GEOLOGY OF THE MOUNT MANSFIELD QUADRANGLE VERMONT

By

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ABSTRACT

The rocks in the Mount Mansfield quadrangle were deposited in a Cambrian eugeosyncline principally as graywacke, siltstone and shale with minor amounts of volcanic, calcareous and carbonaceous material. Rocks in the western part are coarser and include local beds of quartz conglomerate, "shale conglomerate," and many coarse graywackes associated with graded beds. Eastward across the quadrangle the rocks become finer grained. Within this sequence older strata containing lava and volcanic detrital rocks belong to the Tibbet Hill unit and younger rocks containing carbonaceous material and limestone belong to the Ottauquechee formation. The remaining portions of the sequence belong to the Camels Hump group which could not be differentiated.

These rocks have been metamorphosed principally to metagraywacke and phyllite on the west where metamorphism was less intense and to quartz-albite-muscovite schist on the east where metamorphism was more intense along the Green Mountain anticlinorium. All the rocks belong to the greenschist facies, mostly to the chlorite and biotite zones but some rocks along the anticlinorium may be in the lower part of the almandine zone. Amphibolitic greenstone and feldspathic greenstone are two distinctive metamorphosed lavas in the Tibbet Hill unit. In the Ottauquechee formation the rocks are principally graphitic phyllite and schist with minor amounts of quartzite and recrystallized black limestone.

The ridge along the crest of Mount Mansfield closely coincides with the axis of the Green Mountain anticlinorium. Here, the average bedding schistosity dips gently to the south, corresponding with the plunge of the anticlinorium. The drag folds and fracture cleavage indicate that the anticlinorium is slightly overturned to the west. Farther to the west

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the rocks are isoclinally folded so that the Cambridge syncline is recognized principally by the Ottauquechee formation in its core and the Fletcher anticline is recognized by the Tibbet Hill volcanic rocks.

It is postulated that the area suffered one principal deformation and that the direction of the deforming force may have shifted from S.70°E. to east-west.

INTRODUCTION

General Statement

Despite the geographic prominence and easy accessibility of Mount Mansfield little detailed geologic work has been done in its vicinity. Probably the principal deterring factors have been the uniformity of rock types, the apparent complexity of structures of the metamorphosed rocks and the lack of economic deposits.

The present study was undertaken to fill in this void of geologic information and to obtain fundamental data on the Green Mountains in northwestern Vermont. A particularly interesting problem is the lack of correlation of the rocks on the west side of the Green Mountains with the well-established sequence of rocks on the east side.

Location of the Area

The Mount Mansfield quadrangle is located in northwestern Vermont (Fig. 1) between 44° 30′ and 44° 45′ north latitude and 72° 45′ and 73° 00′ west longitude. It includes parts of Chittenden, Lamoille and Franklin counties and portions of Bakersfield, Cambridge, Essex, Fairfax, Fairfield, Fletcher, Jericho, Johnson, Morristown, Stowe, Underhill, Waterville and Westford Townships (Pl. 1).

Regional Geologic Setting

The axis of the Green Mountain anticlinorium extends across the southeastern part of the quadrangle so that, except for the southeastern corner, the area lies on the western flanks of the Green Mountains. The Green Mountains are part of the Taconic allochthone in which the rocks are postulated to have reached their present position by thrusting from the east (Cady, 1945). The location of the roots of the thrusts is not known, although Cady's palinspastic map (1945, p. 568) indicates that the rocks may have originated in New Hampshire more than 100 miles to the east. Traces of the thrust planes beneath the allochthone have

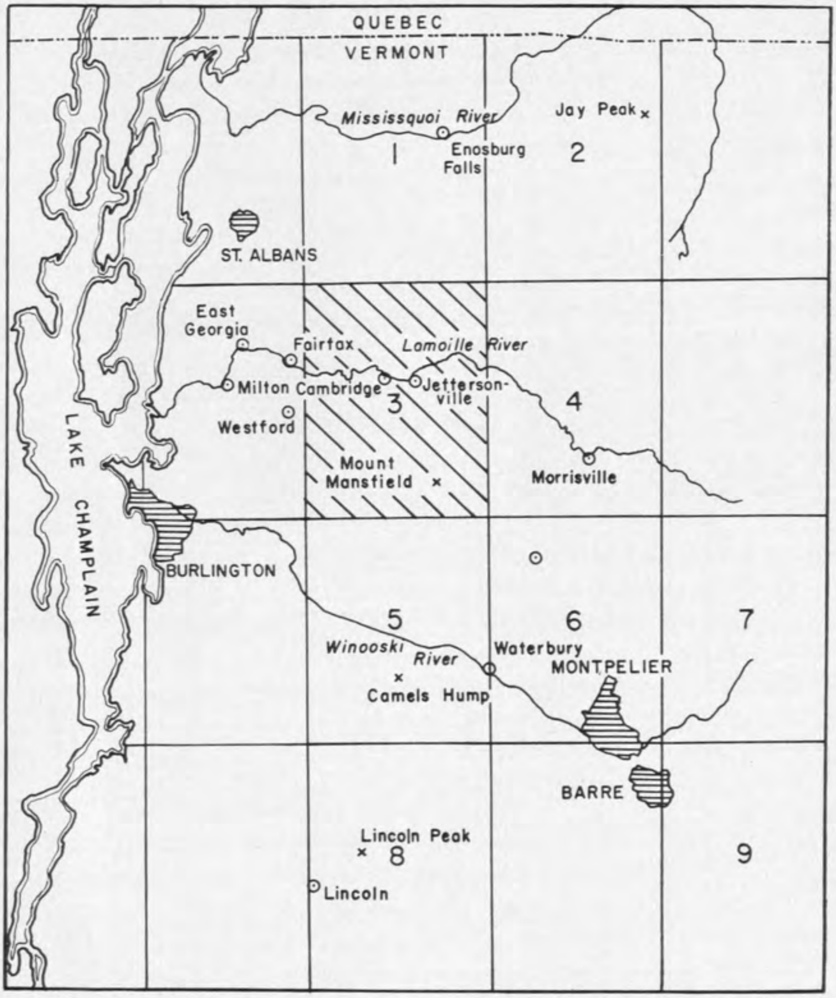

Figure 1. Index map of northwestern Vermont. Mount Mansfield quadrangle is shaded. The quadrangles are numbered as follows:

1. Enosburg Falls
2. Jay Peak
3. Mount Mansfield
4. Hyde Park
5. Camels Hump
6. Montpelier
7. Plainfield
8. Lincoln Mountain
9. East Barre

been mapped one to five miles west of the quadrangle (Stone, 1951). No faults have been recognized in the Mount Mansfield quadrangle and the rocks show no direct evidence of being part of a thrust sheet.

The Green Mountain anticlinorium has a Precambrian core in southern Vermont with the northernmost Precambrian gneisses plunging northward beneath younger rocks near Lincoln, Vermont, about 25 miles south of the Mount Mansfield quadrangle (Cady, personal communication, 1958). North of Lincoln, rocks exposed in the anticlinorium are considered to be Cambrian, although no fossils have been found. East of the Green Mountains the stratigraphic section includes rocks of Cambrian to Devonian age. West of the allochthone, the marbles and dolomites in the Champlain Valley are Cambrian-Ordovician.

Topography and Drainage

The most prominent topographic feature is Mount Mansfield whose summit profile, when viewed at a distance from the east or west, resembles a face. The highest peaks on the ridge of Mount Mansfield, from south to north, are The Forehead, The Nose, The Chin and Adams Apple. The northeastward continuation of the Green Mountains across the quadrangle includes Spruce Peak and the Sterling Range. As aptly described by the name, "Green Mountains," the slopes are heavily wooded except for a narrow zone along the crest and cliffs of Mount Mansfield, particularly at Smuggler's Notch. At the highest elevations, the vegetation consists of sparse, scrubby brush. An elongate "scar" on the west side of Mount Mansfield marks the position of a landslide which occurred in August 1955.

The principal drainage feature of the quadrangle is the Lamoille River which extends east-west across the quadrangle and more or less divides the quadrangle into a northern third and a southern two-thirds. In the southern two-thirds of the quadrangle, west of Mount Mansfield, a series of ridges and valleys trends north-south. The tops of these ridges are generally at 1000 to 1400 feet, except for Flynn Hill at 1860, and the elevations of the valleys range from 800 to about 1200 feet. These valleys are subsequent, although they owe their present form as much to glaciation as stream erosion.

The northern third of the quadrangle has a less well-developed north-south valley system and is more mountainous. Although these mountains display a general north-south trend, they are too equidimensional to be called ridges. The highest of these are Gilson and Fletcher Mountains at elevations of 1940 and 2150 feet, respectively. The drainage is less regularly developed and the presence of Halfmoon, Metcalf and Kings Hill ponds attests to the effects of glaciation, both in the scouring of basins and the deposition of drift.

The Lamoille River transects the north-south structural elements, cutting across the axis of the Green Mountain anticlinorium east of Jeffersonville near East Cambridge. The Lamoille River, as well as the Missisquoi River to the north and the Winooski River to the south which also flow east-west across the north-south mountains, is presumably superposed from an earlier erosion surface of which no other evidence persists (Jacobs, 1938). The dam at Fairfax Falls controls the temporary base level of the river at about 420 feet elevation. This is only slightly higher than the natural baselevel as the dam is built on bedrock at the falls. Downstream from the falls the river is at about 340 feet, the lowest elevation in the quadrangle. Except at Fairfax Falls, the Lamoille River meanders on a well-developed flood plain impinging alternately against opposite valley walls. A number of ox-bow lakes are present. One, south of Sand Hill, was reportedly formed in the flood of 1830 when the Lamoille River changed its course.

Previous Work

The earliest geologic studies of the northern Green Mountains were made by Edward Hitchcock (1861) and C. H. Hitchcock (1884). In subsequent years most of the geologic work in this area has dealt with the general history of glaciation and the structure and stratigraphy of the Champlain valley to the west.

More recently, descriptions of Mount Mansfield and surrounding areas were made by Jacobs (1938) in somewhat of a reconnaissance nature. The structural features were only briefly described and no geologic map accompanies the report. Chidester (1953) gives an excellent detailed geologic description of the talc deposits in a 60-acre area at Sterling Pond on Spruce Peak. A popularized account of the geology of Mount Mansfield State Forest was written by Christman (1956). An unpublished Master's thesis on the geology of the northwestern portion of the Mount Mansfield 7½′ quadrangle was presented to Cornell University by Konig (1956). The rocks in this area are very similar and many of the problems encountered could not be solved within the limits of the area and the six-week time limit. Senior theses were written at Cornell University under the author's direction by Watts (1957) on the garnets of Mount Mansfield and by Radinsky and Stifel (1958) on the abandoned talc deposits in the northeast corner of the quadrangle.

In recent years, considerable geologic work has been done in adjacent quadrangles. The Hyde Park quadrangle to the east and the Montpelier quadrangle to the southeast have been studied by Albee (1957) and Cady (1956), respectively, of the U. S. Geological Survey. As the west side of Albee's geologic map is shown as undifferentiated Camels Hump group, no formational contacts extend to the boundary between the Hyde Park and Mount Mansfield quadrangle. The Milton area to the west was mapped by Stone (1951) and contains the complex structural problems of the thrust sheets. As Stone designated the eastern portion of his area as "pre-Gilman" without differentiation between the older rocks along the boundary, no formation contacts extend into the Mount Mansfield quadrangle. The area to the north was partially mapped by Booth (1950) and is currently being completed by John Dennis of the Vermont Geological Survey. The Camels Hump quadrangle to the south presently is being studied by Christman and Donald Secor.

Present Study

The writer became familiar with the area during two months in the summer of 1955 when he was employed by the Vermont Geological Survey and the Department of Forests and Parks to prepare a popularized account of the geology of the Mount Mansfield State Park. The objective of this initial field work was to traverse all of the trails and to describe and explain the obvious geologic features of the area. Subsequent study of the quadrangle was made during the summers of 1956 and 1957 and two weeks in 1958.

All work was done on the 7½′ U. S. Geological Survey topographic maps published in 1948 at a scale of 1:24,000. In the wooded areas particularly along ridges or streams, an aneroid barometer was used to determine elevations for locating positions. Aerial photographs were used to check locations.

Acknowledgments

Foremost, I wish to acknowledge my gratitude to Dr. Charles G. Doll, State Geologist, who suggested the project, arranged the finances, and made valuable suggestions regarding the field geology. It has been a pleasure to have worked for the Vermont Geological Survey on this project.

During the months of August, students from Cornell University served as field assistants in return for the experience and the opportunity to do field work on their senior theses. These include Robert Watts in 1956, Leonard Radinsky and Peter Stifel in 1957 and Gerald Nalewaik, employed by the Vermont Geological Survey in 1958. Their stimulating interest and companionship were most appreciated.

I was very fortunate in having a number of geologists visit me in the field. Particular thanks are due to Dr. Wallace Cady and Dr. Alfred Chidester of the U. S. Geological Survey who freely passed on information from their studies and were most generous with their time. Their assistance in familiarizing the author with some of the regional problems is most gratefully acknowledged. Dr. E. H. Muller of Cornell University spent a day looking at the glacial features. Mr. Ronald Konig joined me for field conferences a number of times, as did John Dennis, working in Enosburg Falls quadrangle and Donald Secor, working in the Camels Hump quadrangle.

All of the Vermont residents, whom we met, were most kind and considerate and made our summers in Vermont very enjoyable. Particularly, I wish to thank Charlie Lord of the Mount Mansfield Company who was helpful in many ways.

Finally, I wish to thank Dr. Muller and Dr. C. M. Nevin who have read and critically reviewed portions of this manuscript.

METAMORPHOSED SEDIMENTARY AND VOLCANIC ROCKS

General Statement

Most of the rocks in the Mount Mansfield quadrangle originated in a eugeosyncline of probable Cambrian age. They originally were coarse- to fine-grained graywacke, siltstone, and shale with minor amounts of volcanic material which have been metamorphosed to metagraywacke, schist, phyllite and greenstone. In the western part of the area the effect of metamorphism is slight; the massive graywacke retains most of its original minerals with only a minor development of mica or chlorite and the fine-grained rocks have been metamorphosed to phyllite. Eastward, across the quadrangle, metamorphism increases; its effect is accentuated by decrease in average grain-size of the rocks in this direction. The rocks along the crest of the Green Mountains on the east side of the quadrangle are predominantly quartz-albite-muscovite schist with garnet occurring only in some of the rocks on Mount Mansfield.

Inasmuch as deposits of eugeosynclines are characterized by thick sections of similar-appearing rocks, it is not surprising that mappable rock units were difficult to establish. Only three mapping units have been recognized in the quadrangle; these are the Tibbet Hill schist, the Camels Hump group and the Ottauquechee formation.

Tibbet Hill Schist

General Statement

The type locality of the Tibbet Hill schist is on Tibbet Hill in Quebec (Clark 1934) where the rock is described as being a greenstone of Precambrian age. Booth (1950) mapped the Tibbet Hill schist in the Enosburg Falls quadrangle and described it as being a greenstone, sometimes gray in color, often amygdaloidal and locally containing knots of epidote. Although he considers that most of the rocks originally were lava flows, he recognized some detrital beds as probably originating as tuffs. A very early Cambrian or late Precambrian age was assigned to the Tibbet Hill schist. Booth mapped the southernmost occurrence about three miles south of Enosburg Falls, but stated that the rocks reappear to the south in anticlinal inliers. Dennis (personal communication, 1958) has extended the limits of the Tibbet Hill schist to the edge of the Mount Mansfield quadrangle in mapping an area of greenstone in the Enosburg Falls quadrangle just north of Metcalf Pond.

Both Clark and Booth refer to the unit as the Tibbet Hill schist without formally calling it a formation. Because of the manner in which it has been mapped by Booth and the author, for all practical purposes, it may be considered a formation.

In the Mount Mansfield quadrangle, the Tibbet Hill schist occurs in the core of the Fletcher anticline (Pl. 1). The greenstones in the unit were mapped in detail because of their distinctive lithology. The most distinctive and continuous layers are amphibolitic greenstone and feldspathic greenstone. Also in the Tibbet Hill unit are calcareous greenstones which grade into green calcareous metagraywacke and into gray or brown metagraywacke. Because the latter are indistinguishable from the rocks in the overlying Camels Hump group, the upper contact of the Tibbet Hill schist was placed at the top of the uppermost amphibolitic or feldspathic greenstone. Very little calcareous greenstone was found above these units, so this position for the contact seems fairly good. No lower contact is exposed because the schist and greenstone occur only as the oldest rocks in the core of the anticline.

Amphibolitic Greenstone

The amphibolitic greenstone is so-named because it contains porphyroblasts of an amphibole, tentatively identified as actinolite, as long as one-third inch (Pl. 4). The porphyroblasts are dark green, irregular in shape and somewhat flattened to give the rock a weak foliation but no lineation. Upon weathering, the porphyroblasts may become brownish green in color and may stand out slightly in relief to give the rock a characteristic irregularly mottled surface. Because the foliation is only weakly developed, the rocks may occur as massive rounded outcrops or as massive ledges. Locally knots of epidote as large as one by three feet occur within the greenstone. Except for the actinolite (?) and epidote concentrations, the other minerals are too fine grained to be identified megascopically, although the color and sheen of the rock suggests that chlorite and sericite are present in some specimens.

Although it is possible that the amphibolitic greenstones may have formed as sills, it is more likely that they were lava flows. Where observed, they are conformable with the overlying and underlying detrital beds. In places the amphibolitic greenstone appears to grade into calcareous greenstone which, in turn, grades into metagraywacke. This sequence is interpreted as being a lava flow overlain by tuff of similar composition with succeeding beds becoming less tuffaceous and containing more detrital material. To the north Booth (1950) and Dennis (personal communication, 1958) have identified primary structural features indicative of lava flows. The long narrow outcrop pattern (Pl. 1) and the occurrence of the greenstones at about the same horizon over a distance of eighteen miles also suggests that they represent lava flows. The average thickness of these greenstones is only about fifty feet. Outcrops of this rock may be examined most easily southwest of Metcalf Pond (Loc. 1 and 2, Fig. 2)[1] and west of Binghamville (Loc. 10, Fig. 2).

