BEDROCK GEOLOGY OF THE
WOODSTOCK QUADRANGLE, VERMONT

By

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and

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PREFACE

The principal geologic mapping for this report was carried out in the field seasons of 1947–1949 by Ping Hsi Chang, then a graduate student at Harvard University, and formed the subject matter of his doctoral dissertation (1950) entitled "Structure and Metamorphism of the Bridgewater-Woodstock Area, Vermont." The area included the southern two-thirds of the Woodstock quadrangle and part of the Rutland quadrangle. After receiving his degree in 1950 Chang returned to his native country, mainland China. Owing to the barriers erected by international politics we have been unable to communicate with him for the past ten years, hence it has been necessary to prepare his part of this joint work for publication without any opportunity for him to review the manuscript or to dispute the changes we have made in his interpretation. We deeply regret that this is so, and only hope that when and if Chang has an opportunity to read this report he will not be too dismayed by our treatment of it.

Ern mapped the Randolph quadrangle, immediately to the north, during the seasons 1957–59 for the Vermont Geological Survey. His part in the mapping of the Woodstock quadrangle was carried out during the summers of 1959 and 1960 at the request of Dr. Charles G. Doll, Vermont State Geologist, in order to complete the bedrock mapping of the state for the Centennial Geologic Map of Vermont (1961). Ern's activities were principally in the northern third of the quadrangle, not mapped by Chang, with such revision of boundaries farther south as was necessitated thereby.

Thompson's contributions to the map are based on field work carried out sporadically since 1950 in connection with his studies in the Ludlow area adjacent to the south, but with a special effort in the summer of 1960 in order to revise Chang's stratigraphic subdivision and interpretation in accord with the scheme then agreed upon for the state map. This work, partly in cooperation with Ern, involved the location of some lithologic boundaries not mapped by Chang, together with the relocation of others as necessitated by recent mapping in adjacent areas and by revised criteria for locating the boundaries between certain units.

In most essentials, however, Chang's interpretation still stands. What follows below is largely the content of his doctoral thesis with editing by Ern and Thompson in consultation with Prof. Marland P. Billings who was the principal advisor to Chang during the preparation of the thesis.

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Some sections, however, have been extensively revised. Chang's original text, maps and diagrams are in the Harvard University Library and may be seen there. Microfilm copies are also available at cost.

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ABSTRACT

The Woodstock quadrangle, in east-central Vermont, lies partly on the eastern limb of the Green Mountain anticlinorium and partly in the region of domes and recumbent folds of eastern Vermont.

The oldest rocks belong to the Precambrian Mount Holly complex and are exposed in the extreme southwest corner of the area. These are overlain eastward by younger strata. They include the Tyson, Hoosac and Pinney Hollow formations at the base and the Gile Mountain, Standing Pond and Waits River formations at the top. Twelve major units as well as certain of their subdivisions have been mapped. These rocks range in probable age from lowest Cambrian to Devonian and have a total thickness of more than 20,000 feet. Nearly the entire eastern Vermont sequence is thus exposed in the quadrangle.

The deformation is mainly by folding. Thrust-faulting is evident only in the southwest corner. In the southwest part of the quadrangle early folds with east-northeast axes were followed by later folds with axes plunging gently north-northeast. In the eastern half of the area large isoclinal recumbent folds preceded later doming that deformed the earlier axial surfaces.

The rocks were all subjected to regional metamorphism except for some late mafic dikes. Most of the area is in the garnet zone except the southwest corner which is in the biotite zone and the northeast corner which is in the kyanite zone. Textural evidence based on mineral orientation and rotational effects in porphyroblasts indicates that the metamorphism overlapped in time at least a part of each major stage of deformation.

INTRODUCTION

Location and Extent of Area

The Woodstock quadrangle is in south-central Vermont on the east flank of the Green Mountains. It is bounded by the parallels 43°30′ and

43°45′ north latitude and by the meridians 72°30′ and 72°45′ west longitude. The area contains most of Bridgewater, Woodstock and Barnard townships and parts of Sherburne, Stockbridge, Pomfret, Hartland, West Windsor, Reading and Plymouth. (Fig. 1)

Field Work

Chang spent 24 weeks in the field in this area during the summers of 1947, 1948, and 1949. The U. S. Geological Survey topographic map of the Woodstock quadrangle (edition of 1913, reprinted 1942), on the scale of 1/62,500 and enlarged to the scale of 3 inches to a mile, was used as base map. Outcrops were located on the map with the aid of an aneroid barometer. In the heavily wooded regions pace and compass traverses from known points were also made. Field sketches were made of most of the outcrops on which special structural features could be observed. These were used later as a guide in constructing the structure sections.

Ern and Thompson carried out their mapping mainly in 1959 and 1960 (see Preface). Their mapping methods were essentially those of Chang supplemented by aerial photographs providing stereographic coverage. These greatly facilitated problems of location so that pacing was rarely necessary.

Acknowledgments

73°   72°

BURLINGTON

MONTPELIER

44°   44°

| 1 | 2 | 3 |

HANOVER

| 4 | 5 | 6 |

1. Rochester Quadrangle
2. Randolph Quadrangle
3. Strafford Quadrangle
4. Rutland Quadrangle
5. WOODSTOCK Quadrangle
6. Hanover Quadrangle
7. Wallingford Quadrangle
8. Ludlow Quadrangle
9. Claremont Quadrangle

NEW
YORK

RUTLAND

| 7 | 8 | 9 |

43°   43°

0   10
miles

BRATTLEBORO

NEW HAMPSHIRE

73°   72°

MASSACHUSETTS

FIGURE I.   INDEX MAP

Ern's field work was supported by the Vermont Geological Survey. Thompson's was supported in part by the Vermont Geological Survey, and in part by research funds of the then Department of Mineralogy and Petrography, Harvard University. Ern was assisted in the field by B. K. Goodwin of Lehigh University and Thomas Angell of Randolph, Vermont.

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Topography and Drainage

This area lies in the western margin of the New England Upland section of the New England Physiographic Province and in the eastern side of the central portion of the Green Mountain section. Much of the area is drained by the Ottauquechee River, a tributary of the Connecticut. Smaller portions are drained by the White River, Black River and Mill Brook, all tributaries of the Connecticut.

The Ottauquechee Valley, along which runs part of U. S. Highway No. 4, forms the lowest topographic element of this area. It enters from the west near West Bridgewater, whence it takes an eastward course across the central part of the area to the village of Woodstock. The altitude of the valley bottom at Sherburne is about 1,260 feet, whereas at Woodstock it is 750 feet.

The whole area is hilly and the minor topographic features are controlled by the lithology and structure of the underlying formations. A prominent longitudinal range trending north-south through the middle of the area has several peaks more than 2,400 feet in altitude and reaches 2,600 feet at the summit of Long Hill. To the east of this central range the country is chiefly characterized by rolling hills underlain in large part by the Waits River formation, and is partly cultivated as open farms. The altitude of hill tops and the general relief lessen gradually toward the east. The average relief is about 700 feet. In contrast, the region west of the central range is more rugged. The tops of the hills have an average altitude of 2,000 feet with several peaks rising above 2,600 feet. It is in general thickly wooded, and deserted farms with recent over-growth are common. The average relief is about 1,000 feet.

The whole area has been glaciated. Deposits of glacial origin are common, especially along the valleys of the Ottauquechee and its larger tributaries, and along the slopes of the higher ranges. This, combined with the thick cover of vegetation and the recent lumbering operations, makes thorough investigation of the bed-rock geology difficult.

Previous Work

Areal studies of the bedrock geology were made by E. Hitchcock, E. Hitchcock, Jr., A. D. Hager, and C. H. Hitchcock (1861), C. H. Richardson (1903, 1927, 1929) and E. L. Perry (1927, 1929). The stratigraphy of the western marginal zone of this area was discussed by C. L. Whittle (1894a, 1894b) and T. N. Dale (1916), who gave special attention to the problem of the general structure of the main axis of the Green Mountains and the occurrence of the "Algonkian rocks" in Vermont. A. Keith (1932)

and H. E. Hawkes, Jr. (1941) considered the problem of the roots of the Taconic overthrust in west-central Vermont. The granite and dolomite deposits in this area were studied by T. N. Dale (1915, 1923).

More recent studies in adjacent areas include those of C. G. Doll (1944), W. S. White and R. H. Jahns (1950), P. H. Osberg (1952), W. F. Brace (1953), J. B. Lyons (1955) and E. H. Ern (1963).

LITHOLOGY AND STRATIGRAPHY
General Statement

The rocks in this area are chiefly metamorphosed rocks of sedimentary and volcanic origin. Black phyllite, chlorite-sericite-quartz schist and quartzite, with or without porphyroblasts of garnet, biotite, chloritoid or albite, are the most common metasedimentary rocks. Greenstone, chloritic amphibolite, hornblende gneiss, and chlorite-hornblende-garnet schist are the most common types among the metamorphosed volcanics. Premetamorphism intrusives are represented by a few scattered lenticular bodies of serpentine, greenstone, chloritic amphibolite, and metamorphosed albite-quartz porphyry. The whole succession of stratified rocks has a total thickness of over 20,000 feet. They are subdivided into twelve major stratigraphic units (Table 1) on the basis of lithologic character. No fossils have been found in this area, but by correlation with other areas the rocks range in age from Precambrian to Devonian. The stratified series rests upon the Precambrian Mount Holly complex which crops out only in a very small area in the southwest corner of the Woodstock quadrangle.

A small granodiorite body in the Pinney Hollow formation may be younger than much of the deformation. Post-metamorphic mafic dikes are present in all the different formations.

Mount Holly Complex

The rocks in the central Green Mountains were first subdivided by C. L. Whittle (1894a, 1894b). A more intensely metamorphosed, presumably older, group of rocks was called by him the Mount Holly series, and supposedly younger, less metamorphosed rocks of distinctly sedimentary origin were referred to the Mendon series. Whittle, however, published no map and later workers have not been consistent in their use of the name Mendon series. The name has therefore been abandoned. As mapped by Perry (1929) in the Woodstock quadrangle the Mendon corresponds roughly to the lower part of the Tyson formation as described below.

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TABLE 1

STRATIGRAPHIC NAMES AND THICKNESSES

Hitchcock et al 1861	C. H. Richardson 1927–29	E. L. Perry 1929	P. H. Chang 1950	J. B. Thompson 1950	Names in this report Formation	Thickness
	Woodstock schist		Gile Mtn. fm.		Gile Mtn. fm.	2000
Calciferous	Woodstock schist		Standing Pond amphibolite	Standing Pond amphibolite	Standing Pond volcanics	600'
mica	Waits River limestone	Waits River limestone	Waits River fm.	Waits River fm.	Waits River fm.	5000' ±
schist	Memphremagog group, Randolph phyllite	Randolph phyllite	Northfield fm.	Northfield fm.	Northfield fm.	500–800'
				Shaw Mtn. fm.	Shaw Mtn. fm.	0–200'
				Cram Hill fm.	Cram Hill member	200–300'
gneiss	Barnard gneiss		Barnard fm.	Barnard gneiss	Barnard volcanic member	2000–3000'
	Missisquoi group	Missisquoi group	Missisquoi fm.	Moretown fm.	Moretown member	1500–2000'
				Whetstone Hill member	Whetstone Hill member	1500–2000'
Talcose	Bethel schist	Bethel schist	Bethel fm.		Moretown member	300–2000'
schist				Stowe fm.	Stowe fm.	700–2000'
	Ottauquechee schist	Ottauquechee phyllite and quartzite	Ottauquechee fm.	Ottauquechee fm.	Ottauquechee fm.	2600–3600'
	Pinney Hollow schist	Pinney Hollow schist	Pinney Hollow fm.	Pinney Hollow fm.	Pinney Hollow fm.	2600'
	Plymouth dolomite, unnamed schist and quartzite	Older Cambrian group	Grahamville fm.	Grahamville fm.	Hoosac fm.	1200–1300'
	Sherburne conglomerate	Mendon series	Tyson fm.	Tyson fm.	Tyson	600–800'
Green Mtn. gneiss		Mount Holly Series	Mount Holly complex	Green Mountain complex	Mount Holly complex	?

(Missisquoi fm. and Moretown fm. labeled vertically; bracket on right: 5500'–0000')

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The profound unconformity separating the rocks of the Green Mountain core from younger strata to the east was clearly recognized by T. N. Dale (1916), but he also did not publish a map and later workers had such difficulty locating this boundary that its existence was doubted. Mapping by Thompson (1950) in the Ludlow quadrangle adjacent to the south and by Brace (1953) in the Rutland quadrangle to the west has shown conclusively, however, that this unconformity does indeed exist and that Dale (1916) had located it correctly at several points. It is now clear that the Mount Holly was highly metamorphosed, with emplacement of granitic rocks and pegmatite, then deeply eroded, long before the sediments giving rise to the younger formations were laid down.

