feet. The mountains have a subdued outline and are heavily wooded. Extensive areas of bare rock occur only at a few places along the crest of the mountains and on some of the steeper faces, notably on Camels Hump Mountain (Pl. 4). The rounded masses of bare rock north of the Winooski River near Bolton suggest roche montonnees, but the glacial striations are almost normal to the direction expected from the shapes of the outcrops. Some of the rock masses probably formed by exfoliation whereas others appear to be controlled by joints.

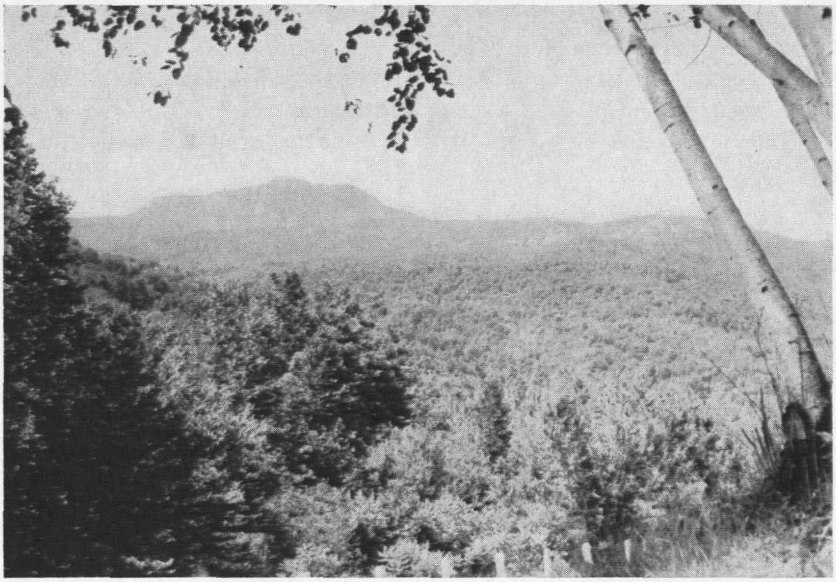


Plate 4. Camels Hump Mountain as seen looking south from north of the Winooski River near Bolton Falls. The rocks exposed near the skyline on the right have flat dips and are located approximately on the axis of the Green Mountain anticlinorium.

The Winooski River forms the principal drainage, flowing about N. 70° W. and dividing the quadrangle into two nearly equal parts. It is the only east-west transportation route through the mountains in this area. As the river transects and completely cuts through the Green Mountains, it presumably is superimposed from an earlier erosion surface of which no other evidence persists (Jacobs, 1938). A smaller pass, without roads, occurring in the northern part of the quadrangle at Nebraska Notch at an elevation of 1880 feet, probably represents the site of a superimposed stream which has been abandoned and is now a wind gap. The Lamoille River and Missisquoi River, other principal rivers north of the quadrangle are also superimposed and have a similar east-west course which transects the Green Mountains.

## *Previous Work and Present Study*

No detailed work has been done previously in the Camels Hump quadrangle. The earliest geologic studies of the northern Green Mountains were made by Edward Hitchcock (1861) and C. H. Hitchcock (1884). More recently, Jacobs (1938) studied these mountains in reconnaissance and briefly described the rocks and structure. No geologic maps of the quadrangle were published.

In recent years, considerable geologic work has been done in adjacent quadrangles. Reports have been published in the Mount Mansfield quadrangle to the north (Christman, 1951), the Hyde Park quadrangle to the northeast (Albee, 1957), the Montpelier quadrangle to the east (Cady, 1956) and parts of the Burlington and Middlebury quadrangles to the west and southwest (Cady, 1945). The Lincoln Mountain quadrangle to the south has been mapped by Cady (1959) who was kind enough to give the authors advanced copies of his map which is in press.

The present study was done during the summers of 1958 and 1959 by the authors. The mapping was done on 7½′ U.S. Geological Survey topographic maps published in 1948 at a scale of 1:24,000. In wooded areas, particularly along streams and ridges, an aneroid barometer was used to determine elevations for locating positions. Aerial photographs were used where possible.

The senior author has been responsible for much of the mapping, interpretations and for the preparation of this report. Secor mapped various parts of the quadrangle but particularly was responsible for the Ottauquechee and Stowe formations in the southeast part of the area. Also, he prepared a Master's thesis (Secor, 1959) on the joints of the area which is briefly summarized in this report.

### *Acknowledgments*

The authors wish to express their gratitude to Dr. Charles G. Doll, Vermont State Geologist, for his support and interest in the study. Working for the Vermont Geological Survey under him was a pleasure and the authors appreciate the progressiveness of the State of Vermont which made this study possible.

Special thanks go to Dr. Wallace Cady and Dr. Alfred Chidester of the U.S. Geological Survey who visited the authors in the field and discussed some of the regional problems. Also, the authors are appreciative of the interest and advice given them by Dr. Charles Nevin of Cornell University by discussions in Ithaca and during several days in Vermont.

A number of Cornell students aided the project as field assistants and the authors are particularly indebted to them for their help and stimulating observations. At various times during 1958, these were Gerald Nalewaik, James Rather and Paul Rockman; during 1959 Stephen Clement and Marco Einaudi kept us running.

The laboratory investigations and preparation of this report were done

in the Geology Department at Cornell University and incidental expenses were covered by the Department. A special faculty research grant covered the cost of two chemical analyses. Dr. H. H. Hess kindly allowed the senior author to use the x-ray equipment at Princeton University.

Finally, the authors wish to thank Dr. D. P. Stewart of Miami University who discussed with them the geomorphology and kindly allowed some of his data to be used.

## METAMORPHOSED SEDIMENTARY AND VOLCANIC ROCKS

### *General Statement*

Most of the rocks of the Camels Hump quadrangle were deposited in a eugeosyncline of early Paleozoic age. The original sediments were coarse- to fine-grained graywacke, siltstone and shale which locally contained minor amounts of volcanic material. These were metamorphosed to metagraywacke, schist, phyllite and greenstone with the rocks in the western part of the quadrangle being least metamorphosed.

The Camels Hump group comprises most of the rocks in the quadrangle. This group was defined by Cady (1956) as the strata which lie above the Precambrian and below the Ottauquechee formation and typified by the rocks occurring along the Winooski River west of Waterbury, Vermont, in the Camels Hump quadrangle. The group is roughly equivalent to the "Bonsecours" as mapped in Quebec (Cady, 1960). The group may be divided into the Pinnacle, Underhill, Hazens Notch and Pinney Hollow formations. Inasmuch as deposits of eugeosynclines are characterized by thick sections of similar-appearing rocks, it is not surprising that the contacts between the formations of the Camels Hump group are gradational and only approximately located.

In the southeastern part of the area, younger rocks belonging to the Ottauquechee formation of Cambrian age and the Stowe formation of Ordovician age cut diagonally across the corner of the quadrangle (Pl. 1).

### *Pinnacle Formation*

#### General Statement

The Pinnacle formation was named by Clark (1934) for a 400-foot series of graywacke near the bottom of the Cambrian section at Pinnacle Mountain in southern Quebec. The formation was mapped in northwest Vermont by Booth (1950) who described the Pinnacle formation as containing "typically greenish-gray to light gray graywacke, often containing beds of sandy slates, black and green slates, quartzitic schist and various types of conglomerate". The Pinnacle formation is underlain by the Tibbit Hill schist and is overlain by the White Brook dolomite in the areas studied by Booth. Where the White Brook dolomite was absent, he found it impossible to locate the contact between the Pinnacle formation and the overlying West Sutton formation. Pinnacle-type rocks were recognized in the western part of the Mount Mansfield quadrangle (Christman, 1959) but were not mapped as a formation because of the absence of the White Brook dolomite and the gradational nature of the rock types rendered it difficult to establish the top of the formation. Cady (personal communication, 1959) has mapped the graywacke and quartzose rocks lying directly above the Precambrian in the Lincoln Mountain quadrangle to the south as the Pinnacle formation.

### Description of Rocks

The Pinnacle formation, which occurs in two areas in the western part of the Camels Hump quadrangle contains principally thick sections of graywacke with some interlayered siltstone and shale. Metamorphism in this region was weak so that the shale and some of the siltstones were changed to phyllite and some schist, but the coarse siltstone and graywacke were only partially recrystallized to metagraywacke and schist. In these, some of the original quartz grains, often bluish in color, may be detected and in some places graded bedding may be observed (Pl. 1). In the hills south of Nashville, some of the rocks are conglomeratic with flattened pebbles as much as 2 cm. long and 1 cm. thick. In general, the rocks are light to dark gray, rather quartzose in appearance with poorly developed bedding schistosity. Megascopically, quartz, biotite and fine muscovite, or sericite, and small magnetite octahedrons may be visible. The metagraywacke generally forms massive outcrops. South of Jonesville some of the metagraywacke is so massive and even-textured that it is known locally as "granite" (Loc. 3, Fig. 2).[1] It should be noted that the term "graywacke" in this report refers to the composition of the rock without reference to genesis or stratigraphic occurrence.

Microscopically, the metagraywacke contains quartz, plagioclase, muscovite (or sericite), biotite and minor amounts of calcite, chlorite, magnetite, pyrite, apatite, zircon, tourmaline, sphene and epidote (Table 1). The plagioclase is generally small, recrystallized untwinned

---

[1]This notation gives the locality number which is located on Figure 2.

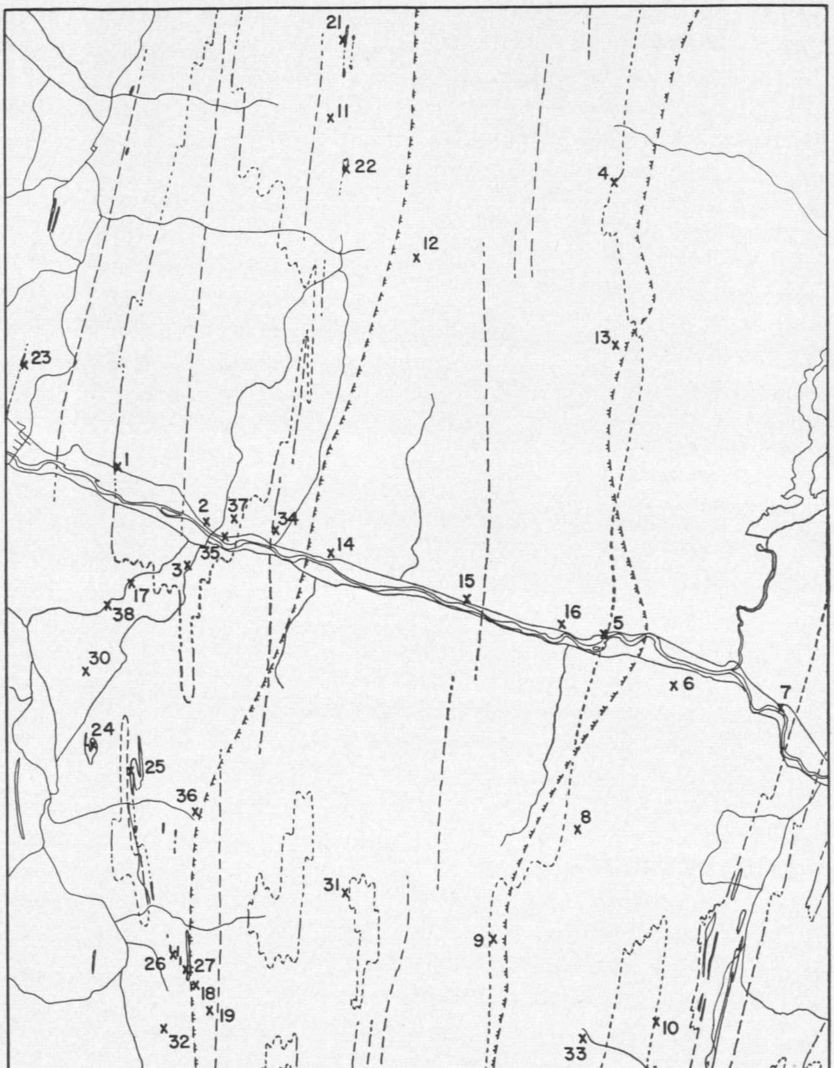



Figure 2. Locality map, Camels Hump quadrangle.

TABLE 1

ESTIMATED MODES AND PETROGRAPHIC DATA FOR ROCKS FROM PINNACLE AND HAZENS NOTCH FORMATIONS

| Locality* | Quartz | Plagioclase | Sericite** | Biotite | Chlorite | Calcite | Other minerals and petrographic data |
|---|---|---|---|---|---|---|---|
| | | | | Pinnacle Formation | | | |
| 1 | 40% | 40% | 10% | 10% | .... | Tr | Magnitite, apatite, zircon and sphene. |
| 2 | 60% | 20% | 10% | 5% | 1% | 2% | Magnetite 1%, zircon, apatite, epidote and tourmaline. |
| 3a | 40% | 40% | 5% | 10% | 3% | Tr | Epidote, zircon, 5ourmaline, apatite, magnetite and pyrite. |
| 3b | 30% | 40% | 15% | 15% | Tr | .... | Zircon 1%, magnetite and apatite. |
| | | | | Hazens Notch Formation | | | |
| 4 | 5% | .... | .... | 1% | 1% | 35% | Epidote 50%, magnetite and rutile (?). |
| 5 | 5% | 40% | .... | 3% | 7% | 25% | Epidote 20%, apatite, spehne and magnetite. |
| 6 | .... | 30% | .... | 1% | 35% | .... | Epidote 15%, amphibole 20% and magnetite; chlorite is anomal. brown, amphibole ext. is 26°. |
| 7 | 65% | 15% | .... | .... | 10% | 10% | Epidote, magnetite, tourmaline and apatite. |
| 8 | 65% | 3% | 20% | .... | 10% | .... | Epidote, apatite 1%, leucoxene and ilmenite 1% and tourmaline; chlorite is anomal. blue. |
| 9 | 45% | .... | 10% | .... | 5% | 30% | Graphite 10%, magnetite 2% and tourmaline; chlorite is anomal. brown. |
| 10 | .... | 40% | .... | .... | 20% | 2% | Epidote 15%, amphibole 20%, magnetite 1%, apatite 1% and sphene 1%. |

*For the location of these localities see Figure 2.
**Sericite includes some muscovite and perhaps unidentified paragonite.

crystals of albite, but most thin sections contain a few detrital grains of more clacic plagioclase which may be recognized by their albite twinning. The amount of plagioclase is so great in some of the rocks that they should be classified as arkosic graywacke and a few as arkose. Similar rocks were studied in greater detail in the Mount Mansfield quadrangle and their modes and discussion of terminology is given (Christman, 1959).

### Distribution and Age

The Pinnacle formation crops out extensively in the core of an anticline in the northwest part of the quadrangle and in a small area in the southwest (Pl. 1). North of the Winooski River the formation forms a broad area of outcrop which becomes narrow and pinches out when traced either to the north or south. The good development of the formation at Jonesville and west of Jonesville and the complete absence of quartzose rocks along strike to the south in the vicinity of Gillett Pond clearly indicate that the formation terminates in this direction. However, as the actual contact is difficult to trace, the pattern shown on the map is somewhat diagrammatic and the interfingering probably is more complex. Although it is possible that the distribution of the Pinnacle formation results entirely from a facies change, it is more probable that the structure of the area is important and that the broad anticline north of the river plunges both south and north so that the Pinnacle formation becomes less exposed in these directions.

The Pinnacle formation which crops out in a small area in the southwest part of the quadrangle also represents older rock being exposed in the core of an anticline (cross section B-B′, Pl. 1). It is presumed that this anticlinal structure is related to the larger one to the north, but the exact relationship is not known.

Across the strike, the contact of the Pinnacle formation north of the Winooski River is sharp on the west and indistinct on the east side. The contact west of Jonesville is easily detected because of the abrupt change from quartzosic to phyllitic rocks. These phyllites occupy the synclinal area west of the Pinnacle formation with the Pinnacle formation coming to the surface again at about the western margin of the quadrangle. According to Cady (personal communication, 1959), who mapped the eastern part of the Burlington quadrangle after the field work for this report was completed, the contact closely approximates the edge of the quadrangle and may actually lie just inside the Camels Hump quadrangle.

The indistinct contact on the east side of the formation results from a facies change in this direction, the rocks becoming finer grained to the east. A cross section through this area (A-A′, Pl. 1) shows that the rocks along the axis of the Green Mountain anticlinorium are low in the stratigraphic section and, if it were not for the facies change, the Pinnacle formation would be expected in the core of the anticlinorium. As the rocks at Bolton Notch and eastward are not particularly quartzose, it must be assumed that the facies change occurs in about this position. The change is probably more gradual than shown on the cross section and much more interfingering probably occurs, so that the eastern contact is somewhat diagrammatic. The problem is complicated by increasing metamorphism to the east so that fine-grained graywacke is not easily recognized.

A similar sequence of graywacke and quartzose rocks in the west and finer-grained rocks in the east was noted in the Mount Mansfield quadrangle (Christman, 1959) so that a similar facies change is postulated for that area.

The Pinnacle formation is considered to be Cambrian because of its stratigraphic position in Canada. In the Lincoln Mountain quadrangle, Cady (personal communication, 1959) shows that the formation directly overlies the Precambrian. Cady (1960) assigns the lower part of the Camels Hump group to the "Cambrian (?)", placing it below definite Lower Cambrian formations.

## *Underhill Formation*

### General Statement

The Underhill formation is named here for rocks which lie above the Pinnacle formation and below the Hazens Notch formation in the Camels Hump group. However, in part, it is equivalent in time to the upper portions of the Pinnacle formation because of changes in facies. To the north, the Underhill formation is partly equivalent to the West Sutton formation, but the latter is poorly defined in northwest Vermont and the Gilman quartzite which marks the top of the formation cannot be recognized except farther to the west and north. Also, it is similar to the "Bonsecours" formation in Quebec but the exact relations are not known. To the south in the Lincoln Mountain quadrangle, the Underhill formation is overlain by the Mount Abraham schist (Cady, personal communication, 1959).

The type locality for the formation is Underhill township in the northern part of the Camels Hump quadrangle and in the southern part of the Mount Mansfield quadrangle. The formation consists principally of moderately fine-grained quartzo-feldspathic and pelitic sedimentary rocks and some basic volcanic rocks which have been metamorphosed to the greenschist facies.