From microscopic study of the amphibolitic schist, the essential minerals were identified as actinolite (?), chlorite, epidote, and albite. Estimations of the percentages of each of these, as well as other petrographic data, are given for rocks from 12 localities in Table 1.

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1. This notation gives the locality number or numbers which are located on Figure 2.

Plate 4. Amphibolitic greenstone with actinolite (?) porphyroblasts, 0.6 miles S.W. of Metcalf Pond (Loc. 1, Fig. 2). Photograph is parallel to the foliation surface except for block at base which shows the foliation surface.

The amphibole is tentatively identified as actinolite on the basis of a spectrographic analysis (Table 2), even though some of its optical properties vary slightly from normal actinolite, and it may contain too much Al_2O_3. Using the principal constituents reported in the analysis and assuming that the balance of the material is SiO_2, the molecular proportions were determined on the basis of 8 parts SiO_2. The result of 1 CaO (plus 1/10 Na_2O and 1/10 K_2O), 3 MgO, 1 Fe_2O_3 and 8 SiO_2 compares quite well, assuming some of the Al_2O_3 has replaced SiO_2, with the standard formula of $2\ CaO \cdot 3\ MgO \cdot Fe_2O_3 \cdot 8\ SiO_2\ (OH)_2$, considering that the analysis is only semi-quantitative. The mineral is negative, length-slow, and may be slightly pleochroic with X = colorless, Y = light green, and Z = light greenish blue. The beta index of 1.642 fits actinolite but the extinction angle of 24–28° is too large by about 10° and the 2V of 75° is too small by about 5° (Winchell, 1951). The optical properties

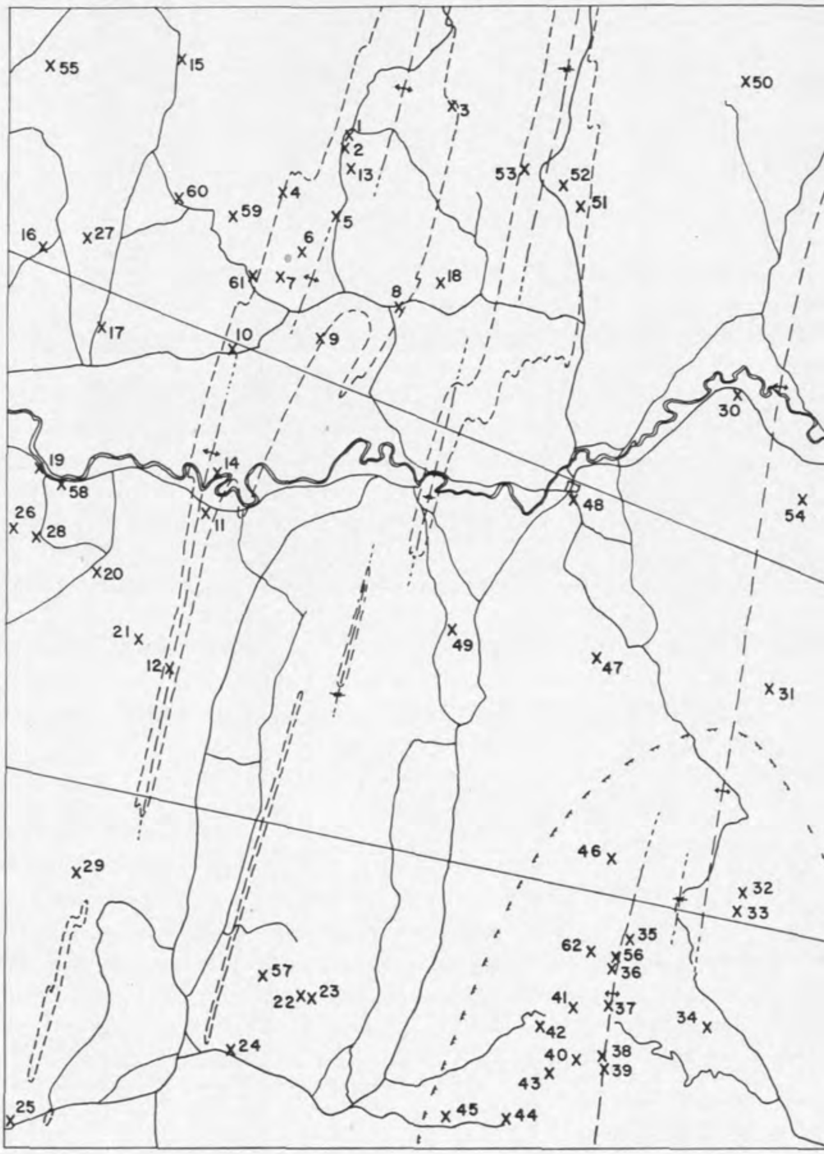

Figure 2. Locality map, Mount Mansfield quadrangle, showing location of rock specimens and structural features described in the text. Contacts, folds, roads, etc. from Plate 1.

most closely fit Eckermannite, a soda-hornblende, but the spectrographic analysis does not show enough sodium. Because the presence of a soda amphibole would indicate that the lavas were spilitic, a chemical analysis of a similar amphibole from the Camels Hump quadrangle will be made.

The actinolite (?) which makes up 10–40 percent of the rock usually occurs as large porphyroblasts with irregular edges and inclusions. It is usually partly altered to chlorite. The chlorite which makes up 10–30 percent of the rock is characterized by having anomalous brown or purple interference colors. All the sections have 5–10 percent epidote and one has as much as 35 percent. The mineral is pleochroic green to yellow-green, has a negative sign, a large 2V with strong dispersion, generally $r < v$. The groundmass of the amphibolitic greenstones consists of 10–50 percent small anhedral grains of albite, mostly untwinned, which are difficult to distinguish from quartz. Quartz was identified only in one specimen and here it may be secondary.

Other minerals occurring in the amphibolitic greenstones are sphene (with leucoxene), yellow-orange biotite, sericite (or muscovite), calcite, apatite, pyrite and magnetite. In some sections sericite comprises as much as 30 percent of the rock and in many rocks sphene and leucoxene account for 5 percent of the rock (Table 1).

Absence of quartz in the amphibolitic greenstones suggests that the original lavas were basic. The mineral suite of actinolite, albite, epidote and chlorite is characteristic of the greenschist facies resulting from low-grade metamorphism of basic volcanics.

Feldspathic Greenstone

The feldspathic greenstone is so-named because it contains lath-like, remnant phenocrysts of plagioclase feldspar as much as one and one-fourth inch long (Pl. 5). These phenocrysts are seen most easily on weathered surfaces as they weather to white to light green in contrast to the fine-grained dark green groundmass. The phenocrysts have no apparent preferred orientation.

The feldspathic greenstone has limited occurrence in the quadrangle. In the northern part it was found only as a discontinuous bed south-southwest of Metcalf Pond (Loc. 13, Fig. 2). North of the Lamoille River (Loc. 14, Fig. 2) just north of the road, about 200 feet of feldspathic greenstone is exposed beneath the glacial cover and south of the river other exposures occur west of Beaver Brook (Loc. 11, Fig. 2). From this latter point discontinuous outcrops occur for five miles to the south with

TABLE 1

ESTIMATED MODES AND PETROGRAPHIC DATA FOR AMPHIBOLITIC GREENSTONES

Locality*	Amphibole	Chlorite	Epidote	Albite	Other minerals and petrographic data
1	35%	20%	10%	30%	Sphene and leucoxene 4%; mica 1%; amphibole ext. 26°; chlorite is anomal. brown.
2	40%	10%	5%	15%	Muscovite 25%; sphene and leucoxene 5%; amphibole ext. 24° 2V= 75°; chlorite is anomal. brown.
3	5%	10%	35%	40%	Calcite 5%; sphene and leucoxene 2%, pyrite and magnetite; amphibole ext. 20°.
4	15%	20%	10%	40%	Sericite 10%; sphene and leucoxene 5%; amphibole ext. 24°; chlorite is anomal. purple.
5	30%	30%	5%	30%	Sphene and leucoxene 2%; amphibole ext. 17°; chlorite anomal. purple; very fine-grained rock.
6	20%	30%	10%	30%	Sphene and leucoxene 10%; amphibole ext. 27°; chlorite is anomal. purple.
7	25%	30%	10%	25%	Biotite 5%, sphene and leucoxene 5%; chlorite is anomal. purple.
8	10%	20%	10%	50%	Biotite 5%; sphene and leucoxne 5%; amphibole ext. 20°; chlorite is anomal. brown.
9	25%	20%	20%	25%	Sericite 5%; biotite 2%; sphene and leucoxene 2%; chlorite is anomal. purple.
10	15%	30%	10%	10%	Sericite 30%; sphene and leucoxene 5%; amphibole ext. 20°; chlorite is anomal. brown.
11	35%	20%	15%	25%	Quartz 5%; calcite 1%; sphene and leucoxene 1%; amphibole ext. 18°; chlorite is anomal. purple.
12	10%	20%	10%	50%	Sphene and leucoxene 10%; biotite 2%; amphibole ext. 28°; chlorite is anomal. blue.

*For the location of these localities see Figure 2.

TABLE 2

DATA ON AMPHIBOLE FROM AMPHIBOLITIC GREENSTONE IN THE TIBBET HILL SCHIST (LOC. 2, FIG. 2) SOUTHWEST OF METCALF POND, MOUNT MANSFIELD QUADRANGLE, VERMONT

Semi-quantitative spectrographic analysis,* reported as oxides of elements listed.				Extinction
Fe	15 %	Zr	0.05 %	$Z \wedge C = 24°$
Mg	12	V	.03	
Al	12	Cr	.02	2V measured with
Ca	5	Ba	.02	Universal Stage
K	1	Co	.01	
Na	0.9	Ni	.008	$2V = 75°$
Ti	.75	Sr	.004	
Mn	.20	Cu	.001	Index of refraction
		Ga	.001	
Si	Not determined			$Y = 1.642$

*Analysis by American Spectrographic Laboratories, Calif.

the southernmost occurrence northwest of McLean Hill. In all places, the greenstone is conformable with the surrounding rocks. In general, the unit is less than fifty feet thick.

Microscopically, the feldspar laths are mostly altered to a mixture of quartz, calcite and sericite so that very little of the original grains remain. In a few places, however, patches of unaltered material show albite twinning with an extinction angle of 18° and an index of refraction higher than balsam. This indicates that the original grains must have been at least as calcic as andesine. The groundmass of the rock is a fine-grained mixture of albite, quartz, epidote, anomalous blue chlorite and yellow-orange biotite, sericite, sphene and leucoxene.

No other rocks in the quadrangle were found with lath-like crystals of plagioclase of intermediate composition. The feldspar in the graywacke is clearly detrital and the plagioclase in the metamorphosed rocks is equidimensional and is albite. Plagioclase of intermediate composition is not formed by low-grade metamorphism (Turner, 1958). Thus, the altered plagioclase phenocrysts in the feldspathic greenstone may be taken as evidence that the rock was of igneous origin and probably represents a lava flow which has been partially metamorphosed.

CALCAREOUS GREENSTONE, PHYLLITE AND METAGRAYWACKE

The less distinctive calcareous greenstone, phyllite, and metagray-

Plate 5. Feldspathic greenstone with weathered plagioclase phenocrysts (Loc. 11, Fig. 2).

wacke comprise the bulk of the Tibbet Hill in the Mount Mansfield quadrangle. As these three rock types are all gradational with each other, they probably represent detrital sediments with the greenstone having a higher content of tuffaceous material.

The calcareous greenstone is characteristically green to dark green with white streaks of calcareous material. The rock generally is so fine-grained that only calcite, or ankerite, and occasional octahedra of magnetite can be megascopically identified with certainty. However, the sheen of the rock may indicate that sericite and chlorite are present and the characteristic yellow-green color of some specimens indicates the presence of epidote. The rock weathers to various shades of green-brown and the streaks of calcareous material weather to leave a brown pitted surface or a series of holes, parallel to the bedding. The pitted appearance of the weathered greenstone is very characteristic of the rock. Microscopically, the rock essentially is composed of a fine-grained mixture of chlorite, epidote, calcite, albite and quartz in various proportions.

As the amount of detrital quartz increases, the calcareous greenstones grade into metagraywackes and as the amount of clay increases, they grade into phyllites or schists. In some of these rocks quartz occurs as rounded light-blue grains of coarse sand size which probably represents original detrital material. The average composition of four metagraywackes from the Tibbet Hill was 35 percent quartz, 20 percent plagioclase, 20 percent anomalous blue chlorite, 10 percent epidote and minor amounts of calcite, sphene with leucoxene, muscovite, zircon, tourmaline and magnetite.

Distribution and Age

In the Mount Mansfield quadrangle the outcrop of the Tibbet Hill unit becomes very narrow to the south. This may be partly due to the southward plunge of the Fletcher anticline (Pl. 1). However, an additional factor is that a decrease in the amount of volcanic material probably results in a facies change to the south so that the map unit is more limited in area. East and west of the anticline in the east-central part of the area discontinuous bands of calcareous greenstone were mapped as part of the Camels Hump group because they occur without associated volcanic rocks within a thick sequence of phyllite and metagraywacke. As they have the same physical appearance and mineral composition as the Tibbet Hill calcareous greenstone, they may represent tongues or remnants of the Tibbet Hill unit. On the other hand, tuffaceous sediments of slightly different ages may look exactly alike after metamorphism, so that such a correlation cannot be made positively.

A few small isolated patches of greenstone occur on the east side of the Cambridge syncline (Pl. 1). Most of these are calcareous greenstone, but at two localities, north of Jeffersonville and west of Smiley school, beds of amphibolitic greenstone occur similar to those in the Tibbet Hill unit.

In preparing the cross sections, it became apparent that if the Tibbet Hill schist and greenstone were the oldest rocks in the quadrangle and the Ottauquechee were the youngest, the stratigraphic thickness of the intervening Camels Hump group, as measured between the Fletcher anticline and the Cambridge syncline was not sufficient to prevent the Tibbet Hill rocks from being exposed in the core of the Green Mountain anticlinorium. As rocks of volcanic origin were not identified on the anticlinorium, it must be postulated that either the Camels Hump group becomes much thicker to the east so that the Tibbet Hill rocks occur at depth or that the Tibbet Hill unit is lost by a facies change to the east.

Although both of these explanations are possible, two lines of evidence suggest that the Tibbet Hill unit has undergone a facies change and is no longer present. The position of the isolated outcrops of greenstone on the east side of the Cambridge syncline is about the same distance east of the syncline as the Tibbet Hill unit is west of the syncline. Thus, it is suggested that these greenstones are tongues of the Tibbet Hill exposed on the east side of the syncline and that they represent all that remains of the Tibbet Hill volcanics. The second line of argument, even more tenuous, is that, if the Green Mountain anticlinorium is to have a core of resistant Precambrian rock as postulated later in this report, there may not be room for the Tibbet Hill rocks to occur at depth.

The Tibbet Hill unit, therefore, is considered to be a series of volcanic rocks with associated detrital sediments. The map unit is lost to the south and to the east because of a facies change involving a decrease in volcanic material and an increase in detrital material. The Tibbet Hill rocks are considered to be Cambrian and represent the oldest rocks in the western part of the quadrangle but in the eastern part the rocks exposed in the Green Mountain anticlinorium are probably slightly older. A logical corollary of this interpretation makes the Tibbet Hill rocks a unit within the Camels Hump group. Because the evidence in the Mount Mansfield quadrangle is only suggestive, this correlation cannot be made formally until more is known about the rock units to the north.

Camels Hump Group

General Statement

The Camels Hump group was named by Cady (1956) for the strata which lie above the Precambrian and below the Ottauquechee formation, as typified by the schist, quartzite and gneiss occurring along the Winooski river west of Waterbury, Vermont, in the Camels Hump quadrangle. When traced to the south these rocks are equivalent to the Pinney Hollow schist of Perry (1928) and to the Granville and Monastery formations, as mapped by Osberg (1952). When traced to the north these are partly equivalent to the Pinnacle and West Sutton formation as mapped by Booth (1950). Neither Cady (1956) in the Montpelier quadrangle nor Albee (1957) in the Hyde Park quadrangle divided the Camels Hump group into formations, although Albee mapped three special rock types of limited occurrence. Cady (personal communication, 1958) has tentatively subdivided the group into formations on the basis of unpublished detailed work in the Jay Peak and Lincoln Mountain quadrangles. Although the author is familiar with Cady's proposed formations, he was not able to use them as mapping units in the Mount Mansfield quadrangle.

In general, the rocks of the Camels Hump group in the Mount Mansfield quadrangle grade from metagraywacke and phyllite on the west to schist, quartzite and gneiss along the anticlinorium in the east. This gradation represents an eastward increase of metamorphism and a decrease in grain size of the original rocks. Within the quadrangle the following rock types can be identified: shiny green quartz-chlorite-mica schist similar to the Pinney Hollow; quartz-chlorite-mica schist with porphyroblasts of albite as found in parts of the Monastery formation; graphitic schist with associated quartzites which partly defines one of Cady's proposed formations; brown to gray, fine- to coarse-grained graywacke similar to those in the Pinnacle formation; and fine-grained phyllitic rocks similar to those in the West Sutton formation. Although these different rock types may be identified, none could be mapped as formations because of their gradational nature and apparent lack of continuity. A study of nearly eighty thin sections from the Camels Hump group failed to improve on the generalizations given at the beginning of this paragraph.

In order to facilitate the description of the rocks in the Camels Hump group, the less metamorphosed rocks in the western half of the quadrangle will be described separately from the more metamorphosed rocks in the eastern half.

