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ABSTRACT

The Woodsville quadrangle, in east-central Vermont and west-central New Hampshire, was studied primarily to determine the stratigraphic and structural relations between two contrasting sedimentary sequences, one in western and central New Hampshire, the other in eastern and central Vermont. The Vermont rocks in this quadrangle consist of three formations, probably Middle Ordovician, with a total thickness between 15,000 and 20,000 feet. A fault separates these rocks from the New Hampshire rocks, which are pre-Silurian, Silurian, and Lower Devonian and total approximately 20,000 feet thick. Volcanic rocks, mostly pyroclastic, constitute about 10 per cent of the stratified rocks in the Woodsville quadrangle. Plutonic and hypabyssal rocks, probably late Devonian, occurs as dikes, sills, large intrusive sheets, and small stocks.

The quadrangle is divided into four major tectonic units by three faults, which trend north-northeast. The Ammonoosuc fault dips about 36°W. and is younger than the folding and regional metamorphism. The Northey Hill and Monroe faults are steep and are older than most of the metamorphism. Because of the faults the major folds are fragmentary, preserved only as single limbs or as parts of synclines. Countless minor folds are present throughout the region.

Progressive metamorphism is well displayed in the quadrangle. Whereas the rocks in a central north-south belt belong in the chlorite zone, the metamorphism increases toward the east and west margins of the quadrangle, where the rocks lie in the staurolite zone. The sillimanite zone has been attained locally around plutonic rocks. Retrograde metamorphism is especially strong along the Ammonoosuc thrust.

The folding, thrusting, metamorphism, and igneous intrusions were all phases of the late Devonian Acadian orogeny.

Special attention was given to the origin and chronology of the cleavage and minor folds in the Vermont part of the quadrangle, where there is evidence of two stages of deformation. During the earlier stage, schistosity formed parallel to the axial planes of contemporaneous minor folds. Because of the later deformation, these earlier folds are inconspicuous. During the later stage, slip cleavage formed parallel to the axial planes of contemporaneous minor folds. In many places these later folds are very conspicuous and if interpreted as ordinary drag folds may lead to an incorrect interpretation of the stratigraphy and major structure. Toward the west the slip cleavage becomes schistosity. The concept of two stages of deformation has important applications to the geology of New England and other metamorphic terranes.

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INTRODUCTION

The Woodsville quadrangle, occupying 210 square miles in east-central Vermont and west-central New Hampshire, lies between 44°00′ and 44°15′ north latitude and 72°00′ and 72°15′ west longitude (Fig. 1). Billings mapped the geology of the New Hampshire part of the quadrangle during parts of the summers of 1934 and 1935, supplemented by brief visits in 1936 and 1937. White mapped the Vermont portion of the quadrangle during the summers of 1937, 1938, and 1939. In 1940, 1943, and 1946, he spent several months in near-by parts of Vermont in the course of investigations for the U. S. Geological Survey, and this work contributed much to his understanding of the regional setting of the Woodsville area. Billings was assisted during 1934 by Allen Waldo, C. Wroe Wolfe, and Samuel Bowditch. White was assisted for one summer each by E. R. Breed, R. F. Story, J. H. Eric, and Ralph Hornblower, Jr. The Associates in Science of Harvard University helped defray part of the field and laboratory expenses.

Problems

Billings undertook the mapping of the New Hampshire portion of the quadrangle in order that J. B. Hadley, who began mapping the Mt. Cube quadrangle in 1934, might have a better correlation with the stratigraphy of the Littleton-Moosilauke area (Billings, 1937; Hadley, 1942). A topographic map of the Vermont part of the Woodsville quadrangle was not then available, and consequently the mapping ended at the Connecticut River.

During the field work in western New Hampshire between 1931 and 1935, it was recognized that the strata in east-central Vermont differ in many respects from those in west-central New Hampshire. It was tentatively believed that the strata in Vermont underlay those in New Hampshire. The Vermont portion of the Woodsville quadrangle was believed to be a favorable place to study these relations. A topographic map became available about this time and White undertook the investigation. The area was not as satisfactory as hoped, because a fault separates the two contrasting lithological types. On the other hand, since White initiated his study 13 years ago, many additional data bearing on this problem have been obtained in adjacent regions.

The investigation inevitably developed into a study of the stratigraphy and structure of the entire area. The results thus constitute a contribution to a long-range program in central New England to elucidate the paleogeography, structure, metamorphism, and petrology of a complex orogenic region. Moreover, the Vermont portion of the area proved to be a particularly fruitful one in which to study two stages of deformation. The earlier stage was characterized by obscure minor folds and a widespread schistosity that is commonly parallel to the bedding. During the later stage of deformation, the bedding and earlier schistosity were thrown into minor folds associated with a second generation of cleavage.

Stratigraphy

General Statement

The stratified rocks of the Woodsville quadrangle range from Middle Ordovician (?) to Lower Devonian (Table 1). In general, the rocks are progressively younger from west to east, and the youngest rocks are in the extreme southeast corner of the area (Pl. 1). The total thickness of the sedimentary column approximates 35,000 to 40,000 feet, but extensive folding makes it difficult to obtain a precise figure.

Seven formations have been assigned to the Ordovician (?): from oldest to youngest they are the Waits River, Gile Mountain, Meetinghouse, Orfordville, Albee, Ammonoosuc, and Partridge formations. Paleontological evidence indicates that the Waits River is Middle Ordovician and that the Partridge is Pre-Silurian.

Two Silurian formations, the Clough quartzite and Fitch formation, which elsewhere overlie the Partridge formation, are absent in this quadrangle because of the Northey Hill thrust. Part of the Devonian Littleton formation is exposed in the extreme southeast corner of the area.

The Orfordville and the formations above it, although extending a short distance into eastern Vermont (Fig. 2; Pl. 1) are largely confined to New Hampshire; they may be termed the "New Hampshire sequence." The Meetinghouse and the formations west of it, although extending into northern New Hampshire, are far more extensively developed in eastern Vermont; they may be termed the "Vermont sequence." These words are put in quotation marks because it is not intended that they should be used as stratigraphic terms.

Although the structure and metamorphism will be described in detail on later pages, the presentation here of two important facts will greatly facilitate the description of the stratig-

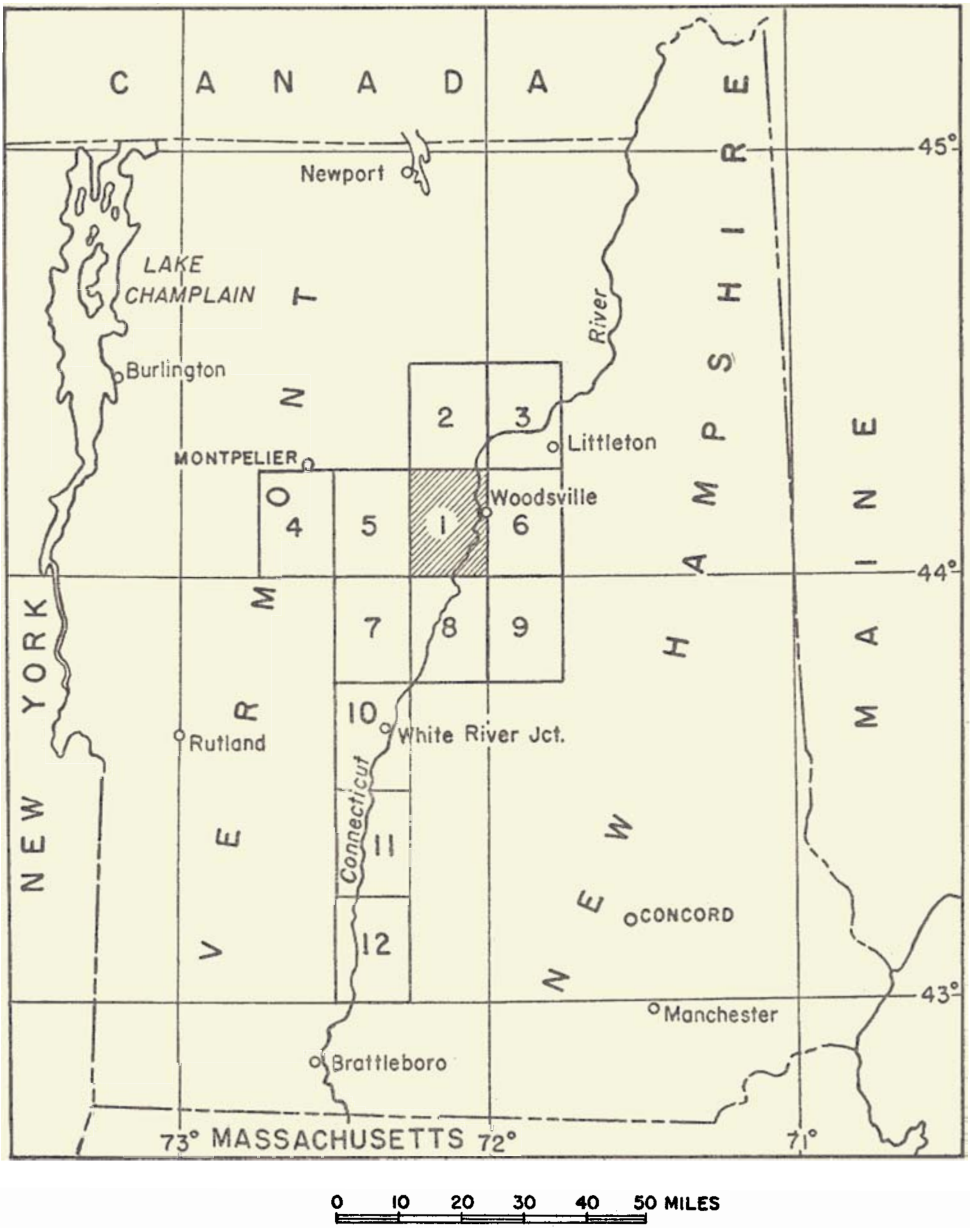


FIGURE 1.—INDEX MAP, VERMONT AND NEW HAMPSHIRE

Woodsville quadrangle is shaded. Quadrangles numbered as follows: 1. Woodsville, 2. St. Johnsbury, 3. Littleton, 4. Barre, 5. East Barre, 6. Moosilauke, 7. Strafford, 8. Mt. Cube, 9. Rumney, 10. Hanover, 11. Claremont, 12. Bellows Falls.

raphy. Three large faults divided the area into four tectonic units; from east to west they are the Northey Hill, Ammonoosuc, and Monroe faults (Fig. 2; Pl. 1). Secondly, the intensity of the metamorphism is not the same throughout the area. A belt of low-grade metamorphism (chlorite zone) extends from the south-central edge of the quadrangle to the northeast corner. The metamorphism increases westward and the biotite, garnet, and staurolite isograds

TABLE 1.—STRATIGRAPHY OF THE WOODSVILLE QUADRANGLE

	Approximate Thickness in feet
Lower Devonian	
Littleton Formation: Gray mica schist, quartz-mica schist, and micaceous quartzite; biotite and garnet porphyroblasts, which in many places are chloritized.	5000
Silurian Clough and Fitch formations missing because of Northey Hill thrust	
Upper Ordovician?	
Partridge Formation: Black slate, with some thin beds of fine-grained, light-colored quartzite	0–2000
Ammonoosuc Volcanics: Fine-grained massive to schistose soda-rhyolite, soda-rhyolite volcanic conglomerate, and chlorite schist.	2000
Albee Formation:	
West of Ammonoosuc thrust: Phyllite, feldspathic phyllite, quartzose phyllite, argillaceous quartzite, quartzite, and feldspathic quartzite. Contains biotite porphyroblasts in north-central part of quadrangle.	
East of Ammonoosuc thrust: Mica schist, mica-quartz schist, micaceous quartzite, and quartzite; near French Pond granite contains sillimanite schist. The Piermont member contains rocks similar to main part of formation, but is dominantly dark-gray staurolite-garnet schist, mica-quartz schist, and mica schist.	5000±
Middle Ordovician?	
Orfordville Formation:	
West of Ammonoosuc thrust: Dark-gray to black phyllite with thin beds of gray micaceous quartzite. Sunday Mountain volcanic member at top is gray soda-rhyolite and chlorite schist. Hardy Hill quartzite 1000 feet below top.	
East of Ammonoosuc thrust: Dark-gray to black mica schist and quartz-mica schist, with some thin quartzite beds. Sunday Mountain volcanic member at top is biotite gneiss and amphibolite. Hardy [illegible] op.	5000±?

TABLE 1—*Continued*

	Approximate Thickness in feet
Lower part of Orfordville formation and possibly some unknown formations missing along Monroe fault	
Meetinghouse Slate: Thinly-bedded dark-gray to black slate (southern part of quadrangle) or gray fissile schist (northern part of quadrangle), with some light-gray micaceous quartzite.	2500
Gile Mountain Formation:	
East of biotite isograd: Dark-gray to black phyllite and light-gray micaceous quartzite.	
West of biotite isograd: Dark-gray to black mica schist, mica-quartz schist, gray quartz-mica schist, and light-gray micaceous quartzite. Biotite, garnet, and staurolite porphyroblasts west of appropriate isograd. Also some calcareous beds at south end of quadrangle.	6500
Waits River Formation: Bluish-gray granular quartzose marble, gray calcareous quartz-mica schist, dark-gray to black quartz-mica schist, and mica schist; small amounts of light-gray or buff quartzite and micaceous quartzite.	10,000

appear in orderly succession (Pl. 1). The rocks are classified into metamorphic zones as follows: chlorite zone—east of biotite isograd; biotite zone—between biotite and garnet isograds; garnet zone—between garnet and staurolite isograds; staurolite zone—west of staurolite isograd.The significance of the actinolite isograd (Pl. 1) will be considered more fully later; for the present it will suffice to consider it an isograd of secondary importance lying within the staurolite zone. Metamorphism also increases eastward from the central belt of low-grade rocks, but here some of the metamorphic zones have been eliminated because of thrusting along the Ammonoosuc fault, to the east of which the rocks lie in the staurolite and sillimanite zones. The lithological description of each formation is consequently more complicated than in an area of uniform metamor-

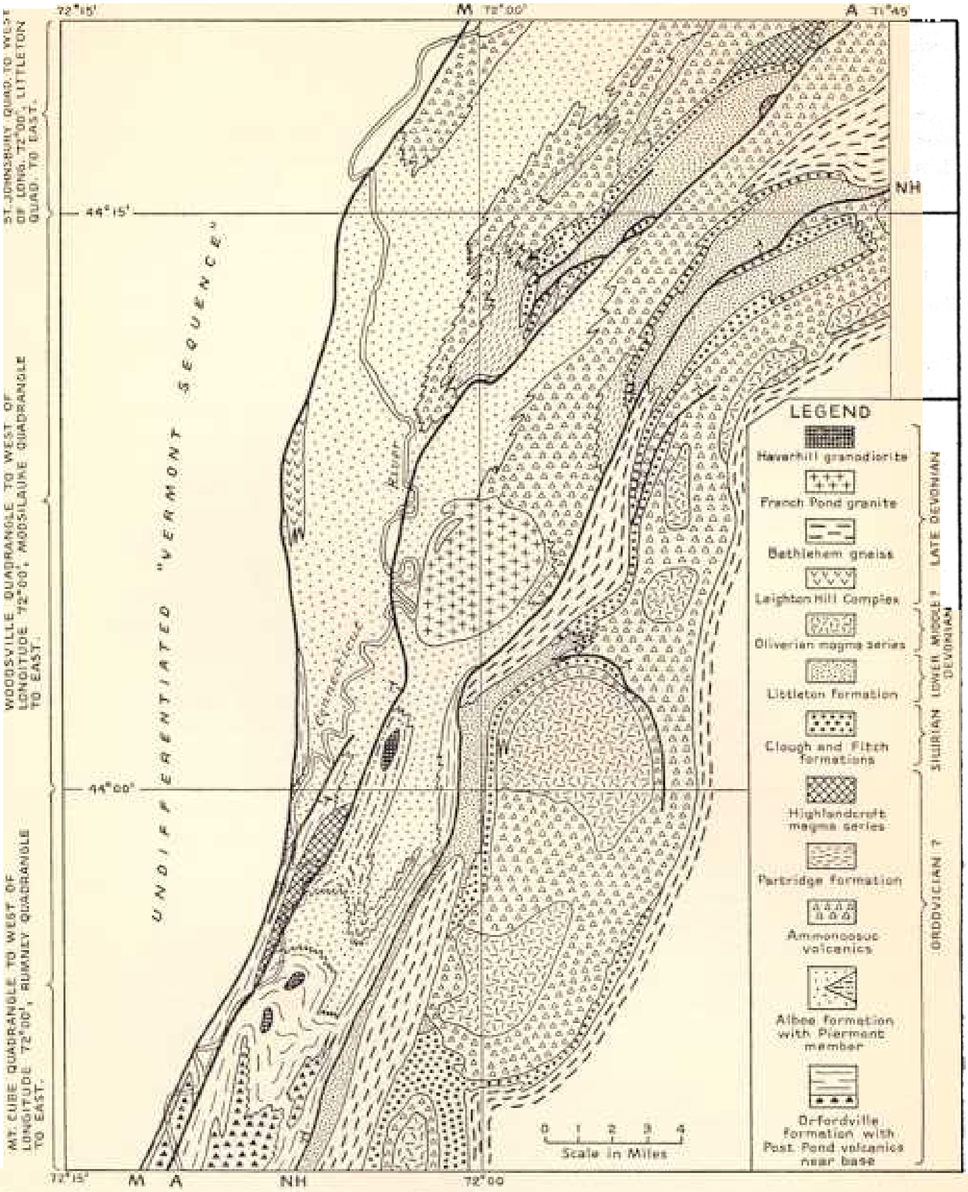


FIGURE 2.—GEOLOGICAL SKETCH MAP OF NEW HAMPSHIRE PART OF WOODSVILLE QUADRANGLE AND ADJACENT AREAS

Faults shown by heavy lines. M = Monroe fault; A = Ammonoosuc thrust; NH = Northey Hill thrust; T = overthrust side.

phism and must be presented separately for each metamorphic zone.

Waits River Formation

Distribution.—The Waits River formation underlies most of the western third of the Woodsville quadrangle (Pl. 1), and forms a belt 2 to 25 miles wide extending from the northern to the southern boundary of Vermont (Hitchcock *et al.*, 1861, map; formation called "calciferous mica schist").

Lithologic character.—The Waits River formation (Currier and Jahns, 1941, p. 1491) is

TABLE 2.—ESTIMATED MODES OF THE WAITS RIVER FORMATION

	1	2	3	4	5	6	7	8	9	10	11	12	13
Number of sections	2	5	5	4	1	1	5	1	2	1	1	2	1
Quartz	76	41	41	32	59	17	27	48	40	10	88	44	70
Plagioclase	1	x	5	x	5	50	6	15	14	30			10
Carbonate	3	54	31	46			40		15	10			t
Muscovite	6	2	4							30	x	35	2
Biotite	12	2	18	10	3	20		8		15	10	17	15
Chlorite		x	x								1	1	3
Actinolite				12*	5	10	3	10	7*				
Clinozoisite					8	2	7	2	12	5			
Diopside							17	2	8				
Garnet									4		1	1	
Microcline					15		x	12					
Tourmaline		t	t		x		t				x	t	x
Sphene		t	t	t	x	1	x	1	x		x	t	x
Apatite			t					x			x	t	
Zircon		t		t	x		t	x	t		x		x
Carbon	2	1	1	x	3		t	1				2	
Metallic opaque		x	x	t	2	x	x	1	t		x	x	x
Per cent of anorthite in plagioclase	50	30	30–60	35–50	46	50	29–95	35	40–90	66			20

* Hornblende in one section
Grain size: Matrix, 0.1–0.5 mm.; porphyroblasts, 0.5–4.0 mm.
x—present in 50 per cent or more of thin sections
t—present in less than 50 per cent of thin sections
Blank—not seen in any of thin sections

1. Calcareous micaceous quartzite
2. Quartzose magnesian marble
3. Calcareous mica schist
4. Lime-silicate rock
5. Lime-silicate rock
6. Lime-silicate rock
7. Lime-silicate rock
8. Lime-silicate rock
9. Lime-silicate rock
10. Lime-silicate rock
11. Quartzite
12. Quartz-mica schist
13. Feldspathic quartzite

made up of interbedded calcareous and noncalcareous schists. The calcareous rocks, which constitute from 40 to 50 per cent of the whole formation, range from bluish-gray granular quartzose marble to gray calcareous quartz-mica schist. These rocks weather dark brown. The noncalcareous rocks are dominantly dark-gray to black quartz-mica schist and mica schist, with subordinate light-gray or buff micaceous quartzite and quartzite. The calcareous rocks are interbedded with the non-calcareous rocks in all proportions; some sections several hundred feet thick are dominantly calcareous, whereas others have little or no carbonate. Individual beds of either calcareous or noncalcareous rock range from a fraction of an inch to over 10 feet thick; the distinctive beds of quartzose marble are typically 1 inch to 2 feet thick.

Over most of the area, the dominant minerals in the formation are quartz, calcite, dolomite, biotite, and muscovite (Table 2). The calcareous rocks locally contain porphyroblasts of actinolite in the western part of the quadrangle, and diopside, clinozoisite, and anorthite are prominent porphyroblasts in highly metamorphosed marbles immediately adjacent to granodiorite in the northwest corner of the area. The lime-silicate rocks contain little or no dolomite, but the ratio of calcite to dolomite was not determined in the rocks that contain both.

The rocks are all medium- to coarse-grained and generally schistose. Schistosity is best developed in the more micaceous rocks, in which microscopic layers of muscovite and biotite separate discontinuous layers or lenses of quartz or quartz and calcite. Dimensional elongation of all minerals parallel to the banding is very good, but the crystallographic orientation of scattered porphyroblasts of actinolite and mica is practically random. In the more massive quartzose marble, schistosity is defined by a rude orientation of scattered mica plates and by dimensional elongation of calcite and some quartz grains.