The name Mount Holly complex is used here in the original sense of Whittle's Mount Holly series. Outcrops of the Mount Holly complex are limited to the southwest corner of the quadrangle. They consist mainly of layered biotite gneiss with extensive development of secondary chlorite and epidote. Sheared pegmatites are common. Granites, quartzites and mica schists are exposed in the parts of the Ludlow quadrangle immediately adjacent to the south. The rocks of the Mount Holly complex are overlain by dolomitic graywacke of the Tyson formation, locally conglomeratic near its base. The Mount Holly complex is extensively exposed in the Green Mountains to the west and southwest of the Woodstock area.

Tyson Formation

The name Tyson formation was given by J. B. Thompson (1950) to the rocks immediately above the Mount Holly complex and below the albite schist of the Hoosac formation.

This formation occupies the floor of the Plymouth Union Valley and the steep slope along its western side in the southwest part of this area. The whole formation dips gently toward the east and lies directly upon the Mount Holly complex. It is poorly exposed. Continuous outcrops across the strike can only be found along the small brooks.

The rocks of the lower part of this formation are chiefly metamorphosed arkose and graywacke with minor amounts of conglomerate and quartzite. They are generally medium to coarse grained, schistose and pale green to greenish brown. The chief constituents, in variable proportions, are quartz, chlorite, sericite and albite with a minor amount of dolomite. The rocks are commonly thick-bedded to massive, with occasional intercalated thin bands rich in chlorite or dolomite, or both. Slip-cleavage and crinkling on the plane of schistosity are common in these thin bands.

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Although the general appearance of the rocks is fairly persistent, the lithology in detail within this part of the formation (the lower member) shows some variation along the strike. The conglomerate that marks the base farther south in the Ludlow quadrangle is only locally present.

As seen in the sections exposed along the small brooks on the west side of Plymouth Union Valley the lower part of the Tyson formation is overlain conformably by white, massive orthoquartzite in beds up to a foot and a half thick. This becomes thinner bedded upwards where it gives way to carbonaceous black phyllite with pyrite cubes and, locally, porphyroblasts of albite. Parts of the phyllite are calcareous and it contains interbeds, generally a few centimeters thick, of fine-grained carbonaceous quartzite.

The quartzites and phyllites are overlain by greenish white albite schists with calcite and dolomite marbles in beds up to 3 feet thick, and discontinuous, massive beds of orthoquartzite, locally several tens of feet thick. Some of the quartzites are conglomeratic with the largest pebbles about the size of a robin's egg. This zone is poorly exposed in the Woodstock quadrangle but is well exposed in the adjacent parts of the Rutland quadrangle between Black Pond and Woodward Reservoir. When W. F. Brace was mapping the Rutland quadrangle these conglomerates were interpreted by him and one of us (JBT), as the basal conglomerate of the Tyson, the lower units being mapped separately as the Salt Ash formation. Further field work, however, and the new road cuts on U. S. Highway No. 4 north of West Bridgewater, have made the earlier position untenable, and all of the rocks shown as Salt Ash on Brace's map are now included in the Tyson Formation.

The highest unit of the Tyson formation is a massive yellow to pink weathering dolomite with rare intercalations of sandy dolomite and dolomitic quartzite, locally containing pebbles of quartz or microcline as much as one centimeter in diameter. Some of these sandy beds, and some of the albitic schists interbedded with the lower carbonates, are impregnated with magnetite, specular hematite, or both, and were once mined on a small scale as iron ore (Hager, A. D. in Hitchcock *et al*, 1861 p. 731–732). At the very top of the formation, at the contact with the overlying Hoosac formation, there is commonly a layer of specular hematite several centimeters thick. This may possibly represent a metamorphosed *terra rosa*.

The basal graywackes and conglomerates were referred by Perry (1929) to the Mendon series, and the upper units were the lower quartzite and dolomite of his "Older Cambrian" group. The Tyson formation is of

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Table 2

ESTIMATED MODES OF THE TYSON AND HOOSAC FORMATIONS

	Tyson formation			Hoosac formation			
	1	2	3	4	5	6	7
Porphyroblasts							
Chlorite	..	..	2	..	..	..	..
Biotite	..	..	..	..	5	..	5
Albite	15	..	40	..	..	..	..
Groundmass							
Quartz	30	45	..	30	35	78	42
Albite	..	..	..	30	13	5	..
Microcline	..	..	..	7	10	2	3
Sericite	35	35	50	25	35	..	40
Chlorite	5	20	..	..	..	..	5
Biotite	..	..	5	5	..	15	3
Epidote-zoisite	5	..	..	..	..	..	..
Calcite	..	..	..	3	..	..	..
Dolomite } 10		..	..	..	..	..	..
Ankerite		..	..	..	..	..	..
Rutile	..	tr	..	..	..	..	..
Titanite	..	..	..	tr	..	..	..
Ilmenite	..	..	..	..	..	..	} 2
Magnetite	tr	tr	2	..	2	tr	
Tourmaline	tr	tr	1	tr	..	..	..
Zircon	..	..	..	tr	..	..	tr
Apatite	..	tr	..	tr	..	..	..
Grain size in mm.							
Porphyroblasts	1.5		1		.3		
Groundmass	.04–.2	.1–.2	.1–.3	.05–.1	.01–.1	.1–.2	.1–.2
Texture	S*	S	S	S	S	S	S

1   Arkose
2   Graywacke
3   Albite schist
4–5 Sericite-albite-quartz schist
6   Laminated quartzite
7   Quartzose phyllite

S*—schistose

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variable thickness and to the south in the Ludlow quadrangle it is locally absent. In the Woodstock quadrangle it has an estimated thickness of 600–800 feet. The interbedded quartzites and schists are 250 to 300 feet thick, and the dolomite at the top is about 200 feet thick.

Hoosac (Grahamville) Formation

The name Grahamville was used by Thompson (1950) for the rocks overlying the Tyson formation in the Ludlow quadrangle adjacent to the south. Subsequent mapping farther south (Rosenfeld, 1954; Skehan, 1961) has shown that the rocks then called Grahamville were substantially equivalent to the Hoosac schist of western Massachusetts (Emerson, 1917). The designation Hoosac formation is used here in the interest of conformity with the recently published state geologic map (Doll, *et al*, 1961).

The Hoosac formation crops out in the southwest part of the area, east of the Plymouth Union valley. Its thickness here is not more than 1,200 to 1,300 feet, but it covers a fairly broad tract in the Woodstock quadrangle owing to structural repetition. The lower 300 to 400 feet of the Hoosac formation is a distinctive schist containing abundant porphyroblasts of clear albite in grains up to 5 mm in diameter. These typically comprise 15 to 20 percent of the rock, locally as much as 30 percent. The remainder is quartz, muscovite, biotite, minor magnetite, calcite, dolomite, pyrite and rare microcline. The basal foot or so, directly overlying the dolomites of the Tyson formation, with the iron-oxide layers referred to above, is commonly rich in magnetite. This may possibly represent in part a residual soil developed on the older Tyson formation.

The central part of the Hoosac formation is characterized by quartzite, thin-laminar feldspathic quartzite, gray to black mica schist and minor albite schist. The feldspathic quartzites are typically somewhat schistose and are conspicuously layered with laminae that may be only a few millimeters thick. Many of these quartzites are dolomitic and thin prisms of tourmaline up to a centimeter long are not uncommon on the foliation surfaces. Higher in the section the quartzites are purer and more massive with beds of orthoquartzite up to 50 feet thick occurring locally.

The upper part of the Hoosac is characterized by dolomite, dolomite breccia, dolomitic quartzite, schistose dolomite with carbonaceous partings, carbonaceous phyllite, minor quartzite and albite schist. Carbonaceous albitic schists are common immediately beneath the overlying Pinney Hollow formation. The quartzites and carbonate rocks in the middle and upper portions of the Hoosac formation are exceptionally

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well developed in this area and in the adjacent parts of the Ludlow and Rutland quadrangles. These were mapped as the Plymouth member by Thompson (1950) and have been so shown on the state map (1961). The dolomites attain their maximum thickness (100 feet or more) in the high valley running northwest from Plymouth village toward Grass Pond. The outcrop belt is marked by several old openings and abandoned lime kilns. The dolomites of the Hoosac are generally buff to gray weathering in contrast to those of the Tyson formation which are yellow to pink weathering and more massive. The dolomite breccias, made of light gray fragments in a darker gray matrix, are found only in the Hoosac, and the magnetite-hematite concentrations noted earlier are found only in the carbonate rocks of the Tyson formation.

Pinney Hollow Formation

This formation was first named by Perry (1929) as the Pinney Hollow schist. It lies directly on top of the Hoosac formation. It is best, though not continuously, exposed along the sides of the middle part of Pinney Hollow, whence the name is derived.

The Pinney Hollow formation is composed chiefly of pale-green, fine-grained, chlorite-sericite-quartz schist with a shiny, silvery appearance on the plane of schistosity. The individual beds differ in the proportions of the constituent minerals and in the thickness of the compositional bands. Porphyroblasts of garnet and biotite are common in some beds, but are generally less abundant than in rocks of similar chemical composition elsewhere in this area. In some beds in the lower part of the formation the garnet porphyroblasts are altered to chlorite, which either forms a thin coating on the garnets or replaces them entirely to form pseudomorphs. The schists are highly micaceous and were described by Hitchcock *et al* (1861) as "talcose" schist or "hydromica" schist. These rocks are nonetheless resistant to erosion and form many of the higher summits in the area. In most of the beds, the darker colored minerals, such as chlorite and chloritoid, tend to concentrate along parallel lines on the surfaces of schistosity and form a set of mineral lineations. Quartz veins are common in these schists and range from an eighth of an inch to several inches thick. Some cut across the schistosity; others are parallel to it. Most commonly the quartz veins take the form of highly distorted lenticular bodies only a few inches long.

To the south of the Ottauquechee Valley the base of the Pinney Hollow formation extends along the upper part of the western slope of the ridge connecting Blueberry Hill, East Mountain, Wood Peak and Mor-

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gan Peak. The whole formation occupies a belt trending north-northwest and has a breadth of three-quarters of a mile on both ends and one and one-half miles in the middle.

The lower part of this formation, which occupies a zone about half a mile wide, is characterized by the dominance of chloritoid-bearing, pale green, soft, fine-grained chlorite-sericite-quartz schist (Plate 3, Fig. 1; Plate 6, Fig. 1). Mineral lineations are well marked on the plane of schistosity. Slip-cleavage and crinkling are developed in association with the later minor folds. Both the lithologic character and the minor structural features are best shown in the road cuts on the north side of the Ottauquechee Valley one-half mile east of West Bridgewater and near the west end of Pinney Hollow. The schist near the base of this formation is in many places albitic. The albite forms porphyroblasts 2 to 4 mm in diameter and gives a coarser appearance to the rock. The white mica or sericite in the chloritoid schists and phyllites includes both muscovite and paragonite. In fresh road cuts near the upper end of Pinney Hollow some of the chloritoid schists have a purplish color because of finely divided hematite. Elsewhere small octahedra of magnetite are common.

Beds of greenstone or actinolitic greenstone occur in the lower part of the formation. The more northerly occurrences are free of amphibole and are rich in rusty-weathering carbonate. Though noted by Chang (1950) they were not shown separately on his map. The areas shown on Plate 1 include only localities seen by Ern and Thompson. The greenstones are probably more extensive than shown but clearly constitute a much lesser portion of the Pinney Hollow formation here than in other areas both north and south.

Above the lower chloritoid-bearing schist the rocks are more quartzose, and chloritoid is absent (Plate 6, Fig. 2). Thin bands of fine-grained quartzite and chlorite-sericite quartzite are intercalcated in the pale green phyllitic schists. Porphyroblasts of biotite and garnet are common. Slip-cleavage and crinkling on the plane of schistosity are rare. This part of the Pinney Hollow formation occupies a zone one-third to one-half mile wide.