Rocks in the Western Area

The less metamorphosed rocks of the Camels Hump group in the western part of the quadrangle are similar to those mapped as the Pinnacle and West Sutton formations by Booth (1950) in the Enosburg Falls quadrangle to the north. Booth describes the Pinnacle as typically greenish-gray to light gray graywacke, often containing beds of sandy slates, black and green slates, quartzitic schist and various types of conglomerates. He also states (1950, p. 1141) that south of the Missisquoi River the extreme variation of texture and composition "makes the designation by a single rock type misleading; some parts are more arkosic than graywackish." Although the West Sutton formation is supposed to contain finer grained rocks, Booth writes (1950, p. 1147), "The West Sutton south of the Missisquoi River is so similar to the

Pinnacle that the two can be differentiated only because of the intervening White Brook" dolomite.

In the Mount Mansfield quadrangle, the differentiation between the Pinnacle and West Sutton formations did not seem possible. The White Brook dolomite is not present, although according to its trend on Booth's map it might pass through the northwest corner of the area. In this vicinity, a few local lenses of impure marble were mapped (Loc. 55, Fig. 2), but no beds of the characteristic brown-weathering dolomite were found.

The rocks in the quadrangle undoubtedly are part of the Pinnacle formation. The only exception to Booth's description is that the fine-grained members are phyllites and schists rather than slates. The only reason why the author did not map the Pinnacle formation as a member of the Camels Hump group is that no contact could be found between it and the schists to the east. Although the schist and Pinnacle-type rocks are distinctively different, the change across the quadrangle is gradational. The recognition of a given point in this gradational series is difficult because within a given area the variation in grain size of the rocks causes considerable variation in rock types. Thus, if a contact had been established it would be completely arbitrary and almost impossible to map.

In the western part of the quadrangle it is estimated that about one-half the rocks are medium- to fine-grained metagraywacke. These grade in color from light green to dark brown and grade in grain size from rocks which are metaconglomeratic to phyllitic. Generally, the only minerals which may be recognized megascopically with certainty are the original detrital grains of quartz and feldspar and rarely grains of pyrite. The quartz grains are usually glassy and rounded and many are light blue in color. The sheen and the color of the fine-grained constituents often suggest the presence of sericite, chlorite or biotite. Many of the metagraywackes form massive outcrops in which bedding is absent or obscure.

At a number of localities, graded bedding may be found. One mile south-southwest of Fairfax Falls (Loc. 26, Fig. 2), coarser layers which are light colored and the fine-grained phyllitic layers which weather black give the exposures of graded beds a conspicuous banded appearance (Pl. 6). At least nine pairs of fine and coarse layers, averaging three feet in thickness, indicate that the beds are parallel to the regional structure and are right side up with the tops of the beds to the east. The maximum size of the grains in the coarser layers is about 2 mm. and many of the

Plate 6. Graded bedding, southwest of Fairfax Falls (Loc. 26, Fig. 2). The light bands are metagraywacke and the dark bands are fine-grained, iron-stained phyllite. The beds are at N.15°E. 70°SE., right-side up with the tops of the beds to the east. Photograph is to the south.

grains are weathered feldspar. East of Huntville (Loc. 27, Fig. 2) graded bedding with poorly developed cross bedding (Pl. 7) occurs in a cross-cutting relation to the regional structure. Here, its relation to the fracture cleavage indicates that the rocks are on the west limb of a northerly plunging anticline, which is presumed to be a local fold as it could not be traced.

Locally, conglomeratic rocks occur within the sequence of metagraywacke and phyllite. Some of these are simply coarse-grained metagraywacke in which the size of some of the quartz grains may be about one-half inch in diameter and rarely as much as two inches in diameter. Beds containing this coarse material rarely may be traced indicating that they are lens-shaped. Northeast of Underhill much of the metagraywacke sequence on top of Hedgehog Hill is a coarse metaconglomer-

Plate 7. Detail of graded bedding from near Huntville (Loc. 27, Fig. 2). Photograph is to the SE showing a fracture cleavage surface at N.35 °E. 85 °SE. The bedding is at N.80 °E. 40 °NW. and cross bedding may be seen in the lower left corner.

ate. The distribution of the coarser grained metagraywacke and metaconglomerate is indicated on the geologic map.

In addition to the quartzose metaconglomerates, metaconglomerates composed principally of fragments of phyllite locally occur with the metagraywackes. These may be called "shale conglomerates" or "intraformational conglomerates" and probably represent penecontemporaneous erosion and deposition of the strata in which they occur (Pettijohn, 1948). The phyllite fragments presumably represent slightly metamorphosed shaly fragments which have been flattened and elongated by metamorphism (Pl. 8). This elongation and flattening parallel to the fracture cleavage may be so well-developed that the folded nature of the rock would be undetected in areas of poor outcrop (Fig. 9).

Locally, calcareous beds occur in the metagraywacke sequence. These include the impure limestones in the northwest corner of the area and

Plate 8. "Shale conglomerate" in metagraywacke sequence of the Camels Hump group (Loc. 28, Fig. 2). Note 50-cent coin for scale.

the calcareous greenstones in the east central part of the area. As already mentioned, the latter may be related to the Tibbet Hill greenstones. In addition, many of the metagraywackes may contain a few percent of calcite which may weather away to give the rock a brown color and a porous appearance. Such calcareous layers are frequently found in association with the shale conglomerates.

The mineral composition of eleven representative metagraywackes from the western part of the quadrangle is given in Table 3; coarse-grained metaconglomerates and fine-grained phyllites are omitted. These mineral percentages, as all those given in this report, are visual estimates supported in a few rocks by point count determinations. Because the mineral content of the rock varies from locality to locality, or even from specimen to specimen, accurately determined modes by the point count method did not seem practical. The values given are probably accurate within five percent. The quartz content ranges from 25 to 75 percent with

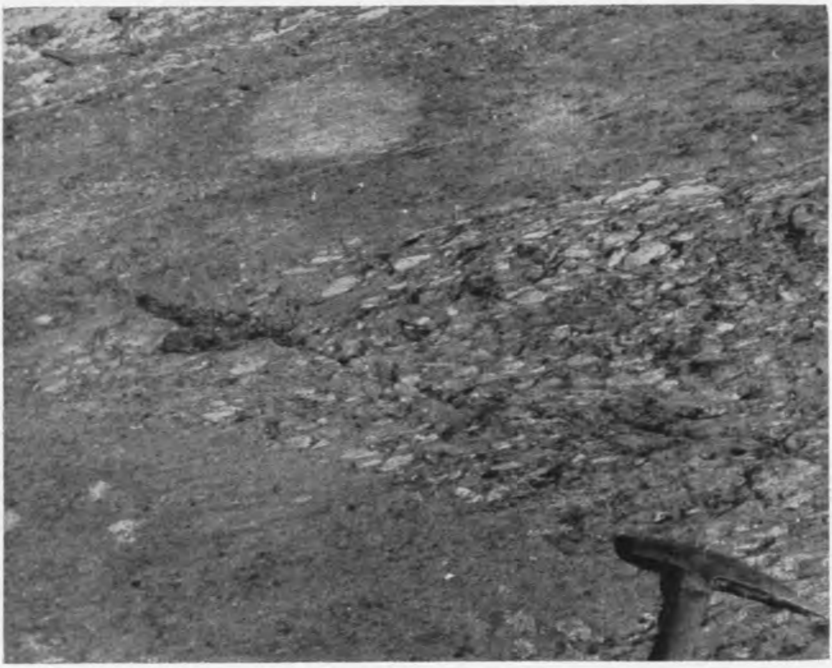

Plate 9. Small fold in bed of "shale conglomerate." The phyllite fragments are flattened and elongated parallel to the fracture cleavage at N.10°E. 75°SE. and do not bend with the fold (Loc. 20, Fig. 2).

an average of about 50 percent and the plagioclase content ranges from 5 to 30 percent with an average of about 20 percent. Most rocks contain varying amounts of sericite, biotite and chlorite. Among the accessory minerals, many contain the characteristic detrital grains of tourmaline and zircon, as well as sphene, epidote, calcite and magnetite, probably titaniferous in all rocks. No potash feldspars were positively identified.

The presence of sericite, biotite and/or chlorite indicates that the rocks have been slightly metamorphosed, as these minerals represent the recrystallization of the clay minerals in the original graywacke. These minerals occur as small oriented crystals interstitial to the larger original detrital grains of quartz and plagioclase. Where the rocks have been more highly metamorphosed the amount of interstitial quartz and albite is greater and the size of the detrital grains is smaller until the two types are indistinguishable. Thus, the process of recrystallization involves dissolving the original grains and crystallization of new crystals. Where detrital grains can be recognized, many have irregular and uneven boundaries as if the grains were being eaten away. Detrital plagioclase usually has albite twinning and its composition may be more calcic than albite, whereas recrystallized plagioclase has no twinning and has the composition of albite. Detrital plagioclase, as recognized by twinning, was observed in all the rocks listed in Table 3, except those from localities 23 and 24 which are farther east in areas of greater metamorphism than the other rocks.

Most of the rocks in Table 3 are arkosic graywackes. The plagioclase content in a few is almost sufficient to warrant calling them arkoses.

In this report the term graywacke is used according to Krynine (1948) and refers simply to the mineral composition without reference to genesis or stratigraphic occurrence, although the rocks do fulfill the more rigorous requirements of Pettijohn's definition of graywacke (A.G.I. glossary). Plotted on Krynine's triangular diagram with the three corners occupied by quartz, feldspar and mica and/or chlorite, most of the rocks in Table 3 fall in the field of "high rank graywackes" and several fall in the field of "low rank graywackes." Because most of the rocks in the Mount Mansfield quadrangle show some evidence of metamorphism, even though very slight, the term "metagraywacke" has been used in this report.

At a few localities in the quadrangle a bright green mica occurs in some of the schists. At Jack Lott hill (Loc. 29, Fig. 2) it occurs in the silica-rich layers of a calcareous greenstone. The mineral is negative with a 2V of about 30° and a Y index of 1.618. Except for the green color these properties fit the mineral Phengite (Troger, 1956, p. 83); however, Rosenbusch (1927) gives this mineral as pleochroic green and occurring in gneisses. An X-ray study of the mineral gave very strong lines at 2.58 and 1.506 and less strong lines at 2.54, 1.77, 1.64, 1.30 and 1.26. These values do not agree with described micas but are fairly close to those of glauconite. Although the mineral does not physically resemble glauconite, composition of phengite and glauconite may be nearly the same, so that one might speculate on what happens to glauconite when it is slightly metamorphosed.

Rocks in the Eastern Area

The more metamorphosed rocks in the Camels Hump group in the eastern part of the area are similar to those mapped by Albee (1957) and

TABLE 3

ESTIMATED MODES AND PETROGRAPHIC DATA FOR SOME METAGRAYWACKES FROM THE WESTERN PART OF THE MOUNT MANSFIELD QUADRANGLE.

Locality	Quartz	Plagioclase	Calcite	Sericite*	Biotite	Chlorite	Other minerals and petrographic data
15	50%	20%	10%	5%	—	10%	Tourmaline, sphene and leucoxene, zircon, and magnetite.
16	45%	30%	—	10%	15%	—	Tourmaline, sphene, zircon and magnetite.
17	25%	5%	—	55%	—	10%	Magnetite 5%; zircon and epidote; Fine-grained rock.
18	45%	20%	—	20%	10%	5%	Epidote, sphene and zircon.
19	40%	20%	2%	30%	5%	Tr	Sphene 3%; zircon and titaniferous magnetite.
20	55%	5%	5%	20%	Tr	10%	Ti. magnetite 3%; pyrite, zircon, tourmaline; chlorite anomal. blue.
21	50%	35%	Tr	5%	—	10%	Tourmaline, sphene and apatite; chlorite is anomal. purple.
22	70%	15%	—	5%	10%	Tr	Tourmaline, zircon and magnetite.
23	75%	10%	—	5%	5%	Tr	Graphite 5%; sphene, zircon, and magnetite.
24	60%	20%	—	15%	—	5%	Tourmaline, zircon and magnetite; chlorite is anomal. blue.
25	50%	30%	Tr	5%	5%	10%	Zircon, sphene, apatite and ti. magnetite; chlorite anomal. blue.

*Muscovite is included with the sericite.

Cady (1956) in adjoining quadrangles. In the Hyde Park quadrangle, Albee describes the rocks adjacent to the Mount Mansfield quadrangle as being "predominantly quartz-sericite-chlorite schist with porphyroblastic albite" with increasing amounts of interbedded graphitic schist and quartzite in the south. In the Montpelier quadrangle, Cady describes the group as consisting predominantly of light-gray, quartz-sericite schist with prominent zones of "rusty weathering carbonaceous quartz-sericite-albite schist" in the area nearest the Mount Mansfield quadrangle.

Along the Green Mountain anticlinorium in the quadrangle, the rocks typically are quartz-albite-muscovite schist. These minerals are present in varying amounts (Table 4) and generally may be identified megascopically. Locally chlorite or biotite may be abundant and porphyroblasts of euhedral to subhedral magnetite and/or garnet may be conspicuous. Delicate needles of black tourmaline were observed at a few localities. In general, the schist is light to dark gray with variations to silver-gray or greenish gray depending upon the amounts of mica or chlorite present. The platy minerals are well aligned to give the rock a well-developed schistosity. Small lenses of milky quartz, which probably were derived locally by a sweating-out process during metamorphism, occur throughout the schist. These are parallel to the schistosity and serve to emphasize the complexity of the small-scale folding. Locally larger concentrations of quartz occur in fractured areas; these pods may contain small amounts of a bright green chlorite.

Estimated modes of 20 schists (Table 4) indicate that quartz, albite, muscovite and chlorite are the most abundant minerals, in that order. Although the highest quartz value is given as 65 percent and the average is 30 percent, some of the sericitic quartzites, not reported here, contained as much as 95 percent quartz. The albite content ranged from 0 to 65 percent. In all cases the albite occurs as anhedral to subhedral porphyroblasts which easily may be confused with quartz inasmuch as albite twinning is absent. Many of the porphyroblasts contain numerous inclusions, giving them a sieve texture, and in some the orientation of fine-grained inclusions show s-curves, indicating that the albite porphyroblasts were rotated during their formation. The muscovite, including some sericite and perhaps unidentified paragonite, ranges from a trace to as much as 70 percent of the schist. Chlorite, usually anomalous blue or purple with crossed nicols, is likewise variable and may comprise as much as 25 percent of the rock. Locally biotite, epidote, magnetite, or graphite may be present in quantities as high as 10 percent. Garnet, sphene, calcite, tourmaline, apatite and zircon occur as accessory minerals in many of the schists. In a few the garnet may be zoned and may show snowball structures. Kyanite, sillimanite and staurolite are not present. Likewise, staining for potash feldspar (Konig, 1956) failed to indicate the presence of orthoclase or microcline.

Chloritoid was identified in thin section from three localities on the

TABLE 4

ESTIMATED MODES AND PETROGRAPHIC DATA FOR SCHISTS FROM MOUNT MANSFIELD.

Locality	Quartz	Albite	Muscovite	Chlorite	Other minerals and petrographic data.
30	50%	25%	15%	Tr	Calcite 5%; biotite and graphite; much graphite in the albitic porphyroblasts.
31	40%	35%	10%	2%	Biotite 10%; epidote 3%; garnet, sphene; chlorite is anomal. purple.
32	30%	30%	25%	5%	Graphite (dust) 10%; tourmaline; highly contorted schist.
33	65%	5%	15%	10%	Calcite 10%; zircon, garnet, tourmaline, apatite; chlorite is anomal. blue.
34	15%	20%	40%	25%	Tourmaline, apatite, garnet, magnetite; chlorite is anomal. blue.
35	40%	—	40%	20%	Titaniferous magnetite, tourmaline, apatite; chlorite is anomal. purple.
36	30%	20%	25%	25%	Zoned garnet, titaniferous magnetite; chlorite is anomal. purple.
37	20%	15%	35%	25%	Zoned garnet 5% with snowball structure; zircon, magnetite; chlorite is anomal. purple.
38	30%	20%	40%	Tr	Magnetite 10%; epidote, garnet, tourmaline.
39	65%	—	15%	10%	Epidote 10%; sphene, apatite, garnet, magnetite.
40	20%	50%	10%	—	Biotite 10%; epidote 5%; magnetite 5%; sphene; porphyroblastic albite with sieve structure.
41	40%	10%	25%	20%	Chloritoid 5%; epidote, magnetite, apatite, tourmaline.
42	45%	30%	5%	15%	Epidote 5%; garnet, zircon, apatite, ti. magnetite, chlorite is anomal. purple; sieve albite.
43	20%	65%	Tr	Tr	Biotite 10%; epidote 5%; apatite, zircon, garnet, chlorite is anomal. purple; sieve albite.
44	Tr	30%	50%	15%	Magnetite 5%; epidote, zircon, tourmaline; chlorite is anomal. blue.
45*	15%	50%	5%	Tr	Biotite 20%; magnetite 3%; epidote 3%; garnet; chlorite is anomal. purple.
46*	15%	15%	60%	5%	Biotite, zircon, magnetite.
47*	25%	25%	35%	10%	Biotite 5%; sphene, apatite, magnetite; chlorite is anomal. brown.
48	5%	10%	70%	10%	Tourmaline, epidote, magnetite; chlorite is anomal. purple.
49*	20%	5%	50%	20%	Magnetite 2%; tourmaline, apatite, sphene, pyrite.

*Estimates from these localities partly based on modes by Konig (1956).