Thickness.—Extreme structural complication within the area underlain by the Waits River formation, and the fact that only one of its stratigraphic boundaries is exposed here, make a reliable estimate of its thickness impossible. In the Barre quadrangle (Fig. 1), where the entire formation is exposed as a simple homoclinal belt, the distance between the upper and lower boundaries, measured normal to bedding, is at least 20,000 feet. Because of repetition of beds in minor folds, this distance may be as much as twice the sum of the present thicknesses of all the beds in the formation (Billings, 1937, p. 475; Currier and Jahns, 1941, p. 1503). The present thickness of the formation, corrected thus for folding, is of the order of 10,000 feet. This present thickness may be considerably less than the thickness at the time of deposition because of thinning of individual beds by rock flowage and shearing during deformation. The amount of such thinning, however, cannot be readily calculated here because of lack of fossils, amygdules, oölites, or other bodies whose original shapes are known.

Correlation.—The calcareous rocks of the western part of the Woodsville quadrangle are correlated with the Waits River formation of the Barre quadrangle (Currier and Jahns, 1941, p. 1491) primarily on the basis of lithologic similarity. The area between the Woodsville and Barre quadrangles is largely underlain by similar rocks, as shown by detailed geologic mapping in the East Barre and Strafford quadrangles (White and Eric, 1944; Doll, 1945; White and Jahns, 1950).

The Waits River formation is younger than the fossiliferous Shaw Mountain formation, which is no older than Middle Ordovician (Currier and Jahns, 1941, p. 1500–1502). Preliminary identification of recently discovered fossils from the Waits River formation near Montpelier (Fig. 1) suggests that it is Middle Ordovician (Cady, 1950). The Waits River formation continues northward into Canada, where it is called the Tomifobia formation and considered to be Middle Ordovician (Clark, 1934, p. 11).

Gile Mountain Formation

Distribution.—The Gile Mountain formation underlies a belt 2 or more miles wide that extends from the southwest corner of the quadrangle to the north-central part. Both the western contact with the Waits River formation and the eastern contact with the Meetinghouse slate are gradational.

Lithologic character.—The Gile Mountain formation is composed of thinly to thickly interbedded dark-gray to black schist (phyllite near the eastern border of the outcrop area) and light-gray micaceous quartzite, either of which may be locally dominant. The schist ranges from almost pure mica schist, with subordinate quartz, to quartz-mica schist. All gradations are found, furthermore, between quartz-mica schist and micaceous quartzite. Feldspathic rocks are not abundant and pure quartzite is rare. Typical outcrops of the formation reveal alternations of micaceous and quartzose beds. Where schist predominates, individual beds are almost everywhere less than 6 inches thick, and are commonly an inch or less thick. Where micaceous quartzite is the most abundant rock, it may occur in beds 3 or 4 feet thick separated by thin schist laminae.

Quartz and muscovite are the principal constituents of the formation as a whole, and small amounts of plagioclase and chlorite are present in most rocks (Table 3). Biotite, garnet, and staurolite are also common in rocks of appropriate metamorphic grade; the degree of metamorphism increases westward, and garnet, for example, is a common minor constituent of the rocks west of the garnet isograd (Pl. 1).

All the rocks of the Gile Mountain formation are schistose. In mica schist and phyllite, quartz and feldspar occur as isolated tabular grains and lenticular aggregates between al-

Table 3.—Estimated Modes of the Gile Mountain Formation

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	
Number of sections	8	4	9	2	6	5	3	2	6	3	5	4	2	1	
Quartz	63	21	64	82	51	22	46	81	51	18	60	84	25	8	50
Plagioclase	2	1	3	1	24	1	2	3	20				11	20	
Muscovite	29	73	20	11	8	58	32	7	7	40	14	1	25	1	3
Chlorite	6	1	4	x	5	8	8	x	1	4	5	1			
Biotite		4	9	6	11	7	10	7	19	24	17	11	30		
Garnet						4	2	1	2	5	2	1			
Staurolite										9	2	2	8		
Sillimanite													1		
Carbonate	t				1			1						70	47
Epidote				x	t				t						
Sphene	x	t	x		t	t	x	x	x	x	x	x	x	x	x
Apatite	t	t	x	x	x	t	x	x	x	x	x	x	x	x	t
Zircon	x	t	t	x	x	t		t	t		t	t	x		
Tourmaline	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x
Carbon	x	x	x		t	x	x	x	x	x	x	x	x	1	x
Metallic opaque	x		x	x	x	t	t	t	x		t	t	t	x	x
Per cent of anorthite in plagioclase	10	10		20	10	20	20–35	5–35	5–30				20–35	60	
Average grain size quartz (mm.)	0.1	0.07	0.2	0.1	0.2	0.05	0.1	0.2	0.2	0.3	0.3	0	0.3	0	0.4

x—present in 50 per cent or more of thin sections
t—present in less than 50 per cent of thin sections
Blank—not seen in any of thin sections

1. Phyllitic quartzite, chlorite zone
2. Phyllite, biotite zone
3. Quartzose phyllite, biotite zone
4. Micaceous quartzite, biotite zone
5. Feldspathic phyllite, biotite zone
6. Mica schist, garnet zone
7. Quartz-mica schist, garnet zone
8. Biotitic quartzite, garnet zone
9. Feldspathic schist, garnet zone
10. Mica schist, staurolite zone
11. Quartz-mica schist, staurolite zone
12. Biotitic quartzite, staurolite zone
13. Mica schist with sillimanite
14. Feldspathic marble, garnet zone
15. Calcareous quartzite, garnet zone

most perfectly oriented plates of muscovite and biotite. The quartz-mica schists are typically banded, with laminae of parallel mica plates separating thin layers or lenses of quartz or quartz and feldspar. In most rocks, this lamination is parallel to bedding, but inasmuch as it clearly cuts bedding in some places, it is considered to be of tectonic origin. In the micaceous quartzites, muscovite and biotite occur as isolated plates, oriented parallel to schistosity, in a granular matrix of quartz and subordinate feldspar; quartz grains in these rocks are much less tabular, as a rule, than quartz grains in the more micaceous rocks.

Garnet, staurolite, and part of the muscovite and biotite form randomly oriented porphyroblasts in the rocks in which they occur. Typically these minerals have replaced the rock bodily; micaceous laminae adjacent to these porphyroblasts are clearly transected rather than bulged out around them.

At the south end of the Woodsville quadrangle, west of Bradford Center, there are three lenses a few hundred feet thick in which 10 to 50 per cent of the rock is quartzose marble and calcareous schist (Pl. 1). Lithologically these calcareous rocks are essentially the same as those in the Waits River formation; as stratigraphic units, they differ from the Waits River in that a high proportion of the interbedded non-calcareous rock is micaceous quartzite. In Plate 1, a narrow tongue of Gile Mountain is shown west of the main body of the formation in the central part of the quadrangle. Whether this represents an attenuated fold, a fault slice, or a stratigraphic tongue is not clear.

Thickness.—The present thickness of the Gile Mountain formation is best measured near Goshen (Pl. 1), where the displacement on the Scotch Hollow thrust is believed to be small, and where the calculation is complicated by a minimum number of minor folds. The beds here are almost vertical, and the breath of outcrop is about 10,000 feet. The sum of the present thicknesses of all the beds in the formation is probably of the order of 6000 or 7000 feet, using a divisor of 1.5 to correct for repetition of beds in minor folds. This divisor is used because minor folds are rare in the easternmost part of the formation, and in the westernmost part the number of minor folds per unit area is comparable to the number in regions in which a divisor of 2 has seemed reasonable (Billings, 1937, p. 475; Currier and Jahns, 1941, p. 1503). As in the case of the Waits River formation, the amount of thinning of beds because of flowage during deformation cannot be calculated, so the original thickness at the time of deposition is not known.

Correlation.—The Gile Mountain formation has been traced by detailed mapping from the Woodsville quadrangle to Gile Mountain in the Strafford quadrangle (White and Eric, 1944; Hadley, 1950; Doll, 1945). The formation, as defined here, includes the rocks at the type locality and most of the rocks mapped by Doll as "Gile Mountain Schist" (Doll, 1945, p. 18–19) in the remainder of the Strafford quadrangle. It does not, however, include any of the rocks exposed in the southeast corner of the Strafford quadrangle east of the Ammonoosuc thrust that Doll mapped as Gile Mountain; these rocks are areally continuous along their strike with the Orfordville formation of the Mt. Cube quadrangle (Hadley, 1942; 1950, geologic maps). The formation is tentatively assigned to the Middle Ordovician.

Meetinghouse Slate

Distribution.—The Meetinghouse slate (Doll, 1945, p. 19) forms a narrow band that extends northward from the south-central edge of the quadrangle. It gradually narrows northward and disappears about a mile south of the north edge of the quadrangle.

Lithologic character.—The Meetinghouse slate is composed of thinly bedded dark-gray to black slate (southern part of quadrangle) or gray fissile schist (northern part), with very minor amounts of thinly interbedded light-gray micaceous quartzite. The slate or schist is dominantly micaceous, and quartz-mica slate or schist is subordinate. Lithologically the various rock types are indistinguishable in hand specimen from their counterparts in the Gile Mountain formation, and the two formations are separated entirely on the basis of the relative proportions of the micaceous and quartzose rocks.

The Meetinghouse typically contains about 10 per cent of arenaceous phyllite and micaceous quartzite, in beds an inch or less thick, interstratified with the darker argillaceous layers. The proportion of the arenaceous layers increases westward to about 40 per cent at the arbitrary western boundary of the formation, and the thickness of these beds increases correspondingly to as much as a foot.

Thickness.—An unknown amount of the Meetinghouse slate is cut out by the Monroe fault, so the total thickness is not known. The breadth of outcrop near Bradford, where the rocks dip very steeply, is about 2500 feet. This probably represents closely the sum of the present thicknesses of individual beds exposed here, as minor folds are uncommon.

Correlation.—The slate belt has been traced southward by detailed mapping to the type locality in the Strafford quadrangle (Hadley, 1950; Doll, 1945). It is tentatively assigned to the Middle Ordovician.

Orfordville Formation

Distribution.—The Orfordville formation is exposed: (1) just west of the Connecticut River at the south end of the quadrangle; (2) just east of the Connecticut River in the south end of the quadrangle; and (3) in a belt 4 miles long and 0.5 mile wide in the southeast corner of the quadrangle.

Lithologic character.—West of the Connecticut River, the Orfordville is dominantly dark-gray to black phyllite with thin beds of micaceous quartzite. Seventy feet west of the contact with the Albee formation, a bed of pebble conglomerate is 1 to 2 feet thick; the matrix is arenaceous, whereas the pebbles are quartzite, impure limestone, and phyllite. Two thousand feet south of Sawyers Ledge, both west of the highway and in an island in the Connecticut River, metamorphosed soda-rhyolite tuffs that may represent the Sunday Mountain volcanics have not been separately designated on the geological map (Pl. 1).

Just east of the Connecticut River in the south end of the quadrangle, the Orfordville is chiefly dark-gray to black slate, with some interbedded quartzite and slaty soda-rhyolite. The Hardy Hill quartzite member, a greenish-gray quartzite and quartz conglomerate, is 60 feet or less thick and lies about 1000 feet stratigraphically below the top of the formation. The Sunday Mountain volcanic member, composed of gray soda-rhyolite tuff and chlorite schist, is at the top of the formation and about 200 feet thick. An unusually good section of the Orfordville exposed along the southern border of the quadrangle is given in Table 4. The thicknesses are only approximate, as it is difficult to evaluate with precision the amount of repetition due to folding.

In the southeast corner of the quadrangle, the Orfordville lies in the staurolite zone. The main part of the formation is dark-gray to black mica schist and quartz-mica schist, with some thin quartzite beds. The Hardy Hill quartzite member occurs about 700 feet below the top of the formation. The Sunday Mountain volcanic member, at the top, is about 300 feet thick and is composed of interbedded fine-grained biotite gneiss and amphibolite; these rocks were originally volcanic tuffs and breccias with the composition of soda-rhyolite, andesite, and basalt. A stratigraphic section of this area is given in Table 5.

Detailed petrographic descriptions are given by Hadley (1942, p. 119–123).

Thickness.—Tables 4 and 5 indicate that about 1500 feet of the Orfordville formation are exposed in the Woodsville quadrangle. In the Mt. Cube quadrangle, where lower parts of the formation are exposed, Hadley (1942, p. 123) estimated a thickness of 5000 feet with a possible error of 30 per cent; the base of the formation was not exposed there.

TABLE 4.—STRATIGRAPHIC SECTION OF ORFORDVILLE FORMATION, WESTERN PART OF TOWNSHIP OF PIERMONT

Top of section is Dartmouth College Road at border of Woodsville and Mt. Cube quadrangles

		Thickness in feet (approximate only)
Albee formation	Greenish-gray slaty quartzite	2000
Orfordville formation	Tuff composed of gray slaty soda-rhyolite matrix, with pebbles of soda-rhyolite 1/2 inch long (*Sunday Mountain volcanic member*)	200
	Gray slate and quartzite	200
	Slaty soda-rhyolite tuff; quartz grains 1/8 inch across in some beds	100
	Gray slate and quartzite	225
	Thin-bedded black slate and quartzite	100
	Gap	200
	Calcareous green schist	30
	Quartz conglomerate (*Hardy Hill quartzite member*)	30
	Black slate	400
	Soda-rhyolite tuff; quartz grains 1/8 inch across	5
	Black slate	10

Base of section is edge of sand plain, 1/3 mile south of B.M. 454, western part of township of Piermont.

Correlation.—The rocks assigned to the Orfordville in the southeast corner of the quadrangle may be traced southwestward into the type locality in the Mt. Cube quadrangle (Fig. 2). The rocks in the vicinity of the Connecticut River in the southern part of the quadrangle are correlated with the Orfordville formation for three reasons: (1) they are stratigraphically beneath the Albee formation, as in the type locality, (2) prior to metamorphism, they were very similar lithologically to those in the type locality, and (3) the Sunday Mountain and Hardy Hill members appear at the appropriate stratigraphic position. This formation is unfossiliferous and in recent years has been tentatively assigned to the Middle Ordovician (Hadley, 1942, p. 124).

TABLE 5.—GENERALIZED STRATIGRAPHIC SECTION OF ORFORDVILLE FORMATION 1.5 MILES SOUTH OF PIKE

		Thickness in feet (approximate only)
Albee formation	Quartz-mica schist with porphyroblasts of biotite, garnet, and some staurolite	
Orfordville formation	Fine-grained amphibolite (*Sunday Mountain volcanic member*)	200
	Dark mica schist with porphyroblasts of biotite and garnet	500
	Quartz conglomerate (*Hardy Hill quartzite member*)	0–100
	Dark mica schist with porphyroblasts of biotite and garnet; actinolite schist, 5 feet thick, 600 feet below top of this member	800
Base of section is Northey Hill thrust		

Albee Formation

Distribution.—The Albee formation occupies much of the eastern half of the quadrangle. Because of a considerable difference in the grade of metamorphism a distinction should be made between a western and an eastern belt. The western belt, lying west of the Ammonoosuc thrust and averaging 4 miles in width, extends from the northeast corner of the quadrangle in a southerly direction to the south end of the quadrangle, where it splits into two tongues that die out a short distance to the south in the Mt. Cube quadrangle (Fig. 2). The eastern belt, east of the Ammonoosuc thrust, extends along the east-central and southeast margins of the quadrangle.

Lithologic character.—The Albee formation west of the Ammonoosuc thrust is in the chlorite zone, except for a long narrow triangular area in the northern part of the quadrangle lying west of the biotite isograd and east of the Monroe fault.

The rocks in the chlorite zone are chiefly phyllite, feldspathic phyllite, quartzose phyllite, argillaceous quartzite, quartzite, and feldspathic quartzite, all six types being about equally abundant. The yellowish-green to bluish-green color of many of these rocks distinguishes this formation from others in the quadrangle.

The phyllites are black, dark-green, or greenish-yellow schistose rocks composed chiefly of quartz, sericite, and chlorite (Table 6, column 1). A variety rich in albite-oligoclase is called feldspathic phyllite (Table 6, column 2). The terms phyllite and feldspathic phyllite are reserved for rocks containing 60 per cent or less of quartz or quartz plus feldspar (Billings, 1937, p. 475). Numerous outcrops are composed exclusively of phyllites or feldspathic phyllites in which the cleavage is well developed but the bedding is difficult or impossible to identify. Microscopic study shows that the sericite and chlorite occur either as small well-oriented, irregular flakes or as matted aggregates in bands parallel to the schistosity. Quartz and feldspar are in small grains scattered through the rock or are concentrated in thin laminae parallel to the schistosity.

Rocks containing 60 to 80 per cent quartz or quartz plus feldspar are termed quartzose phyllites if they posses a good schistosity but are called argillaceous quartzites if they are relatively massive. Such rocks are generally yellowish-green. They may be the only rock present in the outcrop or they may be interbedded with phyllite or quartzite. An average mode is given in Column 3 of Table 6. Some specimens are rich in feldspar, but modes of such rocks have not been prepared.

The quartzites are massive nonfoliated or weakly foliated yellowish-green, buff, or white rocks that contain 80 per cent or more quartz (Table 6, column 4). The feldspathic quartzites contain at least 10 per cent feldspar, the total quartz and feldspar constituting at least 80 per

Table 6. Estimated Modes of Albee Formation

	Chlorite Zone							Biotite Zone				Staurolite Zone					Sillimanite Zone	
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
Number of sections	10	6	2	10	4	1	2	7	2	3	1	2	1	2	1	1	3	1
Quartz	34	24	68	84	63	44	60	45	67	85	78	39	40	72	69	84	45	52
Plagioclase		36		3	23		8	t		1	15		25	5	15	5	9	30
Orthoclase						5												
Muscovite*	55	7	21	9	7	25	7	29	29			30	10	10	7	3	22	
Chlorite	10	28	10	3	6	20	18	15	1	8	3	x	3	7	6	1	1	
Epidote†	x	1	t			4	4	1		1	2							x
Carbonate	1	3	1	1	x			1										
Biotite								8	2	5	1	21	17	2		5	18	2
Garnet												2		x		1	x	4
Staurolite												6	5				x	
Sillimanite																	3	
Magnetite		x	x	x	1	2	2	x	1	x			x	2	3	1	2	
Ilmenite											x	x						
Pyrite						x	1	t			1	2		1				
Rutile						x												
Leucoxene				t	t													
Zircon		x		t				t		x				x				
Apatite		t		t	x	x		x	x	t	x	x	x	1		x		
Tourmaline	t	t	x	x	x		x	t	x			x		x			t	
Sphene	t	1		x	x			1	x	x								
Hornblende																		12
Carbon	t		x							t								
Per cent of anorthite in plagioclase		10		8	10		10				10		40	30	40	20	27	80

* Sericite in chlorite zone
† Includes clinozoisite
x—present in 50 per cent or more of thin sections
t—present in less than 50 per cent of thin sections
Blank—not seen in any of thin sections

1. Phyllite
2. Feldspathic phyllite
3. Quartzose phyllite and argillaceous quartzite
4. Quartzite
5. Feldspathic quartzite
6. Hornfels; Oliverian Brook, 3/4 mile north of Village of Haverhill.
7. Hornfels with porphyroblasts of chlorite retrogressed from biotite; Oliverian Brook, 3/4 mile north of Village of Haverhill.
8. Phyllite with biotite porphyroblasts
9. Mica-quartz schist with biotite porphyroblasts
10. Quartzite with biotite porphyroblasts
11. Feldspathic quartzite with biotite porphyroblasts
12. Staurolite-garnet-mica schist
13. Feldspathic staurolite-mica schist
14. Mica-quartz schist
15. Feldspathic mica-quartz schist
16. Quartzite
17. Sillimanite schist
18. Hornblende-bytownite schist

cent of the rock (Table 6, column 5). The quartzites are in beds an inch to 6 feet thick interbedded with phyllites, quartzose phyllites, and argillaceous quartzites.

In Oliverian Brook, 0.8 mile north-northeast of the village of Haverhill, the Albee has been converted to hornfels (Table 6, columns 6 and 7). In the rock listed in column 7 half the chlorite is in porphyroblasts from 0.4 to 0.6 mm. in diameter that contain numerous needles of rutile. This suggests that the chlorite is altered biotite.

The rocks of the Albee formation (Table 6, columns 8–11) lying west of the biotite isograd and east of the Monroe fault are similar to those described in the preceding paragraphs, except for the presence of biotite porphyroblasts 0.1 to 2 mm. in diameter.

The Albee is more highly metamorphosed east of the Ammonoosuc thrust than to the west. In general, the rocks lie in the staurolite zone and such minerals as biotite, garnet, and staurolite are present. Locally, around the French Pond granite, the rocks are in the sillimanite zone. In a belt nearly 1000 feet wide directly east of the Ammonoosuc thrust the rocks have undergone striking retrograde effects, described in the section on metamorphism.

The Albee formation in the staurolite zone consists of about equal amounts of mica schist, micaceous quartzite, and quartzite. Most of these rocks are white to light gray, rather than yellowish-green as in the lower metamorphic zones. Dark-gray types are distinctly subordinate in the main part of the formation.

The mica schists are light-gray to dark-gray schistose rocks with less than 60 per cent quartz or quartz plus feldspar. Most of these rocks have porphyroblasts of garnet or staurolite or both, and some varieties are feldspathic (Table 6, columns 12 and 13). The mica-quartz schists are light-gray schistose rocks with 60 to 80 per cent quartz or quartz plus feldspar (Table 6, columns 14 and 15). The quartzites, which contain over 80 per cent quartz or quartz plus feldspar, are white to gray massive rocks (Table 6, column 16). Some of the quartzites are crossed by narrow biotite-rich bands(Hadley, 1942, Pl. 3), and in many outcrops it is difficult to decide whether these bands are original bedding or fracture cleavage along which biotite has concentrated.

The Albee along the northwest margin of the French Pond granite and in the large inclusion in the granite lies in the sillimanite zone. The mica schists here are coarser-grained than corresponding rocks elsewhere in the Albee formation; muscovite flakes are not uncommonly 1 to 2 mm. in diameter. Sillimanite is not megascopically visible, but thin sections show that small needles 0.1 to 1 mm. long constitute 2 to 4 per cent of the rocks (Table 6, column 17). Modes of the associated quartzites and mica-quartz schists have not been determined.