Toward the top of the formation the schists become more and more quartzose with some black phyllite and phyllitic quartzite. The best exposures of the uppermost part of the Pinney Hollow formation are in the first two side-brooks on the south side of the Ottauquechee Valley east of West Bridgewater. Each of them shows a transition zone a third of a mile wide between the Pinney Hollow and the Ottauquechee formations. The pale green chlorite-sericite-quartz schist of the Pinney Hollow formation

TABLE 3

ESTIMATED MODES OF THE PINNEY HOLLOW FORMATION

	1	2	3	4
Porphyroblasts				
Ankerite	40	20	60	..
Groundmass				
Quartz	5	20	15	15
Oligoclase	..	4	..	..
Sericite	25	10	10	65
Chlorite	15	40	10	5
Biotite	tr	..	..	..
Epidote-zoisite	5	5	5	..
Chloritoid	..	..	..	15
Magnetite	..	1	..	..
Pyrrhotite Pyrite	10	..	..	..
Grain size in mm.				
Porphyroblasts	.8–1	1	1	
Groundmass	.1–.4	.1–.4	.05–.1	.1–.4
Texture	S*	S	S	S

1–3 Ankerite-sericite-chlorite schist
4 Chloritoid-sericite-chlorite-quartz schist

S*—schistose

is here locally albitic and biotitic. At least three distinct bands of black phyllitic quartzite, each about 200 feet wide, are intercalated in this zone. Similar conditions are found both farther north and south along the strike. The width of this transition zone, however, is not constant, and it may be wholly a zone of infolding of Pinney Hollow and Ottauquechee types This zone narrows toward the south and is not seen in the sections exposed on the hills south of Pinney Hollow, nor in the Ludlow quadrangle, but widens toward the north. Consequently, the upper boundary of the Pinney Hollow formation is rather difficult to locate in the area around the headwaters of Dailey Hollow. The eastern boundary of this transition zone has been mapped as the top of the whole formation.

The thickness of the Pinney Hollow formation is difficult to determine
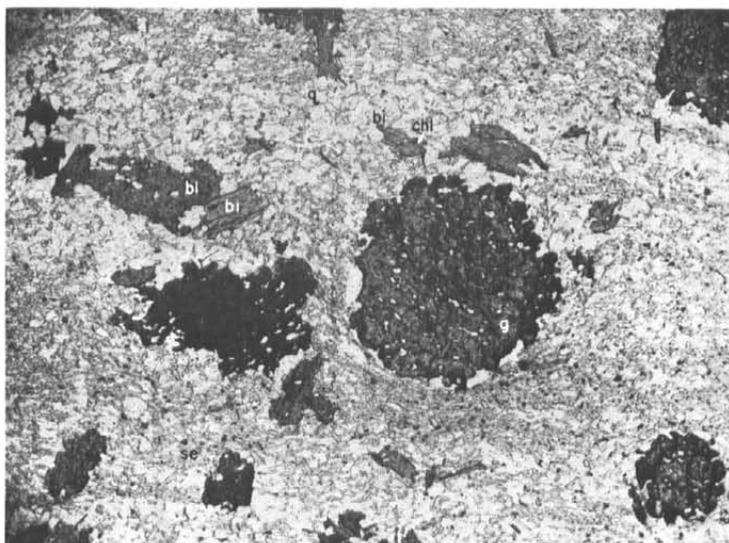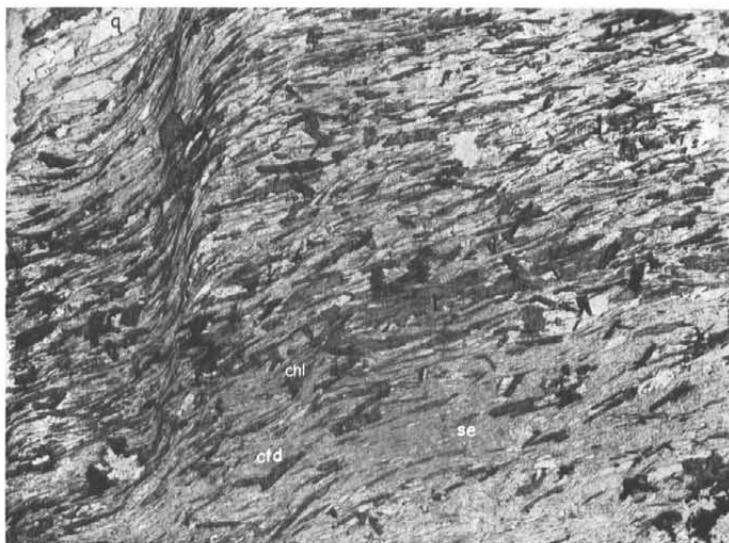
PLATE 3

Figure 1. Chloritoid schist, Pinney Hollow formation: Shows slip-cleavage and the development of chloritoid; ctd, chloritoid; se, sericite; chl, chlorite; q, quartz. (P.H.C.)

Figure 2. Garnet-biotite schist, Whetstone Hill member, Missisquoi formation: Shows typical development of porphyroblasts; g, garnet; bi, biotite; chl, chlorite; se, sericite; q, quartz. (P.H.C.)

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because of minor folds, but it is estimated from the structure sections to be 2,600 feet.

Ottauquechee Formation

This formation was first named by Perry (1929) as the Ottauquechee group. Composed mainly of black phyllite and quartzite, it lies stratigraphically above and immediately to the east of the Pinney Hollow formation, with a width of five-eighths mile at its south end to nearly two miles farther north.

Black phyllite is the predominant rock in this formation. It is fine grained and thinly foliated. The constituent minerals are not all visible to the naked eye, except the comparatively rare biotite and garnet porphyroblasts. Quartzite in beds or lenticular bodies ranging from a fraction of an inch to several feet thick are interbedded with the black phyllite. It is usually coarser grained than the phyllite and ranges in color from pure white to black. Small flakes of biotite and sericite are scattered through the quartzite beds.

In the lower part of the formation quartzite in distinct beds or lenticular bodies up to ten feet thick is abundant. Biotite flakes, either parallel to or across the schistosity, are the only porphyroblasts in the phyllite in this lower part of the formation. Toward the top of the formation the quartzite is in thin bands or layers crinkled and folded with the black phyllite. Near the top of the formation the phyllite contains less interbedded quartzite. Just east of Five Corners, through a zone about 300 feet wide, the black phyllite is calcareous and contains bands and beds of coarse-grained marble up to one foot thick. Porphyroblasts of garnet and biotite occur in the black phyllite in the uppermost part of the Ottauquechee formation.

Lenticular bodies of greenstones occur in the lower and middle parts of the Ottauquechee formation. The foliation, which is weak to well developed, is generally parallel to that in the enclosing phyllite and quartzite. The greenstone bed exposed in the road cut on the south side of the Ottauquechee Valley near the Riverside School is about ten feet thick and is composed of fine-grained epidote, sericite, albite, and quartz. New highway cuts in the Riverside School area show the greenstone to be associated with pale green but locally carbonaceous schist in which granulite layers several millimeters thick alternate with layers of approximately equal thickness containing mainly biotite and garnet. This "bacon-rock" is characteristic of the Ottauquechee in many areas to the south. Similar greenstone occurs on the eastern slope of the hill west of the west end of Dailey Hollow.

TABLE 4

ESTIMATED MODES OF THE OTTAUQUECHEE FORMATION

	1	2	3	4	5	6	7	8
Porphyroblasts								
Biotite	..	..	20	..	..	..	..	..
Garnet	1	1	..	..	..	..	..	..
Hornblende	..	..	..	..	..	..	30	..
Ankerite	..	..	..	..	..	..	20	..
Groundmass								
Quartz	49	52	38	98	30	5	20	5
Albite	..	..	..	..	10	..	4	15
Oligoclase	..	..	..	..	..	10	..	..
Sericite	47	30	20	1	35	..	..	..
Chlorite	..	15	5	..	..	15	20	..
Hornblende	..	..	..	..	..	45	..	45
Epidote-zoisite	..	..	15	..	25	25	5	25
Ankerite-dolomite	..	..	..	..	..	..	..	10
Ilmenite	2	..	2	..	..	..	1	..
Magnetite	..	..	..	..	..	tr	..	..
Pyrite	..	..	tr	..	..	..	..	tr
Carbon	1	2	..	1	..	..	..	..
Grain size in mm.								
Porphyroblasts:	.5–1	.5–1	.6				3	
Groundmass:	.06–.1	.06–.1	.04–.2	.1–.4	.5–1.5	.6–1	.04–.1	.4–2
Texture	S*	S	S	S	S	S	S	S

1–2 Black phyllite
3   Black schist
4   Black quartzite
5   Epidote-sericite-quartz schist
6   Epidote amphibolite
7–8 Ankerite amphibolite

S*—schistose

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Minor folds are common in this formation. However, the slip-cleavage, and crinkling are only locally and weakly developed. The formation is estimated to be from 2,600 to 3,600 feet thick.

Stowe Formation

The Stowe formation (Osberg, 1952) was named by Cady (1956) from exposures near Stowe, Vermont, about 60 miles north of the Woodstock area. It corresponds fairly closely with the unit mapped by Perry (1929), and Chang (1950) as the Bethel schist, following Richardson (1925). The principal difference is that units now assigned to the overlying Missisquoi formation were mapped by Perry and Chang as the upper part of the Bethel. The Bethel-Missisquoi boundary mapped by Perry is approximately the base of the present Whetstone Hill member of the Missisquoi. The revised boundary, mapped here as the Stowe-Missisquoi boundary, has proved much more persistent regionally. Near Hyde Park, Vermont, this boundary is marked by a conglomerate, the Umbrella Hill formation (Albee, 1957), shown as a basal member of the Missisquoi by Doll, *et al* (1961). The Umbrella Hill overlies rocks of the Stowe formation unconformably indicating a distinct break in the stratigraphic succession.

The rocks of the Stowe formation are mainly pale green, quartz-sericite-chlorite schists much like those of the Pinney Hollow formation, but with more abundant and conspicuous porphyroblasts of garnet and biotite. There is also a more marked segregation of the quartzose and micaceous components into discontinuous laminae of comparable thickness.

Numerous lenses of coarse-grained quartz are about one to two inches thick and have boundaries generally parallel to the schistosity. Some have very irregular shapes, presumably the result of strong folding. Crosscutting quartz veins of various sizes are also common.

The Stowe formation occupies the zone immediately east of the Ottauquechee formation and has a width of less than a quarter mile in the south, widening northward to about $1\frac{1}{2}$ miles. The schists of the Stowe like those of the Pinney Hollow are resistant to weathering. Outcrops commonly preserve glacial striae and several high ridges stand in this zone. The contact between this formation and the Ottauquechee is marked by a rather abrupt change upward from black phyllite to pale green, chlorite-sericite-quartz schist. Throughout most of the formation, the rocks are uniformly pale green, although the relative proportion of the constituent minerals and the kind and amount of porphyroblasts may differ locally. The maximum thickness is estimated as about 2,000 feet, but near the south edge of the area the formation is about 700 feet thick.

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TABLE 5

ESTIMATED MODES OF THE STOWE FORMATION

	1	2	3	4	5
Porphyroblasts					
Chlorite	..	..	4	..	..
Biotite	..	..	20	10	1
Garnet	..	..	..	..	5
Groundmass					
Quartz	23	47	20	25	27
Albite	..	10	..	..	5
Sericite	60	35	40	60	30
Chlorite	15	6	15	5	30
Ilmenite	} 2	} 2	..	..	} 2
Magnetite			1	..	
Pyrrhotite	..	..	..	tr	..
Zircon	..	..	tr	..	..
Grain size in mm.					
Porphyroblasts			2	.4–2	1
Groundmass	.05–.1	.05–.1	.05–.1	.05–.1	.05–.1
Texture	S*	S	S	S	S

1–2 Chlorite-sericite-quartz schist

3–4 Chlorite-sericite-quartz schist with biotite-porphyroblasts.

5   Chlorite-sericite-quartz schist with garnet-porphyroblasts.

S*—schistose

Missisquoi Formation

The name Missisquoi group was used by Richardson to include all the metamorphosed sedimentary formations between the base of his Crdovician formations and the top of the Bethel schist in central Vermont (Richardson, 1925, 1927). In the southwest corner of Woodstock township Richardson (1927) mapped a belt of sericite schists, which he included with other rock types such as quartzite, chlorite schist, and hornblende schist, under the name Missisquoi group. The northward continuation of the same belt in Bridgewater township was mapped by Perry (1929). The Missisquoi formation lies immediately east of the area occupied by the Stowe formation and occupies a zone about 2½ miles wide in the south, widening northward to six miles.