Table 5

DATA ON GARNETS FROM MOUNT MANSFIELD

	Garnet from The Forehead	Garnet from Taft Lodge	Garnet from Sterling Pond
Index of refraction:	1.808	1.808	1.808
Spectographic analyses in terms of oxides:			
Si	40.00 %	40.00 %	
Mg	1.	.5	
Al	30.	25.	
Ca	3.5	5.	
Ti	.75	.15	
V	.003	.008	
Cr	.01	.01	
Mn	20.	10.	
Fe	12.	25.	
Co	.004	—	
Zr	.01	.03	
X-ray analyses, position of strong lines:	2.84		
	2.54	2.54	2.55
	1.69		1.60
	1.54	1.53	1.54
	1.265	1.258	1.26
	1.076	1.071	1.076
Size of unit cell (cubic a) determined by indexing the X-ray lines:	11.65 A	11.60 A	11.67 A

west side of Mount Mansfield. Two came from the Half-way House Trail, located approximately as Loc. 41, and the other from the cliffs near C. P. Smith's, located approximately as Loc. 44 (Fig. 2). These sections were loaned to the author by E. C. Jacobs, former State Geologist of Vermont. Inasmuch as chloritoid does not occur in other sections, it is assumed that its limited occurrence is a function of the original composition of the rocks. It is characteristic of mica schists deficient in potash and high in iron (Williams, Turner and Gilbert, 1954).

A study was made of garnets from several localities on Mount Mansfield to determine whether the mineral is almandite or spessartite. Although the results (Table 5) are not conclusive, the data seem to indicate that the mineral is mostly spessartite. Watts (1957) determined that the garnets from six localities on Mount Mansfield had an index of about 1.808, and from calculations based on the spectrographic analyses he concluded that the garnet from The Forehead was 57 percent spessartite, 35 percent almandite and 8 percent grossularite and that the garnet from Taft Lodge was 26 percent spessartite, 60 percent almandite, 11 percent andradite and 3 percent grossularite. As the garnet samples were not completely pure and the accuracy of the analyses is uncertain these calculated compositions are only approximate; later X-ray work does not completely support these results. The size of the unit cell of three garnets was determined to be 11.65, 11.60 and 11.67 which most closely agrees with the size of spessartite. Donnay and Nowacki (1954) give almandite as 11.52 and 11.54 and spessartite as 11.61 and 11.63. The presence of the grossularite molecule would tend to increase the cell size.

On the east side of the anticlinorium, particularly in the vicinity of Sterling Pond, The Pinnacle and the lower ski slopes of Mount Mansfield, many of the schists are graphitic and thin bedded quartzites are common locally. Some of the quartzites may form beds as much as two feet thick. This rock sequence is undoubtedly related to those mapped by Albee to the east and Cady to the southwest. In addition to the graphite and quartzite, many of the quartz-albite schists are characterized by gray to black porphyroblastic albite. The dark color of such porphyroblasts from a locality to the north (Loc. 30, Fig. 2) could be shown to be due to minute graphite inclusions.

On the west side of the anticlinorium, particularly on the ridge west of The Forehead, rocks containing large amounts of white porphyroblastic albite occur in massive cliffs (Loc. 40 and 43. Fig. 2). Because of the paucity of platy minerals these rocks have poorly developed schistosity so that many may be considered gneisses rather than schists. Microscopically, the albite has a sieve structure and is clearly formed as a result of metamorphism.

Farther to the west on the east side of Macomber Mountain, a zone of graphitic schist and phyllite occurs (Pl. 1). This graphitic zone may be traced northward, through a series of discontinuous outcrops, to the northeast corner of the quadrangle where graphitic schist and phyllite are relatively abundant east of Kings Hill Pond. At one place (Loc. 50, Fig. 2) a lense of white marble occurs within the schist. Other calcareous rocks have been found along this trend to the south. South of Jeffersonville near the covered bridge over the Brewster River, these graphitic schists are associated with gray quartzites and recrystallized brown-weathering siliceous dolomites.

This zone of graphitic rocks probably correlates with the more extensive areas of similar rock in the Camels Hump group in the Hyde Park quadrangle. Inasmuch as the section of graphitic rocks and quartzites is much thicker on the east side of the anticlinorium than on the west, a facies change must occur. On the cross section through Mount Mansfield (Pl. 1) the attitude of the beds is such that the graphitic beds at Sterling Pond and Madonna Peak should reappear on the top of Mount Mansfield or its western slopes and the thick section of graphitic rocks in the Hyde Park quadrangle should reappear in abundance on the west side of Mount Mansfield. Inasmuch as these graphitic schists and quartzites do not occur here, it must be assumed that this facies of the Camels Hump group must die out to the west so that the zone of graphitic rocks extending from Macomber Mountain to Kings Hill Pond is all that remains of it on the west side of the anticlinorium.

The isolated outcrops of calcareous and amphibolitic greenstone within the Camels Hump group have already been mentioned regarding their possible correlation with the Tibbet Hill.

Rocks in the Intervening Area

Between the metagraywacke and phyllite in the western area and the schist of the eastern area is an intervening area in which the gradations between these rock types occur. The schists are finer grained and garnet and tourmaline porphyroblasts are not present. The metagraywacke and phyllite are more metamorphosed so that original quartz or feldspar grains are not visible, megascopically. Chlorite, biotite and sericite are common minerals and locally magnetite may be abundant. An unexplained feature in some of these rocks is the occurrence of mica plates which have grown across the schistosity. Although this must represent a late-stage recrystallization, it is not known whether this has a regional significance.

Source and Age of Sediments

The distribution of the sediments in the Mount Mansfield quadrangle strongly suggests that the original source of the rocks was to the west. The metagraywackes and metaconglomerates are more abundant on the west side of the area and the fine-grained rocks are more abundant on the east side. If the graphitic schists and phyllites of the Camels Hump group originally were black shales and siltstones, the thicker sections of these rocks to the east in the Hyde Park quadrangle suggests deeper water sedimentation, in agreement with a proposed western source. Booth (1950, p. 1156) postulates that the source of the Pinnacle and West Sutton formations was "to the west or northwest of the original area of deposition and now lies beneath westerly overlapping higher strata and perhaps one or more of the thrust slices of northwestern Vermont and southeastern Quebec." The arkosic nature of the metagraywackes suggest a nearby source to the west.

This simple picture is destroyed by the presence of the Cambrian-Ordovician limestones and dolomites to the west in the Champlain Valley, so that the relations between these two sedimentary sequences remain one of the major problems in the geology of northwestern Vermont.

The age of the Camels Hump group is Cambrian (Cady, 1956). According to Cady, the group "rests with pronounced angular unconformity on interbedded gneiss, quartzite, and greenstone of pre-Cambrian age in the Lincoln and Ripton townships, 22 miles southwest of Montpelier."

Ottauquechee Formation

General Statement

The type locality of the Ottauquechee formation is in the Ottauquechee River Valley near Bridgewater, Vermont, where it was described by Perry (1928) as consisting of slate-grey quartzite and dark grey or black phyllite. The formation has been traced northward on the east side of the Green Mountain anticlinorium from the Rochester-East Middlebury area (Osberg, 1952) through the Hyde Park quadrangle into Canada (Albee 1957). Albee mapped the formation as occurring on the limbs of the Foot Brook syncline, in the northwest part of the Hyde Park quadrangle, only five miles east of its occurrence in the Cambridge syncline of this report. Albee describes the formation as containing black graphitic quartz-sericite phyllite and schist and containing massive beds of dark-grey quartzite.

Description of Rocks

In the Mount Mansfield quadrangle, the formation consists of the characteristic graphitic schist and phyllite and associated quartzite as well as quartz-sericite schist similar to those in the underlying Camels Hump group. A particularly well-exposed section of the graphitic schist and massive quartzite occurs in a small stream southeast of East Fletcher (Pl. 1). Thin sections from this exposure (Loc. 51, Fig. 2) contain mostly small grains of quartz with about 5–10 percent each of graphite, muscovite, pleochroic orange-brown biotite and pyrite. Extensive areas of graphitic rocks also occur to the south, east of North Cambridge (Pl. 1).

Of particular interest in the quadrangle is the occurrence of thin-bedded graphitic limestone, called "black limestone" but which actually is a low-grade graphitic marble, within a sequence of quartzite, recrystallized dolomite and graphitic phyllite and schist. East of East Fletcher (Pl. 1) these limestones comprise much of the rock in a section about 100 feet thick. This horizon may be traced to the north by its occurrence in two small streams near the northern boundary of the quadrangle. West of East Fletcher additional exposures of the black limestone occur as isolated lenses or as local concentrations in the noses of steeply plunging drag folds. One of the larger bodies of black limestone which was mapped in detail (Fig. 3) appears to have flowed into the crest of a fold as it forms an east-west outcrop pattern in an area in which most of the rocks strike about north-south.

Although the graphitic limestones look as if they should contain fossils, none were found. No evidence of fossil remains was observed in thin sections. Samples broken down with acetic acid from ten different localities, including some from the Camels Hump group, and studied with the binocular microscope contained no recognized microfossils. A thin section of the black limestone east of East Fletcher (Loc. 52, Fig. 2) revealed that the rock was completely recrystallized, containing well-twinned calcite, much graphitic dust and minor amounts of quartz, so it is unlikely that fossils could have survived.

To the south, outcrops of black limestone and associated graphitic phyllite and schist occur extensively east of North Cambridge (Pl. 1) and locally at a few isolated outcrops farther to the south. The southernmost locality where black limestone could be located was west of Buker Hill. The Ottauquechee was mapped farther to the south to near Underhill on the basis of a zone, about 100 feet thick, which contains minor amounts of graphitic phyllite and schist.

Correlation and Age

The rocks in the Cambridge syncline might have been placed in the

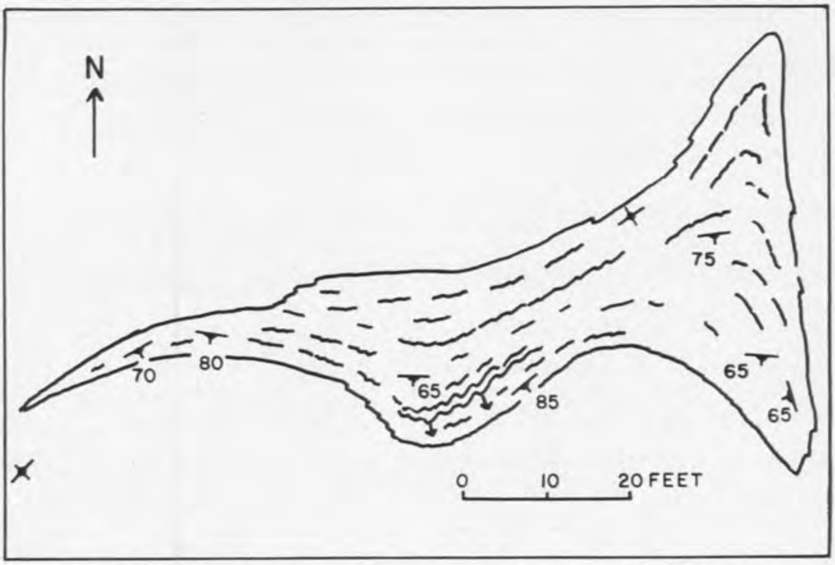

Figure 3. Outcrop pattern of graphitic limestone of Ottauquechee formation, west of East Fletcher (Loc. 53, Fig. 2). The recrystallized limestone occurs in the nose of a fold, plunging south, just west of the axis of the Cambridge syncline.

Sweetsburg formation which occurs to the north in Canada as justifiably as in the Ottauquechee formation. The Sweetsburg is composed of graphitic phyllites and schist but is characterized by the presence of thin-bedded graphitic limestones. As a result of recent work in Quebec, Osberg (personal communication, 1959) has been able to correlate the Sweetsburg formation with the Ottauquechee and Dennis (personal communication, 1958) in the Enosburg Falls quadrangle is tentatively placing the graphitic limestones in the Cambridge syncline in the Sweetsburg formation.

As either formation name appears correct, the author has used the Ottauquechee because of proximity of this formation to the east and because the geologic studies are still in progress to the north so that the continuity of the Sweetsburg from the Mount Mansfield quadrangle to the type locality is based more on hearsay than published information. In addition, as shown in Plate 1, in the Mount Mansfield quadrangle, the quantity of black limestone which is characteristic of the Sweetsburg formation, is quite limited.

No fossils have been found in the Ottauquechee formation. It generally has been called Cambrian (Osberg, 1952, Cady, 1956, Albee, 1957). The contacts, as mapped in the quadrangle, are very generalized and are based on the occurrence of graphite.

INTRUSIVE IGNEOUS ROCKS

The occurrence of intrusive igneous rocks in the quadrangle is limited to three small areas of ultramafic bodies, now altered to serpentinite, talc-carbonate rock and steatite, and to younger basic dikes observed at two localities. The nearest exposed granitic intrusions occur some 15 to 20 miles to the east in the Hardwick and Plainfield quadrangles and at the westernmost edge of the Montpelier quadrangle.

Serpentinite, Talc-carbonate Rock and Steatite

The largest bodies of altered ultramafic rock occur in the vicinity of Sterling Pond (Pl. 1). Because of the economic interest in the talc, a 60-acre area was studied in detail by Chidester (1953). The area was not restudied by the author. Chidester writes that these bodies "formed chiefly by the alteration of ultramafic igneous rocks that were extensively serpentinized prior to steatitization." The age of the bodies is given as Ordovician and the "steatitization probably took place at a late stage in the folding of the Green Mountain anticlinorium." Chidester concludes that the folded lenticular bodies of the talc deposits "are large and of good quality."

About one and one-half miles to the west of Sterling Pond, Jacobs (1938) reports an occurrence of talc on the east slope of Mount Mansfield near Smugglers Notch (along the Hell Brook trail) but its exact position could not be found.

Eight miles north of Sterling Pond, the Rousseau Talc Prospect is located on the south side of the Lamoille River (Pl. 1). A description of this deposit with the data of a diamond-drilling program are given by Chidester, Stewart and Morris (1952). The deposit is a lenticular body composed entirely of talc-carbonate rock and steatite which is presumed to have originated from an ultramafic intrusion. The deposit is about 700 feet long and the talc-carbonate rock is of good quality. This is the southernmost deposit on the west side of the anticlinorium; to the south all talc deposits are on the east side.

In the northeast corner of the quadrangle, another small talc deposit occurs about three miles north of Waterville (Pl. 1). Several prospect

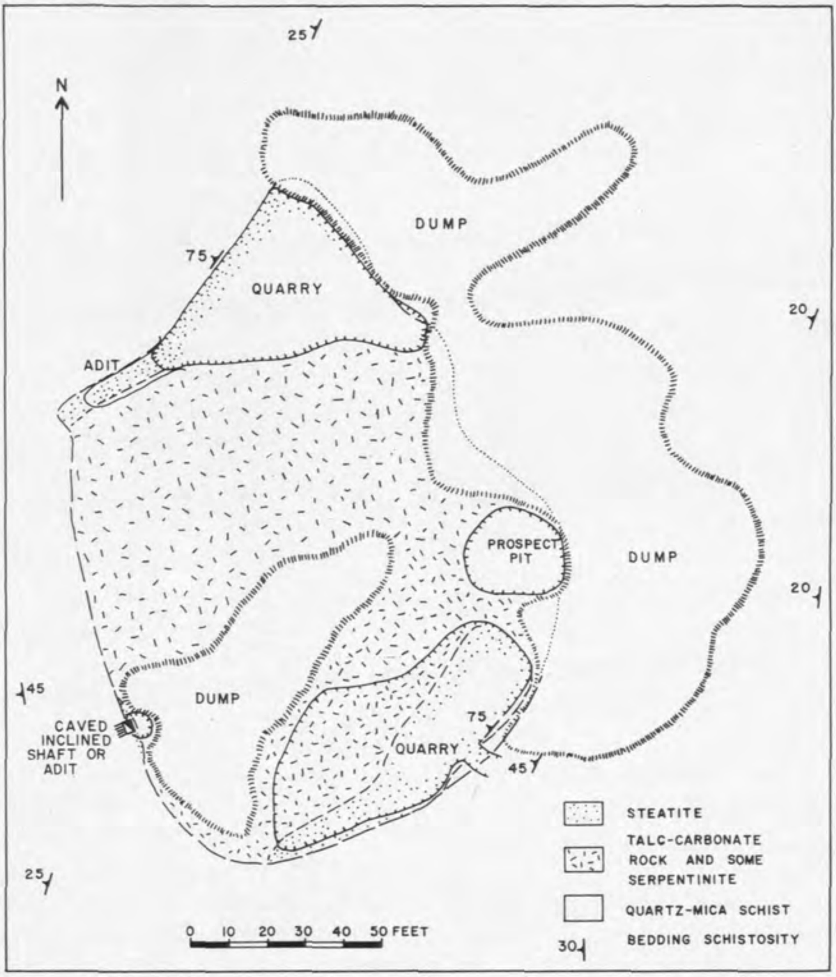

Figure 4. Geologic map of the Waterville Talc deposit, generalized from a detailed map by Radinsky and Stifel (1958).

pits indicate that the original ultramafic body was roughly circular with a diameter of about 110 feet (Fig. 4). This deposit was studied by Radinsky and Stifel (1958), and Figure 4, from their unpublished report, shows the general geology of the area. This deposit is essentially similar to the others, except for the presence of masses of light green actinolite needles in some of the talc.

Basic Dikes

Two basic dikes were observed in the Mount Mansfield quadrangle. A small dark gray, aphanitic dike occurs north of Jericho (Pl. 1) along a joint with an attitude of N. 85° W. vertical. The dike is one foot wide and cannot be traced more than 100 feet. It is clearly later than the metamorphism of the graywacke; the dike contains inclusions of the metagraywacke and small apophyses of the dike extend into the surrounding metagraywacke. Another dike locality occurs on the east side of the Green Mountain anticlinorium in a small stream south of the Lamoille River (Pl. 1). This dike, or dikes, occurs along a joint trending N. 35° E. and is exposed discontinuously over a distance of 3000 feet along the stream. Its maximum thickness is 5½ feet. It is a dark gray aphanitic rock with some specimens having an unusual spotted appearance due to the botryoidal arrangement of the dark fine-grained minerals. The rock is composed essentially of small crystals of sub-calcic augite, brown hornblende and magnetite in a fine-grained altered groundmass of sericite, calcite and feldspar (?) (Loc. 54, Pl. 2).