TABLE 2

ESTIMATED MODES AND PETROGRAPHIC DATA FOR ROCKS FROM UNDERHILL FORMATION, EXCLUDING GREENSTONE

| Locality* | Quartz | Plagioclase | Sericite** | Biotite | Chlorite | Calcite | Other minerals and petrographic data |
|---|---|---|---|---|---|---|---|
| 11 | 40% | 35% | .... | 15% | 2% | 2% | Magnetite 1%, sphene 1%, zircon and apatite; chlorite is anomal. blue. |
| 12 | 5% | 50% | .... | 20% | 15% | 5% | Rutile 1% and hematite (?). |
| 13 | 10% | 35% | 40% | 3% | 7% | .... | Magnetite 2%, tourmaline and garnet albite porphyroblasts have helicitic structure. |
| 14 | 40% | 35% | 4% | .... | 10% | 8% | Magnetite 1%, pyrite, zircon, apatite and garnet. |
| 5 | 60% | 25% | 5% | 3% | 5% | 1% | Garnet 1%, zircon, magnetite and apatite. |
| 16 | 35% | 45% | 7% | 5% | 3% | Tr | Magnetite 2%, garnet, tourmaline, apatite and zircon. Photomicrograph shown on P1. 5. |
| 17 | 15% | 40% | .... | 15% | 10% | 15% | Magnetite 2%, pyrite 1% and apatite; chlorite is anomal. brown. |
| 18 | 45% | 20% | 25% | .... | 10% | .... | Magnetite 1%, tourmaline, zircon and apatite; procholorite occurs as large porphyroblasts. |
| 19 | 75% | .... | 15% | .... | .... | .... | Magnetite altering to hematite 10%, graphite 5%, tourmaline and rutile (?). |
| 20 | 5% | 60% | 20% | 10% | Tr | .... | Magnetite 1%, garnet and zircon; porphyroblasts of albite have sieve structure. |

*For the location of these localities see Figure 2.

**Sericite includes some muscovite and perhaps unidentified paragonite.

## DESCRIPTION OF NON-VOLCANIC ROCKS

In the lower part of the formation in the western half of the quadrangle, the rocks of the Underhill formation are typically phyllite, but some interlayed metagraywacke is common. The phyllite is well exposed in the hills southeast of Richmond, in the hills adjacent to Gillett Pond and in the vicinity of Jericho Center. The rocks contain chlorite,

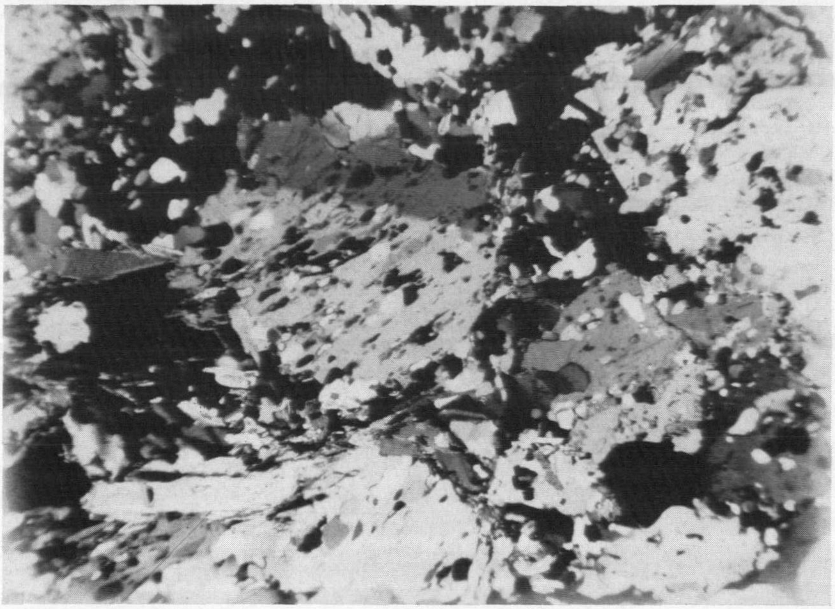


Plate 5. Twinned albite porphyroblasts with sieve structure and ragged outline in albite-quartz-muscovite schist from near North Duxbury (Loc. 16, Fig. 2), 30X, crossed nicols. The twinning is exceptional as most albite porphyroblasts show no twinning.

sericite, quartz and some albite with traces of graphite locally present. The original rocks were probably siltstone and shale.

The interbedded metagraywacke is particularly common near the contact with the underlying Pinnacle formation, suggesting that the difference in the formations resulted from a gradual change in conditions of sedimentation. The older Pinnacle formation characteristically contains the coarser sediments whereas the Underhill formation contains finer-grained sediments. In the western part of the quadrangle where the metamorphism was less intense, original quartz grains and graded bedding may be observed in some of the coarser beds of the Underhill formation (Pl. 1).

As the Underhill formation is traced eastward over the Green Mountain anticlinorium the rocks are more metamorphosed and are typically quartz-albite-muscovite schist with variable amounts of biotite, chlorite, calcite and garnet and variable trace amounts of epidote, magnetite, sphene, tourmaline, apatite, zircon, pyrite and rutile (Table 2). As the

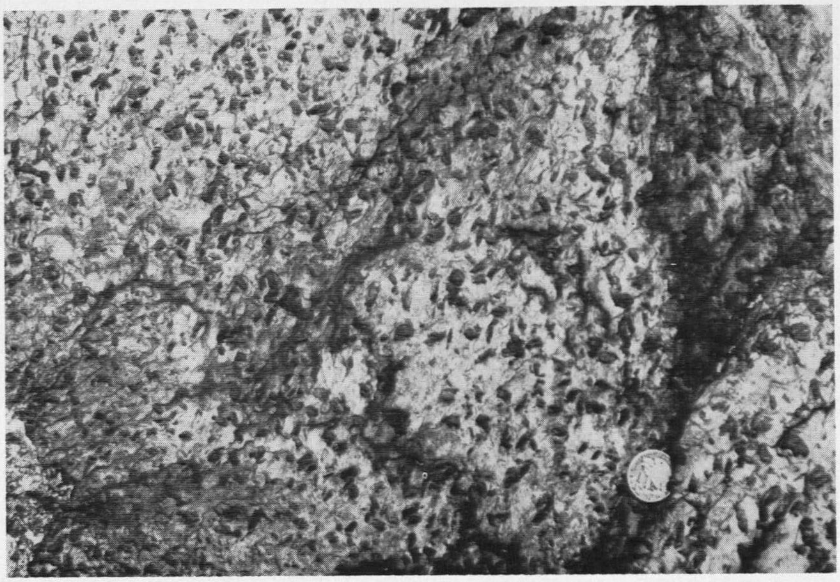


Plate 6. Porphyroblasts of chlorite, east of Huntington Center, (Loc. 18, Fig. 2).

garnet occurs in a wide zone paralleling the Green Mountain anticlinorium with the anticlinal axis about the center of the zone (Pl. 1), it is evident that the distribution of the garnet reflects a higher degree of metamorphism in the core of the anticlinorium. Rutile comprises about one per cent of one of the schists studied (Loc. 12, Fig. 2) and is reported by local prospectors to be present in the schist containing magnetite porphyroblasts in the vicinity of Preston Pond.

In places, the quartz-albite-muscovite schist grades into albite-muscovite schist which contains white porphyroblasts of albite as much as one-half centimeter in diameter. The albite may comprise as much as 75 per cent of the rock, generally has a sieve texture (Pl. 5), and may also show helicitic structure. Near the top of the formation graphite dust occurs in the albite porphyroblasts so that these are gray or black. Because of the large amount of equigranular albite crystals in some of these rocks, the schistosity is poorly developed and the rocks resemble gneiss or granulite. Such rocks are particularly well-developed on the west side of Mount Mansfield in Underhill township.

Because of the metamorphism it is difficult to estimate the nature of the original rock. However, as most of the rocks are not particularly quartzose and as the upper part of the formation grades into graphitic rocks of the Hazens Notch formation, it is likely that the sedimentary trend of increasing finer sediments continued in the eastern part of the quadrangle. The carbonaceous material suggests deeper and farther offshore sedimentation in contrast to the graywacke at the base of the formation.

At five or six localities in the southwestern part of the quadrangle, the quartz-muscovite-albite schist contains as much as 10 per cent unusual porphyroblasts of chlorite. These weather out on the surface and are as much as one and one-half centimeters long (Pl. 6). The porphyroblasts are dark green with only moderately well-developed basal cleavage, so that they might easily be misidentified as amphibole. Most of the schist in which these porphyroblasts occur are near outcrops of amphibolitic greenstone, but a few are found along the strike where greenstone cannot be found. As they all occur at about the same general horizon and in rocks of the same metamorphic grade, it is possible that they represent metamorphosed siltstone that contained volcanic debris from the same source as the greenstone. Such porphyroblasts were not found elsewhere in the quadrangle.

Microscopically, the schist contains quartz, plagioclase, muscovite, chlorite, magnetite, tourmaline, zircon and apatite (Loc. 18, Fig. 2). The chloritic porphyroblasts have sieve and helicitic structures, and the optical properties indicate that the mineral is probably prochlorite. They are olive-green, slightly pleochroic with parallel extinction and polysynthetic twinning parallel to the basal cleavage which is length-slow. The mineral has low birefringence with anomalous blue interference colors, is optically negative with a low 2V which varies between 10 and 30° and has a dispersion of $r<b$. According to table by Troger (1956) the mineral is prochlorite. The x-ray study of the mineral (Table 6) indicated that it was a chlorite and, because of the similarity of d values, it appears to be the same chlorite which occurs intergrown with the amphibole in the greenstone.

## Description of Amphibolitic Greenstone

In the lower part of the Underhill formation in the western half of the quadrangle, particularly in the southwest, discontinuous layers of greenstone occur interbedded with phyllite. Depending upon the megascopic appearance of the rocks, these have been divided into amphibolitic greenstone, calcareous greenstone and feldspathic greenstone.

The amphibolitic greenstone is the most common and is the most

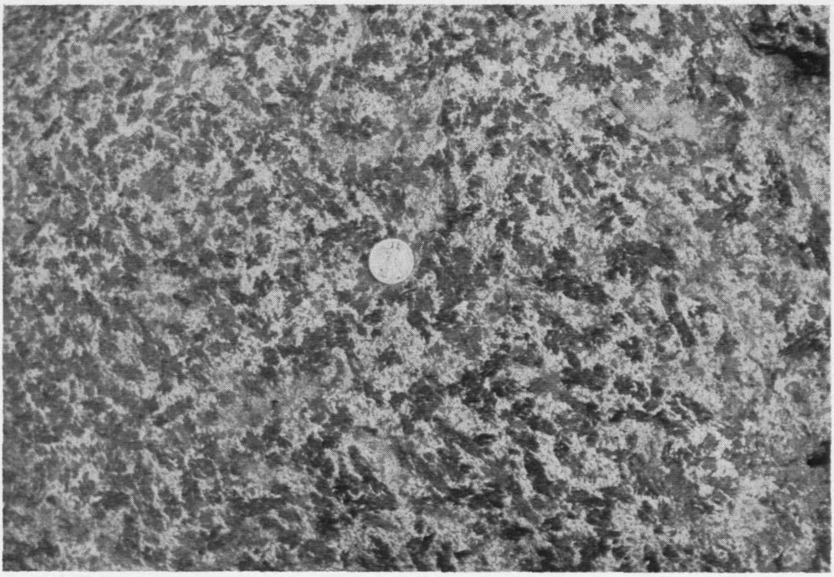


Plate 7. Porphyroblasts of amphibole in the amphibolitic greenstone, east of Huntington Center.

distinctive because it contains porphyroblasts of dark green amphibole which may be as much as two centimeters long (Pl. 7). A similar greenstone was mapped in the Mount Mansfield quadrangle as the Tibbit Hill schist (Christman, 1959) and it was proposed that it represents metamorphosed lavas. Because the rock is distinctive and can be easily mapped, the greenstone in the Camels Hump quadrangle was studied in detail.

A chemical analysis of the amphibolitic greenstone from near Gillett Pond (Loc. 24, Fig. 2) reveals a higher than normal content of sodium which suggests that the original rock may have been a spilite. According to Sundius (1930) spilites are characterized by a deficiency in $Al_2O_3$ and $K_2O$ and high values of $Na_2O$, FeO and $TiO_2$. These characteristics are shown well by the Gillett Pond greenstone and its chemical composition compares well with those of known spilites (Table 3). The only noteworthy differences are that the MgO content is about two per cent higher than the average spilite, the FeO content is considerably higher because the rock contains no $Fe_2O_3$ and the $Na_2O$ content is about one-half per cent lower than average.

The calculated norm of the Gillett Pond greenstone is 1.7% ortho-

TABLE 3

CHEMICAL COMPOSITION OF AMPHIBOLITIC GREENSTONE FROM GILLETT POND COMPARED TO OTHER SPILITES

| | 1<br>Gillett<br>Pond | 2<br>Average<br>Spilite | 3<br>Average<br>Spilite | 4<br>Oregon<br>Spilite |
|---|---|---|---|---|
| $SiO_2$ | 46.94 | 51.22 | 46.01 | 53.15 |
| $Al_2O_3$ | 13.52 | 13.66 | 15.21 | 14.39 |
| $Fe_2O_3$ | 0.00 | 2.84 | 1.35 | 1.28 |
| $FeO$ | 13.38 | 9.20 | 8.69 | 9.33 |
| $MgO$ | 6.61 | 4.55 | 4.18 | 4.74 |
| $CaO$ | 8.65 | 6.89 | 8.64 | 7.04 |
| $Na_2O$ | 4.38 | 4.93 | 4.97 | 4.58 |
| $K_2O$ | 0.24 | 0.75 | 0.34 | 1.01 |
| $H_2O-$ | 0.17 | | | 0.19 |
| | | 1.88 | 2.48 | |
| $H_2O+$ | 1.89 | | | 2.02 |
| $TiO_2$ | 2.54 | 3.32 | 2.21 | 1.50 |
| $P_2O_5$ | 0.23 | 0.29 | 0.61 | 0.19 |
| $MnO$ | ..... | 0.25 | 0.33 | 0.14 |
| $CO_2$ | ..... | 0.94 | 4.98 | 0.10 |
| | 98.38 | | | |

1. Amphibolitic greenstone, Gillett Pond (Loc. 24, Fig. 2), Camels Hump quadrangle, Vermont. Analysis by Ledoux and Company.
2. Average spilite, according to N. Sundius (1930).
3. Average spilite, according to A. K. Wells (1923).
4. Spilite from eastern Oregon, reported by James Gilluly (1935).

clase, 27.8% albite, 16.4% anorthite, 5.1% nepheline, 20.4% diopside, 19.7% olivine, 4.9% ilmenite and 0.7% apatite which places the rock in Class III, Order 5, Rang 3, Subrang 5 in the C.I.P.W. classification. In this group Washington (1917) lists 42 analyses, most of which were called diabase or basalt. Of these, only eleven have a similar sodium content and these were called diabase. Spilites were not calculated by Washington because he considered them to be altered rocks. The normative feldspar is $Ab_{60}Or_4\,An_{36}$.

The original minerals are no longer present because the rock has been entirely metamorphosed to the greenschist facies. The mode, determined by 2700 point counts on seven standard thin sections, is 33% blue-green amphibole, 28.5% albite, 20% chlorite, 19% epidote, 8% sphene and 0.5% apatite (Loc. 24, Table 4). Microscopically, the amphibole, which is described in detail in one of the following paragraphs, forms large porphyroblasts and cannot be an original mineral. No relic remains of other mafic minerals are present. The albite occurs as smaller grains in the groundmass and is generally untwinned. The chlorite is light green with anomalous blue interference colors and occurs intergrown with the amphibole. Epidote occurs as small grains with slightly pleochroic greenish yellow colors. Brownish, cloudy masses of very small grains of a highly birefringent mineral, which appears to be altering to leucoxene, is tentatively identified as sphene. Because of the Holmes effect, the estimates on the sphene may be slightly high.

Because of the variable and unknown composition of the chlorite and epidote, it is not possible to verify the mode by calculating a metamorphic norm from the chemical analysis. Likewise, a comparison between the norm and mode is meaningless because of the differences in mineral suites. The good agreement between the normative albite (27.8%) and the modal albite (28.5%) probably is only coincidental as it requires that the sodium assigned to the nepheline in the norm equals the amount needed by the amphibole in the mode. However, it turns out that if sodium is assigned to the amphibole according to its composition and abundance, the amount of sodium remaining is approximately the amount needed for the albite.

It is assumed in the sections above that the overall composition of the present greenstone is the same as the original igneous rock because no contrary evidence was found. It is known that basic igneous rocks are particularly susceptible to low-grade metamorphism, so that it is not unexpected that the spilite should be completely metamorphosed. Whatever plagioclase was originally present has been recrystallized with the sodic portion forming albite and the calcic portion going into epidote and amphibole. Any original mafic minerals, as pyroxene, have formed amphibole, chlorite and epidote. This recrystallization could occur without the addition or subtraction of material during metamorphism.

The rocks above and below the amphibolitic greenstone show no evidence that material has been added or removed. Such obvious features as small quartz veins or segregations of minerals are absent in the adjacent phyllite, although such features occur farther east where the metamorphism was more intense. In fact, some of the rocks are so little metamorphosed that original quartz grains may be recognized and graded bedding may be distinquished in a few places in the metagraywacke. Thus, although it is possible that some movement of water occurred, it seems unlikely that any wholesale changes as a result of regional metamorphism could have occurred within and restricted to the thin layers of spilite without affecting the adjacent phyllite and metagraywacke.

Evidence collected in the Camels Hump quadrangle is not sufficient to determine the mode of origin of the spilitic lavas. It cannot be stated whether the sodic composition of the rock is primary, is derived from the soda of sea water, or is the result of some type of deuteric action. The question of origin will not be reviewed here as this has been excellently done by Gilluly (1935) and Turner and Verhoogen (1951). However, it should be emphasized that regardless of their origin, spilites "are widespread and voluminous among rocks formed in geosynclines" (Williams, Turner and Gilbert, 1954). Although spilite has not been reported previously in Vermont, its occurrence might have been anticipated.

### Field Occurrence and Distribution of Greenstones

The largest bodies of amphibolitic greenstone occur east of Huntington and Huntington Center, and smaller bodies occur in these same areas and in the northwestern part of the quadrangle (Pl. 1).