Table 6

ESTIMATED MODES OF THE MISSISQUOI FORMATION

	1	2	3	4	5	6	7	8	9
Porphyroblasts									
Chlorite	..	..	2	2	..	..	5	1	5
Biotite	..	14	15	15	8	5	..	..	..
Garnet	1	5	..	15	..	1	10	10	..
Groundmass									
Quartz	10	10	20	38	40	40	45	54	55
Albite	..	?	13	..	..	47	12	..	15
Sericite	69	70	25	30	50	tr	..	15	10
Chlorite	15	tr	..	..	..	5	25	..	..
Biotite	..	..	5	..	..	..	1	20	..
Garnet	..	..	20	..	..	..	..	..	..
Epidote-zoisite	..	..	..	..	..	..	..	..	2
Ilmenite	} 5	..	..	} tr	} 2	..	..	..	..
Magnetite		1	..			2	2	..	3
Carbon	tr	..	..	..	..	..	..	..	..
Grain size in mm.									
Porphyroblasts	.2	.3–1.5	.4–1	1	.5–1	.4–1.5	.4–1	1.2	.6–1
Groundmass	.01–.04	.04–.1	.02–.2	.06–.1	.04–.1	.04–.1	.04–.2	.1–.2	.06–.4
Texture	S*	S	S	S	S	Gr**	S	S	Gr

1 Black phyllite
2–4 Black schist
5–9 Dark-gray quartzose schist

S*—schistose
Gr** granular

Richardson (1927) subdivided the rocks of this formation into two parts, a lower sericite schist and an upper sericite quartzite. This arrangement was followed by Chang (1950) and also by Perry (1929), at least in his description, but the subdivision was not shown on his map. To the north, in Barnard township, Richardson mapped separately a belt of gneiss in the upper part of the sericite quartzite of the Missisquoi

TABLE 6 (*continued*)

ESTIMATED MODES OF THE MISSISQUOI FORMATION

	10	11	12	13	14	15	16	17
Porphyroblasts								
Chlorite	..	..	..	2	..	10	2	..
Biotite	..	5	10	13	5	..	..	10
Garnet	20	..	..	10	..	..	..	..
Hornblende	..	..	..	..	..	..	..	..
Ankerite	..	..	..	..	..	..	..	..
Groundmass								
Quartz	5	22	35	45	45	55	60	65
Albite	8	10	4	15	15	5	10	15
Oligoclase	..	..	..	..	..	..	..	..
Sericite	50	40	40	15	5	30	25	5
Chlorite	2	1	5	..	20	tr	3	5
Biotite	5	..	..	..	5	..	..	tr
Hornblende	..	..	..	..	..	..	..	..
Epidote-zoisite	..	10	1	..	5	..	tr	..
Calcite	2	..	..	..	..	..	..	..
Titanite	..	..	..	..	..	tr	..	..
Ilmenite	..	2	..	..	..	..	tr	..
Magnetite	10	..	5	..	..	..	..	tr
Pyrite	..	..	..	..	..	..	..	..
Apatite	..	..	..	..	tr	..	..	..
Grain size in mm.								
Porphyroblasts	1.5	.6	1	.4	.8–1	.5	.6–1	.4–1.5
Groundmass	.1–.2	.04–.1	.1–.3	.04–.2	.1–.2	.05–.2	.02–.4	.04–.1
Texture	S*	S	S	S	S	S	S	S

10–17 Pale-green sericite-chlorite-quartz schist

S*—schistose

group and named it the Barnard gneiss. He mapped the same gneiss at a similar structural position in Woodstock township and considered it to be a metamorphosed plutonic rock (Richardson, 1927). The Missisquoi formation, as presently delineated in Vermont, and as mapped for this report, includes the Moretown formation of Cady (1956), the Whetstone

TABLE 6 (*continued*)

ESTIMATED MODES OF THE MISSISQUOI FORMATION

	18	19	20	21	22	23
Porphyroblasts						
Chlorite	..	..	20	..	..	..
Biotite	5	..	..	..	..	..
Garnet	1	..	..	2	..	..
Hornblende	..	..	8	10	..	20
Ankerite	..	..	20	5	..	..
Groundmass						
Quartz	68	76	20	3	10	30
Albite	20	2	..	23	10	25
Oligoclase	..	..	5	..	..	..
Sericite	5	10	..	..	..	10
Chlorite	1	5	10	2	30	3
Biotite	..	2	10	..	5	4
Hornblende	..	..	2	30	40	..
Epidote-zoisite	..	..	5	25	3	..
Calcite	..	..	..	..	..	..
Ilmenite	..	..	..	..	2	..
Magnetite	..	3	..	..	..	1
Pyrite	..	tr	tr	..	..	1
Apatite	..	..	..	..	..	1
Grain size in mm.						
Porphyroblasts	.3		1–2	.6		2–5
Groundmass	.04–.1	.04–.2	.04–.2	.04–.08	.1–.4	.02–.2
Texture	S*	S	S	S	S	S

18   Biotite-sericite-quartz schist
19   Pale-green chlorite-sericite quartzite
20–21   Ankerite amphibolite
22   Hornblende-epidote-chlorite schist
23   Amphibolite

S*—schistose

Hill member of the Moretown formation (Thompson, 1950), the Cram Hill formation of Currier and Jahns (1941), and the Barnard gneiss of Richardson (1927). These units have since been relegated to member status within the Missisquoi formation (Doll, *et al*, 1961). The Missisquoi formation is estimated to be about 5,500 feet thick in the south, thickening northward to about 9,000 feet.

WHETSTONE HILL MEMBER

Porphyroblastic dark-gray to black phyllite and schist and interbedded micaceous quartzites and volcanic rocks within the Missisquoi formation were mapped as the Whetstone Hill member to the south, in the Ludlow quadrangle, by Thompson (1950). These rocks contain many porphyroblasts of garnet and biotite (Plate 3, Fig. 2). The included carbonaceous matter is erratically distributed and does not appear to have any stratigraphic significance. Tourmaline is an abundant accessory in these rocks and finely disseminated albite and quartz are present in minor quantities. Good exposures are found along the bottom of Hale Hollow. Here the black biotitic quartzite and carbonaceous phyllitic schist are interbedded, with individual beds ranging from one to three inches thick. Graded bedding is locally developed.

The percentage of garnet and biotite porphyroblasts decreases from south to north through the area and the less quartzose rocks become more phyllitic and less schistose in appearance.

A distinctive bed of light pink, fine-grained garnetiferous quartzite, two to three feet thick, persists along the strike for several miles in the central part of the quadrangle. Above this bed, in the area around Freestone Hill, to the north on Bull Hill, and south of the Ottauquechee River, a pale-green chlorite-sericite-quartz schist carries several beds of garnetiferous amphibolite and hornblendic quartzite. The amphibolite layers are three to six feet thick, and contain prismatic crystals of green hornblende up to one inch long set in a light-gray, fine-grained granular aggregate of quartz and albite. The best exposures are found in the bed of the North Branch of the Ottauquechee River northwest of Bridgewater Center and at the east end of Dailey Hollow. Amphibolite also occurs on the west slope of Richmond Hill and on the next hill south in association with greenstone fragmentals.

Hager (in Hitchcock, *et al*, 1861, p. 731–2) was the first to describe an unusual occurrence of "ironstone" in the lower part of the Whetstone Hill member east of Plymouth. Sporadically exposed and variable in thickness, the unit consists of magnetite, pink garnet, black-weathering

carbonate and variable amounts of quartz. Similar rocks were noted farther south in the Ludlow quadrangle by Thompson (1950) at about the same stratigraphic position. This is also the approximate position of the garnetiferous quartzites mentioned above.

The base of the Whetstone Hill member coincides with the base of the Missiquoi group as mapped by Perry (1929) and of the lower member of the Missisquoi formation as originally mapped by Chang (1950) in the southern two-thirds of the quadrangle.

The distinctive Whetstone Hill lithology occupies the axial portion of the Missisquoi formation throughout most of the Woodstock quadrangle and thins northward. It terminates (Ern, 1963) in the Randolph quadrangle. To the south it thins perceptibly west of the Chester dome (Thompson, 1950). In the Woodstock quadrangle it is estimated to be from 1,500 to 2,000 feet thick.

MORETOWN MEMBER

In the Woodstock quadrangle the Moretown member of the Missisquoi formation is split into an upper and a lower portion separated by the Whetstone Hill member. The most abundant and distinctive rock type is a granulite composed of quartz, chlorite, albite and sericite in which granular layers averaging about half an inch thick are separated by paper-thin, darker-colored, micaceous partings producing a "pinstripe" texture (Plate 4, Figs. 1 and 2; Plate 7, Fig. 1). Quartz and albite make up roughly 90 percent by volume of the granular laminae and chlorite, sericite and epidote constitute the major portion of the schistose laminae. The quartz and albite in the granular laminae are commonly anhedral and are somewhat elongate parallel to the schistosity.

The pinstripe appears to be of primary origin. The pinstripe laminae are concordant with the bedding and in certain outcrops a distinct slip cleavage can be seen to cut across the pinstripes.

Micaceous quartzites are also abundant. Carbonaceous phyllites and greenstones are present only locally. Biotite and garnet porphyroblasts are fairly common in most rock types. The lower part of the Moretown member is estimated to be from 300 (in the south) to 2,000 feet thick, and the upper part is from 1,500 to 2,000 feet thick.

BARNARD VOLCANIC MEMBER

The above name is used on the 1961 edition of the state geological map. Richardson (1927) named the belt of gneissic rocks occurring in Barnard township the Barnard gneiss. This member is composed mainly

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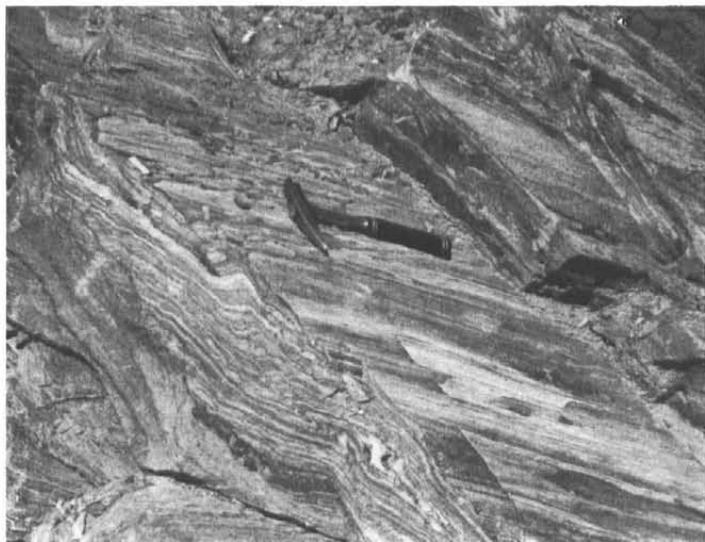
PLATE 4

Figure 1. Moretown member, Missisquoi formation: Dailey Hollow, northeast of Ragged Hill, Bridgewater. Outcrop shows development of pinstripe. (E.H.E.)

Figure 2. Moretown member Missisquoi formation: Thin section shows typical development of pinstripe, also folding; se, sericite; chl, chlorite; q, quartz. (P.H.C.)

TABLE 7

ESTIMATED MODES OF THE BARNARD VOLCANIC
MEMBER OF THE MISSISQUOI FORMATION

	1	2	3	4	5	6	7	8	9	10
Porphyroblasts										
Chlorite	..	5	..	2	20	..	..	5	2	..
Biotite	10	..	..	15	..	..	..	..	5	..
Garnet	1	..	..	1	..	5	..	..	2	2
Hornblende	..	..	..	..	20	20	..	30	20	..
Albite	20	20	..	20	..	..	..	..	..	..
Groundmass										
Quartz	25	35	38	24	14	40	40	30	41	77
Albite	32	40	28	32	20	..	26	10	5	10
Oligoclase	..	..	..	..	..	20	..	..	..	..
Sericite	3	..	3	..	5	2	..	..	tr	5
Chlorite	2	..	3	..	..	2	5	..	15	..
Biotite	..	..	15	..	15	3	20	..	..	3
Hornblende	..	..	..	..	..	..	5	..	..	..
Epidote-zoisite	5	..	10	5	5	8	2	25	10	3
Calcite	..	..	..	..	1	..	..	..	..	..
Ilmenite	..	..	..	..	..	..	2	..	..	..
Magnetite	2	..	3	1	..	..	..	..	..	..
Apatite	..	..	tr	..	tr	..	..	..	..	..
Grain size in mm.										
Porphyroblasts	.5–3	.5–2	.5	.6–1	3	1–3		1–3	3	2
Groundmass	.04–.2	.04–.1	.04–.1	.02–.1	.06–.2	.06–.2	.04–.6	.06–.2	.06–.3	.04–.1
Texture	S*	S	S	S	S	S	S	S	S	Gr**

1–4 Light-colored gneiss
5–6 Hornblende gneiss
7   Biotite-hornblende gneiss
8   Amphibolite
9   Hornblende-chlorite-quartz schist
10   Sericite-epidote quartzite

S*—schistose
Gr**—granular

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of metamorphosed, interbedded light-and dark-colored metavolcanic rocks with a small amount of schist and phyllite of sedimentary origin. It occupies a zone ranging from one-half mile to two miles wide east of the upper part of the Moretown member.