Other post-metamorphic dikes in this part of Vermont have been called diabase approaching camptonite (Cady 1956) and lamprophyre (Albee 1957). Similar dikes are present locally throughout Vermont and they generally are assigned a Mississippian age.

METAMORPHISM

Metamorphism of the rocks in the Mount Mansfield quadrangle is assumed to be due to regional metamorphism because the zone of more intense metamorphism coincides with the axis of the Green Mountain anticlinorium for much of its length. Except for the localized ultramafic bodies and basic dikes, no igneous intrusions occur along the anticlinorium and no evidence, such as granitic dikes, suggests their existence at depth.

The common mineral assemblages resulting from regional metamorphism, in order of increasing metamorphism, are designated (Turner 1958, p. 201) as the zeolitic facies, the greenschist facies, the almandine-amphibolite facies and the granulite facies with the former albite-epidote-amphibolite facies being abandoned. The presence of chlorite and/or biotite in the least metamorphosed rocks in the western part of the quadrangle indicates that the degree of metamorphism is too high for the zeolitic facies, which represents conditions transitional between diagenesis to metamorphism. The presence of albite and the lack of oligoclase in the most highly metamorphosed rocks in the eastern part of the quadrangle indicates that the degree of metamorphism is too low for the rocks to be placed in the almandine-amphibolite facies. Thus, all the rocks in the Mount Mansfield quadrangle are in the greenschist facies.

Within the greenschist facies Turner (1958, p. 218) recognizes three subfacies on the basis 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). In the Mount Mansfield quadrangle the quartz-albite-muscovite-chlorite subfacies is most abundant and the quartz-albite-epidote-biotite subfacies is present but less common because the biotite is restricted in its occurrence. Inasmuch as the garnet seems to contain more of the spessartite molecule than the almandine molecule, it is questionable whether the quartz-albite-epidote-almandine subfacies is present. Because of its occurrence in metamorphic rocks of lower grade, spessartite is not as diagnostic as almandine.

The mineral suite of actinolite (?), chlorite, epidote and albite in the amphibolitic greenstones, likewise, is the expected assemblage for the greenschist facies. Epidote and albite are interpreted as forming in place of more calcic plagioclase typical of higher grades of metamorphism.

Although the rocks on Mount Mansfield do not belong to the quartz-albite-epidote-almandine subfacies, a garnet isograd has been drawn on the geologic map (Pl. 1). The occurrence of garnet is restricted to the axial region of the anticlinorium on Mount Mansfield and it does not extend north of Morses Mill, except near the Rousseau Talc prospect. This latter occurrence may have resulted from conditions related to the alteration of the ultramafic body so that the garnet is not significant in terms of regional metamorphism.

STRUCTURE

Large Folds

The major structural features in the Mount Mansfield quadrangle are, from east to west, the Green Mountain anticlinorium, the Cambridge syncline and the Fletcher anticline. These folds have a general trend of about N. 15° E. and are essentially parallel to each other. None of the axes is continuous from north to south because of plunging and overlapping (Pl. 1). No major faulting has been identified in the quadrangle.

Green Mountain Anticlinorium

The most conspicuous fold, the Green Mountain anticlinorium, forms the Green Mountains which extend the length of Vermont and into Canada. In the Mount Mansfield area the crest of the fold is marked approximately by the ridge-like form of Mount Mansfield. Except where an offset occurs, the anticlinal structure does not have typical complex, large-scale drag folds on its limbs, so it might be considered as a large anticline. However, as the term anticlinorium is established for the structure elsewhere in Vermont, the name will be retained in this report.

The position of the axis of the Green Mountain anticlinorium is approximate and is based on the average attitudes of the rocks. The crest of the fold is rather broad and the dips of the beds show considerable variation due to minor drag folding and warping so that, in many places, estimation of the average attitude is better than specific determination of the attitude at one locality. The zone of variable dips is particularly broad near Mount Mansfield because of the plunging and overlapping of the major axis in the vicinity of Smugglers Notch (Pl. 2). Because the beds plunge to the south, they are horizontal in very few places.

The approximate position of the southern part of the axis can be traced northward from Needles Eye, near the southern border of the quadrangle, to just east of The Chin, passing over The Forehead and just west of The Nose. North of Bear Pond it appears to die out. A poorly developed syncline occurs in Smugglers Notch. Beds on both sides of the Notch dip gently 10–15° towards the synclinal axis. This syncline can be traced only short distances to the north and south. The northern part of the axis of the anticlinorium may be traced northward from a point west of Spruce Peak to north of the Lamoille River where it passes into the Hyde Park quadrangle. The approximate position of the axis as shown by nearly horizontal beds may be seen easily east of East Cambridge Cemetery along State Highway No. 15 which follows the Lamoille River where it transects the anticlinorium.

The overlapping of the axes of the anticlinorium seems to be caused by the dying out of one anticline and the development of a second parallel anticline as the major structure carries through. This seems to be characteristic of the Green Mountain anticlinorium as the trends of the axes consistently show overlap to the east as they are traced northward from the Lincoln Mountain to the Mount Mansfield quadrangle. Cady (personal communication, 1958) reports an eastward offset of the major axis in the Lincoln Mountain quadrangle and incomplete mapping suggests a similar structure in the Camels Hump quadrangle.

A culmination of the Green Mountain anticlinorium lies to the north of the Mount Mansfield quadrangle. Along the crest of the anticlinorium the beds dip gently 10–30° to the south. A slight change in plunge of the anticlinorium occurs east of South Cambridge where locally some of the beds along the crest dip to the north (Pl. 2).

The axial plane of the anticlinorium is inferred to dip steeply to the east at approximately 75–80°, so that the structure is slightly overturned to the west. This is confirmed by minor drag folds (Pl. 10), and by steeper dips on the west flank of the mountain compared to those the same distance from the axis on the east flank. The ledges on Sunset Ridge, which extends west from the Chin, clearly show a gradual and uniform westward steepening of the dip without large-scale drag folding.

Fletcher Anticline

Recognition of the Fletcher anticline, named here, within the area of exposure of the Tibbet Hill rocks in the western part of the quadrangle depends mostly upon stratigraphic evidence. The Tibbet Hill schist has been mapped by Booth (1950) and others to the north as including the oldest rocks exposed in the cores of anticlines. Three-fourths of a mile north of the quadrangle near an area of pillow lavas the eastern side of the anticline is identified by the stratigraphy (Dennis and Cady, personal communication, 1958).

Within the Mount Mansfield quadrangle the evidence for this anticline is meager. Most of the rocks dip steeply to the east, at angles of 60° to 85°, so that the position of the axis of the isoclinally folded anticline cannot be located by reversal of dips, although locally a few beds dipping west were noted on the west side of the anticline. The minor structural features do not seem to indicate the position of the anticline, but they do indicate that the structure of the central and western part of the quadrangle is not simply a homocline on the west flank of the Green Mountain anticlinorium. In many places several features may indicate that the tops of the beds are to the east.

The distribution of the Tibbet Hill volcanic rocks suggests that the anticlinal axis is not continuous but is offset to the east when traced northward (Pl. 1). The position of the segments of the axis in the northern part of the area is arbitrarily placed slightly west of the center between the amphibolitic greenstones which were used to mark the upper contact. The formation narrows to the south and as the duplication of the units is lacking, the position of the axis becomes even more speculative. In the west central area the position of the limy greenstone beds several thousand feet to the east and west suggest that the axis occurs in the vicinity of Duffy Hill. It has been placed between outcrops of feldspathic greenstone which are only about 500 feet apart. No consistently horizontal beds were observed and the repetition of the greenstone is not certain, as they may be different flows. Farther to the south, towards Jericho, the exposure area of the unit becomes so narrow and so discontinuous that the position of the axis is not estimated. Although the axis may occur within the southern segments of the Tibbet Hill unit (Pl. 1), it is also possible that these segments occur on one limb of the anticline and that the volcanics on the other limb were not identified because of thinness of the units and lack of outcrop.

The general trend of the Fletcher anticline parallels the Green Mountain anticlinorium. The fold can be traced to the north through the entire Enosburg Falls quadrangle (Dennis, personal communication, 1959) and it may correlate with anticlinal structures mapped by Cady to the south.

Cambridge Syncline

A syncline is required between the Green Mountain anticlinorium and the Fletcher anticline. The position of this syncline, named the Cambridge syncline at the suggestion of Cady, is based on stratigraphic and structural evidence. It has been mapped in the northern two-thirds of the quadrangle where its presence is indicated by the Ottauquechee formation, which is the youngest in the area. The fold is isoclinal, and the average dip of the beds is nearly vertical.

The trace of the axis is difficult to determine accurately and is generally inferred from the position of the formational contacts. In the northern part of the area near East Fletcher it has been placed arbitrarily about midway between the two outcrop areas of black limestone. Two miles to the south, east of North Cambridge, the projection of this axis passes through an area of complex minor folding with horizontal beds occurring locally. An area of complex minor folding also occurs farther east (Pl. 1 and 2) where a warp in the formation occurs and a small syncline probably is present, though not mapped. Because the Ottauquechee formation does not continue to the south along the same trend, it is probable that the axis of the syncline is offset to the west; therefore, the axis is placed just east of Cambridge within the formational boundaries. Locally, some of the beds dip eastward. As the formation can be traced only a short distance to the south a third offset probably occurs and the axis is placed in the small depression just west of Buker Hill. The syncline, as indicated by the Ottauquechee formation, continues through a small valley about a mile to the south before it is lost.

Although a narrow zone of graphitic rocks which extends southward to Underhill was identified as the Ottauquechee, the thickness of the rocks involved is so small that it would be fortuitous if they represented the synclinal axis. No duplication of this graphitic zone was found.

A possible continuation of the Cambridge syncline has been tentatively mapped along the Winooski River in the Camels Hump quadrangle and Cady (personal communication 1958) has mapped a syncline in a similar position relative to the Green Mountain anticlinorium in the Lincoln Mountain quadrangle farther to the south.

As an alternate hypothesis, the position of the syncline may be considered to lie somewhat farther east, passing through Fletcher Mountain and just west of Jeffersonville. If this were true, the graphite zone on the east side of Fletcher Mountain would lie in the Ottauquechee formation rather than in the Camels Hump. The limited, but identical sections of graphite, dolomite and quartzite, particularly well exposed east of Cambridge and south of Jeffersonville along the Brewster River supports this theory. In addition, horizontal beds were located at various places along the proposed axis. On the other hand, duplication of the black limestone could not be found on the east side, and duplication of a greenstone could not be found on the west side. Because most of the beds in this region have rather uniform dips of 60° to 85° to the west, the axial plane for this isoclinal syncline should dip to the west, an attitude which is not supported by the minor structural features or by the regional picture.

The axial plane of the Cambridge syncline should dip steeply to the east parallel to the proposed dips of the major anticlines, as suggested by the minor structures. If such is the case, the most logical position for the syncline is in the area where the dips off the west flank of the Green Mountain anticlinorium are vertical or overturned, so that the vertical beds are on the east side of the syncline and the eastward dipping beds are on the west side. This position coincides with the highly contorted graphitic beds of the Ottauquechee formation and is one of the principal reasons for mapping the syncline in the present position (Pl. 1).

Bedding

Throughout the area bedding may be recognized by differences in composition and texture. In the western portions of the area where the effect of metamorphism is slight, bedding may be observed at many localities; at a few places graded bedding is well-developed (Pl. 6). In thick metagraywacke sequences in which only slight compositional differences may be detected, the bedding is often obscured by well-developed fracture cleavage. In the eastern part of the area where the rocks are finer grained and metamorphism has produced schist, the bedding may be observed as compositional banding and bedding schistosity, although these may be considerably deformed. As the degree of metamorphism is moderate, metamorphic differentiation is absent or limited so that original difference between layers may still be detected.

The overall trend of many of the mappable rock units in the quadrangle is about N. 20° E.

Bedding Schistosity

Bedding schistosity is the most prominent planar feature in the eastern portion of the quadrangle. On top of Mount Mansfield, for instance, the parallel orientation of mica and chlorite has produced a strong schistosity which is parallel to the compositional differences in the rock. In the western portion of the area, bedding schistosity is developed in the incompetent beds but may be almost completely lacking in the competent beds, such as metagraywacke. In such rocks where the bedding is not parallel to other structural features, it may be shown that bedding schistosity does not occur.

Bedding schistosity is visualized as having formed by slippage along bedding planes during compressional deformation. As the rocks were folded, either on a large or small scale, the folds were made possible by a multitude of small-scale differential movements along the bedding planes. As a result of these movements the platy minerals became oriented parallel to slippage surfaces to produce the bedding schistosity. Thus, the bedding schistosity is an early and continuously developing feature in the metamorphism and deformation of the area.

Because of drag folding, individual bedding schistosity planes may, at many places, deviate slightly from the general trend of the original beds. The average bedding schistosity strikes about N. 10° E. except along the crest of the anticlinorium.

Plate 10. Drag folds, west of The Chin, Green Mountain anticlinorium (Loc. 56, Fig. 2). Viewer is looking north at east-west joint surface. The drag folds strike about north-south, dip steeply to the east and plunge gently to the south.

Drag Folds

As the bedding schistosity developed, these planes often were deformed into drag folds varying from large folds or warps with amplitudes of 50–100 feet to tiny folds observed under the microscope in thin sections. Throughout the area the axial planes of the drag folds have an average strike of about north-south and generally dip steeply to the east at 60° to 85° with an average of about 75° E. (Pl. 10). The plunges of these folds vary greatly; the majority plunge to the south, but many plunge to the north. According to Nevin (1949) such variations are characteristic of minor drag folds. The drag folds are formed by differential movement between beds of different competency. Although such folds occur in all rock types, the more intricate folding is found in the less competent beds.

Cross sections of the drag folds were used to determine the position of

Plate 11. Dextral drag fold in metagraywacke, Hedgehog Hill (Loc. 57, Fig. 2). View is to the south and hammer handle is parallel to the fracture cleavage and axial plane of the fold at N.5°W. 70°NE.

the rocks relative to the major structures. In the eastern part of the structure map (Pl. 2) many of the "tops east" and "tops west" designations represent such interpretations from the drag folds.

The plan views of drag folds have been defined as being dextral or sinistral depending on whether the limbs are displaced to the right or left. White and Jahns (1950) give a detailed account of these terms. Theoretically, an interpretation of the structure should be possible if the plan view and the plunge of the fold are known. As shown on the structural map (Pl. 2) the majority of the plan views in the area are dextral with south plunges. Many observations were made, not all of which are recorded on the map, and it is estimated that nearly 80 percent of the plan views are dextral (Pl. 11). Such a pattern indicates that the folds are on the east side of an anticline. This becomes somewhat anomalous in the eastern part of the area where the dips of the bedding schistosity and the cross sectional views of the drag folds clearly indicate that these folds are on the west side of an anticline.

The outcrop pattern for any of the rock units which could be traced for any distance is also dextral in plan view. Northwest of Fletcher much time was spent in tracing the amphibolitic greenstone whose outcrops appear to be offset from each other. In tracing a layer northward, the general experience was that after following rock outcrops for some distance parallel to the general trend of the bedding schistosity the unit would abruptly end and no occurrences could be found farther along that trend. The next outcrop of greenstone, having the same general appearance and parallel to the bedding schistosity, would generally be found offset to the northeast. At the northern termination of one of the greenstone layers, it could be clearly shown that the greenstone occurred in the nose of a small southward plunging synclinal drag fold. Although the greenstone could not be traced at this locality to the nearest outcrop to the northeast, the attitude of the surrounding rocks supported the conclusion that the greenstone had been folded to give a dextral pattern. The overall trend of the rock unit, therefore, is slightly more to the east than the common trend of its bedding schistosity.

Detailed mapping of a plagioclase lava unit south of Metcalf Pond, where the trend of the rock unit did not match the trend of its individual outcrops, suggested the same type of dextral folding. A somewhat similar fold pattern was mapped east of Buck Hollow school in which the nose of the fold was occupied by recrystallized limestone (Fig. 5). In this case it is likely that the limestone flowed into its present position.

Elsewhere in Vermont, the predominance of one type of plan view of drag folding has been noted. In the Hyde Park quadrangle, directly to the east, Albee's map (1957) shows a dominance of dextral folds plunging to the south (indicating the folds are on the east side of an anticline) even on the east side of Foot Brook Syncline. In the Montpelier quadrangle to the southeast Cady (1956) reports that the drag folds are uniformly dextral in pattern. Farther east in the Plainfield quadrangle Konig (personal communication 1959) reports the majority of the early folds are sinistral and Murthy (1957) had a similar problem in the East Barre quadrangle. He resolved some of the difficulty by distinguishing between first- and second-order folds. White and Jahns (1950) in central and east-central Vermont found that the majority of the early drag folds are sinistral with northward plunges. They interpret this as indicating an early, relatively regional, deformation in which the rocks on the east moved northward with respect to those on the west.