The analyzed greenstone (Table 3), located east of Huntington and south of Gillett Pond, was mapped and studied in detail. Because the greenstone is situated on a woody hillside and has irregular contacts, the size and shape of the body could be determined only by plotting the data of a tape and compass traverse (Fig. 3). It is an odd-shaped body with a maximum outcrop length and width of slightly over 2000 by 800 feet. Flattened amphibole phenocrysts and chlorite have a parallel orientation to give the rock a foliation. This foliation is conformable with the bedding schistosity of the adjacent phyllite. The contact with the phyllite was observed to be sharp in some places and gradational elsewhere.

Although it is possible that the greenstone represents a sill, the evidence seems to indicate it represents a lava flow. Some of the greenstone adjacent to the phyllite, which is interpreted as an upper contact, contains more carbonate which has weathered away leaving holes in the rock surface, suggesting an amygdaloidal top of a lava flow. The gradational contact with the phyllite is not easily explained if the body is a sill, whereas in a lava, it may represent tuffaceous sediments on top of the flow or a reworked contact. The narrow, somewhat continuous layers of greenstone at the same stratigraphic horizon in the Mount Mansfield quadrangle suggested that there the rock represents a flow rather than an intrusion (Christman, 1959).

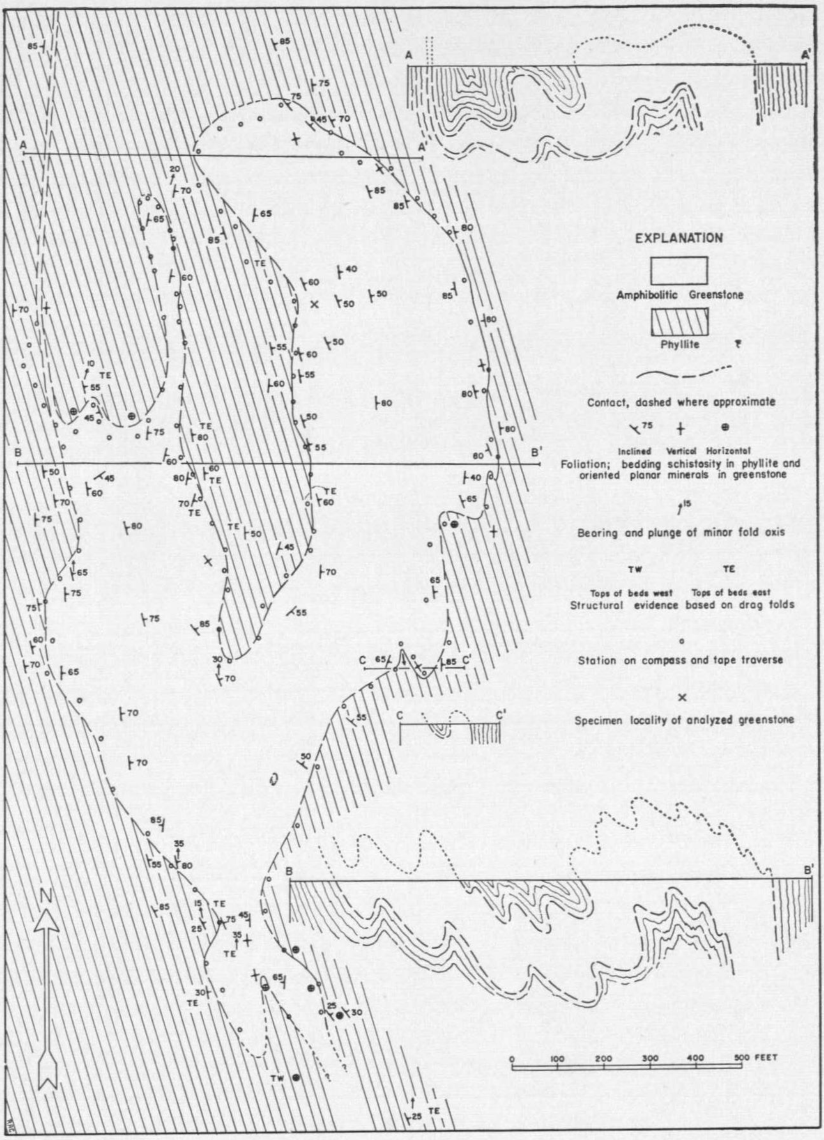



Figure 3. Geologic Map of the Amphibolitic Greenstone, South of Gillett Pond (Loc. 24, Fig. 2).

Because of the isoclinal folding, the structure of the greenstone is not readily apparent as a synclinal drag fold on the east side of an anticline, as shown by the cross sections (Fig. 3). The small drag fold at the locality of cross section C-C' is especially critical as here it could be established positively that the greenstone is below the phyllite. Four hundred feet to the west, the plunge of a tongue of phyllite to the north also indicates that the phyllite overlies the greenstone. At various places irregular dips and horizontal foliation indicate that the rocks have been complexly folded. Most of the drag folds indicate that an anticline lies to the west (shown on Fig. 3 by "T.E.," abbreviating "tops east").

Lack of outcrops prohibits determining the southern extent of the greenstone, but a narrow layer of greenstone can be traced to the north. Here the greenstone is only about 20 feet thick and it cannot be traced more than 1000 feet north of the main body. It is not known how much of the thickness of the greenstone, shown on the cross section as about 150 feet, is due to original thickening of the flow or due to movement of the more easily metamorphosed rock into the noses of the drag folds during the folding and metamorphism. Probably the present size and shape of the greenstone is a function of both.

The large greenstone located about three-fourth miles to the southeast on the east side of the tongue of Pinnacle formation also contains large amphibole porphyroblasts. Although this body was not mapped in detail, it also appears to occur in a large drag fold (Pl. 1) and it extends discontinuously to the south. At one place on the west side of the greenstone, a slightly foliated dioritic-appearing rock occurs adjacent to the amphibolitic greenstone (Loc. 2 , Fig. 2). By thin section study it contains an estimated 15% amphibole, 55% albite, 25% quartz and minor amounts of epidote, sphene, garnet and biotite (Table 4).

The optical properties of the amphibole occurring in this light-colored rock indicate that the mineral is most unusual. It has very strong pleochroism with X being light yellowish brown, Y being a vivid blue and Z being a green to very dark green color. The mineral is negative with a 2V of about 30° with very strong dispersion of $r<v$, and the Y index is approximately 1.705. The orientation is unusual in that Z coincides with b and the extinction angle of Y to c is about 31°. These properties fit none of the amphiboles described by Winchell (1951) and the only amphibole listed by Troger (1956) with a similar orientation is "crossit-bababudanit," but the other properties do not agree. It can only be concluded that the amphibole is probably some unusual soda-rich variety and that the dioritic-appearing rock resulted as a contact effect

TABLE 4

MODES* AND PETROGRAPHIC DATA FOR GREENSTONES FROM THE UNDERHILL FORMATION

| Locality | Amphibole | Max. ext. of amphibole | Albite | Epidote | Chlorite | Sphene | Other minerals, petrographic data and notes. |
|---|---|---|---|---|---|---|---|
| 21 | 30% | 23° | 40% | 15% | 10% | 5% | Magnetite 1%, pyrite and hemantite. |
| 22 | 35% | ... | 40% | 15% | ... | ... | Biotite 10%, magnetite 2%; porphyroblasts of albite and magnetite have sieve structure, amphibole is very fine grained. |
| 23 | 20% | 22° | 40% | ? | 10% | 3% | Sericite 20%, fine-grained alteration material probably albite and epidote 10%, and biotite. |
| 24 | 33% | 29° | 28% | 10% | 20% | 8% | Apatite ½%; in amphibole x = light brown, y = light green and z = light blue with y = b. Analysis of this rock is given on table 3. |
| 25 | 15% | 31° | 55% | Tr | ... | Tr | Quartz 25%, garnet and biotite; description of the amphibole is given in the text. |
| 26 | ... | ... | 45% | ... | 40% | 3% | Biotite 5%, calcite 2% and magnetite 2%; this rock is shown on plate 10. |
| 27 | 68% | 25° | 14% | 8% | 2% | 5% | Calcite 2% and magnetite 2%; analysis and data on amphibole are given in table 5. |
| 28 | 20% | 28° | 45% | 1% | 20% | 3% | Quartz 10% and magnetite 1%. |

*These modes are estimated except for rocks from localities 24 and 27 for which the mode was determined by point count.

of the lava on the country rock, either at the time of eruption or during metamorphism.

A third large mass of amphibolitic greenstone occurs ESE. of Huntington Center along the crest of a north-south hill (Pl. 1 and Loc. 27, Fig. 2). The rock is dense and massive and contains a higher percentage of amphibole and less chlorite than the other greenstones (Table 4). The rock probably has been subjected to a slightly higher grade of metamorphism which is indicated by the smaller content of chlorite and the occurrence of megascopic garnet at one or two places. The amphibole from this rock was studied in detail and is described in the next section of this report.

The other smaller bodies of greenstone shown on the map in the southwestern part of the quadrangle are mainly calcareous greenstones which usually could not be traced more than several hundred feet in their strike direction. Most of these do not contain amphibole. One such folded calcareous greenstone (Pl. 10) is composed principally of albite, chlorite and biotite with minor calcite (Loc. 26, Table 4).

Feldspathic greenstones occur west of Huntington Center and north of Richmond (Pl. 1). The latter locality is on strike with an amphibolitic greenstone containing epidote knots on the west side of Richmond and a porphyroblastic amphibolitic greenstone southwest of Jericho Center. This suggests that the feldspathic greenstone represents a mineralogically different flow which grades into the amphibolitic types. The feldspathic greenstone is similar to those in the Mount Mansfield quadrangle, where they are better developed (Christman, 1959). On the weathered surfaces, outlines of feldspar laths are clearly evident, but in thin section these crystals are completely altered to a mass of albite, epidote and sericite.

Several calcareous and amphibolitic greenstones occur in the north-central part of the quadrangle east of Camp Underhill (Loc. 21 and 22, Fig. 2). These are similar to the other greenstones already described except that biotite is more common in some and amphibole occurs as large porphyroblasts only in a few of the rocks. These localities are on strike with a calcareous greenstone which occurs 4½ miles to the north in the Mount Mansfield quadrangle. It is suggested that the greenstones at both localities are equivalent to the ones on the west flank of the anticline and represent the last appearance of this rock type as the flow becomes thinner to the east.

The amphibolitic, calcareous and feldspathic greenstones in the western parts of the Camels Hump and Mount Mansfield quadrangles are essentially identical, which suggests that they are related to each other and are approximately the same age. As a time unit, these lavas appear to cut the stratigraphic units. In the Camels Hump quadrangle all of the greenstones are in the Underhill formation. In the Mount Mansfield quadrangle some of the greenstone is in the Underhill formation but most of it was mapped as the Tibbit Hill schist, which in the northern part of the quadrangle lies beneath the Pinnacle formation. This relationship suggests that the upper part of the Pinnacle suffers a facies change as it is traced southward so that it is equivalent in time to parts of the Underhill formation. As the Tibbit Hill schist is traced to the north the unit becomes better developed and in Canada it has been mapped as the lowest member of the Oak Hill series and a Precambrian age was suggested at one time (Clark, 1934). In the Mount Mansfield quadrangle Pinnacle-type rocks occur within the Tibbit Hill schist so that it seems unlikely that these volcanic rocks mark the bottom of the section.

## Identification of the Amphibole

As the characteristic minerals of basic rocks metamorphosed to the green-schist facies are albite, actinolite, chlorite, epidote and calcite (Turner, 1958), it would be expected that the amphibole in the amphibolitic greenstone at Gillett Pond, and elsewhere, would be actinolite. However, the mineral cannot be actinolite because a chemical analysis shows too much sodium and aluminum, the extinction angle is too large, the Y index is too high, the 2V is too low and the pleochroic colors are too blue (Table 5). The data collected indicates that the mineral is some type of calciferous hornblende whose occurrence in the greenschist facies is somewhat anomalous with accepted theory. According to Turner (1958) hornblende first appears in a sequence of progressively metamorphosed basic rocks near the top of the greenschist facies in association with almandite garnet and is most common in the amphibolite facies in association with oligoclase or andesine and other high temperature minerals. In the Littleton-Moosilauke area in New Hampshire, Billings (1937) reports common hornblende in the middle-grade and high-grade metamorphic rocks but not in the low-grade ones. In Wiseman's study of epidiorites, actinolite and prochlorite was found in the low-grade rocks, but there was a "distinct change in chemical composition of the hornblendes at the entrance to the garnet zone" (1934, p. 411). He reports that in the garnet zone of the epidiorites common hornblende appears and chlorite disappears.

Although the optical properties of the amphibole have been determined (Table 5), it is not possible to classify the mineral with this information alone because amphibole has so many possible variations in its ten major components that all possible relations between composition and optical properties have not been established. Especially noteworthy of the amphibole in the greenstone is a rather large extinction angle of 25° between Z and c, and a blue pleochroism in the Z direction which in the alkali amphiboles is attributed to the presence of sodium.

A chemical analysis of the amphibole, made by Ledoux and Company,

TABLE 5

DATA ON AMPHIBOLE FROM THE GREENSTONE EAST OF HUNTINGTON CENTER (LOC. 27, FIG. 2), CAMELS HUMP QUADRANGLE, VERMONT

Chemical analysis*

| | |
|---|---|
| $SiO_2$ | 41.66% |
| $Al_2O_3$ | 10.89 |
| $Fe_2O_3$ | 5.22 |
| FeO | 18.72 |
| MgO | 6.94 |
| CaO | 9.69 |
| $Na_2O$ | 3.34 |
| $K_2O$ | 0.49 |
| $H_2O+$ | 0.06 |
| $H_2O-$ | 0.02 |
| $TiO_2$ | 3.28 |
| F | 0.02 |
| | 100.33 |
| Less O for F | .01 |
| | 100.32% |

*By Ledoux and Company

Structural Formula**

| | | | |
|---|---|---|---|
| | K | .09 | |
| (W) | Na | 1.01 | 2.72 |
| | Ca | 1.62 | |
| (X) | Mg | 1.62 | 4.07 |
| | Fe″ | 2.45 | |
| | Al | .52 | |
| (Y) | Fe‴ | .62 | 1.52 |
| | Ti | .38 | |
| (Z) | Si | 6.51 | 8.00 |
| | Al | 1.49 | |
| | OH | .06 | .07 |
| | F | .01 | |

**Structural type taken as:

$W_{2\text{-}3}\,(XY)_5(Z_4O_{11})_2\,(OH,F,Cl)_2$

Trace Elements Determined by Semi-quantitative Spectrographic Analysis***

| | | | |
|---|---|---|---|
| B | 0.01 % | Cu | 0.005% |
| V | .5 | Sr | .01 |
| Mn | .35 | Zr | .01 |
| Co | .01 | Ba | .01 |
| Ni | .003 | Ga | .005 |

***By American Spectrographic Lab.

Optical Data

Y = b and $Z \wedge c$ = max. of 25°
x = light brown, y = green, Z = greenish blue
Y = 1.677, measured with sodium light
2V = 61°, measured with Universal Stage
r > v, with very strong dispersion

shows a moderate amount of sodium (3.34% $Na_2O$) and a surprisingly low water content (0.06%). Because most amphiboles contain about two percent water, the sample was rerun for water and the original value was reaffirmed. The method reportedly used was "(a) loss in ingition and (b) liberation of water by heating the sample to 1000° C. while in a stream of argon with the water collected by anhydrone and weighed". Although the amphibole does not have the optical properties of oxyhornblende, which is low in water, it is probable that some reaction did proceed in that direction and that the original mineral contained more water. If it is assumed that all the 5.22% $Fe_2O_3$ resulted from the oxidation of FeO by water, the original mineral may have contained an additional 0.61% water and a total of 23.4% FeO. One line of evidence suggesting that this oxidation-reduction has occurred is that the analyzed spilitic greenstone (Table 3) contains no $Fe_2O_3$, although the rock contains 33% amphibole. It is possible that the oxidation has not occurred in the amphibole of the analyzed greenstone because this rock is slightly less metamorphosed than the one containing the analyzed amphibole. However, even assuming that $H_2O$ was lost by this reaction, this does not account for a sufficient amount of the (OH, F, C1) radical. Unfortunately a quantitative test was not made for chlorine, but it is unlikely that it is present in large amounts inasmuch as the F content was only 0.02%.

Before classifying the amphibole, perhaps something should be written about the selection and purity of the sample. Originally it was planned to analyze the amphibole from the analyzed spilitic greenstone near Gillett Pond. However, when an X-ray study showed that the prepared sample of amphibole contained sufficient intergrown chlorite to give X-ray lines of chlorite the plan was abandoned. Although the chlorite was known to be present from the thin section study, it had been hoped that most of the intergrown grains might be eliminated by the magnetic separation. As this was not the case, it was decided to analyze the amphibole from east of Huntington Center (Loc. 27, Fig. 2) because its X-ray pattern did not show chlorite lines (Table 6). Also microscopically, the amphibole contains less impurities and associated chlorite.

The amphibolitic greenstone selected was crushed, passed through sieves and decanted. It was found that a heavy mineral separation removed very little material, apparently because a large part of the light minerals was lost in the fines in the process of crushing the sample. Grains of the N. B. S. 40–60 and 60–100 sizes were passed through a Frantz magnetic separator to produce the best concentrate of amphibole possible. The resulting sample was at least 95% pure and probably close to 98% pure; a small amount of albite and chlorite adhering to the amphibole could be observed with the microscope.

The X-ray patterns of three mineral separates of amphibole from two amphibolitic greenstones in the Camels Hump quadrangle and one from the Mount Mansfield quadrangle indicate similar internal structures,

TABLE 6

X-RAY DATA ON AMPHIBOLES AND CHLORITE

| 1 Analyzed amphibole | | 2 Amphibole, Gillett Pond | | 3 Amphibole, Metcalf Pond | | 4 Chlorite | |
|---|---|---|---|---|---|---|---|
| dA | I | dA | I | dA | I | dA | I |
| 1.51 | w | | | | | | |
| 1.59 | w | 1.59 | w | | | | |
| 1.62 | w | | | | | | |
| 1.65 | w–m | 1.65 | w | | | | |
| 2.03 | w | 2.03 | w | 2.02 | M | 2.02 | M |
| 2.17 | w | 2.17 | w | 2.17 | w | | |
| | | | | 2.27 | w | | |
| 2.35 | w–m | 2.34 | w | 2.35 | w | | |
| | | | | 2.40 | w | | |
| | | | | 2.45 | w | | |
| 2.56 | w | 2.55 | w | 2.55 | w | | |
| 2.61 | w | 2.60 | w | 2.60 | w | | |
| 2.72 | M | 2.72 | M | 2.72 | w | | |
| 2.82 | w | 2.82 | w | 2.83 | M | 2.82 | S |
| 2.95 | w | 2.95 | w | | | | |
| 3.15 | v S | 3.14 | v S | 3.13 | S | | |
| 3.29 | w | 3.30 | w | 3.30 | w | | |
| 3.40 | w | 3.40 | w | | | | |
| | | 3.54 | S | 3.54 | v S | 3.54 | v S |
| | | 4.74 | w | | | 4.69 | S |
| | | 7.08 | S | 7.08 | S | 7.08 | v S |
| 8.42 | v S | 8.50 | v S | 8.50 | w | | |

w = weak, M = moderate, S = strong, and v S = very strong.