The commonest rocks of this formation are biotite gneiss, hornblende-sericite-chlorite-quartz schist, hornblende gneiss, chloritic amphibolite, and chlorite-hornblende schist. They are medium to coarse grained, well foliated, and occur in distinct beds. The color of these rocks, depending upon the amount of the dark-colored minerals such as hornblende, biotite, and chlorite, ranges from light gray to green or dark green. The light-colored gneisses, composed of quartz and albite, form bands two to five mm thick, intercalated with much thinner bands of biotite, chlorite or hornblende. Many flakes of sericite and small crystals of garnet and epidote are scattered through the rock. Calcite is present in some of the gneisses. The hornblende-rich gneisses in places pass into amphibolite, but more frequently the amphibolite forms sharply defined beds a few inches to a few feet thick, interbedded with much thicker beds of biotite- or hornblende-gneiss.

A narrow transition zone, a few tens of feet wide at the southern end of the area and up to about 100 feet wide at the northern end, lies between this formation and the underlying Moretown member. The chlorite-sericite-quartz schist of the latter becomes richer upward in thin beds of ankerite-bearing, dark-green, chloritic amphibolite. The amphibolite is fine to medium grained, with thin prismatic crystals of green hornblende and flakes of chlorite on the well-developed schistosity. Rhombohedral crystals of ankerite, many of which are rusty and porous due to weathering, form small spots in the rock. In this transition zone quartzitic layers interbedded with pale-green schists become more and more feldspathic upward in the stratigraphy. The highest strata in the transition zone are made up entirely of interbedded hornblende gneiss, greenish-gray garnetiferous biotite gneiss, and dark-green amphibolite, with or without the ankerite spots.

Generally, amphibolite beds are more abundant in the lower part of the Barnard formation than in the upper part. Excellent outcrops are exposed intermittently along the brook in Curtis Hollow for 1½ miles north of Curtis Hollow School. Near the school, in the river bed, a continuous section about 700 feet long shows the interbedded relationship between amphibolite and gneiss (Plate 7, Fig. 2). The beds are folded and locally sheared, but in none of these exposures is there any evidence of crosscutting relations along the contacts between the light-colored and dark-colored layers.

As calculated from the modes of these rocks, the chemical composition of the amphibolite is very close to that of quartz basalt, and that of the light-colored gneiss is similar to that of dacite or rhyodacite. Relicts of altered phenocrysts of albite and quartz are common in thin sections of the gneisses. A volcanic origin for the rocks in this formation is thus strongly indicated.

The upper part of this formation is poorly exposed through much of its extent. However, the available outcrops in this zone indicate that the amphibolite beds are less abundant here relative to the light-colored gneisses. Some of the light-colored gneisses in the upper part are sericitic and contain few mafic minerals. Except for the scattered larger grains of albite, distinguished from quartz by their chalky weathering, the rock resembles a medium-grained quartzite.

The northern continuation of the Barnard volcanic member in the Randolph quadrangle passes laterally into the Cram Hill member of the Missisquoi formation in the valley of Trout Brook south of the village of Randolph (Ern, 1963). South of the Woodstock quadrangle, the gneisses also pass laterally into Cram Hill schists in the vicinity of Proctorsville (Thompson, 1950). The Barnard volcanic member is perhaps 2,000 to 3,000 feet thick.

CRAM HILL MEMBER

The Cram Hill member has been recognized only in the southern part of the Woodstock quadrangle between the south end of Ohio Hill and the quadrangle boundary. There the rock consists of rusty weathering, locally carbonaceous phyllite and sericitic quartzite. Some of the phyllite has sprays of hornblende, an inch or more long, on the schistosity surfaces.

At Bailey Mills in Reading the Barnard volcanic member overlies the Cram Hill member in a reversal of the stratigraphic order found elsewhere in the Woodstock quadrangle. The Bailey Mills exposures, however, are in a distinct structural belt from the others. Such a relation is not surprising in view of the lateral equivalence of these two members that has been demonstrated elsewhere. The Cram Hill member is not more than 200–300 feet thick in the Woodstock quadrangle.

Shaw Mountain Formation

The type area for the Shaw Mountain formation is near Northfield, Vermont, (Currier and Jahns, 1941). In the Woodstock quadrangle it is mainly quartz conglomerate, quartzite and quartz-mica schist.

The areas of outcrop of the Shaw Mountain are shown on Plate 1. Both

are south of the Ottauquechee River on Long Hill. The total thickness of the formation, where present, is nowhere greater than 200 feet, although in places near the principal summit of Long Hill the width of outcrop is as great as 400 feet because of folding. The formation is undoubtedly more extensive than shown on Plate 1, but is easily missed, owing to its thinness where exposure is sparse. Loose blocks of the distinctive quartz conglomerate were observed on most crossings of the Shaw Mountain horizon, between the south edge of the area and Ohio Hill, but the formation has been mapped only where definite outcrops of the conglomerate or its associated orthoquartzite were found.

The quartz conglomerates occur in beds a few inches to a foot and a half thick and are nearly pure quartz. Cigar-shaped pebbles of vein quartz may be up to two inches long and one half inch in diameter. The long axes plunge gently north-northeast, parallel to the more conspicuous fold axes in their vicinity.

The associated rock types, generally overlying the pebbly beds, range from nearly pure orthoquartzite to a yellow-weathering, quartz-mica schist containing abundant small garnets rarely more than two mm in diameter.

Northfield Formation

This formation was called the Randolph phyllite by Richardson (1923, 1927, 1929). The name Northfield slate was introduced by Currier and Jahns (1941) for rocks of similar stratigraphic position farther north in the Montpelier area.

The Northfield formation in the northern part of this area is in direct contact with the underlying Barnard volcanic member of the Missisquoi formation. South of Ohio Hill it rests upon the Cram Hill member of the Missisquoi formation except on Long Hill northeast of Curtis Hollow school and farther south, on the highest summit, where conglomerates and quartzites of the Shaw Mountain formation intervene.

There is no sharp lithologic break between the Northfield formation and the overlying Waits River formation. The boundary between the two was shown at a markedly higher stratigraphic position by Chang (1950) than on the map accompanying this report. Limestone increases upward in the section both in abundance and in thickness of individual beds. Chang's boundary has been revised downward on Plate 1 to agree more closely with the mapping of White and Jahns farther north where the formations were first established. In practice the boundary is placed so that limestone beds more than a foot thick are nearly all included within the Waits River formation.

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Table 8

ESTIMATED MODES OF THE NORTHFIELD
AND THE WAITS RIVER FORMATIONS

	1	2	3	4	5	6	7	8	9	10
Porphyroblasts										
Biotite	..	..	5	..	..	..	..	..	..	..
Garnet	5	1	1	20	..	..	..	..	..	..
Groundmass										
Quartz	20	10	20	55	73	90	30	45	40	10
Oligoclase	..	..	..	..	..	..	..	3	..	..
Sericite	70	80	70	5	20	2	10	2	8	5
Chlorite	1	6	2	..	2	5	..	..	..	..
Biotite	..	2	..	15	5	3	..	..	tr	5
Zoisite-epidote	..	..	..	..	..	..	..	..	..	10
Calcite	..	..	..	..	..	..	60	50	50	70
Ilmenite	..	..	..	3	..	..	..	..	..	..
Magnetite	2	1	..	..	..	..	..	..	..	..
Pyrrhotite	..	..	2	..	..	..	..	..	..	..
Carbon	2	tr	tr	2	tr	..	..	..	2	tr
Grain size in mm.										
Porphyroblasts	2–3	3	3–5	10						
Groundmass	0.2	0.8	.04–.1	.1–.2	.4	.3–.5	.4	.2–.5	.3–.5	.3–1
Texture	S*	S	S	S	S	S	Gr**	Gr	Gr	Gr

1–4 Phyllite
5–6 Impure quartzite
7–10 Metamorphosed impure limestone

S*—schistose
Gr**—granular

Rocks of this formation are mainly black phyllite with minor amounts of impure quartzite and metamorphosed impure limestone. The phyllite is dark-gray to black and is well foliated. The constituent minerals are sericite, quartz, chlorite, biotite, and garnet. The rock is rich in disseminated, fine-grained, carbonaceous material to which the dark color is due. Garnet forms porphyroblasts ranging in size from two mm to one cm in diameter and is locally abundant. Biotite occurs chiefly in patches on

the plane of foliation. The rock is intensely plicated and is characterized by the development of fine crinkles in the schistosity. Quartz lenses of various sizes are common. The thin quartzites of this formation are typically schistose. Porphyroblasts of garnet and biotite generally occur in a matrix of medium- to coarse-grained quartz and sericite. The metamorphosed impure limestones that are the distinctive rocks of the overlying Waits River formation also occur as rare beds up to several inches thick in the Northfield formation. The limestones are coarse to medium grained granular, dark bluish-gray when fresh, and rusty brown after weathering. Their average mineral composition is estimated as 40 to 50 per cent calcite, 25 to 40 per cent quartz, and 10 to 25 per cent other silicates, such as muscovite, biotite (partly phlogopite), zoisite, and garnet. Because of these impurities and its granular texture, this so-called limestone generally has an arenaceous appearance. It forms beds a few inches thick interbedded with the black phyllite. The thickness of the individual beds is not constant and pinching out of the beds in a short distance is commonly observed.

The apparent thickness of the Northfield formation is between 500 and 800 feet. In the extreme south part of the area near Bailey Mills it has an apparent thickness of only about 250 feet but this probably is the result of tectonic thinning during the emplacement of the Chester dome (Thompson, 1950).

Waits River Formation

The Waits River formation (Currier and Jahns, 1941) occupies most of the eastern half of the Bridgewater-Woodstock area. It is composed of the same rock types as the underlying Northfield formation, but in different relative proportions.

The lower part of the Waits River formation is composed chiefly of thick-bedded, metamorphosed, impure limestone (Plate 5, Fig. 1; Plate 8, Fig. 2) and black garnetiferous phyllite (Plate 9, Fig. 2), the proportion of limestone increasing upward. Beds of metamorphosed impure limestone 10 feet or more thick are common. Quartzite is rather rare in these rocks.

Farther east, near the contact with the Standing Pond volcanics the amount of impure limestone in the black phyllite becomes less and less. Locally, even though thin-bedded impure limestone is still present in small amounts, the rock is very quartzose and rich in garnet porphyroblasts. Some areas of this rock type, more typical of the Gile Mountain formation farther east, have been shown separately on the map (Plate 1).

The area occupied by the Waits River formation is characterized by
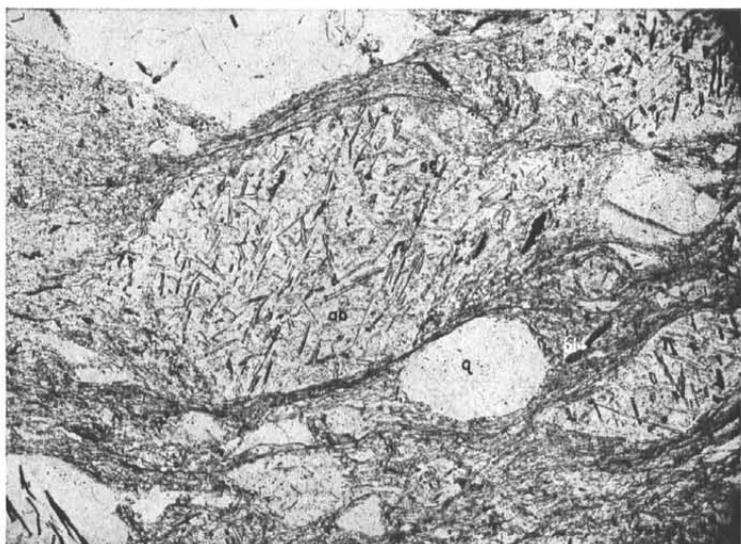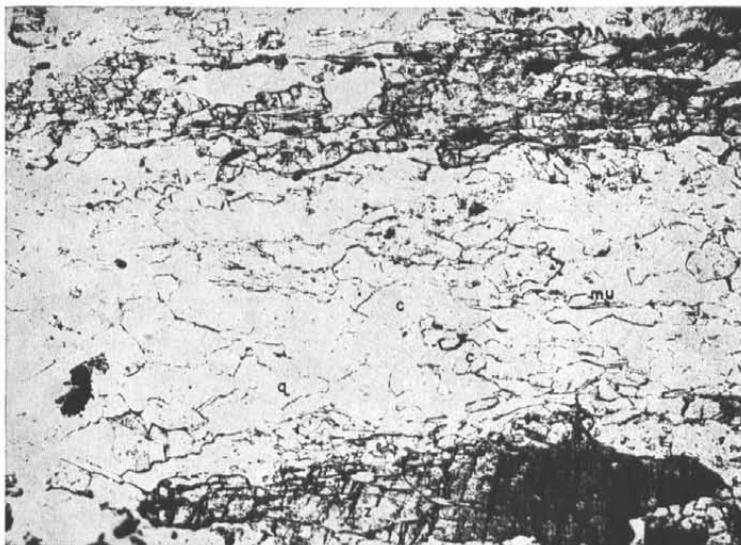
PLATE 5

Figure 1. Limestone with zoisite, Waits River formation: c, calcite; q, quartz; mu, muscovite; z, zoisite. (P.H.C.)