As the drag folds in the Mount Mansfield quadrangle display intricate folding of the bedding schistosity, they must have formed when the

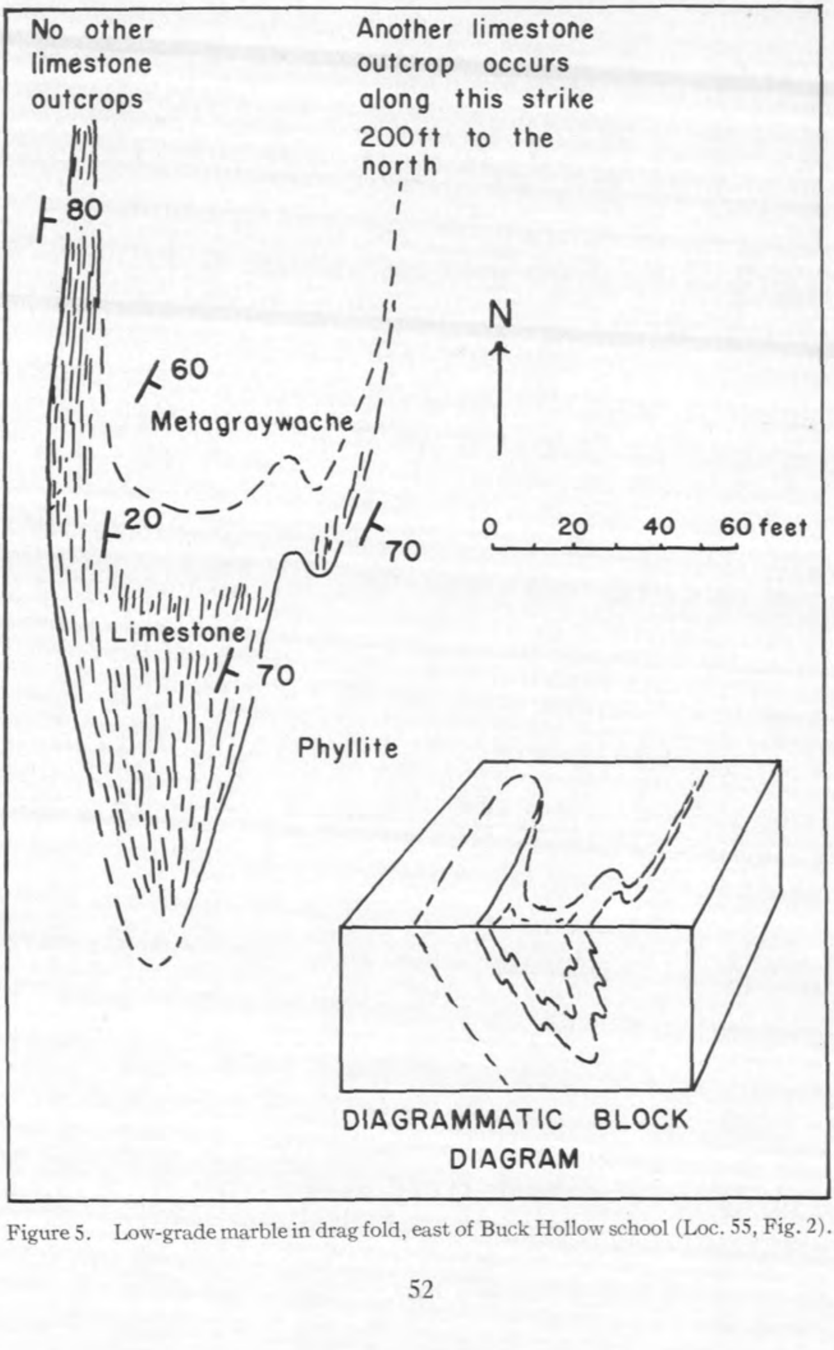

Figure 5. Low-grade marble in drag fold, east of Buck Hollow school (Loc. 55, Fig. 2).

rocks were plastic. Because of this apparent mobility, it seems most likely that they were formed during the height of metamorphism. If the dextral drag folds are postulated to have formed at a later date due to regional shear, as has been suggested, the second deformation would have to be as strong as the first and other evidence should indicate its existence. Thus, it seems unlikely that the dextral folds were formed by a special regional deformation. The reason why dextral folds are more common than sinistral folds is probably related to the entire structural history of the area and will be discussed more fully after the description of the other structural features.

Fracture Cleavage

As used in this report, fracture cleavage refers to the axial plane cleavage of major and minor drag folds along which shear has occurred with little or no development of mineral orientation. Where the competency of the beds becomes less so that shearing is greater or where the amount of mineral orientation along these planes becomes greater, due either to greater metamorphism or to the greater susceptibility of argillaceous rocks to metamorphism, fracture cleavage grades into flow cleavage. Although flow cleavage occurs in the plane of maximum elongation (axial plane) and fracture cleavage occurs in the shear planes, inclined to the axial plane, the angle between them may be so slight that it cannot be used to distinguish them from each other (Nevin, 1949, p. 166). In flow cleavage the original bedding is usually very difficult to recognize.

In much of the geologic literature on Vermont the fracture cleavage of this report is called slip cleavage, which is another name for false cleavage or shear cleavage. The term slip cleavage has been used by many to describe the axial plane cleavage of drag folds formed at the advanced stage of major deformation. In the Mount Mansfield area it can be shown that the cleavage which would commonly be called fracture cleavage in weakly metamorphosed beds grades into what has been called slip cleavage in slightly more metamorphosed rocks. White (1949) in east-central Vermont traces "slip cleavage" grading into "flow cleavage." It appears, therefore, that the feature described in Vermont as slip cleavage may be a transitional feature between fracture cleavage and flow cleavage.

Although the geologists working in Vermont seem agreed on the feature to be called slip cleavage, introduced in Vermont by Dale (1896),

Plate 12. Crinkles in metagraywacke at Fairfax Falls (Loc. 19, Fig. 2). These gently-dipping shallow crinkles are interpreted as representing a poorly developed fracture cleavage related to the second shear direction.

the term may be confusing to others. In some literature, notably where it is called false cleavage, the term implies that this cleavage transects the older structures because it was formed by a later, unrelated deformation. For this reason and because a dividing line between good fracture cleavage and what has been called slip cleavage cannot be established, the term slip cleavage has been abandoned in this report.

In the eastern portion of the Mount Mansfield area, fracture cleavage is common, occurring as closely-spaced cleavage parallel to axial planes of drag folds. However, this cleavage is not always present and rocks with well-developed drag folds may have only faintly developed fracture cleavage (Pl. 10). As the fracture cleavage cuts across the bedding schistosity, it is clearly, in part, a later feature. In many places the fracture cleavage forms minute drag folds which appear as "crinkles" on the bedding schistosity surface.

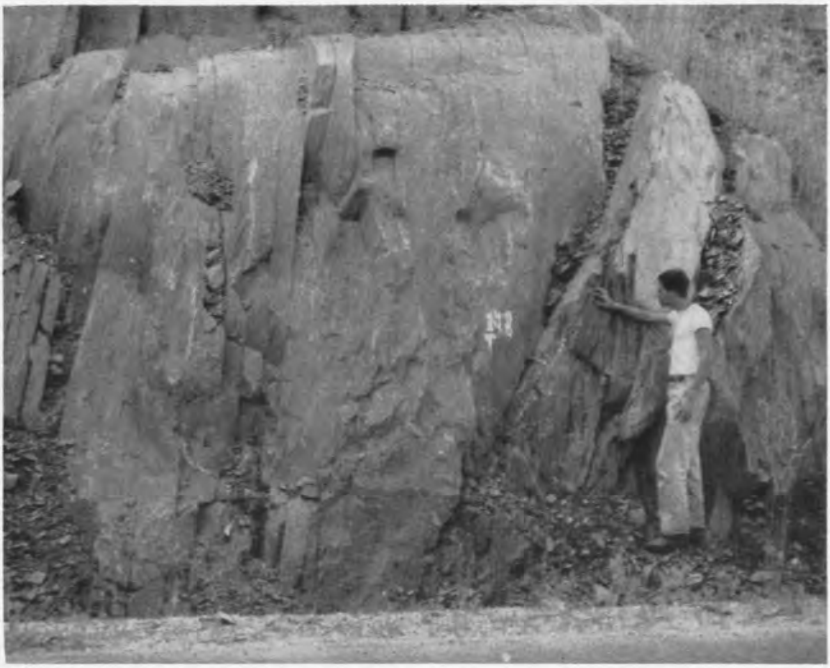

Plate 13. Fracture cleavage in metagraywacke near Fairfax Falls (Loc. 58. Fig. 2) Viewer is looking south. The bedding (near man's hand) is marked by compositional layering and dips less steeply to the east, indicating the area is on the east side of an anticline.

As the attitude of the fracture cleavage and the axial planes of the minor drag folds along the Green Mountain anticlinorium are fairly constant at about NS. 80° E., it appears likely they are essentially parallel to the axial planes of the major folds.

As the fracture cleavage represents a shear plane, a second direction of shear might be expected. Usually only one shear plane develops and except in a few localities only one shear direction was found. At one such locality at Fairfax Falls in the western part of the area, small gently dipping crinkles and "joints" occur at about EW. 25° N. in metagraywacke. The rock has a poorly developed nearly vertical bedding schistosity, which is bent at angles of 5–10° over a distance of several inches to give a series of crinkles. The resulting rock (Pl. 12) has a banded appearance and the crinkles might be mistaken for bedding. These

Plate 14. Fracture cleavage cutting across bedding, east of West Fletcher (Loc. 59, Fig. 2). The bedding (N.60°E.Vert.) is marked by color banding and bedding schistosity is absent. The fracture cleavage (N.30°E. 52°SE.) is slightly warped and is cut by a weak finely spaced cleavage (N.23°E. 55°NW.) of unknown origin.

crinkles and associated fracture planes are interpreted as a poorly developed fracture cleavage complementary to the stronger steeply dipping fracture cleavage in nearby outcrops.

In the western and northwestern part of the quadrangle, fracture cleavage is the prominent planar feature in the metagraywackes. Bedding schistosity cannot develop in massive rocks without bedding planes, so the rock must give by fracturing.

At many localities a cross-cutting relationship between the fracture cleavage and the bedding may be observed (Pl. 13). These were used in the usual manner as clues to the structure and many of the "tops east" and "tops west" designations on the structure map (Pl. 2) represent such interpretations. No uniform trends were delineated so it is believed that the western part of the area consists of many minor folds. Along the flanks of folds where fracture cleavage and bedding nearly coincide,

Plate 15. Bedding cut by fracture cleavage, 0.4 miles SSE. of Halfmoon Pond (Loc. 61, Fig. 2). The bedding trends at about N.80°W. 25°SW. and the fracture cleavage is N.15°E. 75°SE.

movement has caused a parallel alignment of minerals so that it is difficult to determine whether the feature should be called bedding schistosity or fracture cleavage; actually it is both. However, occasionally in nearby areas these may not coincide and it clearly may be observed that bedding schistosity is not present. (Pl. 14 and 15).

At one locality half a mile west of West Fletcher (Loc. 60, Fig. 2) the relations between bedding, bedding schistosity, fracture cleavage and rock type could be demonstrated. Here massive ledges of metagraywacke have widely spaced well-developed cleavage surfaces at N. 20° E. 75° SE. which is parallel to the regional bedding schistosity so that a casual observer would conclude that the bedding and cleavage are parallel. With careful scrutiny, however, the bedding in these metagraywackes is found to be nearly horizontal—a reading of N. 20° E. 10° NW. was obtained in one place—to indicate that the feature is fracture cleavage and not bedding schistosity. A few hundred feet to the west the rock type is an iron-stained phyllite. Here the foliation is formed by shiny mica-chlorite surfaces at N. 25° E. 60° SE. which appears parallel to the compositional differences and was identified as bedding schistosity. This bedding schistosity is minutely folded so that "crinkles" are present on the foliation plane and small drag folds with axial plane cleavage are observed in cross section. This cleavage, which formerly would be called slip cleavage, has an attitude of N. 20° E. 75° SE., which is identical to fracture cleavage in the metagraywacke. Thus, depending on rock type, in a given area competent beds show bedding, no drag folds and widely spaced fracture cleavage and incompetent beds show bedding schistosity, small drag folds and closely spaced fracture cleavage.

Inasmuch as competent beds grade into incompetent beds, all variations in the development of bedding schistosity, drag folds and spacing of fracture cleavage will occur. Where the rocks have been isoclinally folded the bedding will coincide with the fracture cleavage along the flanks and the formation of a type of bedding schistosity will be facilitated. Thus, in the western portion of the area it is often difficult to determine from a single outcrop the exact nature of some of the planar features.

Crinkle Lineation

The intersection of the fracture cleavage on the bedding schistosity surfaces is marked either by faint lines or more commonly by tiny folds called crinkles. These were mapped as a lineation because they could be measured easily on flat outcrops where the attitude of the fracture cleavage or the axial planes of the drag folds was difficult to obtain. The crinkles are especially conspicuous in the incompetent beds and on the flat-lying rocks of the Green Mountain anticlinorium.

Joints

A plot of the poles of 177 joint planes on Mount Mansfield between The Nose and The Chin (Fig. 6) indicate that three prominent joint directions are 1) N. 30° E. 75° NW., 2) NS. vertical, and 3) N. 80° E. vertical (Secor, 1959). The first two sets are approximately parallel to the structural trends and probably are tension joints related to post-metamorphic adjustments. The N. 80° E. Vert. joints may be classified as cross joints.

On the east and west sides of the main ridge of Mount Mansfield, the north-south joints are responsible for some of the cliffs. Some of these

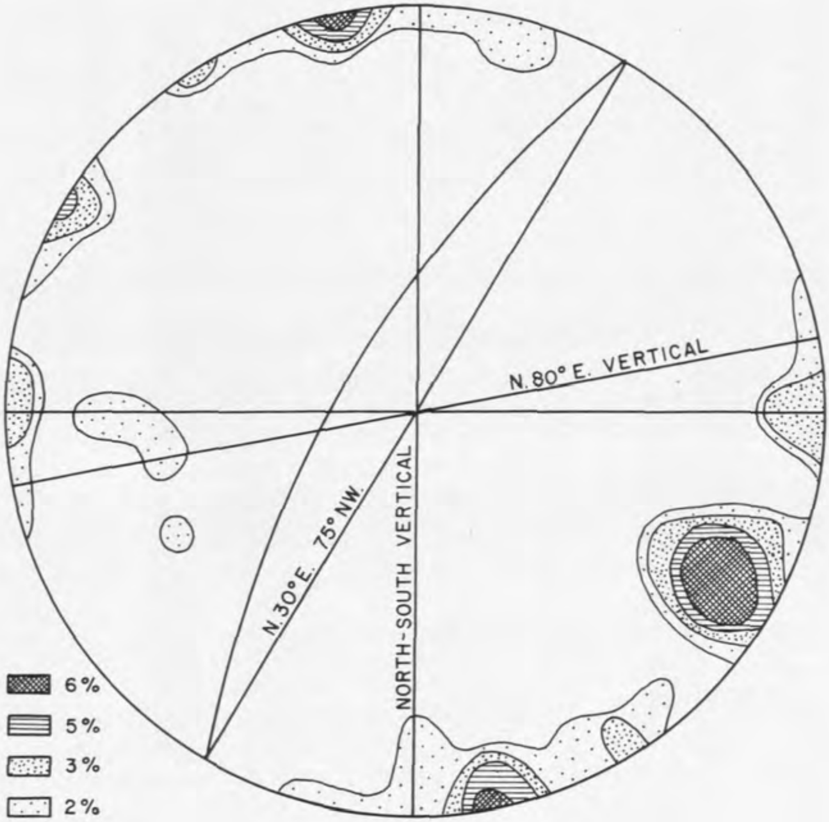

Figure 6. Joint diagram, Mount Mansfield, Vermont. This stereogram projection of the poles to 177 joint planes shows the three principal joint directions, as measured on Mount Mansfield by Donald Secor (1959).

joints have been enlarged by downhill slippage, resulting in narrow steep-walled openings through which trails have been established. The Cave of the Winds was formed by such a joint which has been closed at the top (Pl. 16). It is likely that these joints have been expanded by frost action and those on the east side of the mountain may have been affected by the plucking action of the ice of the Pleistocene glaciation.

Flowage Folds

Minor flowage folds were observed at only one locality, at an elevation

Plate 16. Cave of the Winds viewed from east side of The Chin, Mount Mansfield. Joints in the rock in the foreground probably have been enlarged by downhill movement. The Nose and the Mount Mansfield Hotel are in the distance.

of about 3200 feet on Sunset Ridge (Loc. 62, Fig. 2) west of The Chin. Here chlorite-albite schist with minor quartzose schist layers appears to have flowed towards the synclinal area to the west to form streamlined flowage folds (Pl. 17). The axial planes of some of these is N. 25° W. 30° SE., an attitude which is different from any others observed in the quadrangle.

The schists in the adjacent areas show no evidence of flowage folding. Normal drag folds are present and the bedding schistosity is normal at about N. 5° W. 40° SW. with weakly developed quartz rodding dipping 35° to the S. 65° W. No obvious differences in mineral composition could be noted. Thus, it is difficult to explain why flowage folding occurs in these particular schists but was not observed at other localities.

Quartz Rodding and East-West Folds

At many localities near the axis of the Green Mountain anticlinorium, an east-west lineation is produced by linear concentrations of white, milky quartz, which have been called quartz rodding. These are essentially parallel to the bedding schistosity and occur as small raised ridges

Plate 17. Flowage folds, west flank of Mount Mansfield (Loc. 62, Fig. 2). Viewer is looking N.60°W. at joint surface. Such folds, where the rock appears to have flowed towards the synclinal trough, are not common in the area.

on the bedding schistosity surface, so that they resemble the surface expression of ordinary quartz layers in the folded and warped schist. Many normal lenses of quartz striking in a north-south direction may also be observed.

When the east-west quartz rodding is closely examined, many can be shown to be concentrations of quartz which has migrated into the tension position in the noses of small recumbent east-west folds (Pl. 18). Such folds are recognized only by the outlines of the quartz stringers. Chidester (1953) in his detailed mapping of the Sterling Pond area has identified some "indistinct east-trending open folds and 'rolls'." Cady (personal communication, 1958) reports that larger scale east-west folds occur to the northeast in the Jay Peak quadrangle.