1. Analyzed amphibole from east of Huntington Center (Loc. 27, Fig. 2).
2. Amphibole from analyzed greenstone near Gillett Pond (Loc. 24, Fig. 2).
3. Amphibole from greenstone near Metcalf Pond, Mount Mansfield quadrangle (Christman, 1959, table 2).
4. Porphyroblactic chlorite, shown in plate 6, from east of Huntington Center (Loc. 18, Fig. 2).

suggesting that they are essentially the same. The localities represent slightly different grades of metamorphism and are separated along strike by 30 miles. The three strongest d values, in order of decreasing strength, are approximately 3.14, 8.50 and 2.72 A (Table 6). Although the samples were carefully prepared some amphibole crystals from Gillett Pond and Metcalf Pond contained intergrown chlorite which shows up as strong d lines at 3.45 and 7.08 A.

The Y index of the three amphiboles studied by X-ray is lower where the mineral is associated with chlorite in the less metamorphsed areas. The Y indices are 1.677 for the analyzed amphibole (most metamorphosed), 1.672 for the Gillett Pond amphibole and 1.642 for the amphibole and 1.642 for the amphibole from Metcalf Pond, Mount Mansfield quadrangle (least metamorphsed). It is presumed that with increasing metamorphism the chlorite is converted into amphibole causing slight changes in the properties of the amphibole. However, the maximum Z to c extinction angles of 25°, 29°, and 24° of the amphibole from these three localities, respectively, indicate that the amphibole in the least metamorphosed rock is not actinolite.

The structural formula of an amphibole may be calculated from the chemical analysis. As the type formula, $W_{2-3}(XY)_5(Z_4O_{11})_2(OH)_2$, contains 24 oxygens, the molecular amounts of the elements present can be distributed on this basis. Below are the formulas for 1) the analyzed amphibole (Table 4), 2) the analyzed amphibole assuming that all the $Fe_2O_3$ was originally FeO, 3) the hastingsite end-member, after Sundius (1946), and 4) the ferroedenite end-member after Sundius (1946):

| | W | X | Y | Z |
|---|---|---|---|---|
| 1) | $(Na, K)_{1.1}\ Ca_{1.6}$ | $(Fe'', Mg)_{4.1}$ | $(Al, Fe''', Ti)_{1.5}$ | $Al_{1.5}\ Si_{6.5}\ O_{22}\ (OH, F)_{.1}$ |
| 2) | $(Na, K)_{1.1}\ Ca_{1.6}$ | $(Fe'', Mg)_{4.7}$ | $(Al, Ti)_{.9}$ | $Al_{1.5}\ Si_{6.5}\ O_{22}\ (OH, F)_{.7}$ |
| 3) | $(Na, K)_1\ Ca_2$ | $(Fe'', Mg)_4$ | $(Al, Fe''', Ti)_1$ | $Al_2\ Si_6\ O_{22}\ (OH, F, Cl)_2$ |
| 4) | $(Na, K)_1\ Ca_2$ | $(Fe'', Mg)_5$ | | $Al_1\ Si_7\ O_{22}\ (OH, F, Cl)_2$ |

From the above it is evident that the analysed amphibole contains too little $H_2O$, as previously mentioned, and it does not contain the correct proportions of the XY radical. If it were not for the large amount of Y, the mineral might be considered to lie between hastingsite (3) and ferro-edinite (4). Sodium may either replace the calcium or occur in "vacant positions" so that W may vary between two and three; thus the 2.7 value for W is not abnormal.

The terminology for the calciferous amphiboles in which one Na occupies the vacant position with corresponding substitutions of $Al^{VI}$ for Si and $(Al^{IV}Fe''')Al^{VI}$ for (Mg, Fe'')Si has become somewhat confusing. Billings (1928) divided the hastingsite group into a Mg-rich member called magnesiohastingsite, an Fe-rich member called ferrohastingsite, and an intermediate member called femaghastingsite. This terminology was followed by Buddington and Leonard (1953) and Engel (1959). Sundius (1946) writes that the magnesiohastingsite is nearly the same as paragsite, and he calls the Mg-rich and Fe-rich members pargasite and hastingsite, retaining femaghastingsite for the intermediate member. In contrast, Winchell (1945 and 1951) calls the Mg-rich and

Fe-rich members, respectively, hastingsite and ferrohastingsite, and Boyd (1959) calls them pargasite and ferroparagasite.

The Vermont amphibole has an FeO/MgO ratio of 1.5 by normal calculation and 1.9 by assuming that all the ferric iron was originally ferrous iron, which according to Billings (1928) classifies it as femaghastingsite, in either case. As is common in some metamorphic amphiboles some of the Ca has been replaced by the alkalies (Boyd, 1959, p. 381) so that they cannot be plotted on the type of diagram used by Hallimond (1943), and these amphiboles are difficult to compare with the large number of amphiboles described from igneous rocks.

In conclusion, the amphibole in the amphibolitic greenstone is best classified as femaghastingsite, even though the structural formula is not exactly correct. The mineral definitely is not actinolite, as might be expected for the metamorphic conditions. Because of the similarity in extinction angles, pleochroism and interference figures, it is believed that the amphibole throughout the spilitic greenstones is essentially the same, even though slight variations of opitcal properties, particularly a difference of 0.035 in the Y index, were noted.

## *Hazens Notch Formation*

### General Statement

The Hazens Notch formation is the name suggested by Cady (personal communication, 1959) for the unit of graphitic rocks near the top of the Camels Hump group. The rocks at the type locality at Hazens Notch in the eastern part of the Jay Peak quadrangle are characterized by graphitic phyllite and schist and interbedded dark-colored quartzite. In the Lincoln Mountain quadrangel the formation is easily distinguished (Cady, personal communication, 1959) from the underlying Mount Abraham schist, which contains chloritoid and kyanite, and from the overlying Pinney Hollow formation which contains shiny quartz-sericite-chlorite schist. When these units are traced northward into the Camels Hump quadrangle both the Mount Abraham schist and the Pinney Hollow formation die out so that the Hazens Notch formation overlies the Underhill and underlies the Ottaquechee formations which may contain very similar rocks.

### Description of Rocks

The Hazens Notch formation consists of may different rock types, with the most diagnostic being graphitic schist and phyllite. These contain variable amounts of graphite and some pyrite and the rock, when weathered, has a characteristic rusty color. These graphitic rocks may comprise most of a given section or may occur only as thin layers in non-graphitic rocks. Also prominent in the formation is quartz-albite-muscovite schist in which fine graphite may be present in the albite porphyroblasts, rendering the crystals a gray to black color. In the same area, some layers of schist may be devoid of graphite and the albite porphyroblasts are white. Locally, thin interbeds of blue to dark gray quartzite occur; the coloring is presumably due to graphite. Good exposures of the formation are found around Waterbury Reservoir, particularly below the dam, and along Crossett Brook.

Microscopically, the rocks consist mainly of quartz, albite, and muscovite (or sericite) in varying proportions and lesser amounts of graphite, calcite, chlorite and epidote (Table 1). The albite porphyroblasts are generally untwinned and may show sieve and/or helicitic structures. Most of the rocks contain minor amounts of magnetite and apatite; ilmenite, zircon, tourmaline and sphene are present locally.

Schist which might be called greenstone occurs locally in the formation. It is a dense fine-grained chloritic rock with or without small porphyroblasts of albite. The rock occurs interbedded with and grades into quartz-albite-chlorite schist, and it is too uncommon and poorly exposed to be mapped. Rocks from two localities (Loc. 6 and 10, Fig. 1) contain about 35% albite, 25% chlorite, 20% amphibole, 15% epidote and some calcite and magnetite (Table 1). The amphibole, which is probably also femaghastingsite, occurs as very small elongate crystals, having an extinction angle of as much as 26° and faint green and blue pleochroic colors. Probably the rocks are derived from basic volcanic ash or debris which accumulated locally in the eugeosynclinal basin.

Epidote-rich rocks occur at two places near the base of the formation. South of Lake Mansfield exposed in a stream (Loc. 4, Fig. 2), beds of epidote-rich quartzite with quartz and calcite knots occur interlayered with impure brown limestone with a few lenses of white marble and quartz-muscovite-albite-chlorite schist which locally contains biotite. Near Bolton Falls along Highway 2 (Loc. 5, Fig. 2) some of the albite-mica schist contains as much as 20% epidote (Table 1). As these rocks probably represent additions of basic volcanic material to the eugeosynclinal sediments, it might be possible to use them to trace the lower contact of the Hazens Notch formation with more detailed mapping.

### Distribution and Nature of Contacts

The largest areal extent of the Hazens Notch formation in the quadrangle is on the eastern side of the Green Mountain anticlinorium where it forms an outcrop belt about three miles wide. The upper contact may be estimated by a greenstone horizon in the lower part of the Ottauquechee formation, but this contact is difficult to trace in the vicinity of Waterbury because these greenstones are not present in the Ottauquechee formation. The lower contact of the Hazens Notch formation was mapped at the first occurrence of abundant graphitic schist and phyllite. As the resulting contact closely parallels the garnet isograd and places the localities of epidote-rich rock at the base of the formation, it appears that the presence of graphite is a reliable criterion. The garnet isograd parallels the contact because the rocks in the Hazens Notch formation are not conducive to the formation of garnet.

From the structural cross sections across the Green Mountain anticlinorium (Pl. 1) it may be noted that the Hazens Notch formation should occur on the west side of the anticlinorium and yet only three small areas of the formation may be found. In these areas the rocks are not entirely typical as the amount of graphite is small and the interbeds of quartzite are missing. Because of this the contacts could not be accurately established and are shown rather diagrammatically. In the Mount Mansfield quadrangle the Hazens Notch formation extends over the Green Mountain anticlinorium near Waterville and a tongue extends southward on the west side of the anticlinorium. The distribution of the formation in both the Mount Mansfield and Camels Hump quadrangles clearly shows that a change in facies occurs in an east-west direction and that most of the graphitic rocks are lost as they are traced westward over the anticlinorium.

### *Pinney Hollow Formation*

#### General Statement

The Pinney Hollow formation was named by Perry (1928) for the pale green quartz-chlorite-mica schist occurring at Pinney Hollow in the Bridgewater-Plymouth area. The formation has been mapped in the Rutland area (Brace, 1953), the Rochester area (Osberg, 1952) and in the Lincoln Mountain quadrangle (Cady, personal communication, 1959) where the formation occurs at the top of the Camels Hump group. Although some rocks similar to the Pinney Hollow formation occur locally in the Montpelier quadrangle (Cady, 1956), the formation apparently does not extend this far north.

#### Distribution and Description

In the Camels Hump quadrangle, the Pinney Hollow formation occurs only in the southeast and can be traced only a short distance to the north

(Pl. 1). The rock is typically a quartz-sericite-chlorite schist with little graphite. In contrast to the formations above and below it, which weather to give rusty and dark graphitic surfaces, specimens of the Pinney Hollow formation have clean, shiny surfaces. The lack of such rocks on the east side of Crossett Mountain and the presence of a thick section of graphitic rocks indicate that the formation does not extend northward through the quadrangle.

At the lower contact of the Pinney Hollow formation in the low hills at the southernmost part of the quadrangle, garnet occurs at one locality (Loc. 29, Fig. 2). As this occurrence is somewhat anomalous relative to the occurrence of garnet to the north, it will be discussed more fully in the section on metamorphism.

## *Ottauquechee Formation*

### General Statement

The Ottauquechee formation was named by Perry (1928) for the slate-gray quartzites and dark gray or black phyllites which occur in the Ottauquechee River Valley near Bridgewater, Vermont. The formation has been traced northward on the east side of the Green Mountain anticlinorium from the Rutland area (Brace, 1953) through the Lincoln Mountain and Montpelier quadrangles (Cady, 1956 and personal communication, 1959) into Canada where it has been correlated with the Sweetsburg formation (Osberg, personal communication, 1959). On the west side of the anticlinorium, it has been identified in the northern part of the Mount Mansfield quadrangle (Christman, 1959). A thin tongue of the formation appears to extend as far south as Underhill.

Because the Ottauquechee formation occurs more extensively north-east and south of the Camels Hump quadrangle and has been studied and described by others, the formation was not studied in detail. Its occurrence in the Camels Hump quadrangle is limited.

### Description of Rocks

In the Camels Hump quadrangle, the Ottauquechee formation consists of graphitic phyllite or schist, quartzite, quartz-sericite schist and calcareous greenstone with the graphitic phyllite being the most characteristic of the formation. The amount of graphite varies from sections composed almost entirely of graphitic phyllite to sections composed of less than five percent graphite. In the latter sections, most of the rock is quartz-sericite schist. The presence of graphite, however, cannot be the sole criterion for mapping the formation as both the underlying Hazens

Notch formation and the overlying Stowe formation contain graphitic phyllite.

Gray to blue-gray massive quartzite occurs in thick beds near the top of the formation. Some of these beds are as much as 25-30 feet thick;one such bed is being quarried south of Duxbury. Although it is not possible to trace these beds continuously because of the glacial cover, the thicker beds which were mapped (shown as rows of dots on Pl. 1) appear to be somewhat discontinuous. These form a zone which is considered to approximately mark the top of the formation (Cady, personal communication, 1959). The contact with the overlying Stowe formation is gradational with the principal distinctions being that the Stowe formation contains almost no quartzite and less graphite.

The calcareous greenstone comprises only a small part of the Ottauquechee formation, but because the rock is distinctive, it was mapped in detail to delineate the trend of the rock units. In general, the greenstone forms narrow units which can be traced along their stride with local thinning and thickening. Most of the rock is an epidote-albite-chlorite schist with variable amounts of calcite, quartz, biotite and magnetite. Amphibole is generally absent, but small crystals of femaghastingsite, having the same optical properties as that found in the greenstone at Gillett Pond, occurs in small knots rich in epidote and chlorite at a few localities in the Ottauquechee greenstone. The calcareous greenstone is presumed to represent metamorphosed beds of volcanic ash and debris rather than lavas because of their sedimentary appearance, but this cannot be proven. The lowermost layers of greenstone were used to approximately locate the lower contact of the formation.

#### Distribution and Age

The Ottauquechee formation occurs principally as a narrow belt which cuts across the southeast corner of the quadrangle. In the northwest corner of the quadrangle two small areas of graphitic rocks, which are along strike with those tentatively mapped as Ottauquechee in the Mount Mansfield quadrangle, have been designated as Ottauquechee to agree with the previous work. Although both occur along proposed synclinal areas, this correlation is very tenuous.

The Ottauquechee formation is tentatively assigned to the Cambrian by most writers, although fossil evidence is lacking.

### *Stowe Formation*

The Stowe formation was named by Cady (1956) for the sericite-quartz-chlorite schist with thick interbeds of greenstone which are typically developed in the southeastern part of Stowe township and adjoining areas in the Montpelier quadrangle. It lies above the Ottauquechee formation and below the Moretown formation. The formation has been mapped to the north (Albee, 1957) and to the south (Osberg, 1952). Because of its limited occurrence in the southeast corner of the Camels Hump quadrangle, the formation was not studied in detail, except that the calcareous greenstones were carefully mapped (Pl. 1). Cady assigns a Lower Ordovician age to the Stowe formation.

### *Pinnacle, Underhill and Hazens Notch Formations in the Mount Mansfield Quadrangle*

When the senior author mapped the Mount Mansfield quadrangle, the rocks in the Camels Hump group were not subdivided because the formations comprising this group had not been clearly defined and the differerent rock types in the Mount Mansfield quadrangle appeared too gradational to warrant being used to define new formations. However, with the subdivision of the group in the Camels Hump and Lincoln Mountain quadrangles, it seemed advisable to re-examine the data collected and tentatively subdivide the Camels Hump group. Accordingly, the group was subdivided into the Pinnacle, Underhill and Hazens Notch formations (Fig. 4).

The Pinnacle formation comprises much of the northwestern corner of the Mount Mansfield quadrangle. It appears to overlie the Tibbit Hill volcanics in the northern part of the quadrangle, but Pinnacle-type rocks are also found within the unit of Tibbit Hill schist. The Pinnacle formation occurs on either side of the Fletcher anticline, but the tongue on the east side dies out to the south in what is interpreted as a change in facies. The general distribution of the Pinnacle formation on the 1959 geology map (Pl. 1, Christman, 1959) is shown by the bright yellow pattern of "coarse-grained metagraywacke and metaconglomerate". The only exception is a few areas in the vicinity of English Settlement which cannot be shown to be Pinnacle formation, although it is possible that the formation is brought to the surface by an anticline mapped in the Camels Hump quadrangle which is trending toward this area.