The Piermont member of the Albee formation underlies an area of about 2 square miles south of the village of Haverhill. It consists of rocks identical with those found in the main part of the Albee formation east of the Ammonoosuc thrust, but the proportions are different. The main part of the Albee formation consists chiefly of light-colored rocks, notably quartzite, micaceous quartzite, and mica-quartz schist, with dark-gray staurolite-garnet schist, mica-quartz schist, and mica schist distinctly subordinate. In the Piermont member, the dark-gray types with biotite, garnet, and staurolite porphyroblasts are dominant. The garnet porphyroblasts are small, pink, euhedral crystals that seldom exceed a millimeter is diameter. The staurolite porphyroblasts, from 0.5 to 3 centimeters long, are euhedral, with the prism well developed but the terminal facies lacking; there are some cruciform twins. Although some staurolite crystals are solidly built, others are mere skeletons and include many other minerals poikilitically. The staurolite crystals are diversely oriented and it is apparent that the crystals formed in large part after the deformation had ceased.

Thickness.—Hadley (1942, p. 146) estimates the Albee formation to be about 5000 feet thick, but recognizes that an accurate calculation is impossible because of the intense minor folding.

Correlation.—The Albee formation west of the Ammonoosuc thrust may be traced continuously from the type locality in the Littleton quadrangle (Fig. 2). The belt east of the Ammonoosuc thrust is contiguous to rocks in the Moosilauke quadrangle that have been correlated with the Albee formation because of their lithologic characteristics and stratigraphic position (Fig. 2). The Albee formation is pre-Silurian and in recent years has been tentatively assigned to the Upper Ordovician (Billings, 1937, p. 475). The age is discussed more fully later.

Ammonoosuc Volcanics

The Ammonoosuc volcanics occupy a belt extending northeasterly from Woodsville. The contact with the Albee formation is not always easy to place, as there is a transition zone several hundred feet thick in which Albee and Ammonoosuc lithologic types alternate (Billings, 1937, p. 476). The metamorphism in this belt is low grade; that is, the rocks lie in the chlorite zone.

The principal rocks are fine-grained massive to schistose, and consist of soda-rhyolite, soda-rhyolite with bluish quartz grains an eighth to a fourth inch in diameter, soda-rhyolite volcanic conglomerate, and chlorite schist. These rocks are believed to have been volcanic tuffs and breccias; flows were distinctly subordinate. Slate is locally abundant; an unusually thick bed of slate is found at the base of the Ammonoosuc volcanics a mile N. 10°W. of Woodsville on the steep slopes leading up to Gardner Mountain.

As these rocks are identical with those in the Ammonoosuc volcanics in the Littleton-Moosilauke area (Billings, 1937, p. 475–478), detailed petrographic descriptions are not given in the present paper.

The Ammonoosuc volcanics are estimated to be about 2000 feet thick in the Littleton-Moosilauke area(Billings, 1937, p. 480). The belt of Ammonoosuc volcanics northeast of Woodsville is continuous with the type locality in the Littleton quadrangle (Fig. 2). The Ammonoosuc is pre-Silurian and in recent years has been tentatively assigned to the Upper Ordovician (Billings, 1937, p. 480).

Partridge Formation

The Partridge formation, covering about 0.8 square mile, 2.5 miles northeast of Woodsville, is chiefly black slate, although locally there are thin beds of fine-grained, light-colored quartzite. Inasmuch as the top of the formation is not exposed in the Woodsville quadrangle, the thickness cannot be calculated. Billings (1937, p. 481) estimates the maximum thickness to be 2000 feet, but in many places the formation is absent due to an unconformity at the base of the Silurian formations (Billings, 1937, p. 517–518). These rocks are continuous with rocks in the Moosilauke quadrangle that have been correlated with the Partridge formation of the type locality in the Littleton quadrangle. The formation is pre-Silurian and in recent years has been tentatively assigned to the Upper Ordovician (Billings, 1937, p. 481). The age is discussed more fully later.

Clough Quartzite and Fitch Formation

The Silurian Clough and Fitch formations are not present in the Woodsville quadrangle, but they appear in adjacent quadrangles (Fig. 2). The Clough formation, chiefly quartzite and quartz conglomerate, reaches a maximum thickness of 1200 feet in the Mt. Cube quadrangle (Fig. 2), but thins northward to disappear in the Littleton quadrangle; it is Early or Middle Silurian. The Fitch formation, chiefly limestone, calcareous slate, and dolomitic slate, or the metamorphosed equivalents, is 400 to 700 feet thick; fossils indicate that it is Middle Silurian. In the Littleton-Moosilauke area a major unconformity separates the Silurian rocks from the underlying formations (Billings, 1937, p. 517–518).

Littleton Formation

In the Woodsville quadrangle the Littleton formation is found only in the extreme southeast corner, where it is chiefly gray mica schist, mica-quartz schist, and micaceous quartzite; biotite and staurolite porphyroblasts are common, although in many outcrops they have been chloritized. Detailed descriptions are given elsewhere (Billings, 1937, p. 490–491; Hadley, 1942, p. 133–134). The micaceous quartzite was formerly quarried for whetstone that was fabricated at Pike.

That portion of the Littleton formation exposed in the Woodsville quadrangle cannot be more than 2000 feet thick, but the lower part of the formation, exposed between these rocks and the underlying Fitch formation in the southwest corner of the Moosilauke quadrangle (Fig. 2), is approximately 1500 feet thick. The total exposed thickness in this area would thus be 3500 feet. In adjacent areas, the formation is 5000 feet thick.

The schists in the southeast corner of the Woodsville quadrangle are continuous with rocks in the southwest corner of the Moosilauke quadrangle that have been assigned to the Littleton formation because of their stratigraphic position (Fig. 2). Northeast of the Ammonoosuc thrust in the Littleton quadrangle, the formation has been dated by fossils as Lower Devonian (Billings, 1937, p. 494–495).

Age of Formations

The relative age of all the rocks on opposite sides of the Monroe fault cannot be determined in the Woodsville quadrangle. However, since the mapping of the Woodsville quadrangle was initiated 16 years ago, significant data have been obtained elsewhere. In the Bellows Falls quadrangle (Fig. 1), Kruger (1946, p. 169) mapped what is here called the Gile Mountain formation as the lower part of the Orfordville formation. The Meetinghouse formation is not present. He concluded that the Waits River formation was overlain in orderly succession by the Gile Mountain, Orfordville, Albee, Ammonoosuc, Partridge, Clough, Fitch, and Littleton formations (Kruger, 1946, p. 175–178). Inasmuch as the Waits River is Middle Ordovician and the Clough is Lower or Middle Silurian, he considered the intervening formations to Middle or Upper Ordovician.

Professor John Lyons of Dartmouth College has recently mapped the geology of the Hanover quadrangle. Like Kruger, he has concluded that the Waits River formation is overlain in succession by the Gile Mountain, Meetinghouse, Orfordville, Albee, Ammonoosuc, and Patridge formations. Moreover, he concludes that all these formations cannot be older than Middle Ordovician and are pre-Silurian.

The conclusions of Doll (1943a; 1943b) that the Waits River formation is Silurian and/or Devonian and that the Gile Mountain and Meetinghouse formations are Devonian are here rejected. This problem will be discussed in a later publication.

Plutonic and Hypabyssal Rocks

General Statement

The plutonic and hypabyssal rocks of the Woodsville quadrangle have been assigned to two magma series. The following, placed as nearly as possible in chronological order, are assigned to the Late Devonian (?) New Hampshire magma series: (1) metamorphosed mafic dikes, including the Leighton Hill dike complex; (2) leucocratic dikes; (3) Bethlehem gneiss; (4) French Pond granite; (5) Haverhill granodiorite; and (6) Ryegate granodiorite. A few camptonite dikes are assigned to the Mississippian (?) White Mountain magma series.

Metamorphosed Mafic Dikes and Leighton Hill Dike Complex

Metamorphosed mafic dikes are present throughout the Woodsville quadrangle. East of the Ammonoosuc thrust, they nowhere constitute more than a few per cent of the bedrock. West of this thrust, they make up 15 per cent or more of the bedrock in much of the area underlain by the Albee formation. Areas where they are even more abundant, making up 20 to 50 per cent of the bedrock, are shown on Plate 1 by a special overprint. These areas are located as follows: (1) a belt 2 miles long north of the village of Haverhill; (2) a belt 3 miles long north of the village of Newbury; and (3) a belt 13 miles long that extends from the north edge of the quadrangle along the east side of the Monroe fault as far south as Leight on Hill, whence it trends southeasterly to beyond Town Farm Hill.

In an area of extraordinary dike intrusion extending south from Boltonville, distinguished as the Leighton Hill dike complex (Pl. 1), all the outcrops are metamorphosed mafic rocks, and no exposures of metasediments were encountered. Outcrops a few tens of feet apart generally contain rocks of different structure and texture. Locally contacts between different types may be observed. For this reason the rocks in this area are believed to represent a dike complex rather than a single massive body of metamorphosed igneous rock. The eastern and southern boundaries of the area as mapped are gradational, but the western boundary is abrupt.

The metamorphosed mafic dikes are uncommon west of the Monroe fault. A few were observed in the eastern part of the Gile Mountain formation, but farther west they are rare.

The dikes range in width from a fraction of an inch up to as much as 50 feet. Most of them are parallel to the schistosity of the adjacent country rock. They characteristically cross-cut the bedding. Although a few appear to be folded, many of them truncate the fold axes, or lie parallel to the axial planes of folds (Fig. 3).

The mafic dikes include a wide diversity of rock types, ranging from chlorite schist through actinolitic greenstones to amphibolite; textures are massive to schistose, and colors range from gray or dark-green to black. Billings and White (1950) concluded that all these rocks were originally of basaltic composition, and that the present variety of rock types is the result of differences in the thermal metamorphism and of changes in chemical composition during metamorphism, chiefly variations in the amount of introduced carbon dioxide.

Leucocratic Dikes

About 50 schistose to massive, gray or brown micaceous dikes were seen in the Gile Mountain formation. Most of these dikes are about 4 feet wide, but locally they are over 20 feet wide. They are essentially parallel to the bedding and may be sills. One appears to be folded. Essential minerals are sodic oligoclase, quartz, and muscovite, with subordinate biotite and carbonate. Inasmuch as some of these dikes are schistose, they are either pre-tectonic or syntectonic.

Bethlehem Gneiss

The Bethlehem gneiss covers approximately 0.5 square mile along the east side of the quadrangle a mile north of Pike. This is the southwestern extremity of a body 11 miles long and 1 to 2 miles wide that extends northeasterly into the Moosilauke quadrangle (Fig. 2). The Bethlehem gneiss is a gray, foliated, granulated quartz monzonite or granodiorite, composed on the average of about 30 per cent quartz, 30 per cent oligoclase-andesine, 20 per cent potash feldspar, 17 per cent biotite, 3 per cent muscovite, and traces of apatite, sphene, and pyrite. More complete descriptions and chemical analyses may be found elsewhere (Billings, 1937, p. 504–506).

Between Pike and the southern border of the quadrangle, the Littleton formation is intruded by several sills 1 to 5 feet thick. These sills are composed of a weakly foliated, fine-grained, granulated gray rock that contains biotite flakes a fraction of a millimeter in diameter. Microscopic study shows that the texture is granoblastic, with a grain size of about 0.1 mm. The mode of the one thin section studied is: oligoclase (An_{20}), 50 per cent; quartz, 39 per cent,

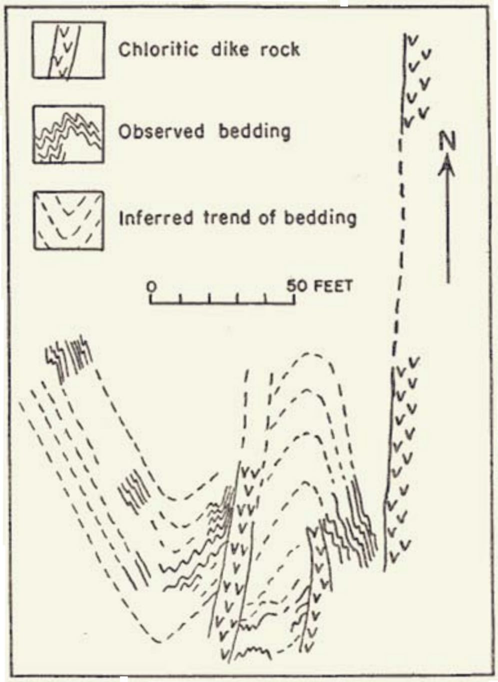


FIGURE 3.—MAFIC DIKES PARALLEL TO THE AXIAL PLANES OF FOLDS
Outcrops on slope 2000 feet south-southeast of Harriman Pond. Minor folds plunge 75°S. 10°E.

biotite, 8 per cent: chlorite, 2 per cent; magnetite, 1 per cent; apatite, tr; allanite, tr. Hadley has described similar rocks immediately to the south in the Mt. Cube quadrangle and considers them to be an early phase of the Bethlehem gneiss (Hadley, 1942, p. 146–147). He has also shown that they cut the amphibolite sills (Hadley, 1942, Fig. 4).

French Pond Granite

The bedrock exposed over an area of about 3 square miles to the north and east of North Haverhill is French Pond granite. Similar rocks extend eastward for 2 miles into the Moosilauke quadrangle (Fig. 2). To the south of the exposures of granite in both the Woodsville and Moosilauke quadrangles, there is a large area, 2.5 miles east-west and 1 to 1.5 miles north-south, completely covered by glacial drift. The distribution of glacial eratics, the structure of the adjacent schists, the distribution of small dikes and sills of French Pond granite in the surrounding schists, and the intensity of the metamorphism give clues, however, to the approximate location of the southern boundary of the granite. A map of the whole stock has been published elsewhere (Billings, 1945, p. 58).

The French Pond granite is a heterogeneous unfoliated rock. The most extensive variety in the Woodsville quadrangle is a porphyritic to coarse-grained biotite granite. The phenocrysts, composed of potash feldspar, are 1 or more centimeters long; locally they are so abundant that the groundmass disappears and the rock becomes a coarse-grained granite. A second and less common variety is pink or gray, medium-grained to fine-grained biotite granite. These two varieties of the French Pond granite, separately designated on Plate 1 have been fully described elsewhere (Billings, 1937, p. 509). A third variety, found in only a few outcrops 0.5 mile due north of B.M. 497, north of North Haverhill, is a medium-grained biotite-quartz syenite with the following average mode; potash feldspar, 68%; biotite, 18%; milky quartz, 10%; sphene, 2%; and epidote, 2%.

Haverhill Granodiorite

The Haverhill granodiorite lies in the southern part of the quadrangle in the vicinity of Black Hill. The main body is oval in plan, with a north-northeast axis 1.2 miles long, and a minor axis 0.4 mile long. Small sills, however, are found to the west and south. Exposures are very good on the west, north, and east slopes of Black Hill, but elsewhere they are poor. There are two abandoned quarries within this body, one on the east slope of Black Hill and the other on the northwest slope.

The Haverhill granodiorite is typically a white granular rock composed of white feldspar, milky quartz, biotite, and muscovite. Although individual grains do not exceed 3 mm. in diameter, even a casual inspection with the hand lens shows that the texture is granoblastic and that the original grains may have been as much as 10 mm. in diameter. Generally massive, the rock locally possesses a weak foliation that strikes N. 10° E. and dips 70° W.

A much finer-grained and grayer facies is exposed on the eastern margin of the body and in sills to the west and south of the main body. In this finer facies the individual grains are

TABLE 7.—ESTIMATED MODES OF HAVERHILL GRANODIORITE

	1	2	3	4	5
Quartz	30	31	26	35	31
Plagioclase	42	43	52	45	45
Microcline	15	15	5	7	10
Biotite	2	4	10	5	5
Muscovite	3	3		5	3
Epidote	3	3	3		2
Sphene	1	1	1		1
Chlorite	4			3	2
Zircon			tr		tr
Magnetite			3	tr	1
Anorthite content of plagioclase	30	30	35	27	30

tr—present but less than 1/2 per cent

1. Main phase, altitude of 840 feet on west slope of Black Hill.
2. Main phase, altitude of 990 feet on west slope of Black Hill.
3. Contact phase, quarry at altitude of 1180 feet on east slope of Black Hill.
4. Sill in Piermont member of Albee formation, altitude of 800 feet on west slope of Black Hill.
5. Average

0.25 mm. to 1.0 mm. in diameter. This phase, particularly well exposed in an abandoned quarry at an altitude of 1180 feet on the east slope of Black Hill, appears to grade westward into the more typical main phase.

Four representative modes are given in Table 7. All four individual specimens fall within the classification of granodiorite. Microscopic study shows that the texture is essentially granoblastic, although locally there is a suggestion of hypidiomorphic granular texture. The quartz grains commonly show strain shadows. The texture thus indicates that the rock has been subjected to post-consolidation deformation, accompanied by granulaton and recrystallization.

Hadley has described Haverhill granodiorite from two small intrusive bodies in the Mt. Cube quadrangle (Hadley, 1944, p. 148).

TABLE 8.—MODES OF THE RYEGATE GRANODIORITE AND ASSOCIATED ROCKS
(Rosiwal analyses, volume per cent)

	1	2	3	4	5	6	7	8	9	10	11	12
Number of sections	5	1	1	1	7	9	2	5	4	3	1	1
Quartz	26	22	24	16	27	25	30	25	27	36	22	14
Plagioclase	51	52	64	47	61	54	36	58	45	35	44	51
Microcline	15	12	9	28	2	13	24	x	14	24	20	23
Biotite	6	12	2	6	9	7	3	10	7	3	12	9
Muscovite	2	2	1	3	1	1	7	6	7	2	2	3
Chlorite	x		x		t	x	t	x	t	t		
Allanite	x	x	x	x	x	x		x	t	x		
Epidote	x	x	x	x	x	x	t	t	t	x		
Sphene	x	x	x	x	x	x	t	x	x	x	x	x
Apatite	x	x	x	x	x	x	x	x	x	t	x	x
Zircon	t			x		t			t	t	x	
Carbonate						t		1	t			
Metallic opaque						t	t	x	t	x	x	
Per cent of anorthite in plagioclase	20	26	24	22	27	26	16	28	23	13	15	30

Grain size: matrix 0.1–0.5 mm.; phenocrysts of quartz and plagioclase, 1.0–5.0 mm.; microcline in porphyry, phenocrysts up to 15.0 mm.

x—present in 50 per cent or more of thin sections.

t—present in less than 50 per cent of thin sections.

Blank—not seen in any of thin sections.

1. Granodiorite, Blue Mountain
2. Fine-grained foliated granodiorite, Blue Mountain
3–4. Porphyritic granodiorite, Blue Mountain
5. Tonalite (quartz diorite), Groton area
6. Granodiorite, Groton area
7. Leucocratic quartz monzonite, Groton area
8. Tonalite dikes in Waits River formation
9. Granodiorite dikes in Waits River formation
10. Leucocratic quartz monzonite dikes in Waits River formation
11. Barre "granite", Rock of Ages quarry (Maynard, 1934, p. 147)
12. Barre "granite", Pirie quarry (Shimer, 1943, p. 1062)

Ryegate Granodiorite

Granodiorite is the most abundant rock in the complex of small plutons here called the Ryegate granodiorite. Associated rocks range in composition from quartz diorite to quartz monzonite.

These plutonic rocks are typically hypidiomorphic granular aggregates of oligoclase (near andesine) and quartz, with subordinate biotite and microcline. A small amount of the granodiorite is porphyritic, with microcline phenocrysts. Muscovite is present in small amounts and sphene, apatite, and zoned allanite are typical accessories. Tourmaline was not seen in any thin-sections. Epidote and chlorite are common secondary alteration minerals. Modes of the rocks, calculated on a Rosiwal stage, are presented in Table 8.

The plagioclase occurs as idiomorphic grains 1 to 5 mm. across, and quartz as allotriomorphic grains almost as large. A finer-grained equigranular aggregate of the same minerals surrounds the larger crystals and contains the biotite, muscovite, and accessories. The larger plagioclase crystals are zoned, and commonly the core contains 10 to 15 per cent more anorthite than the outer zone. Microcline is interstitial to all other minerals, except in the porphyritic rocks. In these it forms large idiomorphic phenocrysts up to 15 mm. in size. The typical interstitial character of the microcline suggests that it is always a late mineral, and it is possible that the phenocrysts are formed by replacement at a late magmatic stage. Plagioclase in contact with microcline is myrmekitic.

Some of the rocks are foliated, and in these the degree of orientation of the biotite determines the perfection of the foliation. This planar element is generally best developed near contacts and is parallel to them, even where contacts cut across the foliation of the country rock. There is no evidence of cataclastic texture or augen structure, and the foliation is therefore interpreted as a primary flow structure.

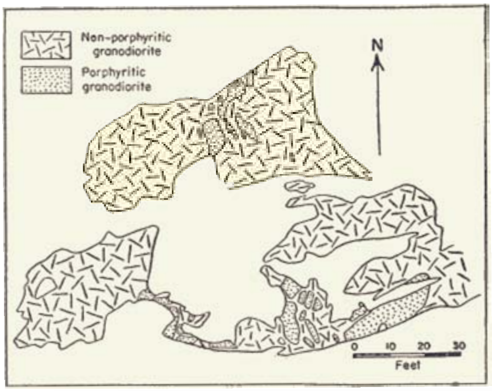


FIGURE 4.—INCLUSIONS OF PORPHYRITIC GRANODIORITE IN NON-PORPHYRITIC GRANODIORITE

Blank areas are glacial drift. Outcrop at altitude 1780 feet on the north ridge of Blue Mountain.

The Ryegate granodiorite forms small plutons in two general areas, distinguished here as the Blue Mountain area and the Groton area. Blue Mountain is underlain by several subconcordant bodies of non-porphyritic granodiorite with subordinate amounts of porphyritic granodiorite. The two types have been separated in Plate 1. The porphyritic rock is found as distinct inclusions in the other (Fig. 4), and is therefore older.

A very poorly exposed complex of calcareous schists and plutonic rocks underlies the large drift-covered area, here called the Groton area, west of South Ryegate (Pl. 1). Outcrops are too few to justify drawing continuous boundaries between plutonic and metamorphic rocks. The greater part of this area is believed to be underlain by massive granodiorite.