Figure 2. Dacite porphyry, intrusive in Hoosac formation: Shows development of schistosity and oriented sericite flakes in plagioclase; ab, albite; q, quartz; se, sericite; bi, biotite. (P.H.C.)

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isoclinal, recumbent folds making an accurate estimate of its thickness difficult. The problem is compounded by the fact that the direction of stratigraphic tops across the Standing Pond volcanics, adjacent to the east, has not yet been established unequivocally. If the Standing Pond *overlies* the Waits River stratigraphically, as we have arbitrarily assumed in the legend for Plate 1 and in the format of this report, the thickness of the Waits River is probably at least 5,000' except in the southern part of the area (Plate 1, Section D-D') where it thins, presumably for tectonic reasons, to about 2,000'. If, on the other hand, the Standing Pond *underlies* the adjacent parts of the Waits River then the latter, as the youngest formation in the area, has no observed roof and may be much thicker than the above estimate.

Standing Pond Volcanics

The name Standing Pond was introduced by C.G. Doll (1944) as the Standing Pond amphibolite member of the Memphremagog (=Waits River) formation. We shall take the option here, however, of designating it as a distinct formation in its own right, following Billings *et al* (1952) and Ern (1963). Where mapped by Doll (1944) the rocks on both sides of the Standing Pond are impure limestones and schists of typical Waits River type, those to the east giving way eventually to relatively noncalcareous types mapped by Doll as the Gile Mountain formation. By Doll's criteria, however, the Waits River–Gile Mountain boundary would have to cross the Standing Pond in the east part of the Woodstock area, making it untenable there to regard the Standing Pond as a member of the Waits River. We have therefore adopted the convention of designating the Standing Pond a separate formation so that the rocks to the west of the Standing Pond belt are arbitrarily Waits River and those to the east are arbitrarily Gile Mountain. This means that both the Waits River and Gile Mountain formations have calcareous and non-calcareous parts as indicated on Plate 1.

The distinctive rocks of the Standing Pond make it a very important horizon marker (Doll, *et al*, 1961). It is characterized by the association of bedded amphibolite and garnet-hornblende-quartz-mica schist. Many of the garnet porphyroblasts in these beds are as much as two inches in diameter. The hornblende also forms large prismatic crystals one to four cm long lying in the schistosity. The associated schists are generally rich in biotite and the grain size of the constituent minerals is much coarser than in the phyllites and schists elsewhere in the area.

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The Standing Pond volcanics crop out along the eastern slope of Mt. Peg and Baylies Hill southeast of the village of Woodstock and continue northward to the southern slope of the hill on the north side of the Ottauquechee Valley one mile northeast of Woodstock. Other occurrences are in the vicinity of Pomfret village and in a small tract in West Windsor. It is extensively exposed farther east in the Strafford, Hanover and Claremont quadrangles. Its thickness in the Woodstock area is estimated to be 600 feet.

Gile Mountain Formation

The name Gile Mountain formation was introduced by C.G. Doll (1944) in the Strafford quadrangle. This formation occupies a small area where it is very poorly exposed east of the village of Woodstock. There is also an area in Pomfret, and another in Barnard. Rocks belonging to this formation in these areas are chiefly garnetiferous black phyllite or schist, micaceous quartzite, and quartzose schist. The garnetiferous schists carry kyanite locally, and rare staurolite. Except for the relative lack of calcareous rocks, the lithologic character of this formation is rather similar to that of the upper part of the Waits River formation. According to J.B. Lyons (1955), the Gile Mountain formation is widely exposed in the Hanover quadrangle and continuous northward to its type locality. The parts of the Gile Mountain formation exposed in the Woodstock quadrangle are at least 2,000 feet thick.

Age and Correlation

No fossils have yet been found in the Woodstock quadrangle hence the dating of the formations is dependent on continuity of mapping (Doll, et al, 1961) with areas to the north where there are several fossil localities in the central and upper portions of the stratigraphic sequence here displayed. The best dated of the stratigraphic units are the Missisquoi and the Shaw Mountain formations.

In the Eastern Townships of Quebec near Lake Memphremagog, and at several localities farther north, black slates referred to the Beauceville series (MacKay, 1921) or Magog formation (Ami, 1900) contain graptolites of late Middle Ordovician age. The data have been summarized recently by Berry (1962), who states that the fossils belong to two graptolite zones indicating Wilderness and Trenton ages, and are therefore correlative with the Normanskill of the Taconic slate belt. These black slates and the associated volcanics and schistose quartzites pass south-

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ward into the Missisquoi formation of Vermont where the Whetstone Hill and Cram Hill members are the principal units containing carbonaceous rocks.

The Shaw Mountain formation contains crinoid columnals near Northfield, Vermont (Currier and Jahns, 1941), and at several other localities between there and the Canadian border (Doll, 1951). The most diagnostic fossils yet found in the Shaw Mountain, however, occur near Albany, Vermont, (Konig, 1961; Konig and Dennis, 1964) where corals and brachiopods (Boucot and Thompson, 1963, p. 318) indicate an age between late Llandovery and early Gedinnian, or, in other words, Silurian to earliest Devonian. Similar rocks near Marbleton, Quebec, contain Silurian fossils (Cooke, 1950; Naylor and Boucot, 1965, p. 160) indicating a Ludlow (late Silurian) age. Also, the Glenbrooke shale and Sargent Bay limestone, overlying the Peasley Pond conglomerate in two isolated synclines near Magog, Quebec, contain Silurian fossils (Naylor and Boucot, 1965). The Glenbrooke and Sargent Bay resemble parts of the Northfield and lower Waits River (Ayers Cliff of Doll, 1951) formations respectively. The Waits River (Cady, 1950; Doll, 1943a) and Gile Mountain formations (Doll, 1943b) have yielded a few fossils but their identification has been disputed. The St. Juste group in Quebec, however, contains Devonian fossils (Clark, 1923; Gorman, 1955; Cady, 1960). The St. Juste is almost certainly continuous with at least part of the Gile Mountain formation. The relative ages of the Waits River and Gile Mountain formations are obscured by structural complexities and changes in sedimentary facies as noted above, but it is clear that both overlie the Shaw Mountain, hence cannot be older than Silurian. The resemblance of the eastern part of the Gile Mountain formation to the Littleton formation of New Hampshire (Billings, 1956), and to the Seboomook formation in northwestern Maine (Boucot, 1961), is striking. Both of these contain early Devonian fossils, but may include some Silurian beds in their lower portions. Taking all of the above evidence into account the Shaw Mountain is most probably Silurian; the Gile Mountain is in part Devonian but may include some Silurian near its eastern border; and the Northfield and Waits River formations are probably Silurian, in part, but may also include some Devonian.

The dating of the formations below the Missisquoi is less direct. There is, however, good evidence that the continuation of the Ottauquechee formation in Quebec (Osberg, 1956, 1965; Cady, 1960) can be traced north and west around the Sutton Mountains anticlinorium into the Sweetsburg slate which is at least in part Middle Cambrian and into the Scottsmore

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formation which is Lower Cambrian (Osberg, 1965, p. 227, quoting G. Theokritoff). The Ottauquechee formation in eastern Vermont has features in common with the West Castleton and Hatch Hill formations of the Taconic slate belt (Theokritoff, 1964; Zen, 1961). These latter rocks are black slates with quartzites and minor carbonates resting on green and purple slates much like a less metamorphosed Pinney Hollow. The West Castleton carries Lower Cambrian fossils and the Hatch Hill carries Upper Cambrian fossils. It thus appears probable that the Ottauquechee includes not only Middle Cambrian strata but some Lower Cambrian and Upper Cambrian as well. The Stowe formation may thus be Upper Cambrian or Lower Ordovician, and the formations underlying the Ottauquechee should be no younger than Lower Cambrian.

For further discussion of the age of the post-Mount Holly rocks of eastern Vermont the reader is referred to Cady (1960).

The Mount Holly complex, cropping out in the Green Mountain core, is overlain on the west with profound unconformity (Billings, et al, 1952; Brace, 1953; and others) by formations containing Lower Cambrian fossils. The Mount Holly resembles, in many of its lithologic features, the Grenville of the southeastern Adirondacks as described by Walton and de Waard (1963).

Igneous Rocks

ULTRAMAFIC ROCKS

Lenticular bodies of serpentine and talc-carbonate rock (steatite) occur mainly in the Ottauquechee formation. Serpentine and talc-carbonate rock in a lenticular body at least 600 feet wide is exposed just west of Plymouth Five Corners. The rock is pale-green to greenish-black and is somewhat brecciated. Long, fibrous crystals of chrysotile are abundant, either in radiating aggregates on the slickensided surfaces or in short intersecting veins. Distorted octahedra of magnetite as much as one centimeter in diameter are common along the border of the serpentine body, where chlorite is also abundant. The exact limits of these serpentine and talc-carbonate bodies are very difficult to determine because of the poor exposures, the recent overgrowth on the abandoned farms, and the lumbering in the adjacent woods. A smaller body of serpentine occurs near the south line of the quadrangle. Talc occurrences, talc-carbonate rocks and soapstone quarries are reported in the Ottauquechee and Missisquoi formations at several localities in the town of Bridgewater by Perry (1929) and Merrill and Chaffee (1957). These and the serpentines in Plymouth

are probably pre-metamorphic ultramafic intrusions as are similar rocks elsewhere on the east flank of the Green Mountain anticlinorium (Doll, *et al*, 1961). None occur above the base of the Shaw Mountain formation hence it is inferred that they are of late Ordovician age.

AMPHIBOLITE

Dark-green, weakly foliated, chloritic amphibolite is exposed in the middle of the Ottauquechee formation. Dark-green, prismatic crystals of hornblende up to three millimeters long are the most conspicuous constituents of this amphibolite. It is probably a pre-metamorphism dike or sill. Other amphibolites reported by Perry (1929) may also be metamorphosed intrusions, but most are more probably metamorphosed mafic volcanics in the Missisquoi formation.

DACITE PORPHYRIES

Dikes and sills of dacitic composition were found at several localities on Soltudus Mountain, on the northwest slopes of Blueberry Hill, and on East Mountain. Thompson (1950) noted similar rocks south of Soltudus Mountain in the Ludlow quadrangle. Some of the sill-like masses are as much as 50 feet thick but most are less than 10 feet thick. All are somewhat foliated and the plagioclases are coarsely sericitized (Plate 5, Fig. 2). They resemble chemically some portions of the Barnard volcanic member of the Missisquoi formation and may be magmatically related.

GRANODIORITE

The small granodiorite body in the Pinney Hollow formation, on the south slope of Morrison Hill one mile due west of the Pinney Hollow schoolhouse, is the only granitic intrusive of mappable size. It is roughly lenticular, about 500 feet long and 100 feet wide, with sharp, locally cross-cutting contacts against the enclosing schist. The rock is coarse grained, granular, and nearly white with few dark-colored constituents. It is homogeneous, practically without any jointing, and extremely tough.

The mode of this rock as reported by Perry (1929, p. 44) is: quartz 30.3%, albite 60.5%, microcline 5.7%, microperthite 1.6%, muscovite 1.7% and apatite 0.1%. The rock also contains a small amount of epidote as small grains between the other minerals. The albite has been partly replaced by sericite.

The rock of this body is exceedingly homogeneous in texture. It lacks any sign of mechanical deformation and locally crosscuts the country

rock. However, foliated rocks of similar composition were found by Thompson (1950) in the Ludlow quadrangle and occur elsewhere in eastern Vermont. It seems likely that the emplacement of these granitic rocks was partly coincident with the waning stages of deformation and metamorphism, and was partly younger.

Diabase and Camptonite

Dikes of diabase and camptonite are found in all the formations in this area. They are usually one to two feet thick and cut across the schistosity of the enclosing rocks. The general appearance of these dike rocks is rather similar. They are mostly fine to medium grained and dark gray to black. It is difficult to distinguish camptonite from diabase in the field except that the former is occasionally porphyritic, containing distinct phenocrysts of black amphibole (barkevikite). Some of the camptonite dikes are somewhat amygdaloidal. The amygdules are partly filled with quartz, calcite, or analcite.

Thin sections of these rocks show that they are all somewhat altered, with the development of such secondary minerals as chlorite, serpentine, and leucoxene. Titaniferous augite is present in most specimens of diabase. Relicts of olivine surrounded by serpentine were also observed. Barkevikite is the characteristic constituent of camptonite and occurs either alone or with olivine. The texture of the whole rock or of the groundmass where it is porphyritic, is diabasic. The plagioclase (An_{50}) occurs in small, lath-shaped crystals.