The Underhill formation comprises much of the Camels Hump group in the Mount Mansfield quadrangle (Fig. 4). It should be noted that in the southwest part of the area the Underhill formation lies over the Tibbet Hill volcanics because the Pinnacle formation is missing. As previously mentioned, it is likely that the coarse-grained rocks of the Pin-

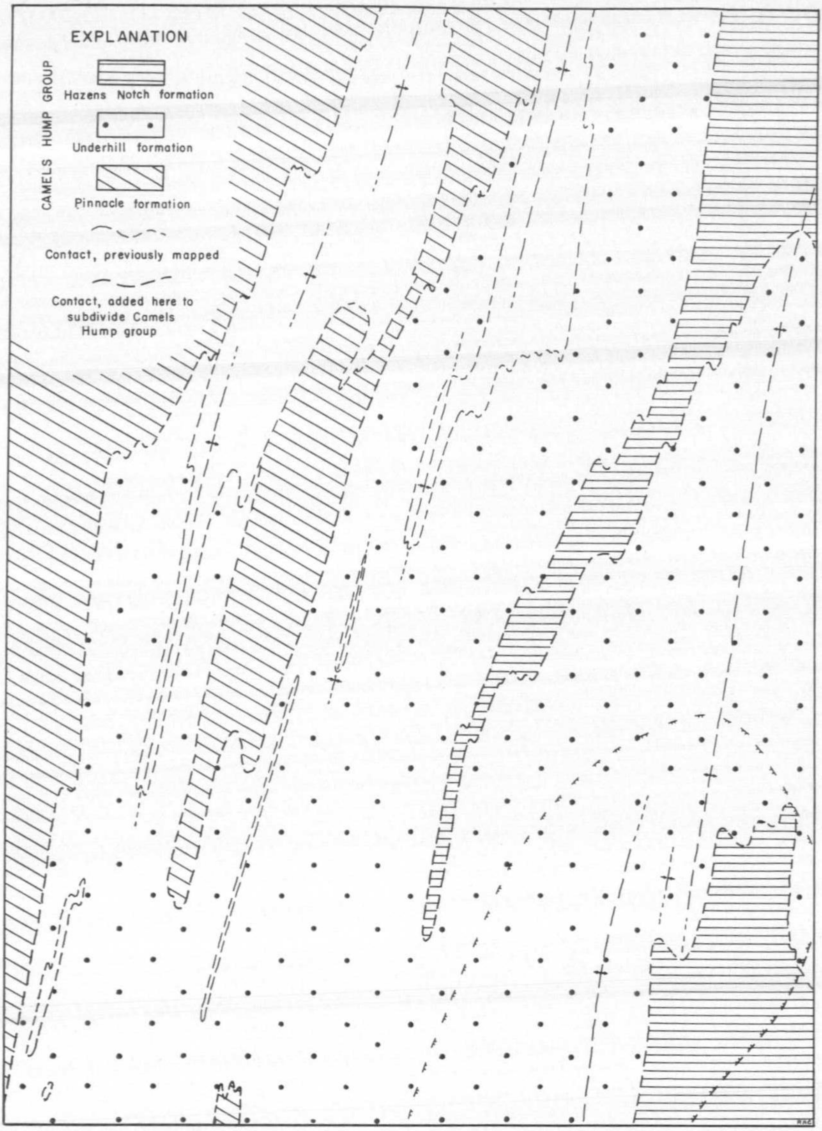



Figure 4. Geologic map of the Mount Mansfield Quadrangle showing the subdivision of the Camels Hump group into the Pinnacle, Underhill and Hazens Notch formations.

nacle formation representing a stratigraphic unit, occur at a greater depth relative to the volcanics which represent a time unit. In the Camels Hump quadrangle it has been shown that the volcanic rocks, similar to the Tibbit Hill greenstones, occur within the Underhill formation.

In the eastern part of the Mount Mansfield quadrangle the Underhill formation extends to the summits of the Green Mountains and includes a variety of rock types (Table 4, Christman, 1959).

The Hazens Notch formation is well developed on the east side of the Green Mountain anticlinorium in the southeast corner of the Mount Mansfield quadrangle. As in the Camels Hump quadrangle, the formation has a facies change as it is traced westward, so that the formation is poorly developed on the west side of the Green Mountain anticlinorium. A tongue of the formation has been recognized as extending southwest from the northeast corner of the quadrangle (Fig. 4). On the 1959 geologic map the distribution of this formation is generally shown by the spots of blue color representing "graphitic phyllite and schist".

## INTRUSIVE IGNEOUS ROCKS

### *Serpentinite, Talc-carbonate Rocks and Steatite*

Ultramafic rocks, which have been altered to serpentinite, talc-carbonate rocks and steatite, were observed in the Camels Hump quadrangle only at three localities where they occur as several small bodies. The two largest bodies, located about one mile northwest of South Duxbury, are associated with calcareous greenstones of the Ottauquechee formation. These are composed principally of serpentine with only traces of talc. Undoubtedly, they are related to the larger bodies of altered ultramafic rocks which occur just to the south, along strike, in the Lincoln Mountain quadrangle. These deposits and other similar deposits which occur in northwestern Vermont have been extensively studied by the U.S. Geological Survey. The rocks are considered to be Ordovician in age.

Other minor occurrences of serpentinite and minor amounts of talc are located slightly over one mile southwest of Phillips School in the Ottauquechee formation and on the east slope of Crossett Mountain in the Hazens Notch formation (Pl. 1). At the latter locality a small amount of talc-carbonate rock is exposed in a prospected area of about 200 square feet. Hitchcock (1861) reported talc on top of Woodward Mountain and steatite at two localities near Waterbury, but these localities were not found in this study.

### *Basic Dikes*

Post-metamorphic dikes were observed at fourteen localities in the Camels Hump quadrangle (Pl. 1). These dikes are dark colored and generally are aphanitic. Most of them are one to three feet wide, although those in the upper part of Cobb Brook are as much as eight feet wide. Except for the dike west of Gillett Pond (Loc. 30, Fig. 2) which strikes N. 10°E. and one in Jones Brook at N.30°E., the dikes have been emplaced in the transverse joint set which strikes approximately east-west. These strike directions are more fully summarized in the section on joints in this report.

Thin sections of dikes from four localities revealed considerable variation in the mineral suite. One (Loc. 30, Fig. 2) contained augite and biotite in a groundmass of glass; another (Loc. 31, Fig. 2) contained mainly pigeonite and chlorite in a groundmass of glass; another (Loc. 32, Fig.2) contained labradorite, augite and alteration products, and still another (Loc. 33, Fig. 2) contained plagioclase, augite and minor amounts of biotite, calcite, chlorite and magnetite. These rocks could be classified as basalts. The authors did not recognize abnormal compositions or pronounced development of euhedral mafic minerals which might warrant calling them lamprophyres.

Similar dikes have been observed throughout much of Vermont and they generally are considered to be related to the White Mountain plutonic-volcanic series. Although they previously had been considered to be Mississippian, recent work (Lyons, et. al., 1957) suggests that they are Permian in age.

### *Sequence of Igneous Activity in Northwest Vermont*

Although igneous activity in the Camels Hump quadrangle is extremely limited, it should be noted that the sequence of igneous activity in northwest Vermont closely follows the broad pattern which has been observed in mountainous areas which have undergone a single orogenic cycle. The general sequence as given by Turner and Verhoogen (1951) is:

1. Eruption of dominantly basic (including spilitic) lavas, during the geosynclinal phase of the tectonic cycle.
2. Injection of ultrabasic and basic plutonic intrusions during the early stages of folding.
3. Development of granodioritic and granitic batholiths during and following the main period of folding.
4. Surface eruption of basalts, andesites and rhyolites during and following elevation of the folded mass, typically at a much later date.

Stage 1 is represented in the Camels Hump quadrangle by the greenstones, some of which are spilitic lavas, in the Underhill and other Cambrian and Lower Ordovician formations of the eugeosyncline. Stage 2 is represented by the ultrabasic rocks, now mostly serpentinite, of Ordovician age which have been highly folded with the country rock. Stage 3 is not present in the Camels Hump quadrangle but may be represented by the Devonian granitic intrusions 12 to 20 miles to the east, and stage 4 may be represented by the basic dike rocks related to the Permian volcanic rocks which occur much farther to the east.

## METAMORPHISM

Metamorphism of the rocks in the Camels Hump quadrangle is assumed to be due to regional metamorphism because the zone of the more intense metamorphism coincides with the axis of the Green Mountain anticlinorium for much of its length in Vermont. As the only igneous rocks in the area are a few ultrabasic bodies and basic dikes, there is no reason to postulate a batholith with depth. The nearest granitic intrusions are exposed twelve miles to the east at the eastern margin of the Montpelier quadrangle.

All of the rocks in the Camels Hump quadrangle are assigned to the greenschist facies. The subfacies of pelitic assemblages, (1) quartz-albite-muscovite-chlorite (chlorite zone), (2) quartz-albite-epidote-biotite (biotite zone) and (3) quartz-albite-epidote-almandine (low grade portion of the almandine zone) (Turner, 1958), are present with the quartz-albite-muscovite-chlorite subfacies being the most common, particularly in the western part of the quadrangle. Despite the presence of hornblende rather than actinolite in the metamorphosed basic rocks, these rocks are placed in the greenschist facies because of the low-grade metamorphism of the surrounding rocks and the presence of albite rather than oligoclase as the stable plagioclase. As concluded earlier, the hornblende must reflect the particular composition of the volcanic rocks rather than an advanced degree of metamorphism.

A garnet isograd is shown on the geologic map (Pl. 1) as paralleling the trend of the Green Mountain anticlinorium and indicates that the degree of metamorphism was slightly higher in the core of the anticlinorium. As Christman (1959) determined that the garnet from the Mount Mansfield, along the same zone, contained more of the spessartite than the almandite molecule, the isograd may not represent a true metamorphic isograd which is defined on the basis of almandite garnet. It is known that spessartite, because of its composition, may occur in metamorphic rocks originating under conditions of slightly lower temperature and pressure than those assigned to the garnet zone. However, as many of the garnet isograds mapped in the past are based on garnet simply inferred to almandite (which has been done in Vermont) the garnet isograd is shown despite its incorrect composition.

As the crystalloblastic development of garnet is a factor of both degree of metamorphism and composition of the original rocks, all rocks within the garnet zone do not contain garnet. Conversely, if the rocks are not favorable for the formation of garnet, the isograd, mapped on the first appearance of garnet, may not always represent the same metamorphic conditions. This problem manifests itself on the east side of the anticlinorium where the graphitic rocks of the Hazens Notch formation do not favor the formation garnet. This is indicated by the fact that the first occurrence of garnet closely parallels the lower contact of the Hazens Notch formation in the Camels Hump quadrangle. To the south in the Lincoln Mountain quadrangle, the Pinney Hollow formation, which apparently is more favorable for the formation of garnet because of its quartzose nature, contains garnet at its lower contact, so that the isograd lies farther east (Cady, personal communication, 1959). At one locality west of South Duxbury in the Camels Hump quadrangle (Loc. 29, Fig. 2) garnet occurs in the Pinney Hollow formation in agreement with the line drawn by Cady to the south. However, as no line could be drawn northward through the barren Hazens Notch formation, it was decided to draw the isograd on the first occurence to the west, even though it is evident that the first occurrence would be farther to the east if the composition of the rocks were different.

## STRUCTURE

### Large Folds

Although the rocks in the Camels Hump quadrangle, particularly those in the western part of the area, have been highly folded, only four folds are shown on the geologic map (Pl. 1). The Green Mountain anticlinorium occurs in the east-central part of the area with a syncline to the west and another anticline and syncline occurring farther to the west. Of these, the Green Mountain anticlinorium is the most distinct as it is defined by broad folding of the beds. The other three structures are isoclinal folds so that their positions are less distinct and are determined partly by the stratigraphy.

#### Green Mountain Anticlinorium

The major structural feature in the Camels Hump quadrangle is the

Green Mountain anticlinorium. Although the structure in this area is a large anticline rather than an anticlinorium, the latter term is used because this name is established elsewhere in Vermont. The Green Mountain anticlinorium has been traced for the length of Vermont and appears to be more complicated in southern Vermont.

In the Camels Hump quadrangle the anticlinorium does not have a single axis but consists of a series of axes which are slightly offset from each other. Individually, the traces of the axes are approximately north-south, but the overall trend of the anticlinorium is about N. 10° E. because these offsets, without exception, are to the northeast when the anticlinorium is traced northward. The dips of the axial planes are presumed to be steep to the east because fracture cleavage and the axial planes of drag folds dip steeply eastward. A slight westward overturning of the anticlinorium also is suggested by steeper average dips west of the axial region.

The summits of the Green Mountains only roughly coincide with the crest of the anticlinorium. At the southern border of the quadrangle, the axis of the anticlinorium lies just east of Burnt Rock Mountain, and it dies out as it is traced northward. Farther east of Burnt Rock Mountain a small syncline is indicated by a series of westward dipping rocks on the east side of the syncline with the axis of the anticlinorium occurring still farther east. The axis of this structure may be traced northward, passing east of Mt. Ira Allen and Mt. Ethan Allen to southeast of Camels Hump where it dies out. Here the structure is again offset to the northeast so that the trace of the axis is almost one mile east of Camels Hump Mountain. In an area of flat dips south of the Winooski River, the axis appears to die out again and is offset to the northeast. This area contains rocks of variable, low dips which form bare exposures, clearly visible from the Winooski River (Pl. 4). The trace of the anticlinorium crosses the Winooski River about one mile east of the village of Bolton and the flat dips of the beds may be seen in the cliffs north of the river and U.S. Highway 2. The axis may be traced farther north, passing west of Bone Mountain, to the west side of Bolton Mountain where it dies out. A small syncline occurs between it and the next portion of the axis which passes through Bolton Mountain and northward through Mount Mayo. In the vicinity of Nebraska Notch, this axis dies out and another picks up to the northeast, passing northward through Dewey Mountain and into the Mount Mansfield quadrangle. At Nebraska Notch an intervening syncline may be present, but it was not recognized because of the many variable strike directions and low dips of the rocks.

The axis of the Green Mountain anticlinorium to the north in the

Mount Mansfield quadrangle and to the south in the Lincoln Mountain quadrangle shows similar offsets to the northeast, when traced northward (Christman, 1959 and Cady, personal communication, 1958).

These offsets of the major anticline appear to be due to the dying out of one anticlinal structure and the development of another parallel to the first but offset from it. No evidence of faulting was found and the presence of an intervening synclinal structure between some of these offsets is a further indication that these are not fault displacements.

Christman (1959) postulates that the region was initially subjected to a deformational force from S.70°E. which broadly folded the original sedimentary rocks into a series of anticlines and synclines trending about N.20°E. The deformational forces are envisioned to have gradually shifted until they were east-west so that north-south trending anticlines and synclines were superimposed on the northeast-trending rock units. It is postulated that the position of the Green Mountain anticlinorium is controlled by a core of more resistant Precambrian rock (not shown on the cross sections of Plate 1) whose trend was determined by the early northeast folds.

Although Precambrian rocks do not occur in the area, the broad fold of the Green Mountain anticlinorium as compared to the isoclinal folds to the west and east suggests that the anticlinorium has a resistant core whereas the other folds do not. About ten miles south of the Camels Hump quadrangle, Cady (personal communication 1959) has mapped the northernmost exposures of Precambrian rocks just east of Lincoln. In his cross sections of the areas south of the Camels Hump quadrangle, Cady indicates that Precambrian rocks probably occur in the core of the Green Mountain anticlinorium.

No culmination of the Green Mountain anticlinorium is apparent in the Camels Hump quadrangle. The rocks along the axial region gently dip both north and south (Pl. 2), indicating that the anticlinorium has no uniform plunge.

### Isoclinal Folds West of Green Mountain Anticlinorium

In the northern part of the quadrangle two synclines and an anticline were mapped west of the Green Mountain anticlinorium while in the southern part only the syncline adjacent to the anticlinorium on the west was recognized.

The position of the syncline west of the anticlinorium is shown by the reversal of dips of the beds (Pl. 2). A distinct zone occurs where the steeply dipping beds on the west flank of the anticlinorium give way to beds dipping steeply to the east. Although such a change in direction of dip might be interpreted as simple overturning, the synclinal structure is postulated because of stratigraphic requirements and the local development of flat-lying beds along this zone. Beds with low dips are best observed east-northeast of Jonesville in the vicinity of Duck Brook and east of Hanksville along Jones Brook. The variation in the position of the zone where the reversal of dips occurs suggests that the syncline is offset to the northeast, when traced to the north, in a manner similar to that of the Green Mountain anticlinorium. The only exception to this order occurs northwest of Bolton Notch where one of the offsets is to the northwest. However, here much of the area is covered so that the evidence for this exception is meager.

In the western part of the quadrangle the occurrence of the metagraywacke of the Pinnacle formation, which is considered to be the oldest formation in the quadrangle, requires that a major syncline be present between these rocks and the anticlinorium. In addition, if no syncline were present, the Ottauquechee and Stowe formations would be expected to occur on the west side of the anticlinorium. The position closest to the anticlinorium where the syncline might occur is in the zone where the reversal of dips occurs, because a position closer to the anticlinorium would require that the syncline be isoclinal with its axis dipping westward. This is highly improbable because minor structures indicate that the axes of the folds are either vertical or steeply dipping to the east.

In conclusion, although the position of the syncline cannot be proven everywhere by the location of flat-lying beds in the axial region, its position in the zone of reversal of dips seems most likely. It cannot occur as an isoclinal fold to the east because of the structural evidence. It could occur as an isoclinal fold to the west, with an eastward dipping axial plane, but the stratigraphy requires that it be as far east as possible.

To the south, the syncline trends towards a syncline mapped by Cady in the Lincoln Mountain quadrangle (personal communication, 1959). To the north it trends towards a synclinal area, named the Cambridge syncline, in the vicinity of Cambridge which could not be traced in the southern part of the Mount Mansfield quadrangle (Christman, 1959).

In the northern part of the Camel's Hump quadrangle an isoclinal anticline was mapped west of the syncline on stratigraphic evidence. The distribution and age of the Pinnacle formation require that it occur in the core of an anticline. As the beds in this area are steeply dipping, with local variations due to drag folds, the position of the anticline cannot be determined by the dips of the beds. Other structural evidence, as the nature of the drag folds and bedding-cleavage relations are too variable to accurately locate the position of the major anticline. In fact, the minor structures indicate that the rocks are complexly folded into a series of small isoclinal folds so that the structure is probably an anticlinorium without a distinct anticline occurring along the culmination. However, as it is necessary to indicate the structure on the geologic map, the anticline was postulated to occur near the center of the area underlain by the Pinnacle formation (Pl. 1). Its exact position could not be determined, but was placed on the map on the basis of a linear topographic low and a few reversal of dips. Although it could not be mapped, the anticline probably extends south of the Winooski River and is probably offset to the west to expose the Pinnacle formation in its core north of Huntington Center (Pl. 1).