Chidester (1953) believes that these east-west features are older structures which formed penecontemporaneously with the development

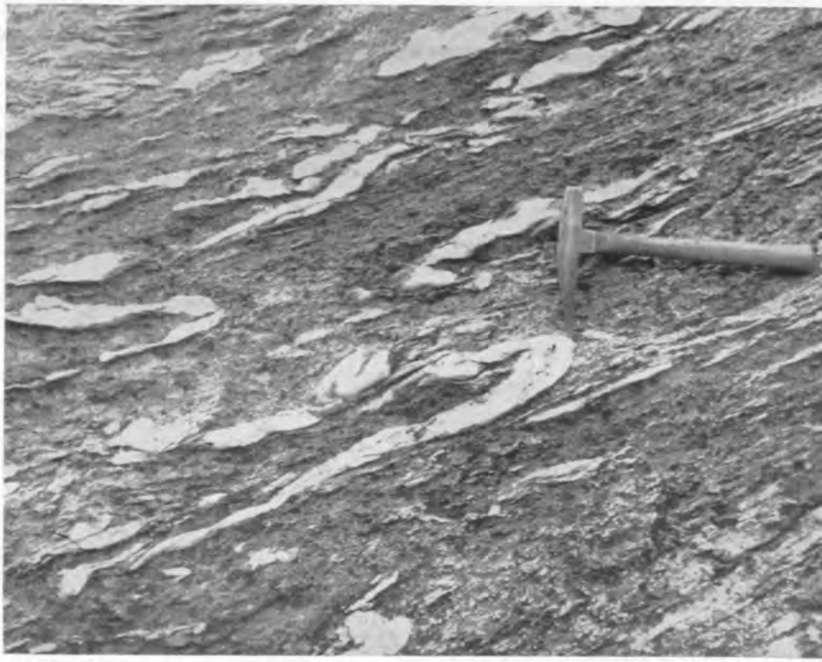

Plate 18. Small recumbent folds marked by quartz stringers (Loc. 62, Fig. 2). Photograph is of a nearly vertical joint surface. The axes of many such folds trend east-west so that the concentrations of quartz in the nose of the folds forms an east-west lineation, called quartz rodding, on the bedding schistosity surface.

of the bedding schistosity prior to the formation of the north-south folds with the folding of the Green Mountain anticlinorium. Albee (1957) writes that both the east-west and north-south structures "result from the differential shear between layers, with slip nearly parallel to the same plane but in different directions." On Mount Mansfield the culmination of the Green Mountain anticlinorium to the north would produce a north-south component of movement between the bedding planes in the axial region which might be sufficient to produce these minor east-west features. No east-west structural features were identified in other parts of the quadrangle.

East-West Crinkle Lineation

At a few localities on the Green Mountain anticlinorium the prominent north-south crinkles transect a weaker east-west set of crinkles to give the bedding schistosity surface a checkerboard appearance. These east-west crinkles are formed by small drag folds which probably originated by the local movement between the bedding schistosity planes related to the formation of minor culminations and sags along the axis of the anticlinorium. Chidester (1953) believes the minor lineation formed in this manner but that the parallelism between the east-west crinkles and the quartz rodding is accidental inasmuch as the quartz rodding is older. No rocks were found which demonstrated conclusively the relationship between the east-west crinkles and the quartz rodding.

Origin of Major Features

Four general observations regarding some of the structural features merit a special discussion of their relations and possible origin. The average trend of the mappable rock units is about N. 20° E. throughout the quadrangle. The average trends of the bedding schistosity and the individual segments of the major folds are about N. 10° E. The average trend of the drag folds and the fracture cleavage is about north-south, except in the western part of the quadrangle where it is more easterly. The plan views of the mappable rock units and the drag folds are predominantly dextral. These generalizations are shown diagrammatically in figure 7 and to a lesser degree by plate 11. If the area consisted simply of plunging anticlines and synclines, some of the rock units should trend east of the average as a result of sinistral folds and the trend of the minor folds and fracture cleavage would be expected to either coincide with or vary in roughly equal amounts on either side of the trend of the major structures. These expected variations were not observed.

The outcrop patterns of the rock units are not those which would be expected for plunging structures. If the Fletcher anticline and the Green Mountain anticlinorium plunge southward, the Cambridge syncline likewise should plunge to the south. The outcrop pattern of the Tibbet Hill rocks in the core of the Fletcher anticline suggests a southward plunging structure, but the outcrop pattern of the Ottauquechee formation in the core of the Cambridge syncline does not suggest a plunging syncline. The outcrop area of the formation should become larger to the south, whereas it becomes smaller. Although this might be explained by a facies change, it is difficult to explain why the only graphitic beds are those near the anticlinal axis. The fact that the graphitic zone in the Camels Hump group can be traced from near Kings Hill Pond to near Macomber Mountain at an average trend of N. 20° E. by a series of

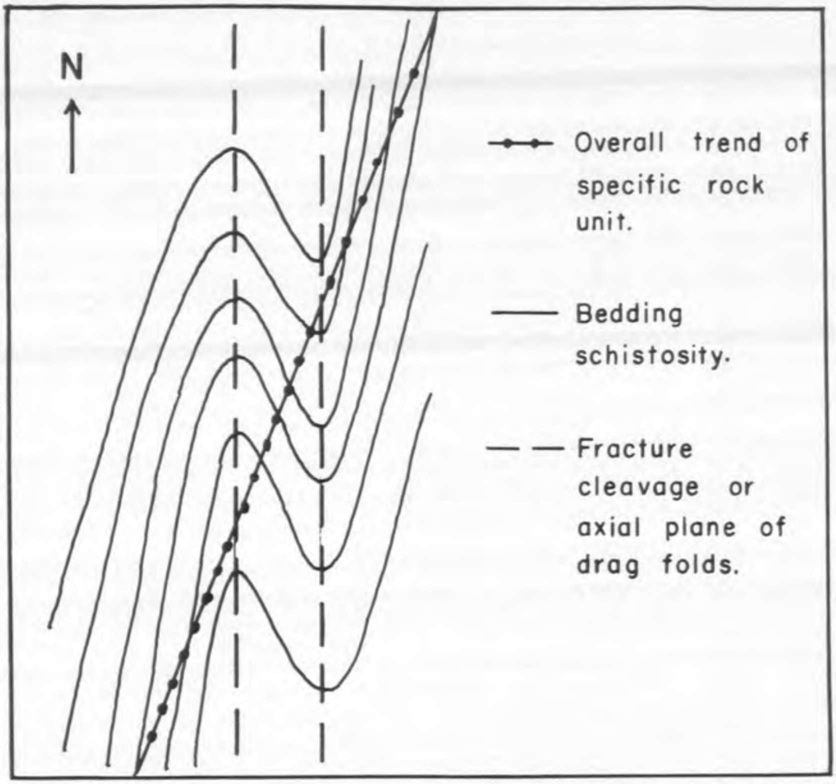

Figure 7. Diagrammatic sketch showing usual structural relationships between rock units, bedding schistosity, and fracture cleavage or the trace of the axial plane of drag folds. Such relations were observed at many scales. The mirror image of this pattern was seldom observed.

outcrops, which suggest a dextral pattern, indicates that something is amiss. If the anticlinorium is plunging to the south, the pattern of the unit should be sinistral and the areas of outcrop in the south should be closer to the axis of the anticlinorium, whereas they appear to be the same distance away or, if anything, slightly farther away. Thus, the fact that the mappable rock units can be traced along a general N. 20° E. trend without "wrap arounds" suggests that the effect of the plunges is minor and not the controlling factor in the distribution of the rock units.

To explain these structural features it is postulated that the deforming compressional forces shifted from S. 70° E. to east-west. It is suggested that such a shift was continuous and probably should not be considered to have been two deformations.

As the rocks are considered to have been deposited in a eugeosynclinal basin, they must have been essentially horizontal prior to the initial deformation. When the deforming forces began to act upon the rocks, they would first be deformed into a series of broad anticlines and synclines. It is suggested that the compressional force was from S. 70° E., so that the resulting structures and rock units would trend N. 20° E. (Fig. 8). Thus, the general trend of the rock units may have been established so that this trend controls their later distribution. Most of the movement associated with this folding probably occurred along the bedding planes, so that the bedding schistosity probably began to develop at this time, and fracture cleavage may have begun to form in the massive rocks.

As the deformation continued and direction of force had shifted, say to S. 80° E., the major folds became better developed and slightly overturned to the east so that the trend of the axial planes was N. 10° E. with a dip of about 80° SE. The bedding schistosity continued to develop, especially near the anticlinorium where the beds were finer grained and the metamorphism was greater. During this stage the drag folds became important and with the development of the drag folds, the fracture cleavage continued to form. The fact that these cut or form wrinkles on the bedding schistosity planes shows that in part they were later.

Near the end of the deformation, the direction of the force may have been nearly east-west, so that many of the axial planes of the minor drag folds and the fracture cleavage have an attitude of NS 70–80° SE. By this stage probably much of the movement was along drag folds or the fracture cleavage. Inasmuch as the direction of deformation was not acting perpendicular to the pre-established bedding schistosity surfaces, a slight horizontal component resulted in which the west side of the bed tended to move northward relative to the east side (Fig. 8). Such a movement would tend to form dextral folds; thus, the dextral drag folds already present would tend to become better developed and more prominent than the sinistral folds as a result of this shear component.

Inasmuch as the folds to the west in the Mount Mansfield quadrangle and those to the east in the Hyde Park quadrangle are all isoclinal, the Green Mountain anticlinorium with flat dips along its crest must represent some special situation. This fact cannot be explained simply by calling the structure an "anticlinorium." As it seems likely that the

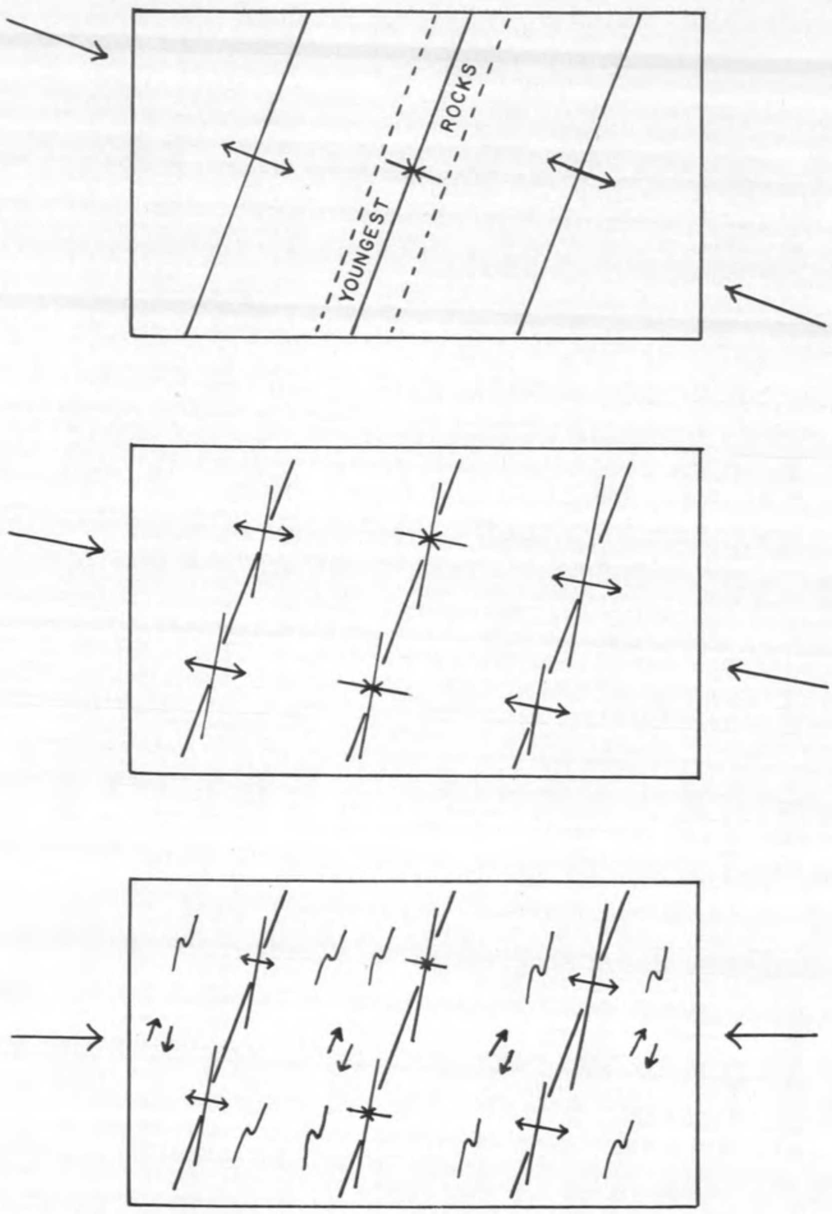

Figure 8. Possible effect of a shifting deformational force from S.70 °E. (upper diagram) to S.80 °E. (middle diagram) to east-west (lower diagram) on the distribution of mapping units and the trends of the structures.

Green Mountain anticlinorium must have a core which resisted isoclinal folding, a core of unexposed pre-Cambrian rock is postulated. The trend of this core probably also reflects the trend of the original N. 20° E. structures.

Age of Deformation

Most geologists who have worked in Vermont seem agreed that the Green Mountain anticlinorium was affected by two orogenies—namely the Taconic near the end of the Ordovician and the Acadian of Middle or Late Devonian age. However, the importance of each relative to the formation of the anticlinorium and associated metamorphism is uncertain. Booth (1950) believes that the strata were strongly folded before thrusting, whereas Cady (1945) believes that the thrusting was Taconic and the principal folds are Acadian.

The Mount Mansfield quadrangle offers no information on the age relationships. The only contribution to the problem is the conclusion that the area was principally deformed and metamorphosed during one deformation. If this deformation was during the Taconic, the effects of the Acadian deformation must have been minor in this area. If this deformation was during the Acadian, it must have obscured the effects of any earlier deformations.

SURFICIAL GEOLOGY

General Statement

As the primary purpose of the geologic study of the Mount Mansfield quadrangle was to study the bedrock geology, the information gathered on the surficial geology is somewhat incidental and by no means complete. However, despite the sketchy nature of observation the features will be briefly described to serve as a basis for future detailed geomorphological studies.

Glaciation

An estimation of the general direction of ice movement may be obtained from the glacial striations on bedrock. Knowing that the ice advanced from the north-west, the directions of the striae of over 200 localities were plotted (Pl. 3). At a given locality, the strike directions of the various striae may vary as much as 20 degrees. As all could not be recorded, an average reading of the deepest striae was made at each station. These, therefore, probably represent the direction of the movement of the last ice sheet. It is probable that many of the less well-defined striations represent variations of movements at the bottom of the ice sheet of the last glaciation; others may represent striations from older glaciations. Where possible the observations were made on the tops of the ridges or on flat areas, as these were believed to be the most representative of the regional glacier movement. Striations near the base of rounded rock knolls varied in direction as though the ice had flowed around the knoll.

The general pattern of the glacial striations in the quadrangle (Pl. 3) shows that the ice moved across the area from the northwest. These observations show a deflection of the ice sheet towards the Lamoille River valley east of Jeffersonville. As this valley represents the only low passage through the north-south trending Green Mountains in this area, it is reasonable to assume that the glacier would flow through this valley. This deflection and the trend of the striae across the Green Mountains demonstrate that the ice lobe in the Champlain Valley was dominant over any which may have occurred in the central part of northern Vermont. South of the Lamoille River the striae seem to indicate that the continental glacier passed over Mount Mansfield with only very slight eastward deflection, as might be expected due to the mechanism of refraction.

The glacial striae on top of Mount Mansfield, noted by many earlier geologists (Hitchcock 1861 and 1904, Hungerford 1868, Goldthwait 1916) strike about S. 45° E. and clearly indicate that the Green Mountains in this region were completely covered by the ice sheet. In addition, erratic boulders are found on Mount Mansfield, as well as throughout the area (Jacobs, 1942). Figure 19 shows two large boulders of garnetiferous quartz-albite-muscovite schist which occur on the Long Trail along the summit ridge of Mount Mansfield (Loc. 3, Pl. 3). Although these boulders probably are of local origin, they are erratics for they rest on a glaciated surface which clearly shows striae passing beneath them.

The sharp cliff on the northeast side of The Nose probably represents glacial erosion, in part "quarrying," along a vertical northwest joint surface. Some of the open joints on the east side of The Chin (Christman, 1956) may represent, in part, the effect of the plucking action of the glacial movement over the mountain.

Inasmuch as local glaciers occurred on Mount Washington in New Hampshire during the Pleistocene, the question has been raised as to

Plate 19. Glacial erratics and striae, Mount Mansfield (Loc. 3, Pl. 3). These erratics, locally called Drift Rock, rest on bedrock showing glacial striations, oriented away from the viewer.

whether local glaciers were ever present on Mount Mansfield. Most writers believe that glaciers did not exist here and agree with Goldthwait's opinion (1916, p. 71) that, "The area above 4000 feet is altogether too narrow to catch snow in any considerable quantity and the absence of long spurs and deep reentrants on the sides of the mountain leaves no pockets where drifting snows could be caught and concentrated into valley glaciers." However, the steep-walled, east-facing slope below the main ridge of Mount Mansfield appears to be suitable for snow accumulation so that the writer suggests that nivation has modified the form of the mountain slope.

The rock basins occupied by Lake of the Clouds and Bear Pond on Mount Mansfield and Sterling Pond on Spruce Peak undoubtedly resulted from the erosive action of the ice sheet, as their locations preclude the possibility of formation by normal processes of erosion by water.

These particular sites may have been areas of weak rock, either as a result of their composition or jointing, so that they were differentially eroded. It is more than coincidental that part of Sterling Pond on Spruce Peak is underlain by a soft talc-carbonate rock. Geomorphically, this pond is particularly interesting because of the delicate balance between its intake and outflow of water. Although it is only 6.8 acres in size and has a drainage area of only 17.7 acres, it has an almost continuous outflow of water to the northwest outlet during the summer (Chidester 1953). If the lake level were to rise approximately 5 feet this lake would have a dual outlet, draining to the south as well as to the northwest.