These dikes are post-metamorphism. The alkalic nature of the camptonite suggests that this group of dike rocks is genetically related to the intrusive bodies of Mt. Ascutney and Cuttingsville, which belong to the White Mountain magma series of probable Mesozoic age.

Metamorphism

The rocks of the Woodstock quadrangle, except for the diabase and camptonite dikes and possibly the quartz diorite in Plymouth, have all undergone a regional metamorphism. Mineral assemblages indicate that much of the area is in the garnet zone. The western border of the area, south of West Bridgwater, is in the biotite zone and the northeast corner is in the kyanite zone. The position of the garnet isograd is approximate because of limited data. Garnet occurs at least as far west along the Ottauquechee River as Riverside school, but is well developed there. In Pinney Hollow it is found as far west as the middle of the Pinney Hollow formation. On East Mountain in Plymouth and on Blueberry Hill it

occurs as far west as the base of the Pinney Hollow formation. Kyanite and staurolite occur in the northeast part of the Ludlow quadrangle within less than a mile of the boundary of this quadrangle, and kyanite with rare staurolite occurs in the northeast part of the Woodstock quadrangle.

Minerals occurring in assemblages with quartz and muscovite, in the biotite zone, include chlorite-biotite-albite and chlorite-chloritoid-paragonite (Plate 3, Fig. 1; Plate 6, Fig. 1). In the garnet zone almandite appears as an extra mineral in each of the above assemblages. Eastward the proportion of chlorite decreases and what survives is paler and more magnesian. Chloritoid is rare east of the Stowe formation but this is probably for compositional reasons. A key assemblage in the schists of the kyanite zone is quartz-muscovite-biotite-garnet-plagioclase-kyanite, with staurolite as an extra mineral in rare occurrences.

Mafic volcanic rocks in the Standing Pond and Barnard volcanics, and in the Whetstone Hill member of the Missisquoi, are epidote amphibolites, and those in the Ottauquechee and in the Pinney Hollow on Blueberry Hill are actinolitic greenstones. Those in the upper end of Pinney Hollow and on Morgan Peak, however, carry chlorite and ankerite but no amphibole.

The dolomites in the Tyson and Hoosac formations contain only rare phlogopite scales and a few talc seams. Both quartz-dolomite and microcline-dolomite appear to be stable assemblages in these rocks. This suggests that talc and phlogopite occur only where Mg exceeds Ca. This is rare in the dolomites and then perhaps due only to metasomatism along those partings and joints where talc has formed.

The carbonate rocks in the Northfield, Waits River and Gile Mountain formations contain zoisite (Plate 6, Fig. 1). Many beds contain weakly birefringent poikilitic garnets that stand out like hobnails on a weathered surface. Less commonly a green calcic amphibole is present but it is developed mainly at the contacts with phyllite interbeds.

The garnet in the calcareous rocks is a calcic almandite (J. L. Rosenfeld, personal communication) and the inclusions in it commonly have a spectacular spiral arrangement. A detailed study of these snowball garnets by Rosenfeld (1960, and ms. in preparation) show that they began growth during early, recumbent, isoclinal folding and that growth during early, recumbent, isoclinal folding and that growth continued through the later rise of the Pomfret and Chester domes. This indicates that these two stages of deformation were not greatly separated in time as both overlap what appears to be a single metamorphic cycle.

This also appears to be true of the two deformations in the western part of the area where mineral orientations can be found paralleling both the earlier and later sets of fold axes. It thus appears that the regional metamorphism here is essentially a synkinematic process that spanned at least two, possibly three, major stages of deformation.

Retrograde metamorphism is not conspicuous in the Tyson and younger formations other than in the local development of chlorite rims about garnet (Plate 9, Fig. 2) or pseudomorphs of chlorite after garnet. In the rocks of the Mount Holly complex, however, there is extensive sericitization and saussuritization of feldspars and chloritization of biotite and hornblende. These rocks have undergone at least two cycles of regional metamorphism, one Precambrian and the other Paleozoic. The Precambrian metamorphism probably reached the amphibolite facies and possibly the hornblende granulite facies. The Paleozoic metamorphism that produced the retrograde, polymetamorphic effects reached here only the green schist facies (biotite zone).

STRUCTURE

General Statement

The western half of the Woodstock quadrangle is on the eastern limb of the Green Mountain anticlinorium and the eastern half is in the area of domes and recumbent folds of eastern Vermont (Doll, *et al*, 1961). Local complexities in the western part are the faults in Plymouth and the anticlinal fold in Curtis Hollow. In the eastern part, certain of the recumbent folds are indicated by the loops made by the Standing Pond volcanics along the extreme east edge of the map (Plate 1). The axial surfaces of these recumbent folds have been arched by later doming, and have been depressed into tight troughs both in the regions between the domes and in the central syncline separating the domes from the Green Mountain anticlinorium. The Chester dome, well displayed in the Ludlow quadrangle to the south (Thompson, 1950) dominates the structure in the southeast part of the Woodstock quadrangle, and the Pomfret dome (Lyons, 1955) has produced the westerly dips along the eastern edge of the quadrangle north of Woodstock.

All the bedding in the sedimentary and volcanic rocks has been modified by the development of secondary foliation. The rocks are generally schistose, the plane of schistosity being marked by the parallel arrangement of typical low- to medium-grade metamorphic minerals such as chlorite, sericite, biotite, and hornblende. The schistosity is generally

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parallel or sub-parallel to the original bedding or the compositional layering. Other kinds of rock cleavage are also developed in the rocks of this area. Slip-cleavage and fracture-cleavage (used only in a descriptive sense) are present and closely associated with the development of the minor folds.

Minor folds of various sizes and shapes are common throughout the area. The nature of the minor folds is an important factor in deducing the local structure.

Linear structures are well developed. They are shown by (1) the axes of minor folds, (2) the orientation of prismatic metamorphic minerals and mineral streaks on the plane of schistosity, (3) the crinkling of the schistosity, and (4) the long axes of deformed pebbles. Lineation of the first type is present in almost all the formations. With few exceptions, it plunges north-northeast or northeast. Lineations of the second type, although of rarer occurrence, are closely related and parallel to those of the first type. Lineation of the third type is present in the more phyllitic rocks and is closely related in origin to the slip-cleavage. In the phyllitic rocks of the Northfield, Waits River, and Missisquoi formations, this type of lineation is generally oriented parallel to the minor fold axes. From the Ottauquechee formation downward stratigraphically, although the general trend and strike of the beds is north-northwest, the minor folds and the associated lineations of the first and second types plunge steeply to the east-northeast. Crinkling on the schistosity of these lower rocks generally plunges at low angles toward the north or north-northwest and thus cuts across the lineation of the other two types. It suggests that at least two distinct stages of deformation have occurred. Lineation of the fourth type is shown by the pebbles in the Shaw Mountain conglomerates. The linear features are plotted on the tectonic map (Plate 2). Where the nature of the lineation has been recorded this is indicated by a letter symbol.

Plymouth Fault Zone

This zone lies in the southwest corner of the Bridgewater-Woodstock area, between West Bridgewater and Soltudus Mountain. It is about seven miles long in a north-northwest direction and is about $1\frac{1}{2}$ miles wide. The rocks have a general north-northwest trend near the southern margin of the zone. Northward the trend changes from north-northwest to north-northeast. The beds dip moderately toward the east-northeast. The rocks in this zone belong to the Tyson and Hoosac formations. East

of the Plymouth Union Valley, the rocks of the Hoosac formation and lower Pinney Hollow formation are duplicated by three high-angle thrusts.

The two high angle thrusts on the west end of Soltudus Mountain extend south into the Ludlow quadrangle where, on Dry Hill, gneisses, schists, and quartzites of the Mount Holly are thrust onto albite schists of the lower Hoosac formation. On Soltudus Mountain the albite schists of the lower Hoosac are thrust over quartzites and schists occurring higher in the Hoosac. According to Thompson (1950) the faults on Dry Hill may be related to inhomogeneities in the underlying Mount Holly basement.

The West Bridgewater-Plymouth thrust extends from the notch east of Soltudus Mountain to West Bridgewater. The rocks on the east side of the fault are dolomite, black phyllite and dark-gray laminated quartzite of the upper Hoosac formation. On the west side, pale green chlorite-sericite-quartz schist of the Pinney Hollow formation is exposed between Plymouth Village and the western slope of Morgan Peak. A thick dolomite between Grass Pond and Plymouth Village is cut out completely, farther south, north of Soltudus Mountain. This thrust is at least seven miles long. It probably dies out both to the north and to the south. The dip of the fault plane cannot be observed directly. The maximum stratigraphic throw is estimated to be about 1,000 to 1,500 feet.

Two generations of minor structures can be distinguished in this zone by their appearance, orientation, and mutual relationships.

Minor structures of an earlier generation include steeply-plunging minor folds and mineral lineation. The former are best shown in the laminated quartzite beds and in the phyllitic rocks rich in quartz lenses. The wave length of these folds is a few inches to a few feet. The horizontal projection of the axes of these folds is essentially parallel to the dip direction of the schistosity. The mineral lineation is present in the pale-green, chlorite-sericite-quartz schist of the Pinney Hollow formation and in the laminated quartzite of the Hoosac formation. Streaks of fine-grained chlorite, chloritoid and, less commonly, biotite are found in the schist of the Pinney Hollow formation (Plate 6, Fig. 1). In the laminated quartzite of the Hoosac formation these streaks contain biotite and, rarely, tourmaline. This mineral lineation is parallel nearly everywhere to the axes of the steeply-plunging minor folds.

Minor structures of a later generation are represented by north-south trending, symmetrical or westward-overturned minor folds, also by slip-
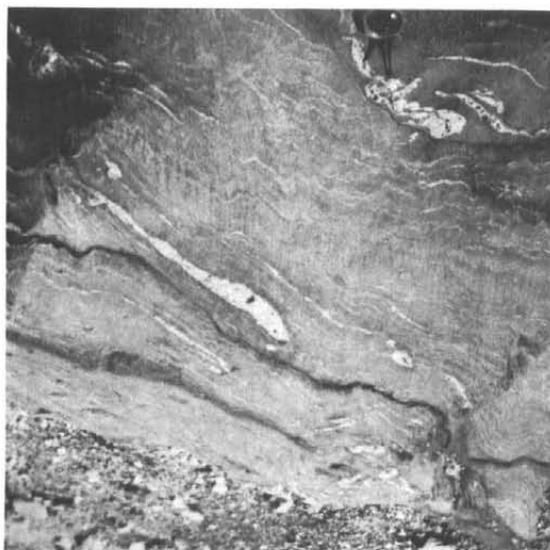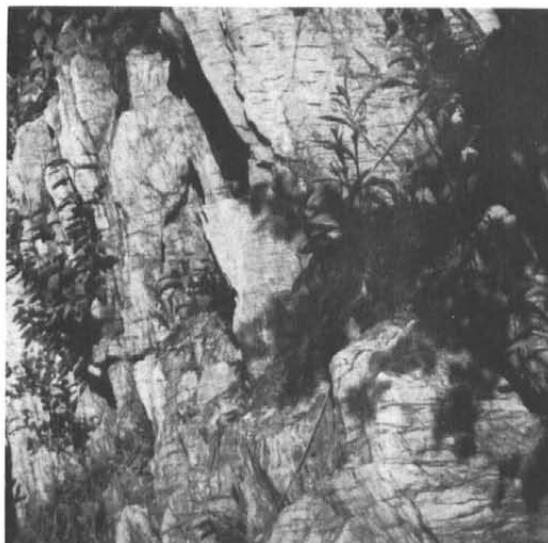
PLATE 6

Figure 1. Chloritoid schist, Pinney Hollow formation: Road cut one-half mile east of West Bridgewater, shows crinkling of schistosity and earlier lineation by streaming of minerals approximately down dip. (P.H.C.)

Figure 2. Sericite-chlorite schist, Pinney Hollow formation: Three quarters of a mile southeast of Wood Peak, shows minor folds, false cleavage, and quartz lenses. (P.H.C.)

cleavage (Plate 3, Fig. 1) and later lineations. They invariably cut across the minor structures of the earlier generation and are in harmony with the major thrusts in this zone.

The strike of the horizontal projection of the axes of the minor folds is nearly north-south, and the axes either plunge gently north or are horizontal. The size as well as the tightness of these folds is controlled by the competency of the rocks. Westward overturning is very common. Slip-cleavage is developed in the less competent rocks. The strike of the slip-cleavage is also nearly north-south and the dip is vertical or steep, either to the east or to the west.

The intersection between the plane of schistosity and the later cleavage is marked by crinkling (later lineation) parallel to the axial direction of the later minor folds (Plate 6, Fig. 1). The earlier lineations (mineral lineations) are folded around the later folds and crinkles. Where slip-cleavage is developed the mineral lineations are dislocated by the minute slip movement.