The third fold west of the anticlinorium is an isoclinal syncline mapped in the northwest part of the quadrangle on the basis of stratigraphic evidence. To the west of the quadrangle, outside of the mapped area, the Pinnacle formation has been identified by Cady (personal communication, 1960) just west of Richmond. The distribution of this formation requires that a synclinal area occur somewhere east of Richmond. Again, the minor structures suggest that the area is highly folded so that the placement of only one synclinal axis is probably an oversimplification of the structure. However, although the exact placement of the synclinal axis is arbitrary, it does indicate the nature of the overall structure. In the northern part of the quadrangle this syncline strikes into the snyclinal area marked by the position of the Ottauquechee formation in the Mount Mansfield quadrangle (Christman, 1959), so that it also may be part of the Cambridge syncline.

The Enosburg Falls-Fletcher anticline lies west of the Camels Hump quadrangle.

### *Bedding and Bedding Schistosity*

Although bedding may be recognized by textural and compositional layering in most of the rocks in the Camels Hump quadrangle, movement and recrystallization has occurred along the bedding planes so that the structure in most places is now bedding schistosity (Pl. 8). Bedding alone may be observed in places in the western part of the quadrangle where the degree of metamorphism is least intense and where the graywacke and quartzose rocks lack the closely spaced bedding planes needed for the development of bedding schistosity. At several

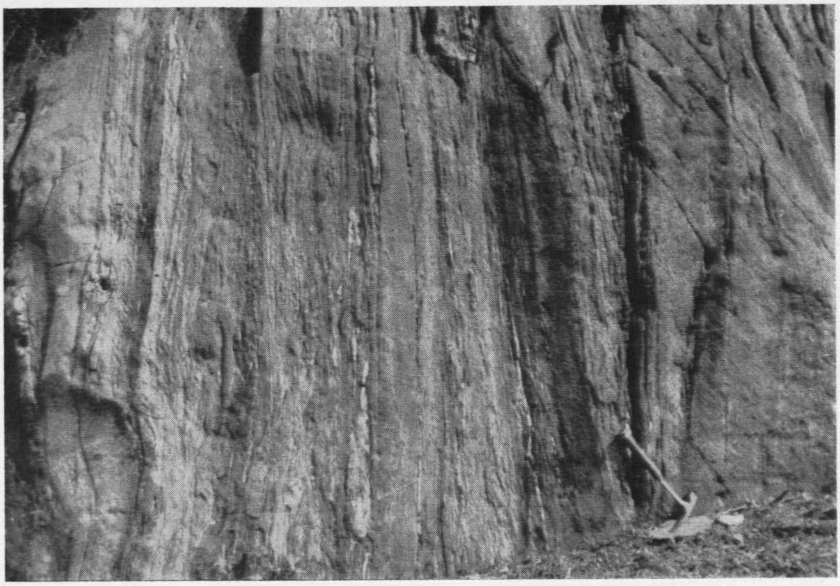


Plate 8. Vertical bedding shown by compositional layering, near Jonesville (Loc. 2, Fig. 2). The rocks have poorly developed bedding schistosity and moderately developed fracture cleavage dipping 45° to the east (parallel to the hammer handle).

places in the Pinnacle and Underhill formations (Pl. 1) in the northwestern part of the quadrangle, graded bedding may be observed. At some of these localities the beds are overturned, indicating that isoclinal folding has occurred. However, the rarity of graded bedding and the variation in direction of the "tops" and "bottoms" within a given series of rock exposures prohibits using graded bedding to unravel the structure of the area. In the central and eastern parts of the quadrangle where the metamorphism is more intense, bedding may still be recognized by the compositional layering and textural differences. The grade of metamorphism is not sufficiently high that metamorphic differentiation masks the original layering.

In most of the area, bedding schistosity is the most prominent planar feature. It results from the parallel alignment of chlorite and mica which have formed by recrystallization and whose orientation is related to slippage along the bedding planes. During deformation of the area, the compressional forces were resolved by a multitude of small-scale slippage movements along bedding planes in the formation of the large and small folds. Thus, the development of bedding schistosity began with the initial

Plate 9. Bedding schistosity with dextral drag folds, east of Jonesville (Loc. 34, Fig. 2). Rock is mica-chlorite schist with quartzose and calcareous layers. Folds are at N.5°E. 60°SE. plunging 20° N.15°E.

stages of folding and continued to develop throughout the period of deformation where folding occurred. In places where the planes of bedding schistosity are intricately folded (Pl. 9), the final development of bedding schistosity probably took place during the maximum period of metamorphism.

The structural map (Pl. 2) shows that on the average the bedding schistosity strikes north-south and dips steeply to the east or west, except along the axis of the anticlinorium. It is presumed that the data collected on bedding schistosity are representative, but local variations due to drag folding requires estimating average values in some areas. If along a given traverse seventy-five per cent of the beds had a similar strike and dip, usually steep, this reading would be recorded in spite of local variations. Although many of the variations are due to small drag folds and are not significant, others may represent the narrow axial regions of isoclinal folds which would be significant if they could be recognized. Thus, the structural map which represents the average attitudes of the bedding schistosity suggests that the rocks are monoclinal in their structure, whereas, in truth the rocks probably are isoclinally folded in a complex manner, as is suggested by the distribution of the mappable rock units such as the greenstones.

### *Drag Folds*

Drag folds are common in the Camels Hump quadrangle. They range in size from microscopic folds observed in rock thin sections to folds with amplitudes of six to ten feet observed in single rock exposures. Undoubtedly, larger drag folds exist, as indicated by the variations in dips of beds on the limbs of the larger folds, but none could be defined exactly because of the lack of continuous exposures. The distribution and pattern of the amphibolitic greenstones east of Huntington suggest drag folding on a large scale (Fig. 3).

Drag folds are visualized as forming when compressional forces become too large to be absorbed by the multitude of movements along the planes of bedding schistosity so that adjustment takes place by the movement of a competent bed, or a series of such beds, resulting in the crumbling and folding of the intervening less competent beds. The axial planes of these drag folds, theoretically, are in the shear positions relative to the major fold, but in practice this position is often subparallel to the axial planes of the major folds (Nevin, 1949, p. 166). For this reason the attitude of the axial planes of the major folds may be determined from the drag folds. Throughout the Camels Hump quadrangle the axial planes of the drag folds strike approximately north-south and dip steeply to the east. These folds may be particularly well developed in some of the calcareous greenstones (Pl. 10).

The relation of the long and short limbs of the drag folds, as seen in cross section, should indicate the position of the major structure. On the structural map (Pl. 2) the letters, T.E. and T.W., abbreviating "tops east" and "tops west," indicate where interpretations from cross-sectional views of drag folds were recorded. East and west of the axis of the Green Mountain anticlinorium the drag folds gave the correct structural information regarding the anticlinorium, as do the dips of the beds, so that many of the observations in these areas were not recorded. However, in the western part of the quadrangle the T.E. and T.W. data are not consistent. Southeast of Richmond (center of extreme west portion of plate 2) drag folds indicated seven changes in the directions of the tops of beds along an east-west traverse of only slightly over two miles. This implies that the rocks are complexly folded and that no one major fold is in control. A careful study, not shown on the map, of all the drag folds along the Winooski River in the western part of the area

Plate 10. Overturned drag folds in calcareous greenstone east of Huntington Center (Loc. 26, Fig. 2). View is to the north, so that the axial planes of the folds dip about 60° to the east.

resulted in so many conflicting "tops east and tops west" determinations that it was deemed impossible to determine the overall structure by this means.

Theoretically, in a complexly folded area drag folds in plan view, termed dextral and sinistral depending upon the position of the long limb, (White and Jahns, 1950) should be about equally abundant and together with the plunge, it may be possible to determine the position of the drag fold relative to the major folds. In the Camels Hump quadrangle the majority of the folds observed were dextral (Pl. 2) with southward plunges, indicating that the majority of the rocks are on the east

Plate 11. Bedding schistosity, dextral drag folds and fracture cleavage in plan view near Jonesville (Loc. 2, Fig. 2). Note that the fracture cleavage is axial plane cleavage to the drag folds.

side of an anticline. This interpretation cannot be correct for large portions of the quadrangle. Elsewhere in Vermont it has been noted that within large areas the plan views of drag folds are predominantly one type (Christman, 1959, p. 51). Various explanations have been offered. White and Jahns (1950) suggest that in northcentral Vermont the pattern results from an early, relatively regional deformation in the form of a shear. However, Christman (1959) in noting that the folds are predominantly dextral in the Mount Mansfield quadrangle suggested that the intricate folding can occur only when the rocks are highly mobile, logically during the height of metamorphism. For this reason, he postulated that the folds are intimately related to the major deformation and that their uniformity of pattern is related to a change in direction of the deforming force. The later forces, acting at a slight angle to the earlier deformed rocks, might result in a slight shear movement which would facilitate the formation of dextral folds and would minimize the formation of sinstral folds.

### *Fracture Cleavage*

As used in this report, fracture cleavage refers to axial plane cleavage of major or minor folds along which shear has occurred with little or no development of mineral orientation. This fracturing may develop during low-grade metamorphism in those rocks which are too competent to absorb the force by movement along bedding planes. Many graywackes fail almost entirely by such fracturing because bedding planes and associated seams of argillaceous material, where slippage might occur, are too widely spaced. In the same area, competent rocks may exhibit well-developed fracture cleavage without small-scale drag folds, whereas less competent rocks may exhibit well-developed bedding schistosity and small-scale drag folding. Many rocks are intermediate between these types with some development of bedding schistosity, drag folds and fracture cleavage (Pl. 11).

As metamorphism increases, rocks which previously had deformed by movements along bedding planes may develop drag folds with fracture cleavage forming along the axial planes. With greater metamorphism, the movement and recrystallization along these shear planes becomes more intense so that fracture cleavage may grade into flow cleavage. With the development of flow cleavage, bedding becomes increasingly difficult to recognize and the grade of metamorphism is often higher than the greenschist facies. Flow cleavage was not recognized in the Camels Hump quadrangle.

The term, slip cleavage, has been used in Vermont for fracture cleavage occurring in highly schistose rocks where some mineral orientation has developed along shear planes. White (1949) in east-central Vermont shows that "slip cleavage" grades into flow cleavage. Christman (1959) in the Mount Manfield quadrangle shows that fracture cleavage grades into "slip cleavage." He suggests that the term, slip cleavage, not be used because of possible confusion to those who consider that it is synonymous with false cleavage.

In the western part of the quadrangle where metamorphism was less intense, fracture cleavage is well developed in the competent rocks with

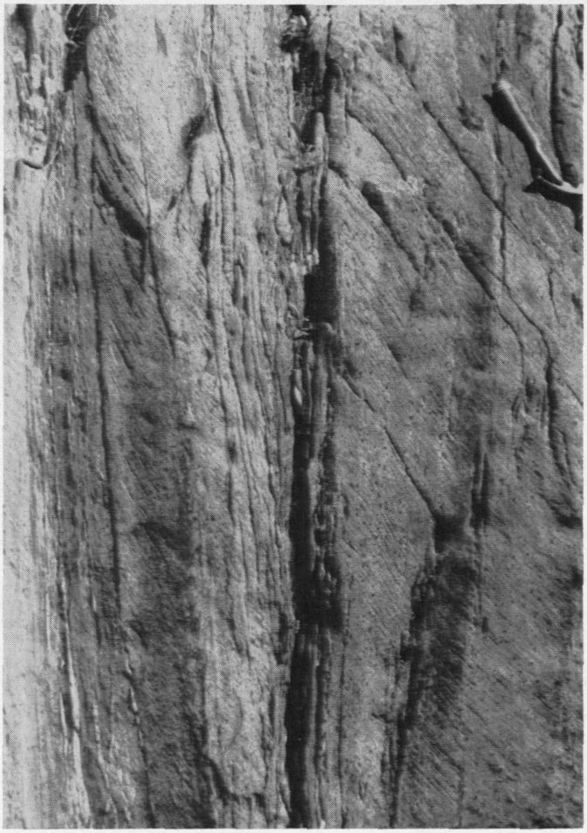


Plate 12. Fracture cleavage cutting vertical bedding, near Jonesville (Loc. 2, Fig. 2., same as Pl. 8). Note magnetite crystals in the mica-chlorite phyllite which grades into quartzitic phyllite in places. The cleavage is at N.5°W. 45°NE.

bedding schistosity being predominant in the less competent rocks. Fracture cleavage is especially well developed in the quartzose rocks of the Pinnacle formation in the vicinity of Jonesville (Pl. 12). On the average the fracture cleavage strikes from about north-south to N.15°W. and dips between 45° and 60° to the east (Pl. 2). At a few localities, the other shear position was observed where the fracture cleavage dips to the west.

In the central and eastern parts of the quadrangle the fracture cleavage is less conspicuous because the original rocks were finer grained and metamorphism was more intense. Although fracture cleavage may be present as axial plane cleavage in small-scale folds, it may be completely lacking in some drag folds (Pl. 9 and 10). Where it is developed in the schist, particularly along the Green Mountain anticlinorium, some micaceous minerals may be aligned parallel to the fracture. It is this feature which previously has been called slip cleavage.

Flow cleavage or slaty rocks were not observed in the Camels Hump quadrangle. Apparently none of the original rocks were sufficiently argillaceous for the development of slate. Instead the rocks were metamorphosed to phyllite and schist and the degree of metamorphism was not great enough for these to develop flow cleavage.

### *Crinkle Lineation*

The intersection of fracture cleavage with the bedding schistosity produces a lineation which is marked by small crinkles, or tiny folds, or more rarely by a series of faint lines. In the western part of the area where fracture cleavage and bedding schistosity may be developed in the same rock, crinkles may be conspicuous on the planes of bedding schistosity (Pl. 13). As the bedding schistosity dips steeply, the crinkle lineations strike in a general north-south direction similar to the strike of the bedding. The lineations plunge gently either to the north or south.

Along the axis of the Green Mountain anticlinorium, crinkles may be especially conspicuous in the flat-lying beds, even though the fracture cleavage does not form a recognizeable break. Most of these lineations are formed by the alignment of the crests of the crinkles which are parallel to the trend of the major and minor folds. These lineations plunge gently north or south without consistant variation.

In a few places, crinkle lineations strike east-west. These are attributed to north-south movements along the bedding schistosity caused by adjustments related to the formation of local culminations along the Green Mountain anticlinorium. The number of such east-west crinkles observed was much smaller than in the Mount Mansfield quadrangle to the north.

### *Quartz Rodding*

Quartz rodding, having an east-west strike, was recorded only at about ten localities. Apparently the rodding results from the concentration of quartz in the tension position of east-west folds, which would be undiscernible if it were not for the quartz. The origin of these structures is not definitely known (Christman, 1959), but their origin may be similar to that of the east-west crinkles.

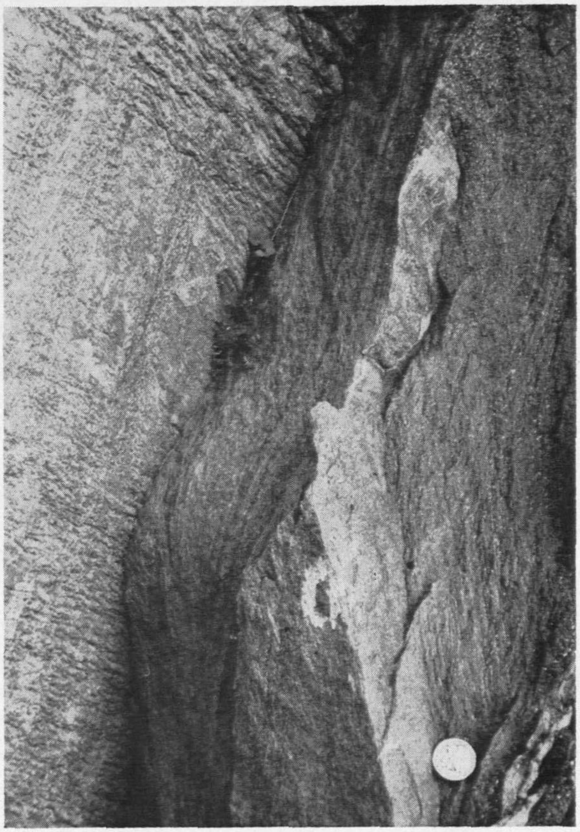


Plate 13. Crinkles on bedding schistosity plane, east of Jonesville (Loc. 35, Fig. 2). The fracture cleavage dips about 70° east (near coin) whereas the bedding dips steeply west, as shown by the compositional banding. The crinkles are exposed on the underside of the bedding plane (left side of photograph) and plunge gently north.

## *Joints*

### General Statement

A study of joints made in the Camels Hump area during the 1958 field season (Secor, 1959) showed two regional conjugate joint sets symmetrical with the trend of the major folds. Although joints cut all structural and metamorphic features, indicating that they developed after the main phase of orogeny, they are related to compressional forces from the same direction as those which caused the major folding.

Data for this joint study were obtained from field measurements and aerial photographs.

## Definitions and Theory

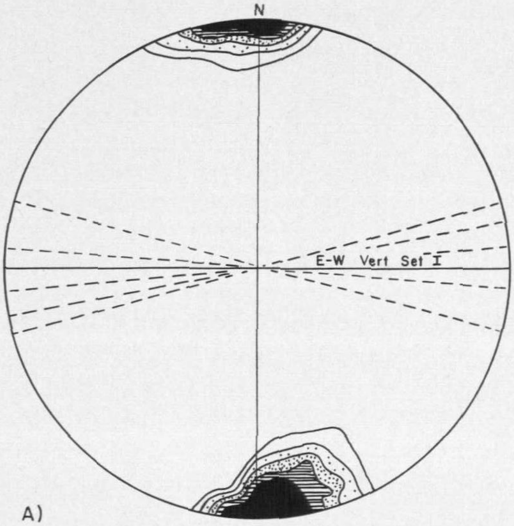
The term, joint, as used in this report, refers to a fracture in which the direction of primary movement at the time of rupture was normal to the plane of break. Joints should develop perpendicular to the direction of easiest relief, paralleling the direction of greatest compression. Apparently some shear movement, present along the joint planes, is influencial in developing the conjugate set. However, the joints in a conjugate set are still nearly perpendicular to the direction of easiest relief, and most of the movement at the time of rupture was away from the plane of the break.

In the past, widely spaced fracture cleavage and small faults sometimes have been incorrectly classified as joints. Fracture cleavage and faults are fundamentally different because they represent fracture by shear movement in which the movement is parallel to the plane of rupture. Thus, no matter how small the displacement these features should not be called joints.