The plutonic rocks of the Groton area include several distinct types, but little is known of their areal distribution because of the paucity of outcrops. The oldest intrusive rocks are coarse, slightly foliated granodiorite and quartz diorite. The commonest rock is a massive coarse-grained to medium-grained granodiorite or quartz diorite, identical in appearance with the nonporphyritic rocks of Blue Mountain. Locally the massive rock cuts the foliated, but the two types may well be penecontemporaneous. In both types, all gradations are found from quartz diorite to a granodiorite near quartz monzonite in composition.

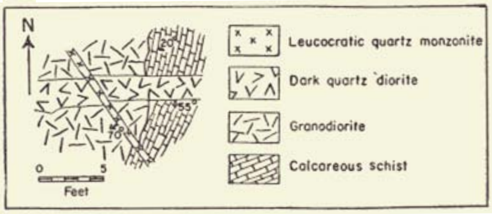


FIGURE 5.—TYPICAL RELATIONS OF GRANODIORITE AND ASSOCIATED DIKES IN THE GROTON AREA

Outcrop in the bed of the South Branch of the Wells River at altitude 1000 feet. Bedding dips 20°ESE. Arrows with numbers give direction and amount of dip of dikes.

The coarse-textured rocks are cut by dikes, up to 5 feet wide, of dark fine-grained quartz diorite, and these in turn are cut by very light-colored quartz monzonite dikes (Fig. 5).

Dikes closely related to these plutonic rocks in mineral composition, texture, and structural relations are found throughout the Waits River and Gile Mountain formations in the Woodsville quadrangle. Most of them are less than 10 feet wide, but one body about 500 feet wide was found a mile east of Ricker Mills. Areas in which these dikes are particularly abundant are distinguished by an overprint in Plate 1.

The close similarity between the Ryegate granodiorite of the Woodsville quadrangle and the Barre "granites," exposed in the northwest part of the East Barre quadrangle (Fig. 1), is suggested by the comparative modal analyses presented in Table 8. Spectrographic analysis of minor elements by Shimer (1943, p. 1055, 1057, 1060) shows an even closer correspondence between rocks from the two areas. That the two are comagmatic seems beyond serious question, inasmuch as both immediately followed the principal orogeny of the region (Balk, 1927, p. 68).

Camptonite

Nine post-metamorphism lamprophyre dikes were found in the area. All are less than 3 feet across. Six were intruded into granodiorite. All but two strike within 8° of N. 68°E., and all dip within 20° of vertical.

Four are camptonites, in which anhedral crystals and aggregates of augite occur in a finer-grained matrix of idiomorphic grains of brown hornblende and plagioclase. Magnesite (possibly dolomite) and serpentine occur as pseudomorphic aggregates after olivine, and as scattered irregular interstitial material throughout the rock

Camptonite dikes in the Littleton-Moosilauke area were considered by Billings (1937, p. 517) to be related to the White Mountain magma series; the dikes of the Woodsville area are undoubtedly the same.

Relative Ages of Plutonic and Hypabyssal Rocks

The arrangement of the plutonic and hypabyssal rocks in chronological order is based primarily on their relation to metamorphism and the extent to which they have been deformed. Only rarely do the rocks belonging to different map units come into contact. The metamorphosed mafic dikes are considered to be oldest because they are older than the metamorphism. The leucocratic dikes are also among the oldest because they are schistose. Whether they are older or younger than the metamorphosed mafic dikes is uncertain. The Bethlehem gneiss is younger than the metamorphosed mafic dikes (Hadley, 1942, Fig. 4, p. 147) and is considered to be syntectonic (Billings, 1937, p. 536–538). The French Pond, Haverhill, and Ryegate are late tectonic or post-tectonic, as they are relatively undeformed, but their age relative to each other is uncertain.

Structure

General Statement

Inasmuch as plutonic rocks are distinctly subordinate to stratified rocks in the Woodsville quadrangle, the structural pattern is dominated by folds and faults. Three master faults—the Northey Hill, Ammonoosuc, and Monroe—trend north-northeasterly and divide the area into four major tectonic units (Fig. 2; Pl. 1).

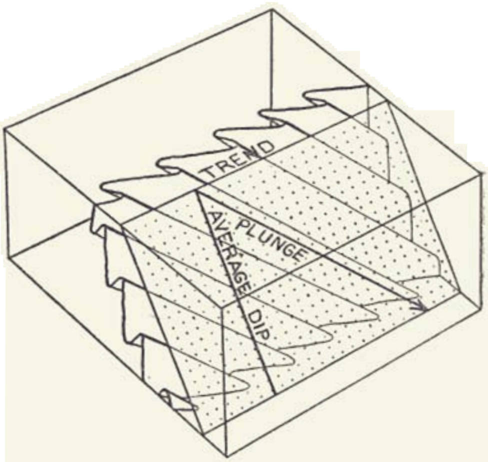


FIGURE 6.—STEREOGRAM OF FOLDED BEDDING
Sketch illustrates the use of terms "trend" and "average dip" in the description of folded bedding.

Because these three faults are so significant, they will be described first. The structure within each of the three easterly tectonic blocks is much more easily understood if a larger region is studied and for this reason Figure 2 has been prepared. Although minor folds and cleavage are present throughout the quadrangle, they were extensively studied only west of the Connecticut River.

Terminology

Trend and average dip.—In highly folded rocks the bedding strikes in various directions in different parts of an outcrop, and the only significant measurement that can be made is on the "trend" of the bedding (Balk, 1936, p. 702). The trend of bedding, as used here, is the strike of a plane that is tangent to the noses of the minor anticlines shown by a single bed. The dip of this plane is here called the "average dip" (Fig. 6). Many of the symbols for the attitude of bedding in Plate 3 actually represent trend and average dip, where no uniform strike could be measured.

Dextral and sinistral fold patterns.—The "pattern" of minor plunging folds refers to the zig zag trace of a folded bed in an exposure (Fig. 7). Where opposite limbs of folds are of unequal length, two types of fold pattern are possible, and each type is the mirror image of the other. It is proposed to distinguish the two by the terms "dextral" and "sinistral" (Fig. 7). In asymmetric folds with long and short limbs, the short limb may be regarded, for the purpose of definition, as an offset in the general course of bedding on the long limbs. If, as the bedding is followed away from a given point, the bedding is offset to the right, the fold pattern is dextral; if the offset is to the left, the pattern is sinistral. This terminology was suggested by Anderson's (1942, p. 55) proposed usage for transcurrent faults. If the terms should be used for fold patterns as viewed in cross-section, the orientation of the section and the point of view must be stated. The pattern in plan is meant here unless a section is specified.

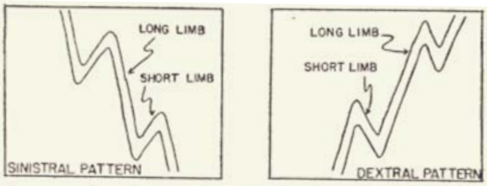


FIGURE 7.—PATTERNS OF FOLDED BEDDING IN PLAN

Illustrates use of "dextral" and "sinistral" patterns.

Cleavage.—"Cleavage" is used throughout this paper to refer to the general capacity of rocks to part along essentially parallel surfaces of secondary origin. It does not include parallel joints of the order of a foot or more apart, nor bedding fissility, nor the primary flow structure of igneous rocks, which are included in the broader terms "foliation," or "foliate structures." Two prominent types of cleavage, schistosity and slip cleavage, are recognized in the Woodsville quadrangle; "cleavage" is used to refer to both collectively.

Schistosity.—"Schistosity" here means the "tendency to a common parallel orientation of the crystal elements," an arrangement that "imparts to the rock a schistose or fissile property" (Harker, 1939, p. 194). The rocks here called "schists" commonly display also a lamination of more and less micaceous layers that may generally be of tectonic origin (see "gneissic structure" of Swanson, 1941, p. 1447); but, as used here, "schistosity" refers primarily to the planar parallelism of individual mineral grains.

Slip cleavage.—"Slip cleavage" (Dale, 1899, p. 209) is a direction of splitting, transverse to schistosity, that is due to the tiny crinkles or microfolds on the schistosity surfaces of foliate rocks. Rocks may cleave along the limbs of these microfolds, on surfaces parallel to the axial planes, whether or not the limbs are actually faulted; but in its most characteristic development, slip cleavage follows the tiny faults or foliation planes that form where the limbs of microfolds are sheared off. The nature of this cleavage in east-central Vermont, and the reasons for using "slip cleavage" instead of "fracture cleavage" for the phenomenon, have been discussed elsewhere (White, 1949).

Major Faults

Northey Hill thrust.—A large fault extends north-south through the village of Pike. The schists in the southeast corner of the quadrangle belong to the Littleton formation, because they overlie the Clough and Fitch formations exposed in the southwest corner of the Moosilauke quadrangle (Fig. 2). A little more than a mile west of the east edge of the Woodsville quadrangle the Albee formation is exposed. East of the Albee formation are rocks that have been assigned to the Orfordville formation. A large fault, with a stratigraphic throw of 12,000 feet, must lie between the Orfordville and the Littleton formations. It is not easy to locate this fault, however, because the schists of the Orfordville and Littleton formations are lithologically similar. The fault is placed between black schists on the west, characteristic of the Orfordville formation, and gray schists to the east, characteristic of the Littleton formation.

About 0.7 mile north-northwest of Pike, the fault coincides with the northwest contact of the Bethlehem gneiss. The northwest contact of this intrusive body may thence be traced northeasterly for 11 miles (Fig. 2). Here, as a direct continuation of the northwest contact of the Bethlehem gneiss, a large fault emerges and may be traced northward to its type locality at Northey Hill in the Moosilauke quadrangle

(Billings, 1937, p. 530–531). The large fault in the southeast corner of the Woodsville quadrangle is therefore considered to be the Northey Hill fault.

The relatively straight course of the Northey Hill fault in the Woodsville quadrangle, despite a relief of over 500 feet, implies that it is essentially vertical. Where observed at one locality in the Moosilauke quadrangle the dip is 75° NW. (Billings, 1937, p. 530). The fault is considered to be older than the metamorphism, because the metamorphism is of equal grade on both sides. Moreover, in the Moosilauke quadrangle, staurolite porphyroblasts within a few feet of the fault are undeformed. Brecciation, mylonitization, and silicification have not been observed along the fault in the Woodsville quadrangle.

Ammonoosuc thrust.—The Ammonoosuc thrust may be traced southwesterly from a point on the east side of the quadrangle 3 miles northeast of Woodsville to the south edge of the quadrangle 2 miles south of the village of Haverhill. Throughout most of its extent the fault is on the New Hampshire side of the Connecticut River, but in two plaecs, one 1.5 miles south of Woodsville and the other 1.5 miles west of the village of North Haverhill, it is on the Vermont side of the river.

The fault was first recognized in the Littleton and Moosilauke quadrangles because of a sharp stratigraphic discontinuity where strata high in the Littleton formation are in contact with strata well down in the Ammonoosuc volcanics. There the stratigraphic separation is of the order of 7000 feet. A great deal of mechanical evidence for the fault was also discovered there; not only was the fault plane observed in several localities, but locally there is considerable mylonitization and silicification. The silicified zones were especially helpful in tracing the fault. Moreover, a sharp discontinuity in the grade of metamorphism characterizes the trace of the fault.

In the Woodsville quadrangle, stratigraphic evidence for the fault is found only in the area 2.5 miles northeast of Woodsville, where the top of the Ammonoosuc volcanics is in contact with rocks well down in the Albee formation. South of Woodsville, the Albee formation lies on both sides of the fault and stratigraphic evidence is lacking. It is therefore necessary to use the other criteria to trace the fault.

Especially pertinent data are found along the valley of Burton Brook 1.5 miles northeast of Woodsville. The valley of Burton Brook has been excavated to a depth of 100 to 200 feet for a mile along the trace of the fault. A third of a mile north of Lake Gardner, the footwall of the fault has been silicified for 2000 feet along the strike. The thickness of the silicified zone ranges from 6 to 30 feet. The average strike of the top of this silicified zone is N. 30°E. The dip of the top of the silicified zone, presumed to be the fault plane, was observed at four different localities to be 32°, 36°, 38° and 40° toward the northwest; the average is 36° NW. Near the southwest end of the silicified zone, near the road, the top of the silicified zone is a wavy surface; the wavelength of these gentle rolls is 2 to 3 feet, and the amplitude is half a foot. These rolls, plunging directly down the dip of the surface, indicate that the displacement along the fault was essentially a dip-slip movement with no strike-slip component. The evidence that the hanging wall has moved up relative to the footwall has been discussed elsewhere (Billings, 1937, p. 529).

For 3.5 miles south of Woodsville, the trace of the fault can be located only approximately, chiefly because of glacial deposits and alluvium along the Connecticut River. On the Vermont side of the river, for 1.5 miles south of Horse Meadows, the trace of the fault has been accurately located. Although the rocks on both sides of the fault belong to the Albee formation, the rocks on the east side of the fault are in the staurolite zone whereas those of the west side are in the chlorite zone.

From a point 1.8 miles southwest of the village of North Haverhill, the fault can be readily traced southward for 2 miles. The rocks east of the fault are in the staurolite zone; those on the west side are in the chlorite zone. Two miles southwest of North Haverhill, a large silicified zone occurs along the fault. This silicified zone, the north end of which lies 0.2 miles south of the Dartmouth College Road, strikes N.15°W., is 150 feet long, and 75 feet wide. The east side of the zone is a network of quartz veins, and the whole zone is apparently a silicified breccia.

For nearly 0.5 mile to the south, the trace of the fault lies in the bottom of a narrow valley. For 2 miles south of the railroad north of Oliverian Brook, the trace of the fault is buried under glacial drift. The southernmost mile of the thrust in the Woodsville quadrangle is accurately located because of the contrast in metamorphism on opposite sides of the fault.

The retrograde metamorphism associated with the fault will be described below.

Monroe fault.—The rocks of the Albee and Orfordville formations are separated from those of the Meetinghouse slate and Gile Mountain formation along a fault that crosses the quadrangle from north to south. This fault has been named the Monroe fault (Eric, White, and Hadley, 1941). Although the strike of the bedding and schistosity next to the fault are essentially parallel to it and none of the outcrops of the contact suggest a structural break, there is evidence to prove that the boundary between the formations is a fault.

First, in the St. Johnsbury quadrangle (Fig. 2), Ammonoosuc volcanics on the east side of the fault are in contact with the Gile Mountain formation on the west. In the southern part of the Woodsville quadrangle and in the Mt. Cube quadrangle, the Orfordville formation on the east side is next to the Meetinghouse slate on the west. Moreover, the lower part of the Orfordville formation is missing at the contact. In most of the Woodsville quadrangle, phyllite of the Albee formation on the east is the dominant rock in contact with the Meetinghouse slate and Gile Mountain formation on the west, but locally, at Boltonville, quartzite predominates. Second, Eric has found detailed evidence of truncation of bedding in the Vermont portion of the Littleton quadrangle (Eric, White, and Hadley, 1941). Lastly, the biotite isograd is slightly offset in the vicinity of Boltonville.

The trace of the fault has been continuously mapped for 72 miles and probably extends much farther. Throughout most of the Woodsville quadrangle, the Orfordville formation has been cut out along the Monroe fault. Consequently the stratigraphic throw is here equal to the thickness of the Orfordville formation, approximately 5000 feet. The west side is upthrown, but the possibility of a reversal of movement subsequent to the metamorphism is discussed later.

Inasmuch as the strike of the bedding and the schistosity on the two sides of the fault are almost parallel to the strike of the fault, it is inferred that the dip of the fault is equally concordant with the dip of the bedding and schistosity (Pl. 2). Such parallelism is actually tectonic, because, on a regional scale, the contact is crosscutting. If the fault dips parallel to the bedding, the dip is 70°W. at the extreme north end of the area, vertical in most of the north half of the area, and 60°E. at the south end.

Structure East of Northey Hill Thrust

The Owls Head dome is a large dome in the southwest corner of the Moosilauke quadrangle and the northwest corner of the Rumney quadrangle (Fig. 2). The core of the dome is occupied by the Owls Head granite of the Oliverian magma series. On the west flank of the Owls Head dome, the Ammonoosuc, Clough, Fitch, and Littleton formations dip westward. The Littleton formation in the southeast corner of the Woodsville quadrangle is part of the west flank of this dome.

The Bethlehem gneiss a mile north of Pike is the southwest end of a body 11 miles long and 1 to 2 miles wide that extends northeasterly into the Moosilauke quadrangle (Fig. 2). This body, termed the Green Mountain pluton, is a concordant sheet, in the northern half of which the contacts and foliation are vertical. Toward the south, however, the contacts and foliation dip about 75°NW; in the Woodsville quadrangle the foliation and presumably the contacts dip about 60°NW. The pluton thus appears to be a huge sheet that is vertical in its northeastern half but toward the southwest gradually rolls over to dip 60°NW.

Structure between Northey Hill and Ammonoosuc Thrusts

That portion of the quadrangle east of the Ammonoosuc thrust is underlain chiefly by the Albee formation, except in the area occupied by French Pond granite. In the southern part of the quadrangle, however, the appearance of the Orfordville formation between the Albee formation and the Northey Hill fault indicates that the trough of a syncline lies somewhere to the west of the Northey Hill fault. The geological map of the Mt. Cube quadrangle (Fig. 2) (Hadley, 1942, Pl. 4) shows that the trough of the syncline lies about a mile west of the Northey Hill fault; in the southern end of the Woodsville quadrangle it would lie between Morris Brook and Catamount Ridge. The Piermont member of the Albee formation, exposed over a large area on the west limb of the syncline, pinches out eastward and does not appear on the east limb.

In the Moosilauke quadrangle this fold has been called the Salmon Hole Brook syncline.

The regional map-pattern of this fold in the Mt. Cube, Woodsville, and Moosilauke quadrangles (Fig. 2) suggests a syncline plunging toward the northeast. However, the minor folds in the Moosilauke quadrangle plunge steeply southwest, indicating an inverted plunge (Billings, 1937, p. 522). Throughout most of the Woodsville quadrangle the minor folds in this syncline plunge northwest, west and southwest; the axial planes dip steeply northwest and west. The average of all readings indicates a steep plunge to the west; thus, in general, the plunge in the Woodsville quadrangle appears to be inverted. But to the south, in the Mt. Cube quadrangle, the plunge becomes northeast and is thus normal. In the southeast corner of the Woodsville quadrangle, the Orfordville formation plunges southward beneath the Albee formation.

The Black Hill pluton, composed of Haverhill granodiorite, is a concordant lens 1.2 miles long and 0.4 mile wide that dips steeply to the west. The eastern contact, well exposed in a quarry at an altitude of 1180 feet on the east slopes of Black Hill, strikes N. 30°E. and dips 54°NW. parallel to the foliation in the rocks on both sides. The foliation throughout most of the pluton dips 75°W. This is true even near the western contact, and indicates that this contact dips at a similar angle. For 100 to 200 feet west of the main body, sills of Haverhill granodiorite 5–10 feet thick intrude the schists of the Piermont member of the Albee formation. On the ridge 1 mile south-southwest of Black Hill, there are sills of Haverhill granodiorite in the schists; this suggests that the pluton may fray out at its southwest end into a series of sills.

The French Pond granite forms a large cylindrical pluton with a vertical axis. A detailed map has been presented elsewhere (Billings, 1945, p. 57).

Structure between Ammonoosuc and Monroe Faults

Major structure.—The outcrop pattern of the formations between the Monroe fault and the Ammonoosuc thrust in the northern part of the Woodsville and southeastern St. Johnsbury quadrangle (Fig. 2) defines two major northward-plunging synclines and an intervening anticline. The Ammonoosuc volcanics, overlain by the Partridge formation, outline the nose of a northward-plunging syncline, the Walker Mountain syncline (Billings, 1937, p. 519–521), just north of the village of Woodsville. An oval-shaped area of Ammonoosuc, bounded on the west by the Monroe fault in the southeast ninth of the St. Johnsbury quadrangle, reveals a northeasterly plunging syncline called the Monroe syncline. The anticline between these synclines, the Gardner Mountain anticline, has its crest in a broad area underlain by the Albee formation. The axial planes of all these folds may be projected for a short distance into the central part of the Woodsville quadrangle, but cannot be readily traced through the area of complexly folded Albee formation that occupies the whole space between the Monroe fault and the Ammonoosuc thrust southwest of Woodsville.

Near the southern border of the quadrangle, just east of the Connecticut River, the Orfordville formation is exposed in the core of a northward-plunging anticline that Hadley (1942, p. 154) has called the Piermont anticline. The Orfordville is also exposed in a narrow belt that extends along the east side of the Monroe fault south from West Newbury. There is presumably a tight syncline, therefore, in the Albee formation between these two areas of the Orfordville.

Although the Hardy Hill quartzite is well exposed on the southeast limb of the Piermont anticline, it does not appear on the northwest limb. Locally the lower slates of the Orfordville formation—those beneath the Hardy Hill quartzite—are in contact with the Albee formation. These relations indicate that the northwest limb of the anticline is broken by a fault, which is considered to be a thrust dipping steeply to the southeast.