Although the origin of Smugglers Notch cannot be positively established, it must have been partially formed or modified by glaciation. Smugglers Notch occurs at an elevation of 2170 feet as the drainage divide between the Brewster River flowing north and the West Branch of the Waterbury River flowing south (Pl. 1). The divide is narrow with high cliffs on either side and is partially filled with large talus blocks derived from these cliffs. As it is narrow and lacks a U-shape, it is unlikely that the Notch owes its present shape to glacial erosion. Jacobs (1942) reports north-south striations along the Hell Brook trail somewhere between the Lake of the Clouds and Smugglers Notch. He interprets them as proving that an ice tongue flowed through the Notch. It is probable that ice did flow across the divide between Mount Mansfield and Spruce Peak before the Notch was eroded to its present level. As it does not display linear elements, it is unlikely that the sinuous shape of the pass is due to a structural weakness such as a fault or joint system. The shape of the pass suggests that it was formed by stream action, but the depth of the pass is too great for it to have formed simply by the headward erosion of the Brewster and West Branch Rivers.

At an elevation of about 1950 feet on the north side of the Notch, a local deposit of well-sorted sand (Loc. 1, Pl. 3) indicates that water must have been impounded at this elevation at one time. North of the Notch at an elevation of about 1750 feet a swampy area without rock outcrops in the headwater valley of the Brewster River suggests that surficial deposits accumulated abnormally at this high elevation. These two facts suggest that during the waning stages of glaciation, perhaps when the continental ice sheet still lapped onto the western side of the Green Mountains but had melted away from parts of the eastern side, the Notch served as a marginal channel for melt waters from the glacier. This stream probably originally flowed across a low point in the divide between

Spruce Peak and Mount Mansfield, eventually cutting a V-shaped valley. Later when the ice had melted back, the divide may have served as a barrier and a small lake was formed on the north side for a brief time. Eventually, the ice retreated completely and the headward erosion of the Brewster and West Branch Rivers may have modified the Notch and developed the present drainage system.

A heavy mineral analysis was made of two samples of sand from Smugglers Notch in an attempt to establish a foreign origin for the sand. However, the samples contained only heavy minerals which are found locally and its mineral suite was similar to that of two other samples of locally derived material (Loc. 2 and 3, Pl. 3).

Glacial and Recent Deposits

Glaciofluvial and glacial deposits occur throughout the area, particularly in the valleys and on the lower slopes of the Green Mountains. The distribution of these deposits is shown on Plate 3, although no attempt was made to accurately map the contacts or to distinguish the Pleistocene deposits from those of recent origin. Thus, the map unit indicates only where bedrock is generally lacking and the surficial cover appears to be thick.

The area contains many gravel and sand pits; their locations are shown on Plate 3. Many of these are either kames, kame terraces or glaciolacustrine deposits.

One esker, north and northeast of Underhill, (Loc. 4, Pl. 3) was traced discontinuously for about 8000 feet. The local drainage history of this particular area is probably complex. Gravels in a nearby pit are crossbedded to indicate that the depositional currents flowed from west to east.

Near Halfway House on the west slope of Mount Mansfield at an elevation of about 1900 feet, the torrential rains of 1955 produced an interesting exposure along the main stream (Loc. 2, Pl. 3). At the base of the exposure was about 20 feet of horizontal laminated fine-grained sand and clay with local lenses of pebbles. This was overlain by about 30 feet of sand in which bedding could not be discerned, although a few rounded pieces of schist were scattered through the deposit. The sand was overlain by about 15 feet of bedded sand and gravel. This sequence probably represents marginal lake deposits which were formed when the glacier was just west of this point, filling the Champlain Valley but not covering Mount Mansfield. The rock ridges to the north and south of this locality were probably a factor in the formation of this small proglacial lake.

North of Smugglers Notch along the banks of a tributary to the Brewster River, several deposits of gray till were observed (Loc. 5, Pl. 3). The boulders and pebbles, up to seven inches in diameter, in this till were rounded and were composed mainly of quartz and quartzite; no schist was observed.

A laminated fine-grained sand deposit occurs along the Seymour River near Pleasant Valley (Loc. 6, Pl. 3) at an elevation of 640 feet. This is interpreted as a standing water deposit of proglacial origin.

Lake Lamoille Deposits

The sand and clay deposits occurring at low elevations along the Lamoille valley are fine-grained and well-sorted suggesting glaciolacustrine origin. The distribution of these deposits in the quadrangle is shown in Plate 3. Most of the deposits occur at elevation below 600 feet. At Sand Hill sand is found at an elevation of 800 feet, but at this locality the sand deposits may represent dune material rather than a former lake level. In places, the sand deposits form terraces on either side of the valley; elsewhere they form irregular sand patches with features typical of beach deposits with the exception that moving dunes were not found. The sand is well sorted and composed of relatively pure quartz.

Locally, laminated sediments are exposed in recent road and stream cuts (Pl. 3). Although some of these laminated deposits contain fine-grained sand, most are composed of rhythmically alternating silt and clay. They were observed at elevations from 440 to 540 feet and were usually overlain by sand. Particularly good exposures were observed north of the Lamoille River about two miles east of Fairfax Falls in a new road cut (Loc. 7, Pl. 3). Here laminated clays at an elevation of 440 feet are both overlain and underlain by sands and gravels. Although most of the beds are horizontal, some have primary dips as high as 20° where the varves appear to lie on unconformable scoured surfaces cut in sand and gravel. At this locality in a vertical section of seven feet, 47 silt-clay pairs were measured, an average thickness of about 1¾ inches per laminated pair.

Just northeast of the town of Jeffersonville, varved clays and sands are exposed in a 100-foot section along the Brewster River (Loc. 8, Pl. 3). Here two small landslides have occurred in which the overlying sands have slid down over the underlying clays.

The sorted sands and varved clays clearly indicate that some time in the history of the Lamoille Valley, it was occupied by a lake with an elevation as high as 600 feet. In this paper the deposits from this lake have been simply designated as "Lake Lamoille deposits." This body of water may have existed as an independent pro-glacial lake when the ice blocked normal drainage to the west or it may have been a branch of Lake Vermont when the Lamoille valley was connected with the Champlain valley.

Among the early writers who speculated on the origin of these sand deposits was Hitchcock (1906) who believed that they represented "ordinary terraces made when the Lamoille River confined its labors to its own proper limits." When the ice blocked the normal drainage to the west, he postulated that the Lamoille River flowed eastward at one time to join with the waters from Glacial Lake Memphremagog to flow southward through the Stowe strait. This view has been supported by many writers including Bigelow (1932) who worked in the Stowe area.

Merwin (1908) in his extensive study of the history of the Lamoille Valley recognizes several different stages of Lake Lamoille. The First and Second Lake Lamoille occurred east of the quadrangle and passed into the Lake Mansfield stage when the drainage was south through the Stowe valley to the Winooski river. When the waters of this lake dropped to cause a division of the waters at the divide between the Winooski and Lamoille valleys north of Stowe, Third Lake Lamoille formed and was responsible for various sand and clay deposits at elevations of about 650 feet between Jeffersonville and eastward to Morrisville. At this stage, given as part of the Coveville stage of Lake Vermont, Merwin suggests that the ice front may have been in the vicinity of Jeffersonville. Presumably, the deposits at lower elevations between Jeffersonville, Cambridge and Fairfax Falls occurred at some later stage (Fourth Lake Lamoille?) when the ice had retreated westward down the valley.

Chapman's maps (1942) showing the positions of the ice sheet and Lake Vermont in the Champlain valley, postulate that the ice blocked the mouth of the Lamoille valley as late as the Coveville stage of Lake Vermont. When the ice retreated further north, the Lamoille River developed an outlet east of Fairfax Falls south past Westford into the Winooski valley. Still later when the gorge at East Georgia was opened, even though the water level dropped, Chapman believes that the Lamoille valley was probably flooded to a depth of 200 feet during the Fort Ann Stage of Lake Vermont, so that the delta into Lake Vermont at this time was "far east of Fairfax." He also has shown that the waters of the marine invasion of the Champlain trough extended only as far east in the Lamoille valley as south of East Georgia, thirteen miles west of Jeffersonville.

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PLATE I

MOUNT MANSFIELD QUADRANGLE
VERMONT
15 MINUTE SERIES (TOPOGRAPHIC)

TOPOGRAPHY BY U. S. GEOLOGICAL SURVEY 1948

GEOLOGY BY ROBERT A. CHRISTMAN

SCALE 1:62500

CONTOUR INTERVAL 40 FEET
DATUM IS MEAN SEA LEVEL

APPROXIMATE MEAN DECLINATION, 1948

BEDROCK GEOLOGY, MOUNT MANSFIELD QUADRANGLE, VERMONT

EXPLANATION

MISSISSIPPIAN (?)

Md — BASIC DIKE ROCKS

ORDOVICIAN

Osts — SERPENTINITE, TALC-CARBONATE ROCK, AND STEATITE

CAMBRIAN

€o — OTTAUQUECHEE FORMATION
Graphitic phyllite and schist with quartzite and some graphitic limestone.

€ch — CAMELS HUMP GROUP
Metagraywacke and phyllite in west; quartz-albite-mica schist in east.

€th — TIBBET HILL SCHIST
Greenstone, phyllite and metagraywacke.

MINOR ROCK TYPES

Outcrop areas of some distinctive rock types

Graphitic phyllite and schist

Amphibolitic and feldspathic greenstone

Calcareous greenstone

Limestone and/or dolomite

Coarse-grained metagraywacke and metaconglomerate

Garnet isograd

STRUCTURAL SYMBOLS

Contact, dashed where indefinite or gradational

Anticline — Syncline
Axis of major fold

15 Inclined — 70 Graded — 80 Overturned graded
Strike and dip of bedding

25 Strike and dip of bedding schistosity

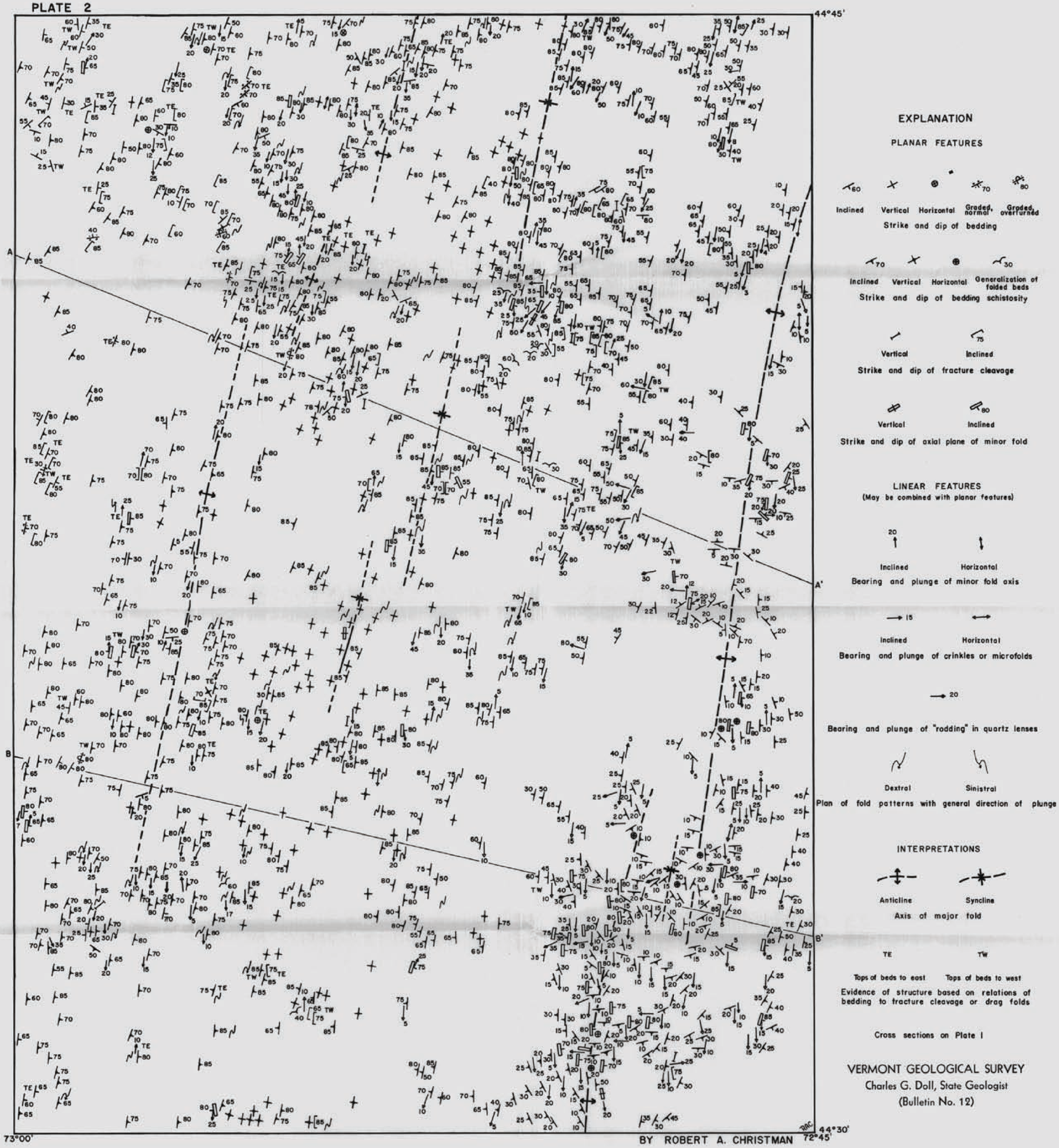
Strike of vertical bedding schistosity

SECTION A—A'

SECTION B—B'

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 12)

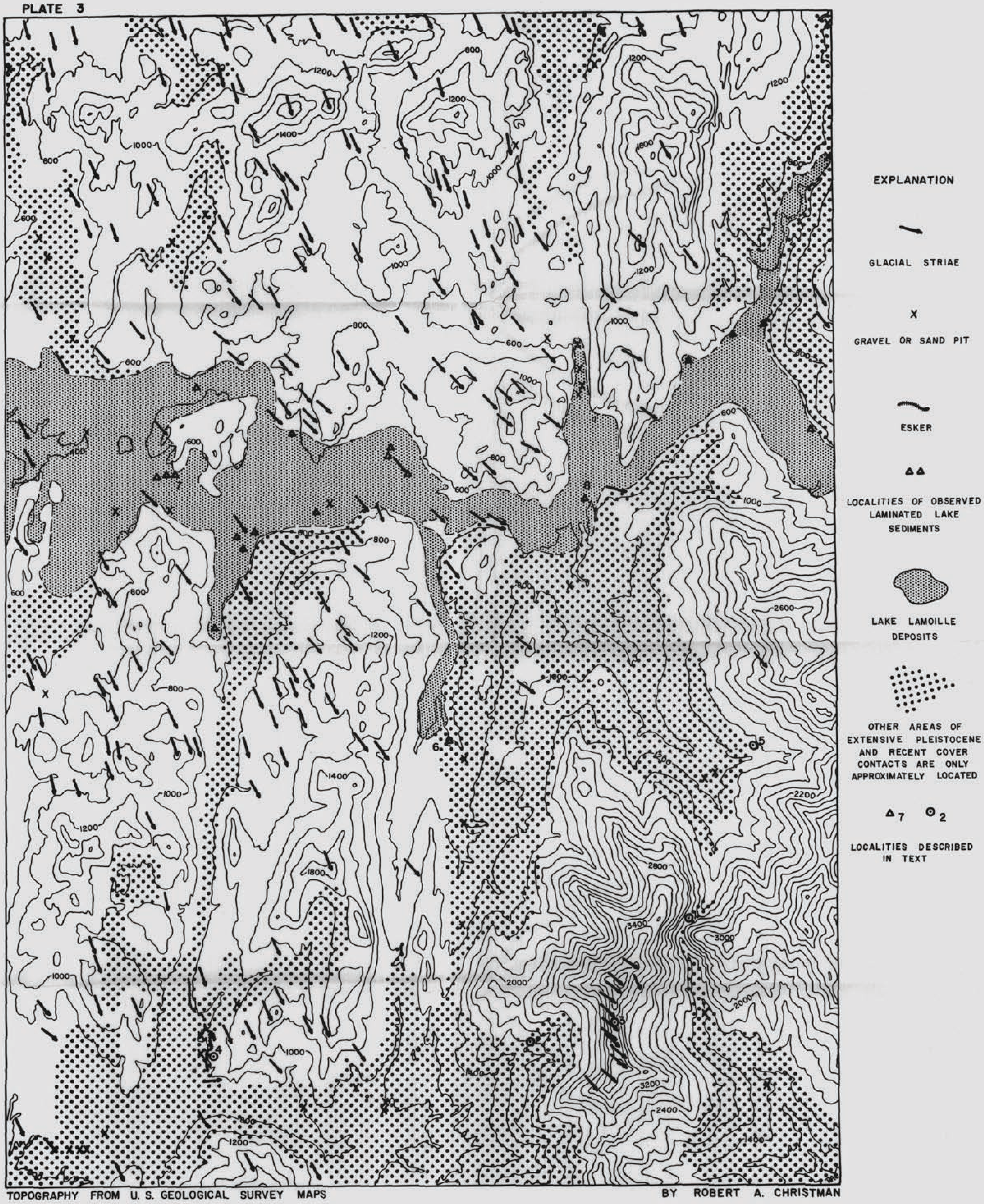
PLATE 2



STRUCTURAL MAP, MOUNT MANSFIELD QUADRANGLE, VERMONT

0 1 2 3 4 MILES

PLATE 3



TOPOGRAPHY FROM U. S. GEOLOGICAL SURVEY MAPS

BY ROBERT A. CHRISTMAN

SURFICIAL GEOLOGY, MOUNT MANSFIELD QUADRANGLE, VERMONT

CONTOUR INTERVAL IS 200 FEET

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 12)