## Bedding Schistosity and Strike Joints

The influence of bedding schistosity on the development of joints is important because in certain cases it tends to direct the formation of the joint. Throughout much of the Camels Hump quadrangle, the bedding schistosity dips steeply and strikes parallel to the N.10°E. structural trend. If forces tending to produce strike joints had been present, in most cases, these joints would be parallel to the bedding schistosity and would be indistinguishable from slight separations along the bedding schistosity planes. Obviously, strike joints will be recognizable only in rocks in which the bedding schistosity is poorly developed or at an angle to the joints. Thus, strike joints, when observed, are often irregular and appear to be less numerous than transverse joints.

In the Camels Hump quadrangle evidence of a regional strike joint set is found in quartzite lenses where bedding schistosity and fracture cleavage are poorly developed or absent. These quartzites also show dip joints. In the Mount Mansfield quadrangle, strike joints are beautifully developed on the summit of Mount Mansfield and in the Smugglers Notch area where the bedding schistosity is nearly horizontal along the crest of the Green Mountain anticlinorium. These joints form a compound set striking N.-S. and N.30°E. (Christman, 1959, p. 59).



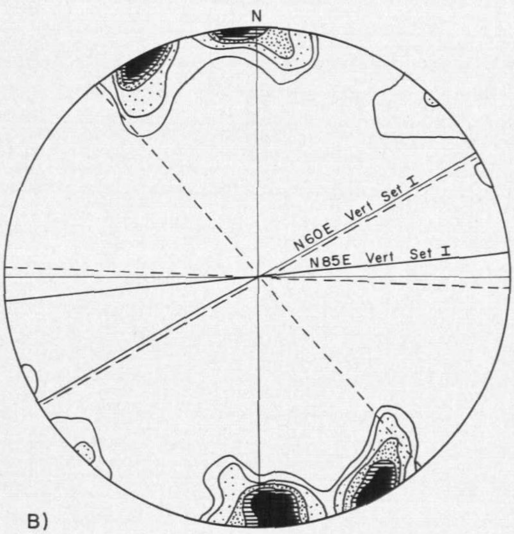




Figure 5. Joint diagrams for (A) east of Huntington (Loc. 36, Fig. 2) and (B) near Jonesville (Loc. 37, Fig. 2). Poles to 63 (A) and 97 (B) joints are plotted by projection on a lower hemisphere and contoured at 15–12–9–6–3 %. Bedrock is mica schist with attitude of (A) N.5°W. 60°SW. and (B) N.10°E. 85°SE. Broken lines are strike directions of lineations observed on aerial photographs at these respective localities.

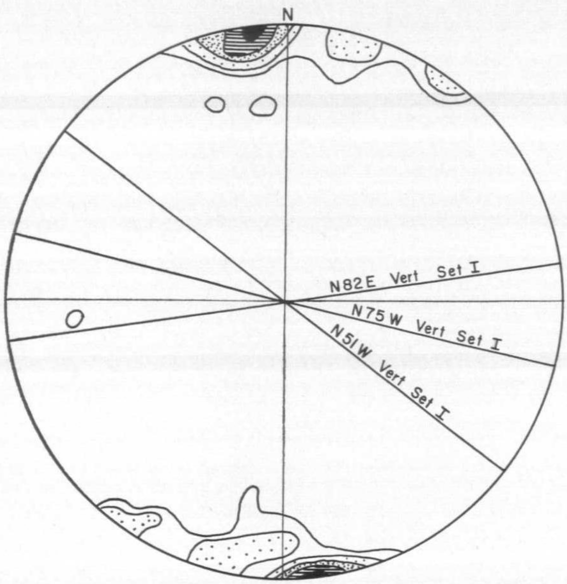



Figure 6. Joint diagram, Huntington Falls (Loc. 38, Fig. 2). Poles to 25 joint planes are plotted by projection on the lower hemisphere and contoured at 32–24–16–8–4 %. The bedrock is metagraywacke with an attitude of N.5°W. 65°NE.

## Local Joint Pattern

Before describing the regional joint pattern, something should be said about the patterns at particular localities. Because of variations in the orientations of joints, several dozen readings should be taken at the same locality with a conscientious effort to take a fair sampling. In general, the joints are spaced a few yards apart, so that readings were taken on each joint. However, in some places where the jointing was strong, they may be a few inches apart, so that it becomes impractical to record each joint. As exposures of bare rock are limited, detailed studies were made at only seven localities. Of these, diagrams of three representative localities are reproduced in this report (Fig. 5 and 6). Although some variation between these diagrams may be noted, they all show a single vertical joint direction transverse to the N.10°E. structural trend. Frequently a second, and sometimes a third group of vertical joints occur which are also transverse to the structural trend and which are inclined to the first with angles of ten to twenty degrees. It is not known how much of this divergence is due to local variations of the primary direction.

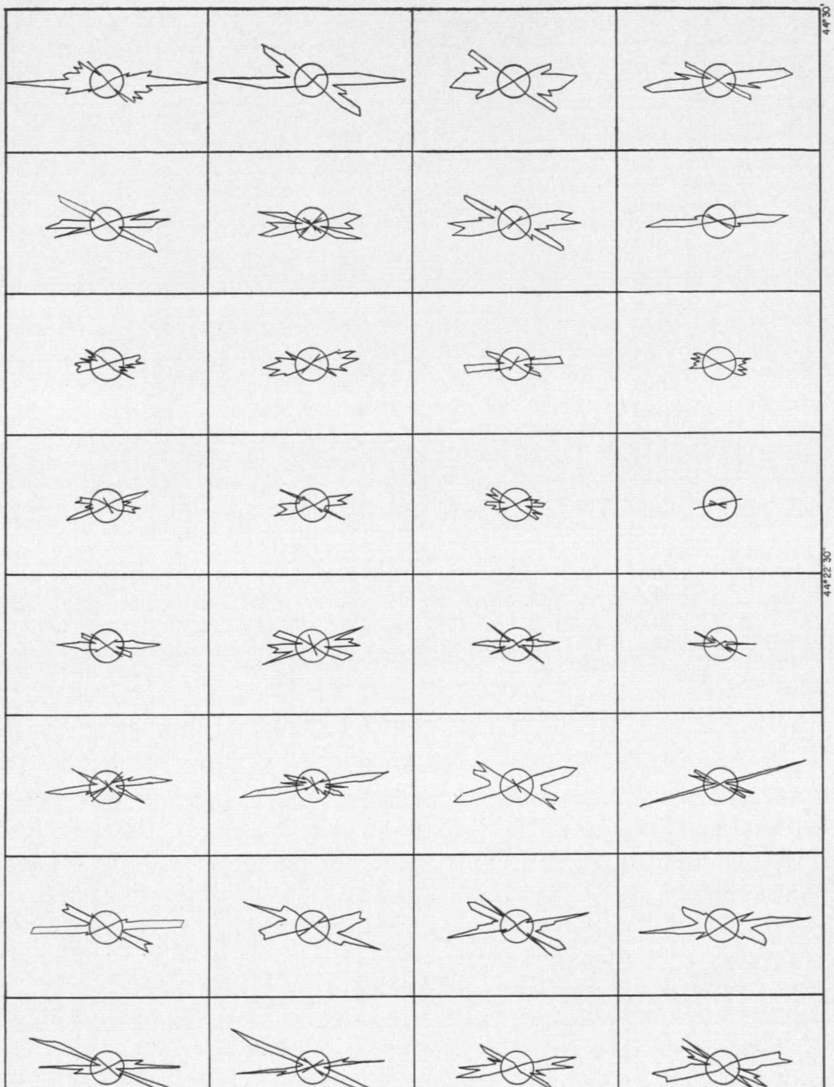
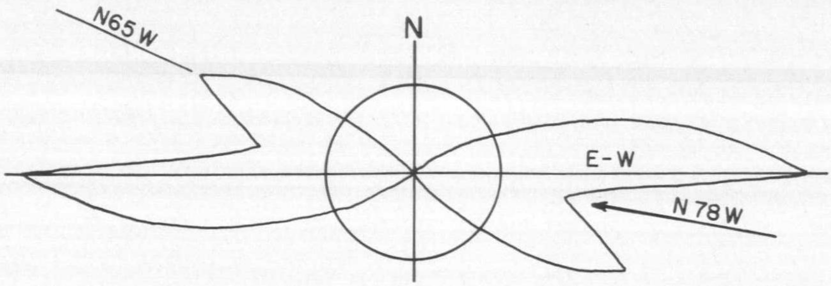
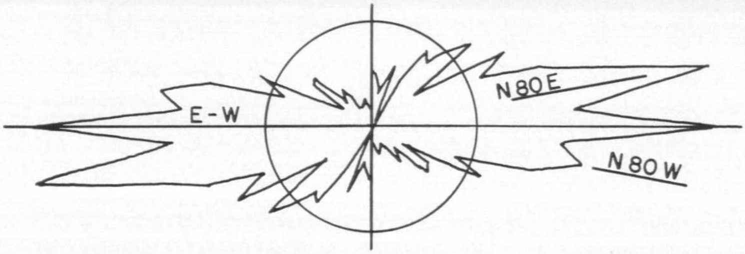



Figure 7. Joints from aerial photographs, east one-half of the Camels Hump quadrangle. Each circle has a radius which represents five joint lineations, and each of the 32 patterns records the joints measured in equal area in the eastern half of the 15′ quadrangle. The relative position of the areas is shown but the size and shape of the areas are distorted in this illustration.




A. Circle radius is 200 lineations




B. Circle radius is 20 joints

Figure 8. Composite joint diagrams: (A) 6381 joint lineations measured on aerail photographs, and (B) 666 joints measured during field work.

In the study of aerial photographs, the quadrangle was divided into 64 equal parts and the linear features within each section were recorded and plotted, omitting those features attributed to bedding schistosity. The resulting diagrams (Fig. 7 shows 32 diagrams for the eastern half of the quadrangle) indicate that considerable variation exists within the quadrangle, but no recognizable trend in the variations could be detected. It is likely, therefore, that these are local variations which are observed on the photographs because of enlargement by weathering and selective glacial plucking.

## Regional Joint Pattern

The composite diagrams of all joints measured indicate that the joints occur in a 50 degree range which is transverse to the structural trend at N.10°E. As the composite diagram of the joint lineations measured on the aerial photographs represents over 6000 measurements from the entire quadrangle, this diagram probably presents the best overall picture (Fig. 8). It shows that transverse joints occur as a compound set at N.65°W. and E.-W. with the acute angle bisected by a line striking N.78°W., which is nearly perpendicular to the structural trend. This compound set is designated as Set I.

The composite diagram based on 666 joints measured in the field (Fig. 8), although probably more biased, is quite similar to the one obtained from the photographs. In this, three principal joint directions are N.80°W., E.-W., and N.80°E.

As previously discussed, in addition to Set I, a major strike joint set is also present, similar to the one measured on Mount Mansfield, but is largely concealed by the bedding schistosity. This conjugate set is designated as Set II.

The distribution of basic dikes intruded along joints appears random and gives no indication of the relative ages of the joints. In the Camels Hump quadrangle the following vertical, or nearly vertical, dikes were observed: one at N.65°W., one at N.80°W., three at N.85°W., one at E.-W., three at N.85°E., three at N.80°E, one at N.75°E., two at N.70°E., two at N.60°E., one at N.30°E., and one at N.10°E. This data indicates that most of the dikes occur in Set I but two dikes occur in Set II.

Many workers (Balk 1926; Osberg, 1952; Brace, 1953; Eric and Dennis, 1958; and Konig, 1959) have reported joint patterns in the metamorphic and igneous rocks of central and northern Vermont which are similar to those in the Camels Hump quadrangle. Transverse joints perpendicular to the north-south structural trends are prominent and probably belong to Set I. Osberg also reports strike joints parallel to the structural trend; these probably belong to Set II. Balk describes joints which locally controlled the emplacement of the Bethel, Barre and Woodbury granites in north-central Vermont. Dennis (1956, p. 50) describes a northwest-striking joint which forms one of the boundaries of a granite in the Lyndonville area.

### Origin

The intricate folding and regional metamorphism to the greenschist facies suggests that the deformation occurred at considerable depth which is not a favorable environment for the development of joints. Furthermore, the joints cut bedding schistosity and drag folds without warping, indicating that they developed after the major deformation and probably after erosion had removed some of the overburden. However, the joints appear to be related to forces acting from the same direction as those which caused the major deformation. The acute angle between the conjugate components of Set I is bisected, approximately by the

N.80°W. direction of compressional stress, suggesting that Set I joints are primarily tension fractures resulting from the horizontal tension action normal to the direction of compression.

The conjugate components of Set II are transverse to the N.80°W. compression force and may have developed as a result of elastic expansion of the rocks as the compressive forces dissipated. Although compression might have acted from the north or south, bisecting the acute angle between the components of Set II (Christman, 1959), Sets I and II are apparently both closely related to the same deforming force and form a single system.

If the transverse fractures bounding granitic bodies in central Vermont are joints, the joint system may be older than the intrusions which are generally considered to be Devonian (Murthy, 1957; Konig 1959). As both sets are intruded by basic dikes, the joints are definitely older than the dikes which are considered to be Permian.

## Age of Deformation

The age of the deformation of the Green Mountains is not exactly known. It is generally believed that the Green Mountain anticlinorium was formed during both the Taconic orogeny near the end of the Ordovician and the Acadian orogeny of Middle or Late Devonian age. Cady (1945) believes that the thrusting occurred during the Taconic orogeny and the principal folds formed during the Acadian orogeny, whereas Booth (1950) believes that the strata were strongly folded before thrusting. If the intrusion of ultrabasic rocks coincides with the initial stages of deformation of a geosyncline as suggested by Hess (1938), then, the dating of the ultrabasics as Ordovician suggests that the earliest deformation occurred during the Taconic orogeny. In the Plainfield quadrangle, the north-south folds, which are probably the same age as the north-south folds of the Green Mountain anticlinorium, deform rocks which are Silurian and the folds, in turn, have been deformed by the intrusion of acid igneous rocks of Lower Devonian age (Konig, 1959).

The study of the rocks in the Camels Hump quadrangle contributed little to solving this problem, except that the authors found no structural or mineralogical evidence for two deformations. However, it is possible that the first deformation was weak and a second stronger deformation may have completely obliterated evidence of the earlier one. If thrusting has occurred west of the quadrangle so that the rocks in the quadrangle are part of a thrust sheet, the prevailing steeply eastward-dipping fracture cleavage which displays the same movement sense as would be expected of the thrust, suggests that the Green Mountain anticlinorium was formed first. With the overturning of the Green Mountain folds, the eastward dipping fracture cleavage became well developed. During renewed or a later deformation, movement along certain of the fracture cleavage planes, which would be planes of weakness, might have resulted in thrusting, particularly in the western areas where the temperature of metamorphism was probably too low for the rocks to be mobile.

## SURFICIAL GEOLOGY

Inasmuch as Dr. D. P. Stewart is currently (1959–1960) studying the glacial geology of the Camels Hump quadrangle, this subject will be dealt with briefly. The distribution of glacial, lacustrine and fluvial deposits are shown on Plate 3 and are modified from those shown on a detailed and unpublished map of the surficial geology which was prepared by Stewart and kindly loaned to the authors. The areas of cover, principally till, were located approximately by the authors to indicate those areas in which bedrock could not be found. Till is much more widely distributed than is indicated and occurs as a thin cover over much of the area. Glacial striae were recorded by the authors wherever they could be observed on bedrock. Each represents an average strike direction at a particular locality, and the direction if ice movement is assumed to be away from the Champlain Valley, which presumably was the site of a major ice lobe.

The trend of the glacial striae indicates that the Winooski Valley acted as an east-west passageway for some of the ice movement through the Green Mountains. The striae along the Winooski River show a remarkable parallelism to the valley walls, whereas those elsewhere in the quadrangle show a northwest-southeast orientation (Pl. 3), which is also the prevailing direction in the Mount Mansfield quadrangle (Christman, 1959) and the general direction in northwest Vermont (Flint et. al., 1959). On the west side of the Green Mountains on the north side of the Winooski River, the change in strike of the striae suggests a diffraction of the bottom ice into the Winooski Valley and on the east side of mountains the diffraction is away from the valley. A similar change in direction is suggested by the striae to the north where the ice apparently followed the Lamoille River valley where the river transects the Green Mountains (Christman, 1959). As the east-west movement of ice along these valleys differs from the overall movement of ice, as recorded along the crest of the mountains and flatlands, these striae probably represent the direction of movement of an initial or final ice lobe moving up the valley when the tops of the mountains were clear of ice. The presence of such a lobe of ice in the early stages of glaciation can only be inferred, but the successive glacial deposits formed at various levels in the valley strongly suggests deposition in front of the retreating ice lobe in these valleys during the waning stages of glaciation.

The area in the vicinity of Waterbury Reservoir once was occupied by glacial Lake Mansfield (Bigelow, 1932) in which sandy varved clays were deposited. Some of these beds are well exposed around the margin of the reservoir and locally clay concretions of various shapes may be found. Nalewaik (1959) studied some of the varved sequence but neither exposures nor time were available to make a comprehensive study of the sequence of varves. A thick sequence of varved clays is reported to have been exposed during the excavations for the Waterbury Dam (Doll, personal communication).

Extensive glacial deposits are located in the northwestern part of the quadrangle, with well-developed kame topography occurring locally south of Jericho Center. Also, along the western margin of the area, lacustrine sands related to various stages of Lake Champlain are well developed. Other extensive glacial deposits occur to the south along the Huntington River Valley.

The straight-line topographic low which is occupied by Gillett Pond and extends northeast to the Winooski Valley and southwest to Huntington cannot be definitely explained. Because weaknesses in the form of faults, joints or easily eroded rock types are completely absent, and as the valley appears to be transverse to all geologic structures, it can only be assumed that the valley represents the position of an old superimposed stream which drained the Winooski River to the southwest at sometime during the complex glacial history when the mouth of the river was blocked by ice. The drainage is now through the Huntington River gorge and the flow is northward into the Winooski River.

Although Bolton Notch probably has been modified by glacial action, no evidence was found to indicate that it is primarily a glacial feature. The position of the valley was probably determined by structure and the presence of a graphitic zone which was easily susceptible to glacial and fluvial erosion.

# REFERENCES CITED


ALBEE, A. L., 1957, Bedrock geology of the Hyde Park quadrangle, Vermont: U. S. Geol. Survey, Geol. Quadr. Map, GQ 102.

BALK, R. A., 1926, A contribution to the structural relations of the granitic intrusives of Bethel, Barre and Woodbury: Vt. State Geologist 15th Rept., p. 39–96.