Correlation of the anticline and syncline at the southern border of the quadrangle with the

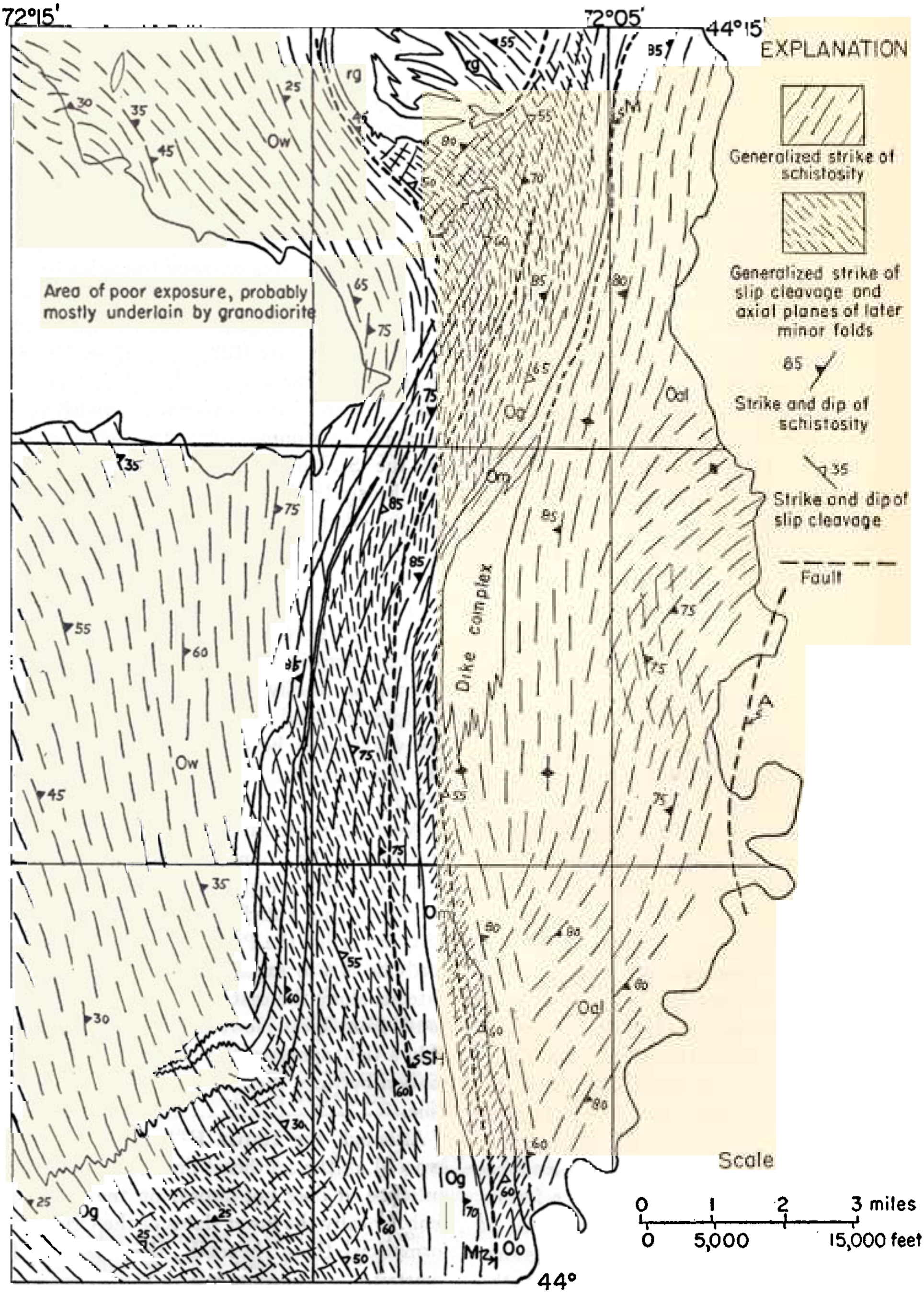


Figure 8.—Generalized Strikes of Cleavages

Lines based on data of Plate 3. Lithologic units and faults identified by letter symbols as follows: Ow, Waits River formation; Og, Gile Mountain formation; Om, Meetinghouse slate; Oo, Orfordville formation; Oal, Albee formation; rg, Ryegate granodiorite; SH, Scotch Hollow thrust; M, Monroe fault; A, Ammonoosuc thrust.

folds in the northern part is not established. If the axial planes of all the major folds are more or less parallel to the trend of the schistosity in the Albee formation (Fig. 8), the Walker Mountain syncline may be the same as the syncline west of the Piermont anticline. Under this interpretation, the Monroe syncline and Gardner Mountain anticline die out or are cut off towards the south by the Monroe fault, and the Piermont anticline is cut off by the Ammonoosuc thrust somewhere near Newbury. Under any alternative correlation of major folds, the schistosity in the Albee formation must be assumed to cut across the axial planes of the major folds, and would presumably, therefore, be later than the major folds.

Minor structural features.—Minor structural features of the rocks include minor folds, schistosity, slip cleavage, and shear planes. Most outcrops of the Albee formation contain minor folds that range from minute crenulations in the bedding of phyllite to folds 10 feet or more in amplitude in massive quartzite. In phyllite, the folds are nearly isoclinal and of the type called similar folds; their axial planes are essentially parallel to the schistosity in the vicinity. In massive quartzite, the folds are more open (Fig. 9) and approach parallel or concentric folds in shape; their axial planes may depart by as much as 30° from the attitude of the schistosity in the vicinity.

The axial planes of minor folds (Pl. 3) strike north to northeast, and dip vertically or steeply east; northwest dips are less common. The axes of minor folds have uniformly steep plunges, either northward or southward. Relatively few folds plunge at angles less than 60°; a large number plunge directly down the dip of their axial planes; and none were observed to have horizontal or near-horizontal axes. Both north and south plunges may be found in the same or adjacent outcrops, and at one place a single curved fold axis was seen to plunge northward at the top of a small cliff, vertically in the central part, and southward at the bottom. Inasmuch as the regional plunge is northward, as revealed by the outcrop pattern of older and younger formations in the major folds (Fig. 2; Pl. 1), and inasmuch as all evidence suggests that the minor folds reverse their direction of plunge through the vertical rather than the horizontal, the southward-plunging axes are believed to be overturned. Minor folds that appear to be southward-plunging anticlines are actually synclines in the sense that they have younger rocks in their cores than on their flanks. The word "overturned," as applied here to fold

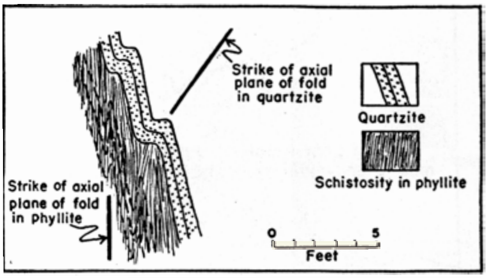


FIGURE 9.—EFFECT OF LITHOLOGY ON THE ORIENTATION OF MINOR FOLDS IN THE ALBEE FORMATION

Diagrammatic plan, folds plunge vertically. Heavy dashed lines are bedding in phyllite.

axes, is intended only as a descriptive term, because it is not at all clear whether these fold axes have actually been rotated from a subhorizontal to a steep or overturned position, or whether the folds were originally formed with essentially their present attitude, perhaps in response to differential movement parallel to the strike of the rocks.

The patterns of minor folds west of the Connecticut River are shown in Figure 10. A slight majority of those found east of the Monroe fault are sinistral.

The dominant cleavage in the rocks is schistosity. In thick phyllite beds, the schistosity is due to nearly perfect orientation of the micaceous elements, and is parallel to the axial planes of minor folds. The quartzites of the Albee formation have no cleavage, although locally there are fan joints parallel to the axes of folds and more or less normal to bedding. Where quartzite and phyllite (or schist) are thinly interbedded, the relation of schistosity to bedding differs in the various parts of folds. In the noses of folds, the schistosity is parallel to the axial plane where the phyllite beds are more than 3 or 4 inches thick; it approaches parallelism with the limbs away from the noses of folds (Fig. 11), and, in outcrops with no folds, the schistosity is essentially parallel to bedding. The micas within 2 or 3 millimeters of quartzite

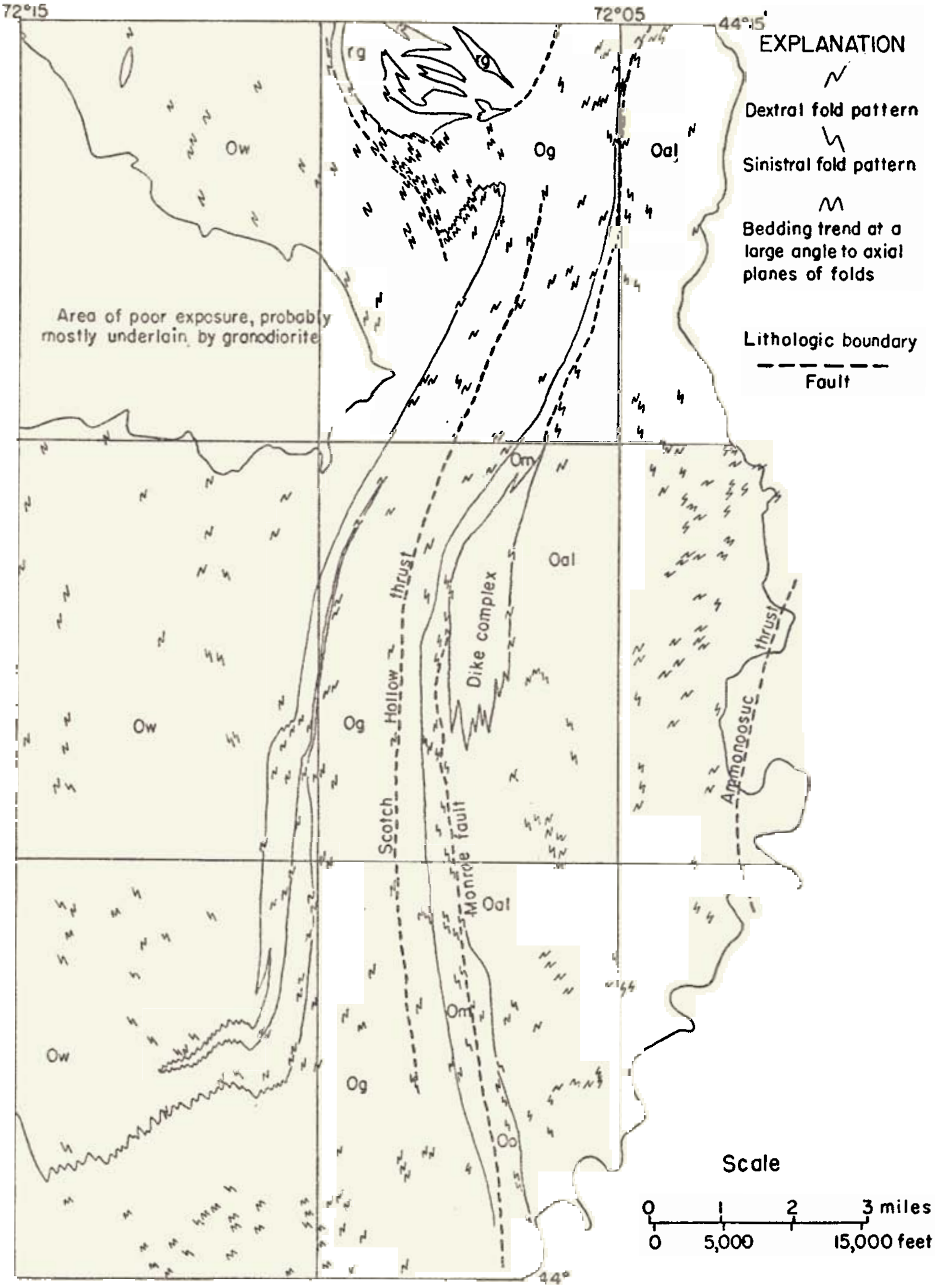


FIGURE 10.—PATTERNS OF MINOR FOLDS OTHER THAN THOSE SHOWN IN FIGURE 14

Lithologic units identified by letter symbols as follows: Ow, Waits River formation; Og, Gile Mountain formation; Om, Meetinghouse slate; Oo, Orfordville formation; Oal, Albee formation; rg, Ryegate granodiorite.

beds are parallel to bedding in all parts of folds, as are the micas in thin micaceous partings between quartzite beds.

Slip cleavage is found in rocks of both the Albee and Orfordville formations in the vicinity of the Monroe fault in the southern half of the quadrangle (Fig. 8; Pl. 3). Near the fault, schistosity and bedding strike parallel to the fault and dip steeply east. Slip cleavage, transecting both schistosity and bedding, strikes northeast and dips steeply southeast. It is parallel to the axial planes of minor folds that differ from the more common type described; in the minor folds associated with slip cleavage, both schistosity and bedding wrap around the noses. These folds plunge steeply east to southeast (Pl. 3).

Two possible explanations of the slip cleavage and associated minor folds are offered here. Both are suggested by the fact, emphasized by trend lines in Figure 8, that the schistosity near the Monroe fault in the southern part of the quadrangle is parallel to the fault and seems to lie athwart the northeastward trend of schistosity farther east, whereas the slip cleavage seems to lie parallel to the projected trend of this northeast-striking schistosity. First, the schistosity that is parallel to the fault may be slightly older than the dominant schistosity farther east, and perhaps due to the movements on the fault itself. Later movements that gave rise to the dominant northeastward-trending schistosity would then have acted, in the vicinity of the fault, on rocks already schistose, and formed slip cleavage by redeforming them.

An alternative, and perhaps less likely explanation involves the assumption that an earlier ubiquitous schistosity, oriented more or less parallel to the Monroe fault, has been destroyed by a later northeast-trending schistosity, except in the vicinity of the fault. Near the fault, the movements that produced the later schistosity must have been capable only of superimposing a slip cleavage on the earlier schistosity.

Shear planes with small to large displacement are characteristic of much of the Albee formation. They are particularly conspicuous in large outcrops of interbedded phyllite and quartzite (Fig. 12), where they are revealed by offsets in the bedding. These shear planes are generally parallel to the schistosity in strike and dip, and may be inconspicuous where both walls are phyllite. The movement on some of these shear planes must have taken place after the full development of the schistosity, because in

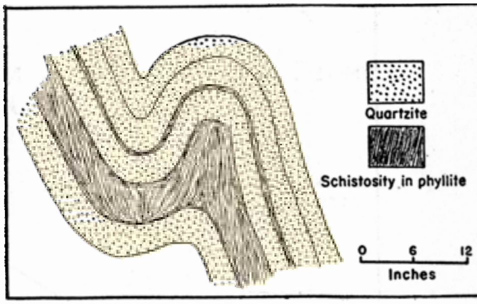


FIGURE 11.—RELATION OF SCHISTOSITY TO QUARTZITE BEDS IN MINOR FOLDS IN THE ALBEE FORMATION

Diagrammatic plan, folds plunge vertically.

a few places the schistosity, intersecting the shear planes at a low angle, is noticeably bent at the intersection. Movement on these shear planes may have been an important part of the total deformation of the rocks, particularly in the closing stages of the orogeny.

Structure West of Monroe Fault

General statement.—The outcrop pattern of formation boundaries west of the Monroe fault does not, by itself, provide sufficient evidence for understanding the major structure. Clues to the nature of the major folds and to the sequence of tectonic events are derived chiefly from the attitude and relationships of minor folds and cleavage, and for this reason minor structural features are described first.

Evidence of two distinct stages in the deformation of the rocks west of the Monroe fault will be presented. The term "stage" is used here, without any implication that separate orogenies need be involved, to separate distinct tectonic events that may well have taken place at different times in the same general orogeny. The earlier stage was characterized by the widespread formation of schistosity more or less parallel to bedding and formation boundaries, and by northward-plunging minor folds with a predominantly sinistral pattern; the Ryegate anticline, just southeast of Blue Mountain, was formed during this earlier stage of deformation. The later stage of deformation was less ubiquitous than the earlier, and the structural features associated with this stage show wide variation in type and intensity of development. Cleavage, if present, may be slip cleavage or schistosity; minor folds plunge northward and have predominantly dextral patterns. The Wrights Mountain fold, in the southwest ninth of the quadrangle (revealed by the change in course, from south to west-southwest, of the boundary of the Waits River formation), belongs to this stage of deformation. Superposition of features of the later stage on those of the earlier are very conspicuous over large areas and afford the principal evidence for the distinction between the stages and their sequence.

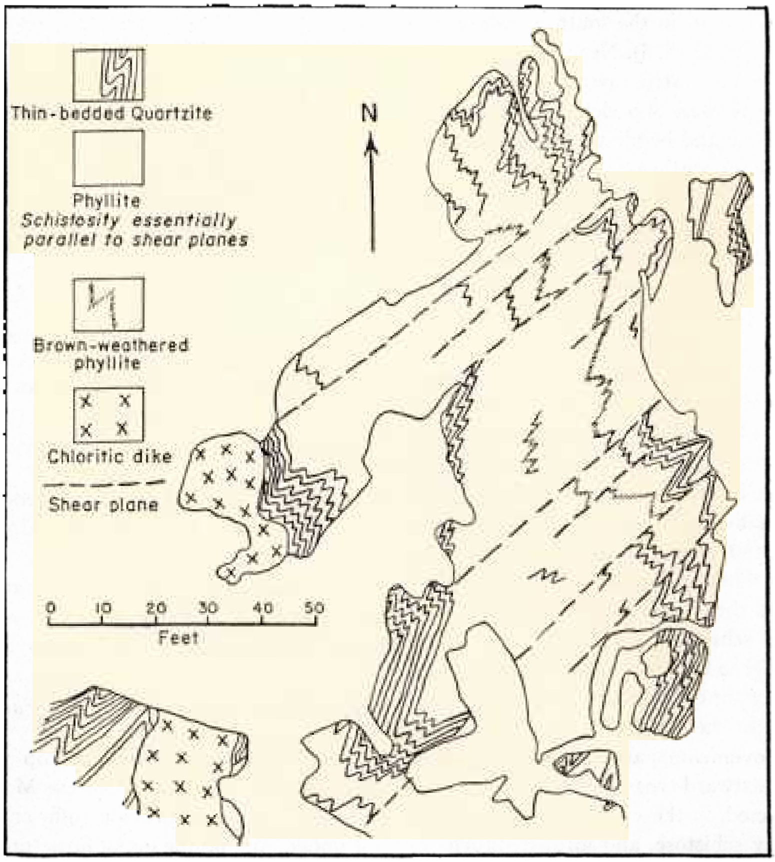


FIGURE 12.—SHEAR PLANES IN ALBEE FORMATION

Schistosity is essentially parallel to shear planes, dips vertically. Folds plunge steeply both towards the northeast and southwest. Outcrop 4700 feet east of highest point on Wallace Hill.

Most of the detailed structural measurements on which the following descriptions and interpretations are based are presented in Plate 3. Figure 8 gives a more generalized view of the trends of the various cleavages observed in the area, and should be studied for proper understanding of the symbols in Plate 3 and their interpretation. It will be noted in Figure 8 that the area west of the Monroe fault may be roughly divided into three parallel belts of unequal width on the basis of the character of the cleavage in each. A central belt contains slip cleavage transecting schistosity. The schistosity in the central and eastern belts is regarded as belonging to the earlier stage of deformation, and is called "earlier schistosity." The schistosity in the western belt, which is parallel to the projected trend of the slip cleavage, is considered to be a product of the later stage of deformation, and is called "later schistosity." Earlier schistosity was not recognized in the western belt, as will be discussed below. It seemed undesirable to distinguish the earlier and later schistosities in Plate 3 because the principal criterion for distinction is attitude; this criterion cannot be applied to all outcrops with complete objectivity in borderline cases.

Minor structural features of the earlier stage.—The characteristic cleavage of the earlier stage is a schistosity that is due primarily to a high degree of parallelism of the micaceous constituents of the rocks. To a lesser extent, it is due to dimensional orientation of tabular grains of quartz, calcite, and feldspar, where present. Schistosity is best developed in the micaceous rocks, and less conspicuous in the more quartzose or calcareous layers, but virtually all the rocks have this property in some degree. There are numerous tiny shear planes parallel to the schistosity, and most of the quartz-mica schists consist of elongate lenses or layers rich in quartz separated by thin micaceous foliae ("gneissic structure" of Swanson, 1941). But in all the rocks, quartzose ormicaceous, mineral parallelism is the dominant cause of the schistosity.

There is no measurable angle between bedding and schistosity in most outcrops. Within the limits of observational errors, the average trend of the schistosity is parallel to the boundaries of formations. For practical purposes, therefore, the schistosity may be termed bedding-schistosity. As used here, the term does not imply that bedding and schistosity nowhere intersect one another, for such intersections can be seen or inferred at many places. The term simply means that, within any small area, the average attitude of schistosity is parallel to the average attitude of bedding and that the trend of bedding does not deviate appreciably from the strike of the schistosity. The general parallelism of schistosity and formation boundaries suggests that the angle between these average strikes, if any, is very small in the quadrangle as a whole.

The attitude of the earlier schistosity and bedding in individual outcrops is shown in Plate 3, and symbols for the two have been combined where reliable bedding was seen to

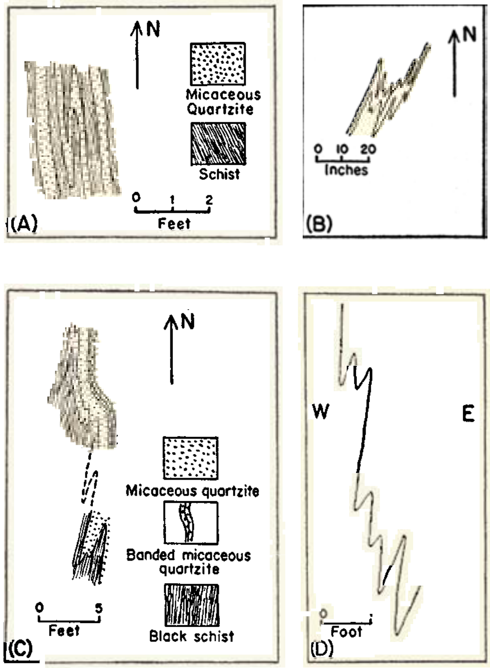


FIGURE 13.—TYPICAL EARLIER MINOR FOLDS WITH AXIAL-PLANE SCHISTOSITY

(A) Idealized earlier minor fold; axial plane steep to vertical, axis plunges 35°N. (B) Folded quartzose marble layer in schist. Outcrop in brook 4600 feet east of West Bradford School. (C) Outcrop 4700 feet northwest of Ryegate village, about 50 feet northeast of road. (D) Trace of a folded bed in a plane normal to the plunge of folds. Outcrop 2500 feet north of West Cemetery, southwest of Blue Mountain.

be parallel to schistosity. It should be re-emphasized that only the schistosity in and east of the belt of later slip cleavage (Fig. 8) is interpreted as earlier schistosity. The general parallelism of earlier schistosity and bedding and formation boundaries is evident from Plate 3, and holds even where the strike of the eastern boundary of the Waits River formation strikes east-northeast, as in the vicinity of Bradford Center.

The earlier minor folds (Fig. 13) are nearly isoclinal. The schistosity just described is parallel to their axial planes, and cuts the bedding at a high angle in their noses. On the limbs of

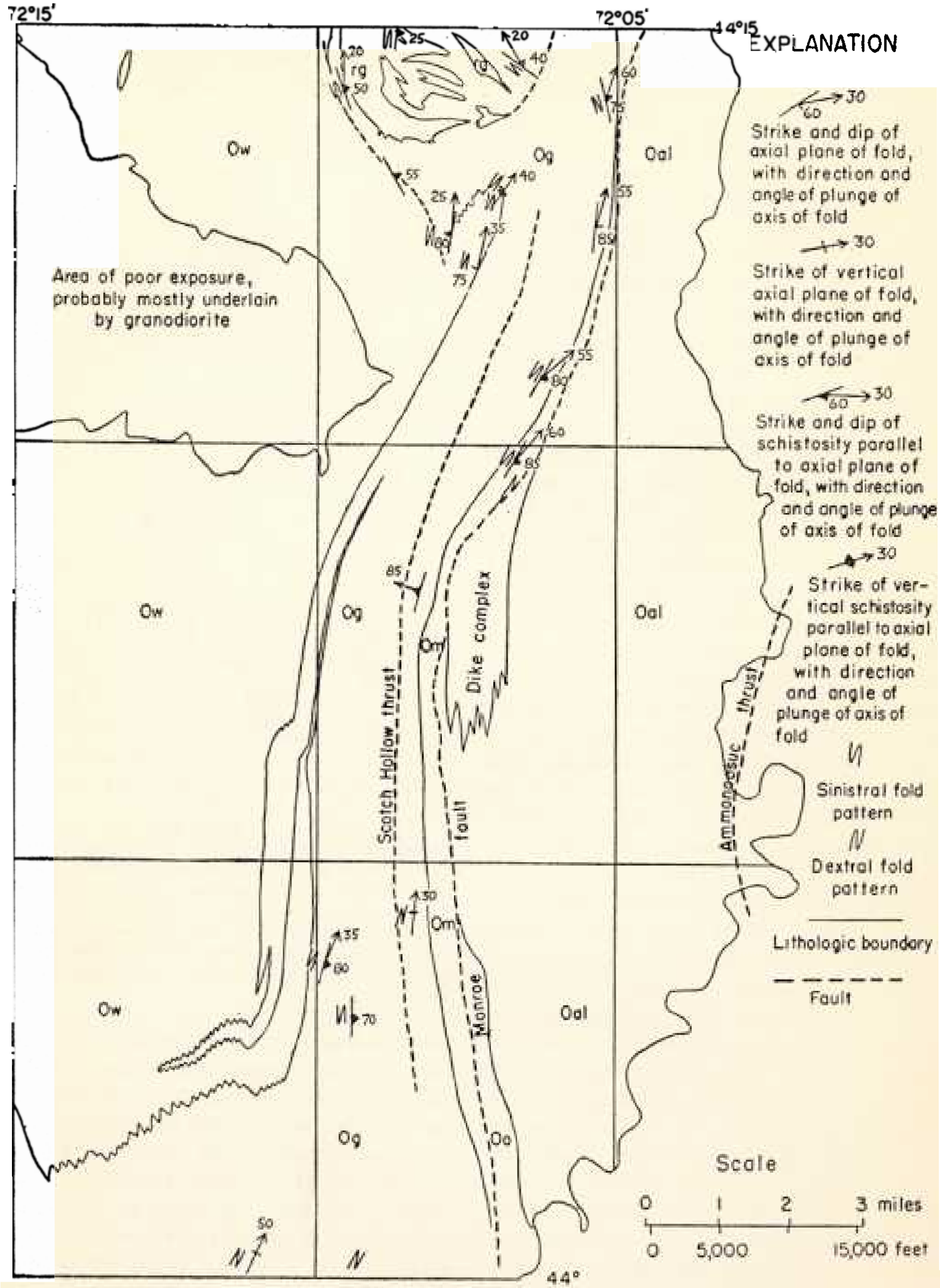


FIGURE 14.—STRUCTURAL MEASUREMENTS ON EARLIER MINOR FOLDS WEST OF MONROE FAULT

Lithologic units identified by letter symbols as follows: Ow, Waits River formation; Og, Gile Mountain formation; Om, Meetinghouse slate; Oo, Orfordville formation; Oal, Albee formation; rg, Ryegate granodiorite.

folds, a few inches to a couple of feet from the noses, the schistosity is typically parallel to bedding.

Clearly recognizable earlier minor folds are not common, and, to avoid confusion with the more abundant data on later minor folds in Plate 3, structural data for these earlier folds have been segregated in Figure 14. The folds, with one exception, plunge northward at an average angle of 32°. The patterns of 11 out of 15 of these folds are sinistral. In the same formations in the Vermont portion of the Littleton quadrangle (Fig. 1), Eric (1942, p. 47) found that 13 out of 15 earlier folds were sinistral. This pattern suggests that during the earlier stage of deformation, rocks on the east rose up with respect to rocks on the west.

Minor structural features of the later stage.—The cleavage here assigned to the later stage of deformation is schistosity in most of the western third of the area, slip cleavage farther east, and is generally absent in the immediate vicinity of the Monroe fault (Fig. 8). The later cleavage, whether slip cleavage or schistosity, strikes northwest over most of the area in which it occurs, and unlike the earlier schistosity, makes a moderate to large angle with formation boundaries. The differences in the later cleavage that are observed as one crosses the area from east to west provide a convenient framework for describing this cleavage.

As stated above, the rocks near and west of the Monroe fault display only the earlier bedding-schistosity, except where locally cut by the northeastward-trending slip cleavage more characteristic of rocks east of the fault. Half a mile to a mile west of the fault, scattered outcrops show minute crinkling of the earlier schistosity. There may be as many as 30 or 40 such crinkles per inch. Farther west, the crinkles are manifestly due to small flexures, up to about 2 millimeters across, whose axial planes are planes of weakness (slip cleavage) in the rock. Small biotite porphyroblasts may lie in and parallel to the axial planes of these flexures, forming by their intersection with the earlier schistosity a distinct mineral lineation parallel to the crinkles. Still further west, the short limbs of many of these minute flexures are faulted off, and well-defined planes of slip are apparent in the rock (Fig. 15B). Although there may be as many as 20 such planes to the inch, the orientation of the earlier schistosity is still preserved as a prominent direction of weakness in the rock, and the orientation of the

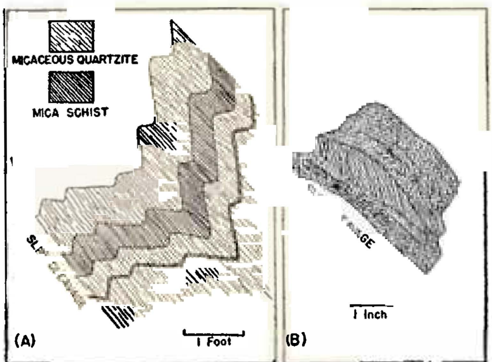


FIGURE 15.—LATER MINOR FOLDS AND SLIP CLEAVAGE

(A) Diagrammatic cross section illustrating ideal later minor folds with slip cleavage parallel to axial planes. (B) Folded schistosity with incipient slip cleavage parallel to the axial planes of folds. Hand specimen, southeast ridge of Tucker Mountain.

micas in the segments between slip planes is readily apparent in hand specimen. Rocks of this description are found as far west as the western boundary of the belt of slip cleavage (Fig. 8), but the commonest rock in the westernmost part of this belt shows a more advanced stage of development. Particularly in the region south of Wrights Mountain, the slip planes are very closely spaced, and the earlier schistosity in the segments between slip planes has been folded and rotated. The original orientation of the schistosity is not readily, if at all, discernible in these rocks. The presence of the earlier schistosity, however, may be detected in thin-section and many hand specimens from the microfolds preserved in the segments between slip-cleavage planes. These rocks possess a microscopic banding, attributable to the concentration of oriented micaceous constituents along slip-cleavage planes; thin quartz lenticles are abundant in the intervening spaces. The only prominent planar element is the slip cleavage, and bedding is difficult to trace.

The above description applies primarily to the more micaceous rocks. Slip cleavage is less prominent in the more quartzose rocks, where the slip planes are more widely spaced and discontinuous (Fig. 15A), and there has been little or no bending of the pre-existing earlier schistosity planes adjacent to the slip planes.

West of the belt of slip cleavage (Fig. 8), micaceous foliae like those that define the slip cleavage farther east are more or less merged, and the micas in the intervening more quartzose lamellae or lenticles are dominantly parallel to the foliation planes. The cleavage here may, for practical purposes, be called a schistosity, although (Fig. 8) it is parallel to the projected trend of the slip cleavage and makes an appreciable angle with such earlier schistosity as can be recognized in contiguous rocks to the east. This schistosity in the western third of the quadrangle is called later schistosity to distinguish it from the earlier bedding-schistosity recognized in and east of the belt of slip cleavage. The earlier bedding-schistosity may very well have been present in this western area before the development of the later cleavage, but, if so, it was more or less completely obliterated or masked by the later.

As described in more detail elsewhere (White, 1949, p. 590–591), the evolution of schistosity as an end-product of the development of slip cleavage and the obliteration of the earlier schistosity by formation of the later, seem to have involved two distinct processes, namely rotation and smearing out of the earlier schistosity, and growth of new mica flakes oriented parallel to the later cleavage. It is generally impossible to tell, however, whether a given group of parallel mica flakes, now parallel to the later schistosity, represents mechanically reoriented earlier schistosity or new micas that have grown with an orientation controlled by the later cleavage direction.

Minor folds related to the later stage of deformation are found in most outcrops west of the Monroe fault. In and east of the belt of slip cleavage (Fig. 8), they are readily identified by the fact that the earlier schistosity wraps around the noses of the folds. They range from a few inches to about a foot across in dominantly argillaceous rocks, but may be several feet across where most of the rock is micaceous quartzite. These folds are fairly open near the Monroe fault and become tighter towards the west. Near the eastern border of the Waits River formation, the minor folds may be nearly isoclinal, although the angle between limbs is generally over 30°. In shape, the folds closely approach the type called similar folds, even in fairly massive micaceous quartzite beds (Fig. 15A). Where slip cleavage is present, it is rigidly parallel to the axial planes of the folds both in the noses and on the limbs.

West of the zone of slip cleavage, tight to isoclinal folds with axial planes parallel to the later schistosity are abundant. One or both limbs of many are sheared off. Most of these minor folds are believed to be products of the later stage of deformation, but insofar as the schistosity here identified as later schistosity may, in part, represent earlier schistosity that has been rotated, a few of these minor folds may likewise represent rotated earlier folds. As in the case of the earlier minor folds farther east, the later schistosity tends to approach parallelism with bedding on the limbs of minor folds. There is, therefore, no simple basis for distinguishing possible earlier from later minor folds in the westernmost part of the quadrangle. Hadley (1950, p. 31–32) independently arrived at an almost identical interpretation of the relation of slip cleavage and later schistosity in the Mt. Cube quadrangle.

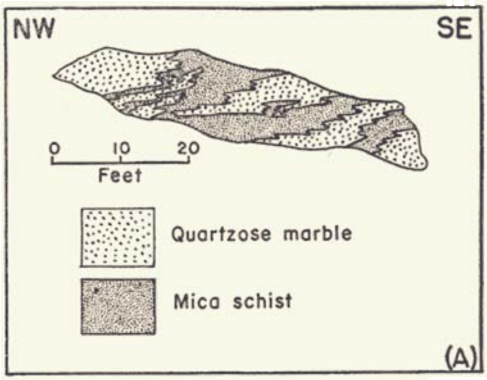
The minor folds in calcareous rocks of the Waits River formation differ, somewhat, from those in the argillaceous rocks of the Waits River and Gile Mountain formations. In general, two types of fold may be distinguished. One type is made manifest by the contacts between dominantly calcareous and dominantly argillaceous beds. The folds brought out by these contacts are mostly 6 inches to a foot across, and range from fairly open to almost isoclinal. Where the folds are nearly isoclinal or characterized by short and long limbs, the schistosity is essentially parallel to the long limbs. At the noses of isoclinal folds and other places where contacts cut across the schistosity there is small scale "interdigitation" (Greenly, 1930, p. 177–178) of calcareous and argillaceous material.

Another type of minor fold is found only in the beds of quartzose marble, and is brought out on weathered surfaces by a faint banding that is due to slight differences in composition of alternate layers. These folds are almost isoclinal, with amplitudes of several feet and widths of 2 to 6 inches between the same layer on opposite limbs. They may occur where the boundaries of the marble bed are perfectly straight, showing no evidence whatever of folding. The axial planes of these folds are parallel to or make a very low angle with the contacts between calcareous and argillaceous beds. Scattered mica flakes in the quartzose marble are fairly well oriented parallel to the axial planes of these isoclinal folds on the limbs and in the noses. This highly attenuated second type of later minor fold is taken as evidence that during deformation the calcareous layers were more plastic than the argillaceous layers.

It is difficult or impossible at a great many outcrops in the Waits River formation to determine the prevailing pattern of the minor folds, although their axes may be very obvious. The limbs of folds are sheared off, and only isolated noses of folds are preserved. Individual calcareous layers cannot be traced across such an outcrop. The prominent banding in such outcrops is large-scale cleavage banding rather than bedding. The preservation of noses of folds in separate cleavage bands is suggested on a highly exaggerated scale and in greatly oversimplified form in the geologic cross sections (Pl. 2).

Some of the folds in calcareous rocks are indescribably complex. A fairly common type is represented by Figure 16A; it is generally not clear in such exposures whether the tapering lenses of quartzose marble are original stratigraphic lenses, disrupted beds, or earlier folds whose axial planes have been deformed during the later stage of deformation. Even more complicated folds that defy analysis (Fig. 16B) are found in many places.

Minor folds of the later stage throughout the area west of the Monroe fault have axial planes that strike north to northwest and dip east to northeast, parallel to the later cleavage. The folds plunge uniformly northward at angles that range, for the most part, from 5° to 55°N. In the area of slip cleavage, the later minor folds are almost exclusively dextral in plan (Fig. 10). This dextral pattern suggests that during the later stage of deformation, rocks on the west moved relatively upward and northward with respect to rocks on the east, a complete reversal of the direction of differential movement indicated by the sinistral earlier minor folds. West of the area of slip cleavage,



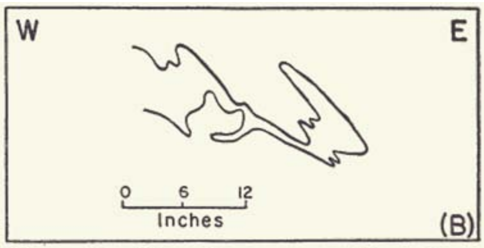


FIGURE 16.—MINOR FOLDS IN THE WAITS RIVER FORMATION

(A) Cross section exposed in a cliff 1500 feet north-northwest of the junction of Pike Hill Brook and Waits River, west of East Corinth. The tapering lens of quartzose marble in the center may be an earlier syncline. The minor crenulations are typical later minor folds. (B) Outcrop just north of road and north of a point at altitude 820 feet in Hedgehog Brook, southeast Topsham. The line represents a single horizon in a thick quartzose marble bed as viewed on the sloping surface of the outcrop.

the average trend of bedding makes a large angle with the axial planes of folds, and both dextral and sinistral patterns are found.

Superposition of later minor structural features on earlier.—Superposition of later minor structural features on earlier is a characteristic feature of the belt that contains slip cleavage, and has various manifestations. Slip cleavage everywhere cuts, and commonly offsets, the earlier schistosity, and the earlier schistosity wraps around the noses of the later minor folds. Such phenomena are fairly common in

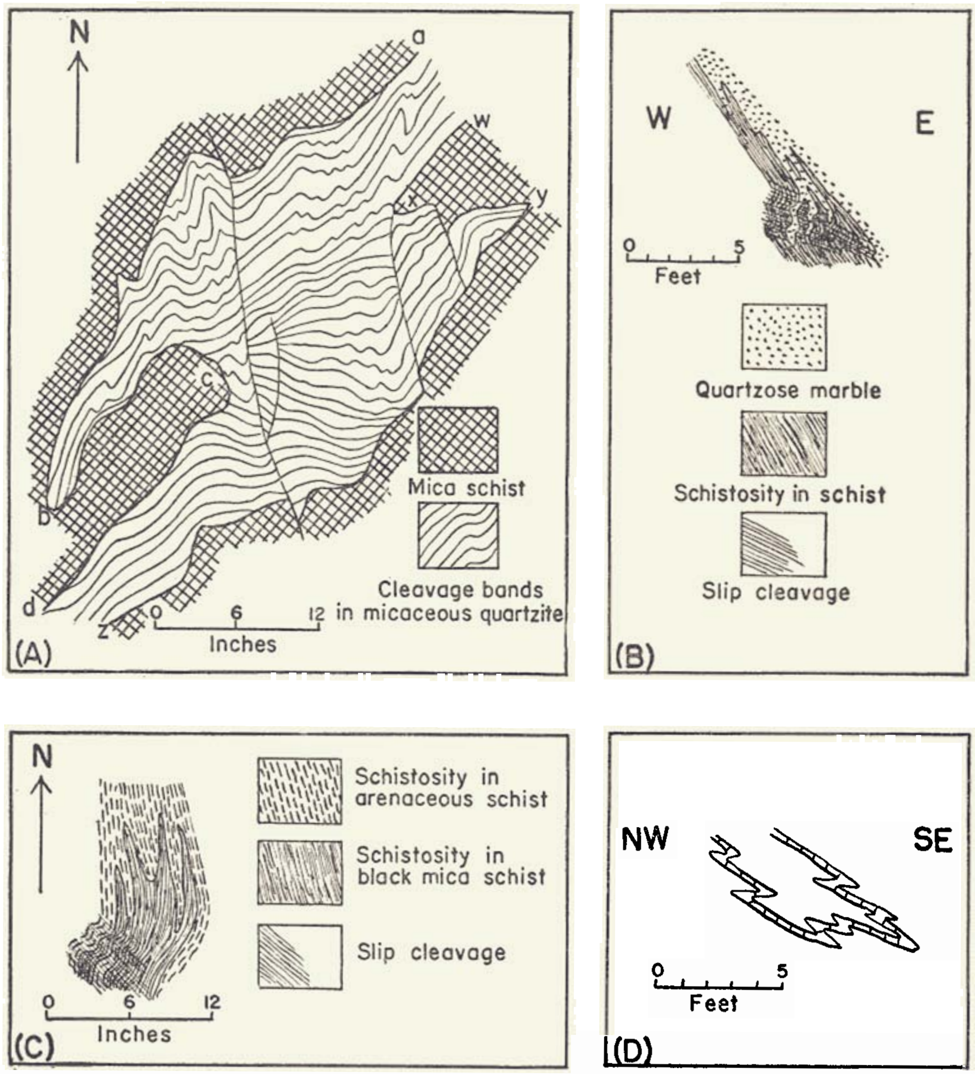


FIGURE 17.—SUPERPOSITION OF LATER MINOR FOLDS ON EARLIER MINOR FOLDS

(A) Micaceous quartzite bed in south-central part of outcrop sketched in Figure 18. Letters refer to points discussed in text. (B) Outcrop in brook just west of boundary between Waits River and Gile Mountain formations, $\frac{1}{3}$ mile north of the Woodsville quadrangle in the St. Johnsbury quadrangle. (C) Outcrop 4000 feet south-southeast of Red School, Ryegate. (D) Folds in a quartzose marble bed in cliff face. Outcrop 2500 feet northeast of School No. 3, Ryegate.

metamorphic terranes, and may be subject to various interpretations (Mead, 1940, p. 1011, 1020–1021; Cloos, 1937, p. 68–69). Less commonly described are the transection of the axial planes of earlier minor folds by later cleavage and sharp folding of the axial planes of earlier minor folds. (Fig. 17).

In Figure 17A, the line *abcd* represents the western boundary of a bed of micaceous quartzite in mica schist, and line *wxyz* the eastern. The wavy lines are cleavage bands that were originally more or less parallel to the axial planes of earlier-stage folds. The axial plane (or axial surface) of a small minor fold of the earlier stage follows one of the wavy lines connecting *c* and *y*. The axial plane of the complementary fold follows a wavy line connecting *b* and *x*. The pattern of this fold-pair is sinistral in plan. Superposed on the northeast-trending axial planes of these earlier folds are later dextral folds whose axial planes strike more or less north. The axial planes of the earlier folds are deformed in the later.

Similarly deformed axial planes are evident in the other sketches of Figure 17. The earlier schistosity of Figure 17B, which dips steeply east to vertical, is parallel to the axial planes of sinistral minor folds of the earlier stage. These folds are revealed by the interfingered marble and schist. The earlier schistosity and the axial planes of the earlier minor folds are both deformed in a later set of dextral minor folds whose axial planes dip gently east, parallel to the slip cleavage. The sketch of Figure 17C is similar except that the earlier minor folds are revealed by the interfingering of mica schist and arenaceous schist, and their pattern is not evident from this small exposure. In Figure 17D, the limbs of an earlier fold with southeastward-dipping axial plane are deformed by small later folds; the axial planes of these later folds have a more gentle apparent dip to the southeast than the axial plane of the earlier fold.

The lime-silicate bands of Figure 18 reveal many earlier minor folds whose axial planes are more or less parallel to the schistosity. A number of such folds with sinsistral pattern are evident in the northeast quarter of the outcrop. In the group of dextral minor folds just south of the center of the outcrop, on the other hand, the schistosity and lime-silicate bands are both folded together. The axial planes of these later folds trend north to northwest, and make a fairly high angle with the schistosity.

The deformed axial planes of the Woodsville quadrangle are significant for two reasons. First, they show clearly that there has been a sequence of tectonic events. One might argue, for example, that the cleavage here called "earlier schistosity" is primarily mimetic after bedding and does not represent a tectonic event. If this were true, of course, the fact that the earlier schistosity wraps around the noses of the later minor folds would have no tectonic significance. The earlier minor folds, however, imply movement as the cause of the earlier schistosity, and, particularly where the axial planes of these earlier folds are deformed, this earlier movement is clearly dated as prior to the later minor folds.