BIGELOW, E. L., 1932, The last lake of the Stowe Valleys: Vt. State Geologist 18th Rept., p. 359–362.

BILLINGS, M. P., 1928, The chemistry, optics and genesis of the hastingsite group of amphiboles: Am. Min. v. 13, p. 287–296.

——, 1937, Regional metamorphism of the Littleton-Moosilauke area, New Hampshire: Geol. Soc. Am. Bull., v. 48, p. 463–566.

BOOTH, V. H., 1950, Stratigraphy and structure of the Oak Hill succession in Vermont: Geol. Soc. Am. Bull., v. 61, P. 1131–1168.

BOYD, F. R., 1959, Hydrothermal investigations of amphiboles: Researches in Geochemistry, Abelson, editor; Wiley and Sons, P. 377–396.

BRACE, W. F., 1953, Geology of the Rutland area, Vermont: Vt. Geol. Survey Bull. No. 6, 124 p.

BUDDINGTON, A. F. and LEONARD, B. F., 1953, Chemical petrology and mineralogy of hornblendes in northwest Adirondack granitic rocks: Am. Min., v. 38, p. 891–902.

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

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

——, 1960, Stratigraphic and geotectonic relationships in northern Vermont and southern Quebec: Geol. Soc. Am. Bull., v. 71, p. 531–576.

CHRISTMAN, R. A., 1959, Geology of the Mount Mansfield quadrangle, Vermont: Vt. Geol. Survey Bull. No. 12, 75 p.

CLARK, T. H., 1934, Structure and stratigraphy of southern Quebec: Geol. Soc. Am. Bull., v. 45, p. 1–20.

DENNIS, J. G., 1956, Geology of the Lyndonville area, Vermont: Vt. Geol. Survey Bull. No. 8, 98 p.

ENGEL, C. C., 1959, Igneous rocks and constituent hornblendes of the Henry Mountains, Utah: Geol. Soc. Am. Bull., v. 70, p. 951–980.

ERIC, J. H. and DENNIS, J. G., 1958, Geology of the Concord-Waterford area, Vermont: Vt. Geol. Survey Bull. No. 11, 66 p.

FLINT, R. F. e. al., 1959, Glacial map of the United States east of the Rocky Mountains: Geol. Soc. of America publication.

GILLULY, J., 1935, Keratophyres of eastern Oregon and the spilite problem: Am. Jour. of Sci., 5 ser., v. 29, No. 171, p. 225–252 and 336–352.

HALLIMOND, A. F., 1943, on the graphic representation of the calciferous amphiboles: Am. Min., v. 28, p. 65–89.

HESS, H. H., 1938, Gravity anomalies and island arc structure with particular reference to the West Indies: Am. Phil. Soc. proceedings, v. 79, p. 71–90.

HITCHCOCK, C. H., 1884, Geological sections across New Hampshire and Vermont: American Mus. Nat. History Bull., v. 1. No. 5.

HITCHCOCK, EDWARD, 1861, Report on the geology of Vermont: vol. 2, 988 p.

JACOBS, E. C., 1938, The geology of the Green Mountains of northern Vermont: Vt. State Geologist, 21st Biennial Rept., p. 1–73.

KONIG, R. H., 1959, Geology of the Plainfield quadrangle, Vermont: Unpublished Ph.D. thesis, Cornell Univ. and Vt. Geol. Survey Bull., in press.

LYONS, J. B. et. al. 1957, Lead-alpha ages of some New Hampshire granites: Am. Jour. of Sci., v. 255, no. 8, p. 527–546.

MURTHY, V. R., 1957, Bed rock geology of the East Barre area, Vermont: Vt. Geol. Survey Bull. No. 10, 121 p.

NALEWAIK, G. G. 1959, Lacustrine deposits of the Waterbury Reservoir, Vermont: Unpublished senior thesis, Cornell University, 42 p.

NEVIN, C. M., 1949, Principles of structural geology: Wiley and Sons, 410 p.

OSBERG, P. H., 1952, The Green Mountain anticlinorium in the vicinity of Rochester and East Middlebury, Vermont: Vt. Geol. Survey Bull. No. 5, 127 p.

PERRY, E. L., 1928, Geology of Plymouth and Bridgewater, Vermont: Vt. State Geologist 16th Biennial Rept. p. 1–64.

SECOR, D. T., 1959, Joints in the Camels Hump area, Vermont: Unpublished Masters thesis, Geol. Dept., Cornell Univ. 35 p.

SUNDIUS, N., 1930, On the spilitic rocks: Geol. Mag. v. 67, p. 1–17.

——, 1946, The classification of the hornblendes and the solid solution relations in the amphibole group: Sveriges Geologiska Undersokn. Arsbok 40 (no. 4) p. 1–36.

TROGER, W. E., 1956, Optische bestimmung der gesteinbildenden minerale: Schweizerbart'sche Verlagsbuchhandlung, 147 p.

TURNER, F. J., 1958, Mineral assemblages of individual metamorphic facies: Geol. Soc. Am. Memoir 73, Chap. VII, p. 199–237.

TURNER, F. J. and VERHOOGEN, J., 1951, Igneous and metamorphic petrology: McGraw-Hill, 602 p.

WELLS, A. K., 1923, The problem of spilites: Geol. Mag. v. 60, p. 62–74.

WHITE, W. S., 1949, Cleavage in east-central Vermont: Am. Geophys. Union Trans., v. 30, p. 587–594.

WHITE, W. S., and JAHNS, R. H., 1950, Structure of central and east-central Vermont: Jour. Geology, v. 58, p. 179–220.

WILLIAMS, H., TURNER, F. J., and GILBERT, C. M., 1954, Petrography: W. H. Freeman and Co., 406 p.

WINCHELL, A. N., 1945, Variations in composition and properties of the calciferous amphiboles: Am. Min. v. 30, p. 27–50.

——, 1951, Elements of Optical Mineralogy: Wiley and Sons, 551 p.

WISEMAN, J. D. H., 1934, The central and south-west Highland epidiorites; a study in progressive metamorphism: Quat. Jour. of the Geol. Soc. of London, v. 90, part 3, p. 354–417.

# PLATE I


TOPOGRAPHY BY U. S. GEOLOGICAL SURVEY 1948

GEOLOGY BY ROBERT A. CHRISTMAN & DONALD T. SECOR, JR.

SCALE 1:62 500

CONTOUR INTERVAL 40 FEET
DATUM IS MEAN SEA LEVEL

APPROXIMATE MEAN DECLINATION, 1948

## BEDROCK GEOLOGY, CAMELS HUMP QUADRANGLE, VERMONT

## EXPLANATION

PERMIAN (?)

Pb

BASIC DIKE ROCKS

ORDOVICIAN

Osts

SERPENTINITE, TALC-CARBONATE ROCK, AND STEATITE

Os Osg

STOWE FORMATION

Sericite-quartz-chlorite schist (Os) with some interbedded calcareous greenstone (Osg)

CAMBRIAN

€o €og

OTTAUQUECHEE FORMATION

Graphitic phyllite and quartz-sericite schist (€o) with interbedded calcareous greenstone (€os) and quartzite (rows of dots)

CAMELS HUMP GROUP

€ph

PINNEY HOLLOW FORMATION

Quartz-sericite-chlorite schist

€hn

HAZENS NOTCH FORMATION

Graphitic schist, quartz-albite-muscovite schist and some quartzite

€u €ug

UNDERHILL FORMATION

Phyllite and some metagraywacke (€u) with minor calcareous and amphibolitic greenstone (€ug) in west; quartz-albite-mica schist and albite-muscovite schist in east (€u)

€p

PINNACLE FORMATION

Metagraywacke and some phyllite

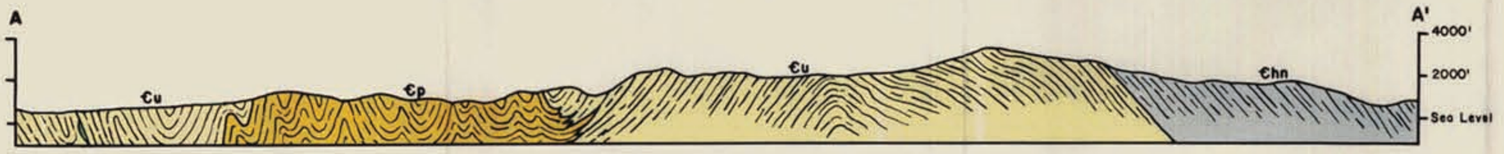
Contact, dashed where indefinite or gradational

Anticline | Syncline
Axis of major fold

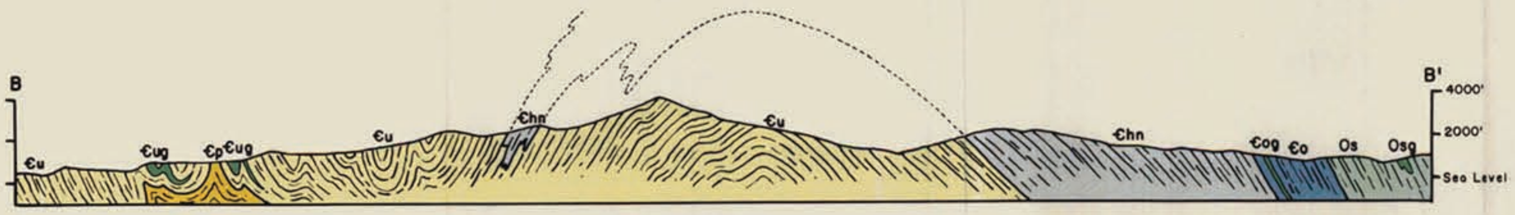
65 | 70
Inclined | Overturned
Strike and dip of graded bedding

55
Inclined | Vertical | Horizontal
Strike and dip of bedding schistosity

Garnet isograd



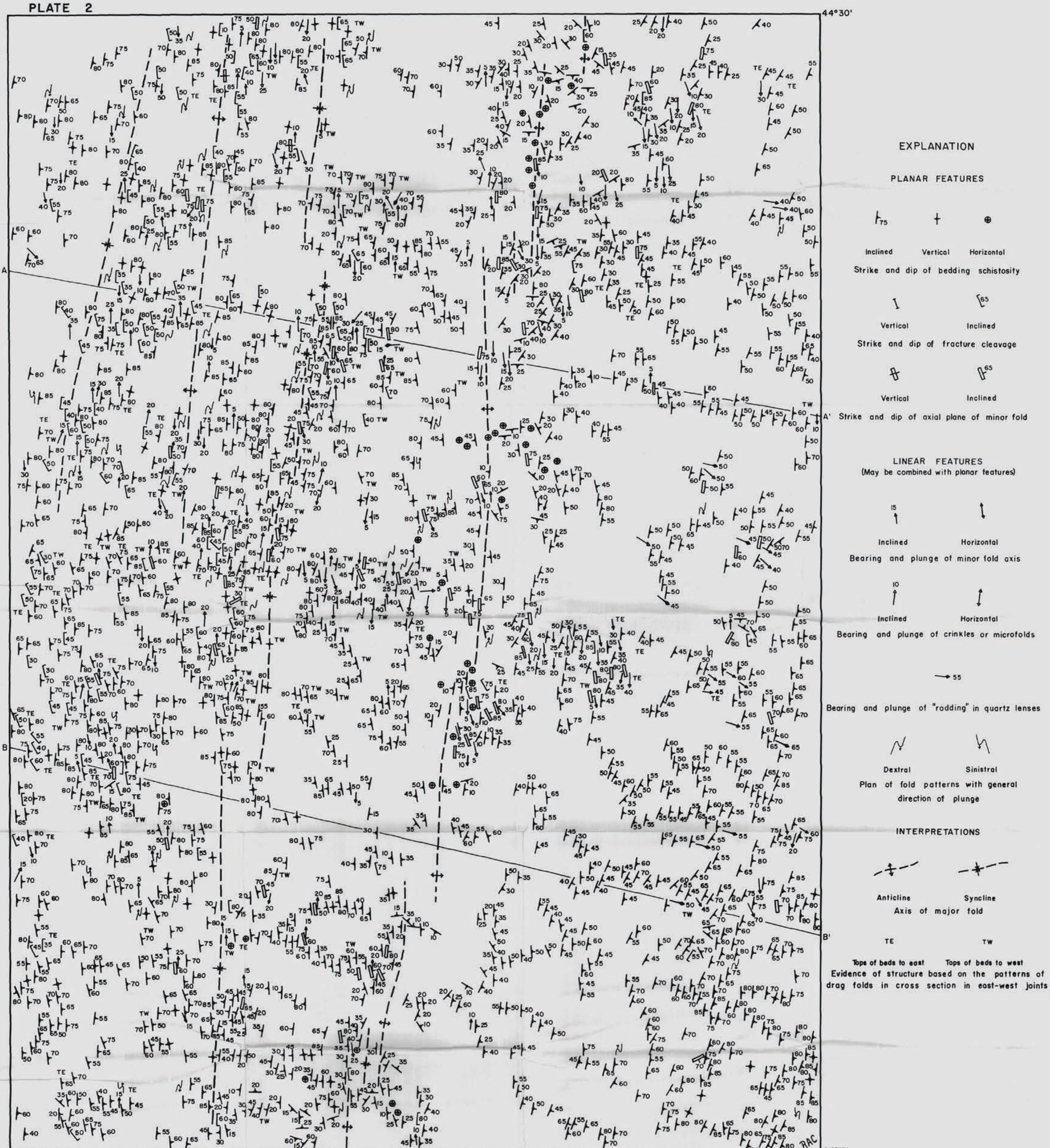

SECTION A-A'




SECTION B-B'

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 15)

PLATE 2

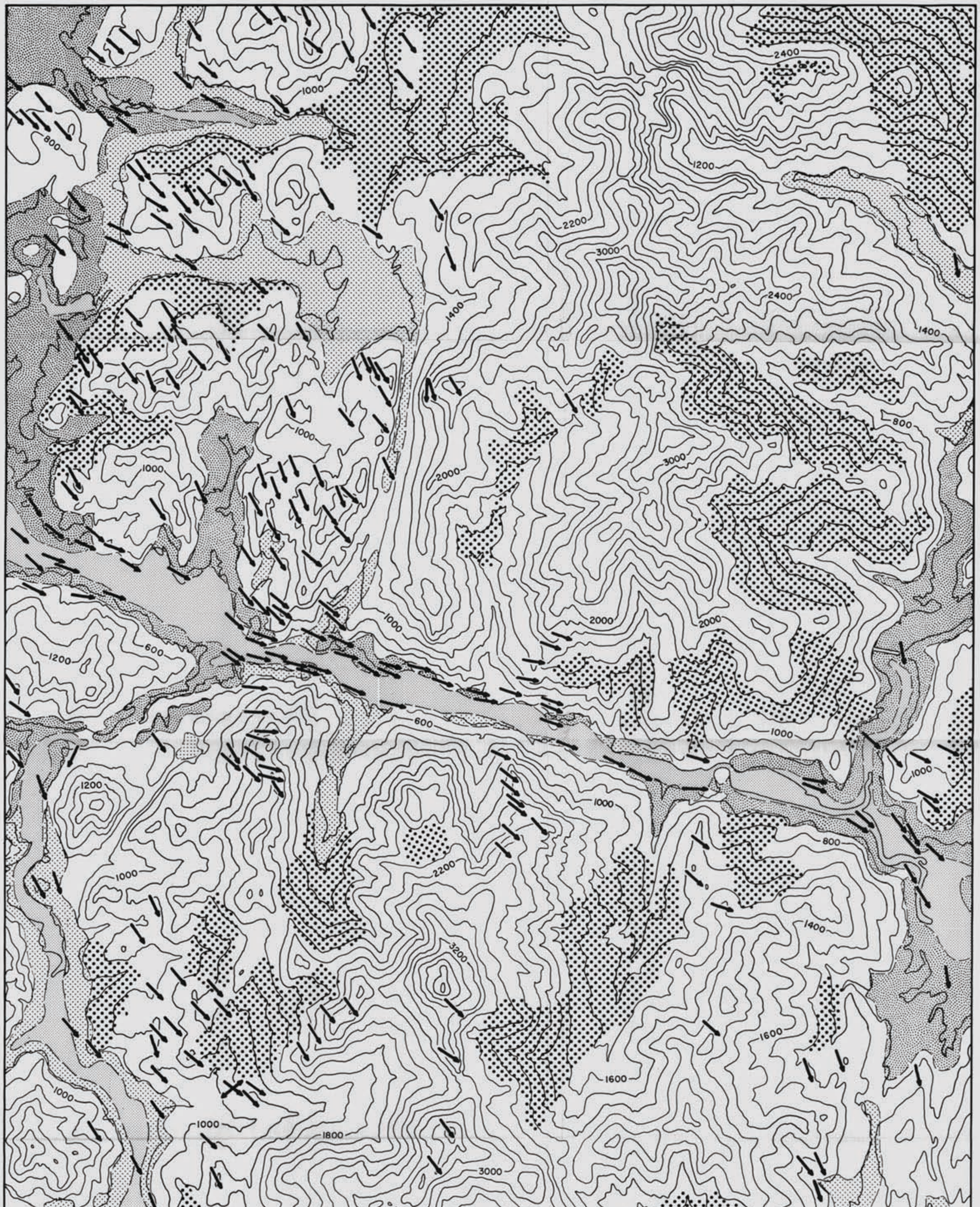



BY ROBERT A. CHRISTMAN AND DONALD T. SECOR, JR.

# STRUCTURAL MAP, CAMELS HUMP QUADRANGLE, VERMONT

0 1 2 3 4 MILES

PLATE 3




EXPLANATION

FLUVIAL DEPOSITS, GRAVEL AND ALLUVIUM

LACUSTRINE DEPOSITS, SILT, SAND AND MINOR GRAVEL

GLACIAL DEPOSITS, KAME MORAINE, KAME TERRACE, AND OUTWASH

OTHER AREAS OF EXTENSIVE COVER, PRINCIPALLY TILL. CONTACTS ARE ONLY APPROXIMATELY LOCATED

GLACIAL STRIAE

TOPOGRAPHY FROM U. S. GEOLOGICAL SURVEY MAPS

BY R. A. CHRISTMAN AND D. P. STEWART

## SURFICIAL GEOLOGY, CAMELS HUMP QUADRANGLE, VERMONT

0 1 2 3 4 MILES

CONTOUR INTERVAL IS 200 FEET