Second, the dominantly sinistral patterns of the earlier minor folds, contrasted with the dominantly dextral patterns of the later, indicate that the events whose order has been established are drastically different. There must have been significant reversal of the direction of differential movement over a fairly large area. This reversal is emphasized and unequivocal where later dextral minor folds are superposed on the axial planes of earlier sinistral folds.

Major folds.—Two major folds, the Wrights Mountain fold and the Ryegate anticline, are revealed by irregularities in the boundary between the Waits River and Gile Mountain formations. Their relationship to the minor structural features described above provides clues to their nature and relative age.

The Wrights Mountain fold is outlined by an abrupt change in the trend of the boundary of the Waits River formation on the southwest slope of Wrights Mountain (Pl. 1). On the northeast limb of this fold, the bedding and earlier schistosity trend north and dip more or less vertically. On the southwest limb, which may be regarded as extending essentially to the western border of the quadrangle, the same structural elements trend N. 70°E. and have an average dip of 15–20°NW. The trace of the axial plane of the fold may be roughly located by drawing a line connecting the points at which the strike of bedding swings from south to west-southwest (Fig. 8, Pl. 3). This line passes just east of Bradford Center, along the east side of the Waits River valley, and strikes about N 30°W. The dip of the axial plane, as revealed in cross sections S-S′ to W-W′, is of the order of 30° to 50°NE. The plan view of this fold is deceptive in that the trace of the axial plane strikes more or less normal to the southwest limb of the fold, and therefore makes this limb look more like a nose than a limb; actually, because of the gentle dip of the axial plane, the solid angle between the plane and each limb is approximately the same, as can be readily determined on a stereographic net. The plunge of the fold is about 20°N.

The axial plane of the Wrights Mountain fold

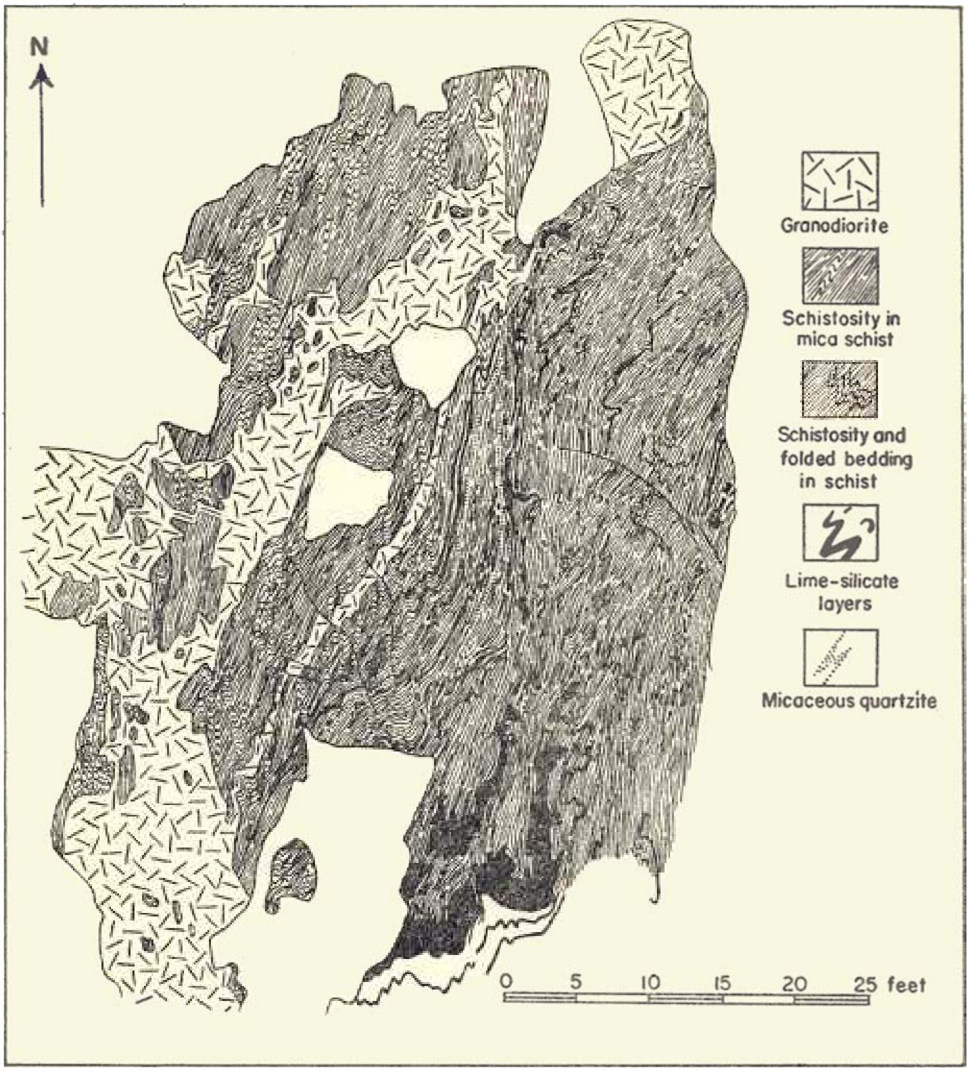


FIGURE 18.—COMPLEX FOLDS OF THE EARLIER STAGE OF DEFORMATION

Later minor folds, superposed on the earlier, have axial planes that trend slightly west of north, and are most abundant in the western half of the outcrop. Bedding and schistosity dip 20°–35°E. Outcrop 800 feet north-northwest of highest point on east-west trail, north side of Blue Mountain.

is more or less parallel, in strike and dip, with the later slip cleavage and the axial planes of the later minor folds (Pl. 3). Its axis is essentially parallel to the plunge of the later minor folds. The earlier schistosity, furthermore, curves with the bedding around the nose of the fold. There can be little question, therefore, that the Wrights Mountain fold was formed during the later stage of deformation.

In the Ryegate anticline, a mile northwest of Ryegate village (Pl. 1), the boundary between the Waits River and Gile Mountain formations is doubled back to form a "V" pointing slightly east of north. The axial plane of this fold strikes about N. 30°E. and is interesected almost at right angles by the axial planes of the later minor folds and slip cleavage (Fig. 8; Pl. 3). The Ryegate anticline is therefore older than the slip cleavage and presumably belongs to the earlier stage of deformation.

This major fold is interpreted as an anticline on the basis of its outcrop pattern and the probable northward plunge of its axis. All but one of the earlier minor folds in the quadrangle plunge northward (Fig. 14), and the average plunge is about 32°. An earlier major fold, such as this anticline, should have a similar plunge. Approximately this same plunge may also be derived from the structural data obtained in the immediate vicinity of the fold (Pl. 3). By projecting the boundary between the formations downward parallel to the axes of minor folds, as was done in preparing the cross sections (Pl. 2, C-C′ and D-D′), one can determine the probable shape and attitude of the fold in depth. The basic assumption involved in such projection is almost certainly valid where the major fold is not isoclinal and where the minor folds all have a fairly uniform direction of plunge, as here, because the axes of all minor folds, regardless of their relative age, must lie parallel to the bedding. The Ryegate anticline, as reconstructed in depth by such projection, has a plunge of 29°N. 11°E.

West of the Ryegate anticline, the contact between the Waits River and Gile Mountain formations trends northwest, and the change in strike at a point 1.5 miles due west of Ryegate village suggests the nose of a syncline. Detailed mapping, however, showed clearly that this northwest-trending boundary between formations strikes nearly at right angles to the trend of bedding in the Gile Mountain formation on the west limb of the Ryegate anticline. The boundary appears to mark a fault.

Interpretation of major structure and relative age of formations.—It will be noted that the Wrights Mountain fold and the Ryegate anticline give contradictory evidence of the relative age of the Waits River and Gile Mountain formations. Both folds plunge north and have the Waits River formation in their cores, but the fold at Ryegate is anticlinal in appearance and the Wrights Mountain fold synclinal (Pl. 2). One or the other is obviously unreliable, and, for lack of other unequivocal evidence for relative age, the sequence of formations must be decided on the basis of an interpretation of the relationship of the folds to one another.

The structural relations evident in the western part of the Woodsville quadrangle may have come about in one of at least two ways (Fig. 19): Under the first hypothesis, the Waits River formation is younger than the Gile Mountain, and during the earlier stage of

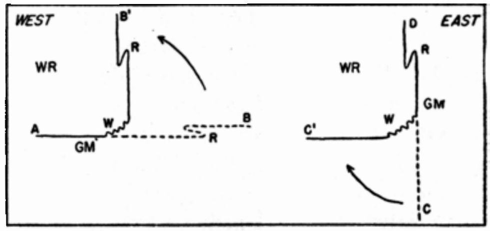


FIGURE 19.—DIAGRAMMATIC CROSS SECTIONS OF STRUCTURE WEST OF MONROE FAULT

Sections illustrate two possible origins, described in text, of present relationship between Ryegate anticline (R) and Wrights Mountain fold (W). Present boundary between Waits River (WR) and Gile Mountain (GM) formations represented by line AWB′ under first hypothesis, and C′WD under second.

deformation is involved locally in a recumbent fold (Fig. 19, R), the synclinal portion of which is here called the Ryegate anticline. The relations at the end of the earlier stage are shown by the partly solid, partly dashed line AB. During the later stage, the rocks on the east are rotated upward to form the northeast limb of the Wrights Mountain fold (W), producing the present relations represented by the solid line AWB′.

Under the second hypothesis, the Waits River formation is older than the Gile Mountain. The rocks are steepened during the earlier stage of deformation, and the Ryegate anticline (R) is formed as part of a very large drag fold on the east limb of an enormous anticline, whose axis lies somewhere to the west. During the later stage, the rocks down the dip from the Ryegate anticline are displaced relatively upward and westward to form the Wrights Mountain fold (W). Relations before and after the later stage of deformation, under the second hypothesis, are shown by lines DC and DWC′, respectively.

The first hypothesis appears offhand to be the simpler and perhaps more reasonable explanation. The strong rotation of the northeast limb of the Wrights Mountain fold offers a ready explanation for the transection of earlier by later cleavage on that limb, and for the apparent reversal of the direction of differential movement. This first hypothesis is not favored here, however. If intensity of development of the later minor folds and cleavage is proportional to the amount of gross deformation of the rocks, it is most logical to infer that the southwest, rather than the northeast, limb of the Wrights Mountain fold has undergone the greater amount of disturbance. As described above, later cleavage and minor folds are practically nonexistent near the Monroe fault, on the northeast limb, and become more and more intensely developed westward; in the southwest corner of the quadrangle, slip cleavage grades into a transverse schistosity, apparently obliterating the earlier bedding-schistosity. The minor structural features of the later stage, therefore, are best explained by the second hypothesis, and the Waits River formation is regarded as older than the Gile Mountain.

The orogenic movements involved in the second hypothesis are best considered in the light of the geologic structure farther to the southwest and west and will not be taken up here. A more regional treatment is given the subject by White and Jahns (1950).

Scotch Hollow thrust.—The Scotch Hollow thrust strikes nearly north-south and can be traced from Goshen village to Ryegate village. South and north, respectively, of these points its presence is not definitely known. The evidence for its existence is three-fold: (1) Along the line of the fault the bedding, schistosity, and slip cleavage dip more gently westward than elsewhere. This gentler dip suggests drag on a fault. (2) A continuous section in a brook crossing the fault 1½ miles north of Newbury Center displays about 100 feet of crushed and broken rock. (3) The fault as mapped on structural evidence alone coincides with the eastern limit of garnet-bearing rocks for 3½ miles south of Ryegate (Pl. 1) and closely follows the eastern limit of biotite-bearing rocks from north of Newbury Center to Goshen. Such a relationship is to be expected along post-metamorphism faults (*e.g.*, the Ammonoosuc thrust), which may bring together rocks of different metamorphic grade.

Where the biotite and garnet isograds are closest and the fault is best developed, in Scotch Hollow, the breadth of outcrop of the steeply dipping Gile Mountain formation is least. Such relations indicate crustal shortening, which is accommodated by thrust rather than gravity faulting. The drag indicates that the thrust is relatively from the west (Pl. 2, Section M-M′). When a bed or schistosity plane is rotated by drag along a fault, its attitude approaches that of the fault plane as a limit. The lowest westward dips of bedding in proximity to the Scotch Hollow thrust are about 40° at Newbury Center. The dip of the fault, therefore, is probably slightly less than 40°W. The post-metamorphism Ammonoosuc thrust dips about 36°W., as described above.

The fault cannot be traced north of Ryegate or south of Goshen. If it is assumed that it dies out beyond these points, the breadth of outcrop of the Gile Mountain formation at Goshen represents the breadth before faulting. At this place the formation is about 12,000 feet wide. The breadth of outcrop is narrowest east of Jefferson Hill, where it is about 8200 feet. The horizontal displacement normal to the strike of the fault is therefore of the order of 3800 feet. If the dip of the fault is slightly less than 40°, the dip slip is about 5000 feet east of Jefferson Hill and decreases to the north and south.

Other faults.—Two other faults mapped in the Waits River and Gile Mountain formations occur in the vicinity of Blue Mountain. One forms the contact between the Gile Mountain and Waits River formations northwest from the Ryegate anticline (Pls. 1, 2). The evidence for the fault is found in the southern half of its trace where it truncates the trend of bedding and earlier schistosity in the Gile Mountain formation (Fig. 8). Furthermore, although stratigraphic subdivisions within the Gile Mountain formation are not clearly defined, rough boundaries between more and less quartzose units can be drawn locally; along the southernmost 1½ miles of the fault trace (Pl. 1), the rocks of the Gile Mountain formation adjacent to the fault are unlike at different places.

The dip of the fault and the direction and amount of displacement are not known. It strikes parallel to the later cleavage, and is assumed, in the cross sections, to have the same dip as this cleavage.

A fault with northeast trend has been drawn from a point 3500 feet southwest of the south end of Lower Symes Pond to the northern boundary of the area east of the pond. On the southeast side of the fault trace the bedding and schistosity strike parallel to the fault and dip steeply or are vertical. On the northwest side of the fault the bedding trends northwest or nearly north. Discordance is suggested, and the fault has been drawn between the areas of different trend. There is no evidence bearing on its dip or the direction and amount of displacement.

Ryegate granodiorite.—The structural relations of the Ryegate granodiorite are best displayed in the vicinity of Blue Mountain (Pl. 1). There the granodiorite was intruded in thick sheets, more or less parallel to the regional structure of the stratified rocks. The map shows that all the sheets join to form a single body on the northwest side of Blue Mountain, but outcrops are few in this section, and possibly there are unexposed schist septa here as well.

The upper contact of the large southwestern sheet is well exposed in a cliff above Gibson quarry and dips 20°NE., parallel to bedding and cleavage in the schists above. Biotite schlieren observed in several quarries in this sheet dip, on the average, about 45°NE. The lower contact is not exposed, but mapping indicates that it has the same strike as the schists. Therefore, the body is believed to dip about 45°NE., parallel to the underlying schists. The divergence of the upper and lower contacts may be due to local irregularity or downward enlargement of the body. Along the south side of the mountain, the lower contact dips more steeply.

The higher intrusive sheets are much less clearly defined. The schist septa are full of large and small dikes and sills of granodiorite and, unlike the lower intrusive sheet, the upper sheets contain large inclusions of schist. The prevailing dips of the schists and observed contacts are gentle northeast, however, and it is believed that the general attitude of all the sheets is similar to that of the lowest one.

Small outlying bodies of granodiorite are subconcordant and tabular like the larger bodies.

Detailed maps of some of the granodiorite-schist contacts bring out several facts very clearly.

(1) Although most visible contacts are essentially parallel to bedding and schistosity, the trend of a contact may be at right angles to the bedding.

(2) There is strong evidence that locally, at least, the granodiorite made way for itself by forcing aside the layers of schist. Figure 20 illustrates the results of such a mechanism.

(3) Most contacts are very sharp, and, even in directions parallel to cleavage, there is little tendency for schist to grade into granodiorite. Feldspathized schist was found only on the hill north of Lower Symes Pond, although injection gneisses were seen farther north in the St. Johnsbury quadrangle.

(4) Inclusions of schist in granodiorite are mostly oriented with their long axes parallel to contacts. The orientation is clearly dimensional, inasmuch as the cleavage within oriented fragments may have random orientation (Fig. 21). The cleavage, however, is normally parallel to the elongation of the fragment.

(5) The granodiorite cuts folds in the schistosity (Figs. 18, 21) and therefore followed some at least, of the folding of the schistosity.

The plutonic bodies in the Groton area are poorly exposed. Most of the area is probably underlain by granodiorite, but schist septa are undoubtedly present. The attitude of the stratified rocks north of the drift-covered area (Fig. 8; Pl. 3) suggests wrapping around a large intrusive, but granodiorite is insufficiently exposed to map as a unit on such meager evidence. All that can be said, therefore, is that there are probably large and small subconcordant intrusives in the area, just as on Blue Mountain.

The granodiorite bodies may have been emplaced in one of three ways: (1) by feldspathization of schists in place, *i.e.*, granitization; (2) by stoping on a large or small scale; or (3) by forceful injection.

The principal argument against granitization, apart from lack of evidence of metasomatism, is the fact that, although the country rock in the Blue Mountain area is argillaceous and

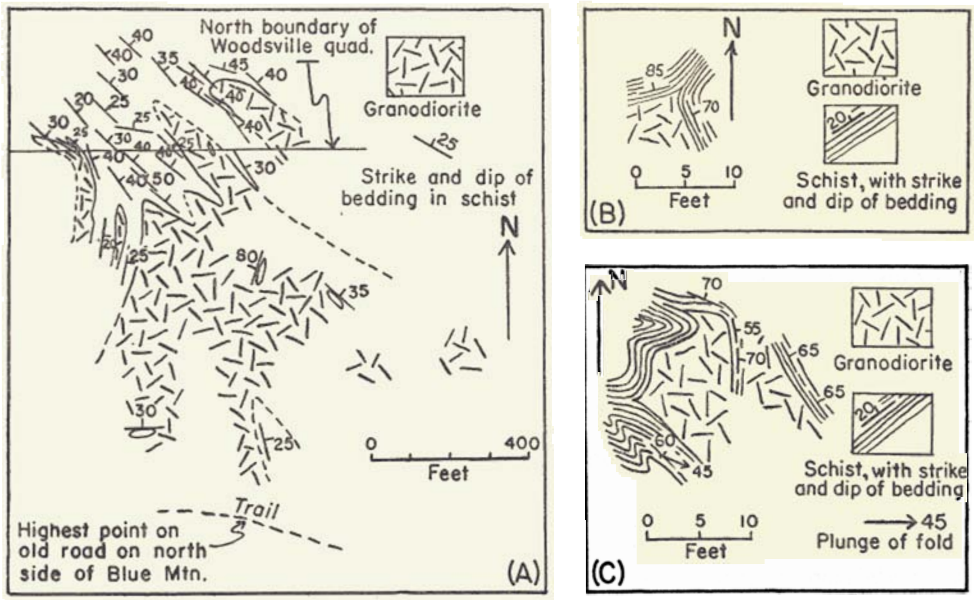


FIGURE 20.—INTRUSIVE CONTACTS OF RYEGATE GRANODIORITE, ILLUSTRATING RESULTS OF FORCEFUL INJECTION

(A) Pace and compass map, north ridge of Blue Mountain. (B) Outcrop beside house 2500 feet south of north border of Woodsville quadrangle on road east of Blue Mountain. (C) Outcrop just south of trail north of Blue Mountain, 500 feet from east end of trail.

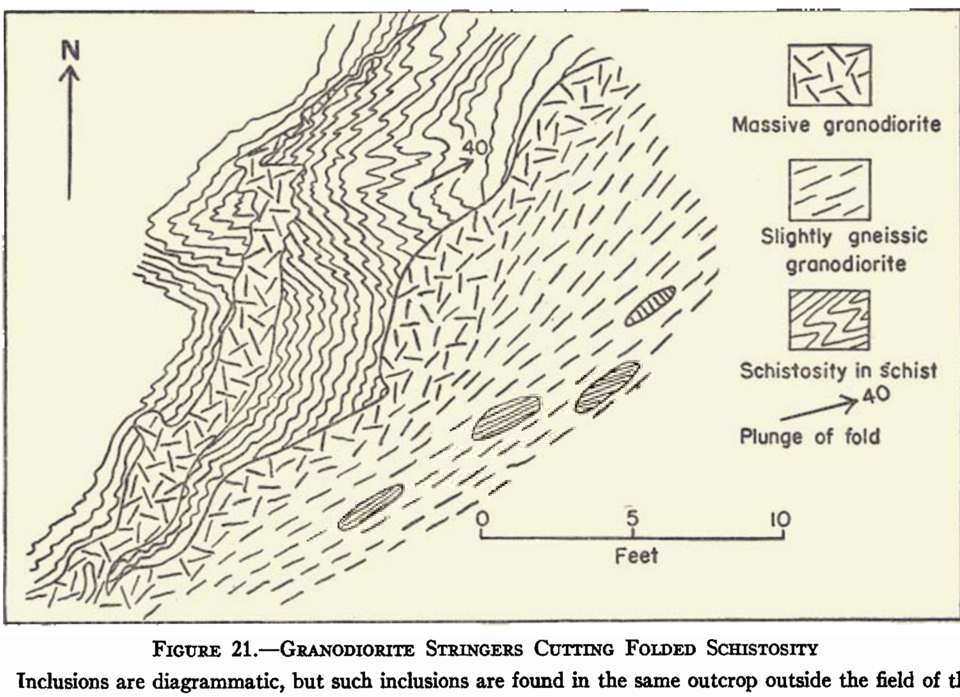


FIGURE 21.—GRANODIORITE STRINGERS CUTTING FOLDED SCHISTOSITY

Inclusions are diagrammatic, but such inclusions are found in the same outcrop outside the field of the sketch. Same outcrop as Figure 20B.

that in the Groton area is calcareous, there is no notable difference in the mineral composition of the plutonic rocks in these two areas.

The bodies lack the characteristic features of those plutons in the region that are believed to have been emplaced by stoping, such as marked truncation of bedding, cleavage, and fold axes (Chapman and Chapman, 1940, p. 201–209).

The remaining mechanism is forceful injection. Evidence that locally the walls have been forced apart is shown in Figure 20. Furthermore, the bodies are lenticular and subconcordant. Schist inclusions are parallel to contacts. Lastly, the trend of the foliation and of the axial planes of later folds diverge around the granodiorite areas (Fig. 8; Pl. 3).

The granodiorite is therefore believed to have been forcefully injected into the schists, mostly along the cleavage planes but partly along fractures of different orientation.

The granodiorite was intruded after most of the folding. Later folds are cut by stringers of igneous rocks (Figs. 18, 21), and no outlying dikes have a secondary foliation, regardless of the angle at which they cut the planes of deformation. A few have a vague foliation parallel to their boundaries, but this is believed to be primary flow structure. And although the granodiorites have unquestionably caused local deformation by their forceful injection (Fig. 20) the dikes in particular are typically cross cutting, with sharp, clean walls, and suggest only that they were intruded into already deformed rocks.

METAMORPHISM

General Statement

Many of the principal facts concerning the metamorphism have already been mentioned. The present section will therefore merely summarize the data already presented but will also consider some special problems that deserve elaboration. The metamorphism of the mafic dikes is discussed in another paper (Billings and White, 1950).

Progressive Metamorphism

It has been repeatedly emphasized in the description of the metasedimentary rocks that a belt of low-grade metamorphism averaging 4 miles wide extends from the south-central part of the quadrangle to the northeast corner. West and east of this central belt, there is an increase in the grade of metamorphism so that rocks in the western and eastern parts of the quadrangle lie in the staurolite zone. Locally, around bodies of plutonic rocks, the sillimanite zone is attained.

In the western part of the area the porphyroblasts that reflect the progressive metamorphism are later than most of the deformation of the later stage. Large crystals of chlorite, muscovite, biotite, garnet, and staurolite generally cut indiscriminately across slip-cleavage planes, replacing rather than displacing the pre-existing fabric. Microfolds associated with the slip cleavage may be traced uninterruptedly through and across the borders of many porphyroblasts; as seen in thin section, the original outlines of these folds are perfectly preserved by lines of unreplaced opaque inclusions. Where the earlier schistosity is curved in these microfolds, outside of the porphyroblasts, individual flakes of mica that define the schistosity are not themselves bent; large numbers of them have evidently recrystallized or grown as straight plates, oriented parallel to tangents of the fold, after bending had ceased (Richey, 1948, p. 208). Evidence of post-crystallization bending, crushing, or rotation of crystals is sufficiently uncommon and local in the area of slip cleavage to indicate that most of the deformation had ceased before the peak of recrystallization. More substantial overlap of the periods of deformation and metamorphism has been noted west of the Woodsville quadrangle (White and Jahns, 1950, p. 209).

Retrograde Metamorphism

Ammonoosuc thrust.—Rather striking retrograde effects are found directly east of the Ammonoosuc thrust both in the Piermont member and in the main part of tne Albee formation. In a belt 500 to 1000 feet wide on the east side of the fault all the biotite has been converted to chlorite. The retrograde origin of the chlorite is evident because much of it contains rutile needles or leucoxene, both of which are considered to have exsolved during the alteration from biotite. Moreover, as one approaches the

TABLE 9.—ESTIMATED MODES OF ROCKS INVOLVED IN RETROGRADE METAMORPHISM

	1	2	3	4	5	6	7
Number of sections	3	1	2	1	1	3	2
Quartz	44	10	26	45	58	46	3
Orthoclase			20	10		27	
Plagioclase	33	50					45
Sericite							2
Chlorite	16	[illegible]		[illegible]	[illegible]	11	30
Epidote*	1	2		25		t	4
Muscovite	4	1		3	20	10	
Garnet	1				1	2	
Staurolite					7		
Magnetite	x		3		2	2	
Ilmenite							1
Pyrite	1			2			
Leucoxene			x			2	2
Apatite	x		x	x			
Tourmaline	t						
Sphene		8					1
Calcite							12
Per cent of anorthite in plagioclase	30	10					10

* Includes clinozoisite.

x—present in 50 per cent or more of thin sections.

t—present in less than 50 per cent of thin sections.

Blank—not seen in any of thin sections.

1. Feldspathic chlorite schist, Albee formation.
2. Feldspathic chlorite-epidote schist, Albee formation.
4. Chlorite-epidote schist, Albee formation.
5. Staurolite-garnet-chlorite schist, Piermont member of Albee formation. Half of staurolite is altered to sericite and chlorite.
6. Feldspathic chlorite schist, Piermont member of Albee formation.
7. Albite-chlorite-calcite rock, retrogressed from amphibolite.

All specimens are from a belt 500 feet wide directly east of the Ammonoosuc thrust.

belt of retrograde metamorphism from the east, biotite crystals in all stages of chloritization may be seen. Staurolite crystals do not begin to alter nor do they disappear as soon as the biotite. Even within 200 feet of the thrust, unaltered cores of staurolite may be found. The altered staurolite has a deep-brown, greasy appearance and forms either a complete pseudomorph or a shell surrounding the amber-colored unaltered staurolite. Thin sections show that the altered material is a fine-grained mat composed chiefly of sericite and what is probably chlorite. Garnet appears to have been more resistant to the retrograde processes than the biotite and staurolite. In specimens where the last two minerals have been completely altered, the garnet may be entirely fresh or chloritized only along a few cracks that cross it. Specimens probably not more than 70 feet from the fault contain relatively fresh garnet. Still closer to the fault, the rocks are cut by irregular quartz veins from 1 to 3 inches wide, and locally a large silicified zone is present.

Amphibolites in a belt several hundred feet wide on the east side of the Ammonoosuc thrust have retrogressed to albite-chlorite-carbonate rocks.

Modes of some of the rocks that have undergone retrograde metamorphism are given in Table 9.

From the descriptions given above, it is apparent that as one goes from the chlorite zone eastward across the Ammonoosuc thrust the first new metamorphic index mineral to appear is garnet, followed by staurolite, and, lastly, biotite. Thus, from such an area one might get the erroneous impression of the sequence in which the index minerals normally appear in progressive metamorphism. It is also important to note that no great amount of shearing appears to have accompanied the retrograde metamorphism. Locally, especially near the thrust, an occasional small fault may be seen in thin sections, but no wholesale shearing has taken place. It is believed, therefore, that the chief instigator of the retrograde metamorphism has been the introduction of aqueous solutions that entered along fractures associated with the faulting.

Hadley (1942, p. 171–173) has described the retrograde metamorphism southeast of the Ammonoosuc thrust in the Mt. Cube quadrangle.

Haverhill-Bradford area.—In Oliverian Brook, 0.8 mile north-northeast of the village of Haverhill, some of the hornfelses contain porphyroblasts of chlorite that are apparently pseudomorphs after biotite. It appears that in this area the biotite zone was attained, but that subsequently the biotite has retrogressed. On a small knob just west of U. S. Route 5, a mile northeast of Bradford, chlorite pseudomorphs after garnet are present in some of the phyllites. The garnet appears to have formed penecontemporaneously with the widespread earlier schistosity and to have been converted to chlorite during or following the development of the later slip cleavage.

There is thus a suggestion that in the Haverhill-Bradford area there was a stage of metamorphism that antedated the progressive metamorphism in the western part of the quadrangle. The earlier metamorphism in the Haverhill-Bradford area may have been contemporaneous with the progressive metamorphism in the eastern part of the quadrangle. If this is true, the climax of the progressive metamorphism in New Hampshire is older than that in eastern Vermont. A problem for future investigation is to determine the relative ages of the progressive metamorphism in the two states.

Occurrence of Minerals

The mode of occurrence of most of the minerals is essentially the same in the Woodsville quadrangle as in other areas with a comparable degree of metamorphism. A few minerals west of the Ammonoosuc thrust raise special problems that require further discussion.

CHLORITE: A porphyroblastic chlorite was observed locally in the biotite, garnet, and staurolite zones. The indices of four specimens range from $\beta = 1.623$ to $\beta = 1.630$; the optic sign is positive in two and negative in two; birefringence is less than 0.004. According to Winchell (1936, Fig. 4) these chlorites are ripidolite and diabantite, with intermediate FeO/MgO ratio and moderately high alumina content. These chlorite porphyroblasts have the habit of chloritoid and occur as large tabular crystals with ragged edges, lying for the most part at high angles to the schistosity. Where observed, they appear to be the latest crystals to form in the rock. Noble and Harder (1948, p. 958–959) describe porphyroblastic prochlorite from Lead, S. D., and believe that it formed as a product of progressive metamorphism of somewhat greater intensity than that required to form garnets; they show a prochlorite isograd in the middle of the garnet zone. The porphyroblastic chlorites of the Woodsville area though mostly encountered in the garnet and staurolite zones, are also found in rocks of the biotite zone. These chlorites are products of neomineralization, and are unlike the shreddy or ragged chlorites noted below that merely replace biotite and garnet crystals; but where they occur in rocks with garnet, staurolite, and biotite, it is reasonable to assume that they did not form in equilibrium with these minerals. It appears that they are retrograde and were formed during the period of declining temperature that immediately followed the peak of the metamorphism.

Chlorite replaces garnet or biotite in many micaceous rocks and is commonly most abundant as small ragged flakes in and adjacent to planes along which late slipping has taken place. It is abundant in a specimen of breccia from the Scotch Hollow thrust, but elsewhere retrograde chlorite of this type appears to have no systematic distribution or structural significance.

DOLOMITE: The occurrence of dolomite in the Woodsville area is very unusual. In most regions, dolomite reacts with quartz to form tremolite in the biotite, or at most the garnet zone (Harker, 1939, p. 257–258). Qualitative chemical tests of the carbonate of the Waits River formation indicate that dolomite is present with quartz in rocks as far west as the actinolite isograd (Pl. 1), which lies several miles west of the staurolite isograd of the aluminous rocks. The actinolite has undoubtedly originated largely from the reaction of quartz and dolomite in the normal manner, but the temperature at which this reaction took place must have been abnormally high. Thin calcareous beds close to staurolite schist along the eastern boundary of the Waits River formation do not contain actinolite. One is faced with the inevitable conclusion that for some reason the normal low-temperature dissociation of dolomite was prevented. The reason may be very high confining pressure that reduced permeability and prevented the escape of carbon dioxide. Any attempt to reconstruct the folds above the area suggests that the present level of exposure was once very deep.

CLINOZOISITE: The abundance of clinozoisite in metasedimentary rocks with diopside or biotite and calcic plagioclase, west of the actinolite isograd, is surprising. Barth (1936, p. 823–825) and Billings (1941, p. 875) believe that calcic plagioclase and epidote may exist in equilibrium. In Eskola's amphibolite facies (Barth, Correns, and Eskola, 1939, p. 352), rocks with diopside and calcic-plagioclase do not contain epidote. In the Woodsville area there is no evidence that the clinzoisite is of retrograde origin and out of equilibrium with the other minerals that were formed at the peak of the metamorphism.

DIOPSIDE: The composition of diopside in any specimen of lime-silicate rock is in direct ratio to the composition of actinolite in the same specimen, as determined from refractive indices. Where $\beta = 1.690$ in diopside, it is 1.635 in actinolite, and $\beta = 1.700$ in diopside corresponds to 1.650 in actinolite. Where β of diopside exceeds 1.700, indicating a hedenbergite content of 40 per cent (Winchell, 1933, p. 226), hornblende is present rather than actinolite. Under the conditions of metamorphism in the Woodsville area, therefore, actinolite with a ferro-tremolite content greater than 40 per cent does not form.

Cause of Metamorphism

The problem of the causes of metamorphism is fundamentally the problem of the source of heat. Although mineral equilibria and the attainment of equilibrium are modified by hydrostatic pressure, stress, and the presence of moving solutions, the surfaces whose traces are the biotite, garnet, and staurolite isograds correspond closely to isothermal surfaces, and for practical purposes the deviations caused by factors other than temperature can probably be ignored where the temperature range is large. Three possible sources of heat are considered below.

The rocks of east-central Vermont are most highly metamorphosed near the crestal region of the cleavage arch west of the Woodsville quadrangle (White and Jahns, 1950), and the metamorphism decreases progressively both east and west from this region. The north-south variation along the crestal region has not yet been studied. Large intrusives have been mapped in this crestal region, and their close relation to the principal metamorphism in time and space is striking. If one postulates that these intrusives are apical stocks of a more extensive batholith extending north-south below the cleavage arch, all the features of the regional metamorphism are readily explained.

Billings (1937, p. 557–559) has shown that in western New Hampshire the progressive increase in the grade of metamorphism toward the southeast is spatially related to the appearance of large bodies of the New Hampshire magma series. Moreover, in the eastern part of the Woodsville quadrangle, the high-grade metamorphism in the sillimanite zone of metamorphism is clearly related to the French Pond granite (Pl. 1).

Friction caused by deformation is probably not an important source of heat. The evidence commonly cited to prove that heat has been largely supplied by friction is the coincidence between intensity of deformation and degree of metamorphism in an area. (*See* Ambrose, 1936, p. 278–284.) Such evidence may involve confusion of cause and effect. The same amount of work will produce more distortion in a more plastic body than in a less plastic, other factors being equal. The increase of rock plasticity with temperature is almost axiomatic among geologists (Harker, 1939, p. 182–184; Leith, 1923, p. 322; Longwell, Knopf, and Flint, 1939, p. 370, 395), although not yet fully confirmed by experimental evidence. Griggs (1942, p. 112) has found that dry rocks are not notably affected by temperatures of a few hundred degrees, but notes (p. 114) a marked reduction in the strength of quartz immersed in sodium carbonate solution at elevated temperature. If plasticity does increase with temperature, other independent variables being approximately equal, the relation of apparent intensity of deformation to temperature may well be that of effect to cause, rather than the reverse, as assumed by Ambrose.

In the Vermont part of the Woodsville quadrangle, intensity of deformation is generally greater with higher temperature, as indicated by the isograds. There is, however, one remarkable exception. Westward from the trace of the axial plane of the Wrights Mountain fold, there is a very marked and abrupt increase in the amount of deformation as recorded by the number of minor folds per unit area. If the metamorphism was caused by heat of friction, there should be a corresponding increase in metamorphism west of this axial plane. Yet the garnet isograd crosses the trace of the axial plane without apparent deflection, and the staurolite isograd, though poorly defined in this vicinity, appears to curve away from, rather than toward parallelism with this line as it is followed south.

In western New Hampshire, Billings (1937, p. 557) could find no evidence that the intensity of deformation was any greater in the high-grade zone of metamorphism than in the low-grade zone.

The writers, therefore, do not consider that heat generated by friction in deformation was an important cause of the metamorphism in the Woodsville quadrangle.

Finally, lateral variations in the degree of metamorphism might result simply from differential uplift of originally flat-lying isothermal surfaces. There is close correlation in central Vermont (White and Jahns, 1950) between a broad area of domical uplift and a high grade of metamorphism. Inasmuch as this domical uplift is at least locally occupied by intrusive granitic rocks, however, the importance of uplift alone is not readily evaluated. Billings (1937, p. 557) finds no correlation in New Hampshire between domical uplifts and metamorphism.

Heat introduced by rising bodies of magma seems to be the most likely source of heat for the metamorphism.

Tectonic Evolution

The major folds and the earlier minor folds were the first product of the tectonic forces. In general, the Woodsville quadrangle is located on the east flank of a complex anticlinorium. Four synclines are present on the flanks of the anticlinorium (Fig 2): the Monroe (southeast corner of the St. Johnsbury quadrangle), Walker Mountain (directly west of the Ammonoosuc thrust), Salmon Hole Brook (directly west of Northey Hill thrust), and Garnet Hill (directly southeast of Northey Hill thrust).

It has been shown that the earlier minor folds evolved during the later stages of the major folding. The later minor folds were considerably younger.

It has long been realized that the major faults were not contemporaneous with one another (Billings, 1937, p. 525–531). The relations in the Littleton-Moosilauke area indicate clearly that the Ammonoosuc thrust is younger than the folding (Billings, 1937, p. 529). Moreover, it is younger than the progressive metamorphism of the eastern part of the Woodsville quadrangle, inasmuch as some of the metamorphic zones have been eliminated along the fault. It has already been shown that the Scotch Hollow thrust is in many ways similar to the Ammonoosuc thrust and thus presumably contemporaneous with it. The Scotch Hollow thrust is younger than the progressive metamorphism in the western part of the quadrangle. Consequently these two faults are considered relatively young tectonic features.

The Northey Hill thrust is older than the Ammonoosuc thrust, for it is older than the progressive metamorphism of the eastern part of the area (Billings, 1937, p. 531). On the other hand, the major folds were well formed before the Northey Hill fault developed, because in the northern part of the Moosilauke quadrangle the fault cuts across the west limb of the Garnet Hill syncline. It is likely that the fault at one time had a more gentle dip to the west and that additional compression rotated it to its present attitude.

The Monroe fault is believed to have formed during the earlier stage of deformation, but movement must have continued beyond the close of the later stage. Evidence bearing on this subject is as follows:

(1) The fault is parallel, in strike at least, to the earlier schistosity of the rocks on either side; it makes a large angle with the later cleavage and axial planes of the later minor folds.

(2) On the west side of the fault, particularly in the southern part of the area, the schistosity is folded. The axial planes of the resulting minor folds trend northeast and dip steeply southeast. Many of these folds have a slip-cleavage parallel to their axial planes. They are believed to be the same as the similarly oriented later folds with slip cleavage in the Orfordville formation across the fault to the east. The prob-

able magnitude of the fault makes it unlikely that folds and cleavage of identical type and orientation have been brought together from widely separated points, and most of the movement on the fault is therefore believed to antedate these later structural features.

(3) No breccia or broken cleavage planes were found along the fault in the Woodsville area, although Eric (1942, p. 65–66) found these phenomena locally along the fault in the Vermont portion of the Littleton quadrangle. Such features indicating late movement are certainly not abundant.

(4) The biotite isograd is displaced along the fault at Boltonville. Although this apparent displacement might be explained as the result of processes other than faulting—such as the greater metamorphism of the rocks on the east side of the fault by heat from the numerous late-tectonic mafic dikes—it is tentatively regarded here as a real mechanical displacement. The horizontal offset is about 3 miles; if the surface whose trace is the isograd dips eastward at a gentle to moderately steep angle (a logical assumption; *see* Noble and Harder, 1948, p. 969), its intersection with the fault plunges gently north, and the vertical displacement on the fault is only a few thousand feet. The east side is upthrown. A displacement of this order of magnitude is clearly inadequate to explain the stratigraphic break, and the amount of post-metamorphism movement must therefore represent but a fraction of the total.

In summary, the Monroe fault is believed to have formed, and most of the displacement on it to have taken place, during the earlier stage of deformation. There was, however, some post-deformation and post-metamorphism movement on it.

The age of the cleavage has been discussed in detail already. Suffice to say that the schistosity east of the Monroe fault is believed to be contemporaneous with the earlier folding, whereas the slip cleavage in the central part of the quadrangle and the schistosity in the western part are believed to be contemporaneous with the later folding.

The chronological position of the plutonic and hypabyssal rocks can be established only approximately. The metamorphosed mafic dikes are older than the metamorphism. The Bethlehem gneiss is syntectonic as it is strongly deformed. The Haverhill granodiorite, the French Pond granite, and the Ryegate granodiorite are late tectonic and are slightly deformed. The camptonite belongs to the White Mountain magma series and is post-tectonic.

The age of the metamorphism has been discussed on previous pages. Clearly, the final stages of the progressive metamorphism in the western part of the area are younger than the slip cleavage. Possibly, the progressive metamorphism in the eastern part of the quadrangle is somewhat older, but conclusive evidence is lacking.

Geologic History

The recorded geological history in the Woodsville quadrangle began in the Middle Ordovician. The Waits River, Gile Mountain, Meetinghouse, Orfordville, Albee, Ammonoosuc, and Partridge formations were deposited prior to the Silurian. Although the thickness of these formations cannot be stated with precision, it is certain that the total accumulation was many thousands of feet, probably at least 20,000 feet.

These rocks were deformed during the Taconic revolution near the end of the Ordovician, but in this area the intensity of this deformation does not appear to have been great. The proof of such a revolution in central New England is shown most convincingly around Littleton, New Hampshire, where Silurian rocks rest unconformably on the pre-Silurian (Billings, 1937, p. 468). Upper Silurian appears to have been a time of non-deposition, but fossils are scarce and rocks of this age may be present. During extensive sedimentation in the Lower Devonian at least 5000 feet of shales and sandstones accumulated (Billings, 1937, p. 468).

The great orogeny and the associated igneous intrusives are Acadian—that is, middle or late Devonian. The most convincing evidence is found in the Franconia quadrangle (Williams and Billings, 1938, p. 1025–1026), where the unmetamorphosed Moat volcanics of the White Mountain magma series rest unconformably on the highly metamorphosed Lower Devonian Littleton formation. The orogeny must be post-Littleton and pre-Moat. The White Mountain magma series is comagmatic with the Quincy granite near Boston, Massachusetts; the Quincy is pre-Pennsylvanian. Thus between lower Devonian and Pennsylvanian the following events occurred in central New England (Williams and Billings, 1938, p. 1040): (1) orogeny, including intrusion of the New Hampshire magma series; (2) erosion; (3) extrusion and intrusion of the White Mountain magma series. The first event is assigned to the middle and late Devonian (?), the second to the late Devonian (?) and early Mississippian (?), and the third to the Mississippian (?).

All the complex tectonic events in the Woodsville quadrangle are considered Acadian. None are believed to be Appalachian (Permian) because the Mississippian (?) White Mountain magma series throughout Vermont and New Hampshire has not been subjected to compressional forces (Billings, 1945, p. 44–53). It is believed that none of the structures described in this paper are Taconic (late Ordovician), because the pre-Silurian and post-Silurian rocks of central New England show the same intensity of deformation.

The youngest bedrock in the area is camptonite, a late member of the Mississippian (?) White Mountain magma series.

The geological history between the Mississippian (?) and Pleistocene is shrouded in mystery. Certainly many thousands of feet of rock were eroded. Pleistocene and Recent events are recorded by glacial, fluvio-glacial, and fluviatile deposits, as well as by features resulting from glacial and stream erosion.

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CALUMET, MICHIGAN; GEOLOGICAL MUSEUM, HARVARD UNIVERSITY, CAMBRIDGE 38, MASSACHUSETTS.

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