Upper Ottauquechee Valley Homocline

This structural unit occupies much of the western half of the Wood-stock quadrangle and is in the area where the Pinney Hollow, Ottau-quechee, Stowe and Missisquoi formations crop out. The western limit is defined by the Sherburne Valley (in the adjacent Rutland quadrangle) and the West Bridgewater-Plymouth fault zone. On the east the homo-cline passes gradually into the Reading-Pomfret folds with an increase in size and abundance of the minor folds.

The formation boundaries in the southern two-thirds of the homocline have a north-northwest trend. This swings to the north and north-northeast in the northern third of the area. A general easterly dip is prevalent throughout. The strike of the schistosity shows local deviations from the general trend of the formations because of minor folds. Such folds are more abundant toward the north as well as toward the east and cause the northward widening of the different formations. This is particularly evident in the western belt of the Moretown member of the Missisquoi formation where it passes into the Randolph quadrangle. The folds are either asymmetric or isoclinal, but the anticlines all have much longer eastern limbs. The axial surfaces strike north-northwest to north, and the fold axes plunge either gently north-northeast or steeply east-northeast. The steep lineations are oriented approximately down the dip of the foliation and are nearly normal to the axes of the minor folds. Some of the steep lineations are early folds, others are a mineral streaming formed by streaks of platy minerals, mainly chlorite and mica, in the plane of

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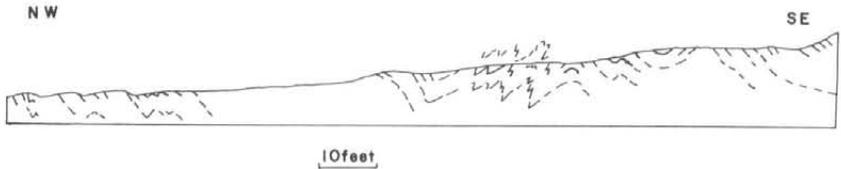
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Figure 2.   Cross section showing folds in the Barnard volcanic member along the brook 1,000 feet east of Curtis Hollow School (compiled from field sketches).

schistosity. The streaming and steeply plunging folds are best developed in the western third of the quadrangle and much less conspicuous farther east.

The strike of the schistosity is very close to the general trend of the formations except along the western margin of this structural unit where the attitude of the strata has been modified locally by later folds related to the slip-cleavage. The dips are mainly between 30 and 50 degrees east-northeast. Deviations from this characteristic attitude are found, however. In the southern and central parts of the homocline the deviation of the strike is almost invariably toward the east. The common "abnormal" strikes are between due north and N 15°E, and the associated dips, though still generally to the east, are at much steeper angles (65 to 80 degrees). Drag folds suggesting overturning of the strata are occasionally observed on the outcrops showing the "abnormal" attitudes. It is likely that all outcrops showing such "abnormal" attitudes are on overturned limbs of nearly isoclinal folds (Figure 3). The pattern of such reconstructed folds is conformable with the other minor structures observed. These "abnormal" attitudes are common in the larger outcrops in the bed of the North Branch of the Ottauquechee River east of Chatauguay where they occur at intervals between the "normal" ones. The change is invariably sharp and the less common strikes and dips apparently represent the short limbs of northeast plunging step-like folds as shown in Figure 4.

In the Pinney Hollow formation the minor structural features are essentially the same as in the West Bridgewater-Plymouth fault zone. Steeply-plunging mineral lineation, together with the axes of the steeply-plunging minor folds, are interesected by later slip-cleavage and crinkling (Plate 6, Figs. 1 and 2). The slip-cleavage in the schists of the lower part of this formation is well shown in the road-cuts east of West Bridgewater where it dips toward the west. The sense of relative movement of the hanging wall of the cleavage plane is up and eastward (Plate 6, Fig. 1).
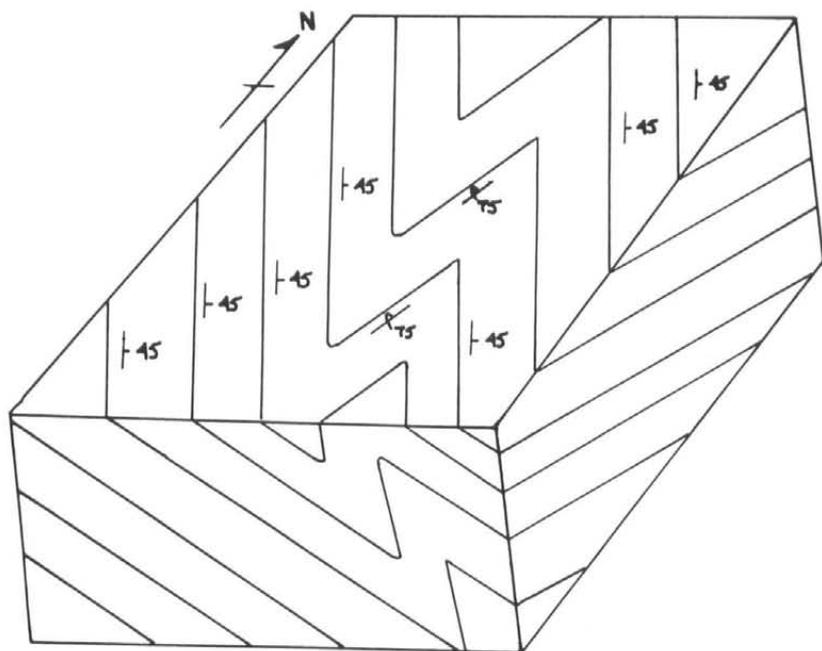
Figure 3. Local change in strike and dip of bedding as related to overturned folds in the Upper Ottauquechee Valley homocline.

5 feet
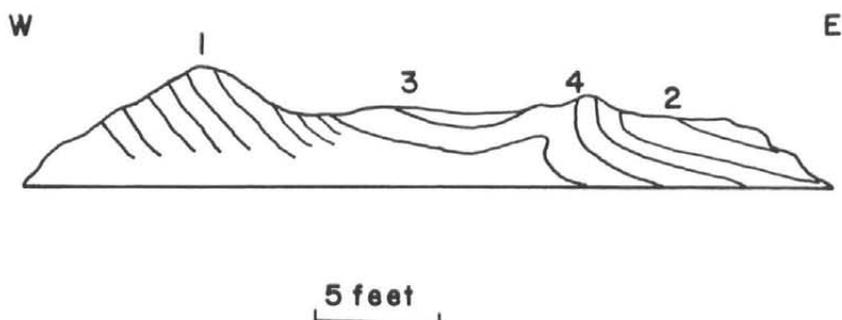
Figure 4. Outcrop of Whetstone Hill member on the north side of the North Branch of the Ottauquechee River 2 miles east-southeast of Chatauguay, showing local variation in strike and dip of bedding.

Attitude at 1: strike N35°W, dip 55°E
Attitude at 2: strike N30°W, dip 40°E
Attitude at 3: strike N55°W, dip 20°E
Plunge of fold axis at 4: 20°N 35°E

Slip-cleavage and crinkling are confined to the phyllitic members of the Ottauquechee formation and steeply-plunging minor folds are common in the quartzites. The minor folds in the overlying formations plunge gently north-northeast. From the Moretown member of the Missisquoi formation eastward, many minor folds are overturned toward the east (Plate 7, Fig. 1). Fracture cleavage is developed locally parallel to the axial surfaces of such folds.

Mineral lineation is well developed in the hornblendic rocks of the Barnard volcanic member where it is shown by prismatic hornblende crystals. This lineation plunges north-northeast parallel to the plunge of the minor fold axes.

Boudinage is developed locally in the amphibolite beds of the Barnard member. The best developed occurrences are found in the thin amphibolite beds that lie between thicker beds of the light-colored gneisses. As seen in the flat outcrops in the river bed near the mouth of Curtis Hollow, the direction of extension during deformation, as represented by the direction of disruption of the amphibolite beds, is parallel to the mineral lineation and minor fold axes. The gaps between the boudins are filled with vein quartz and a small amount of feldspar. Joints striking N 30°E are well developed and cut across lithologic boundaries (Plate 7, Fig. 2).

The consistent pattern of the Upper Ottauquechee Valley homocline is locally disrupted by a large anticlinal fold developed in the Barnard volcanic member in Curtis Hollow to the south and east of Bridgewater. Figure 2 shows the nature of the tight minor folding within this anticline. The outcrop width of the Barnard volcanic member is broadened here. The Cram Hill member and the Shaw Mountain and Northfield formations are offset to the east where some of the larger folds are indicated by east-west strikes. Pebble elongations in the Shaw Mountain formation parallel the axes of these folds.

Reading-Pomfret Folds

This structural unit occupies the entire eastern half of the area where more than 90 per cent of the rocks belong to the Waits River formation. The Standing Pond volcanics and the Gile Mountain formation appear along the northeast, east-central, and southeast margins of the area.

The rocks within this unit are intensely folded. Isoclinal folds overturned toward the east are common. The plunge of the folds is generally toward the northeast at gentle angles except locally around the village of Woodstock where the plunge is to the south.

In the Northfield formation along the crest of Long Hill the rocks are
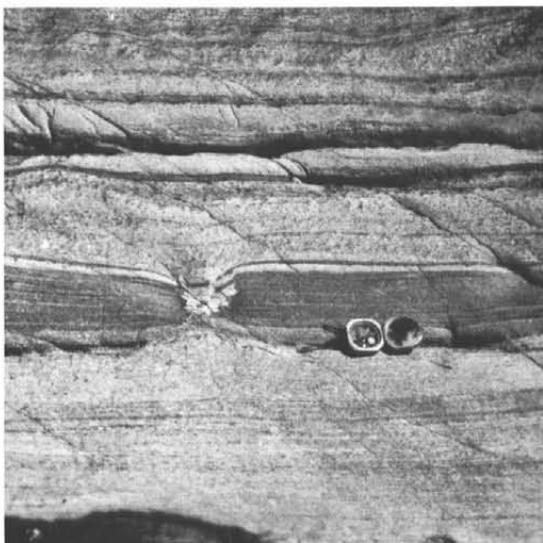
PLATE 7

Figure 1.   Dextral folds, Moretown member, Missisquoi formation: Fletcher Brook, Stockbridge, 0.4 mile upstream from junction with Stony Brook. (E.H.E.)

Figure 2.   Boudinage, Barnard volcanic member, Missisquoi formation: River bed at north end of Curtis Hollow, amphibolite in biotite gneiss. (P.H.C.)

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strongly deformed. Bedding planes are difficult to distinguish from the secondary foliation in the phyllitic rocks. Minor folds may be recognized easily only where quartz lenses are numerous or where the rock contains thin beds of quartzite and impure limestone. They are almost invariably overturned to the east (Fig. 5). These folds are disharmonic. The size and frequency is controlled by the relative competency of the individual beds. Gently dipping cleavage is developed in these folds.

In the central and eastern part of this unit, where the schistosity dips steeply westward, tight minor folds overturned toward the east are observed along certain north-south belts in the metamorphosed impure limestones (Fig. 7 and Plate 8, Figs. 1 and 2). The axial surfaces of these folds generally dip steeply west. With few exceptions, the axes of these folds plunge north. Where the plunges are steep enough the folds are readily observable on the surfaces of flat-lying outcrops.

Similar tight, north-south trending minor folds are found farther south where the schistosity dips gently away from the Chester dome. North-ward-plunging tight folds, from a fraction of an inch to a few feet in amplitude, are developed in the gently northward-dipping strata (Figs. 6 and 8). Slip-cleavage is roughly parallel to the axial surfaces of these folds. In the more quartzitic rocks a north-south, steeply dipping slip-cleavage occurs without accompanying minor folds.

Minor folds with flat-lying axial surfaces are present in thick horizontal beds of metamorphosed limestone in the south part of Woodstock. They are presumably formed by differential gliding within these carbonate beds. The axes of such folds strike generally north-south and are either horizontal or plunge gently toward the north. Many of them pass into north-south trending folds of much larger scale.

The intricate loops of the Standing Pond volcanics along the eastern margin of the quadrangle provide excellent evidence for interpreting the major structures within this unit. The volcanics along the northeast and east-central border outline the noses of two large recumbent folds. The axial surfaces of these folds arch over the Pomfret dome lying to the east in the Hanover quadrangle (Lyons, 1955), and their axes extend north-eastward into the Strafford quadrangle where the Standing Pond volcanics outline four large recumbent isoclinal folds at the north end of the Strafford dome (White and Jahns, 1950, Doll, 1944).

The axial surface of a westward-closing recumbent fold can be followed north from Woodstock in an arc passing just west of Pomfret village and Hewetts Corners. From there it crosses the northwest corner of the Hanover quadrangle and extends on to the vicinity of Strafford where it may be identified as the Old City "syncline" of White and Jahns

W   E

3 feet
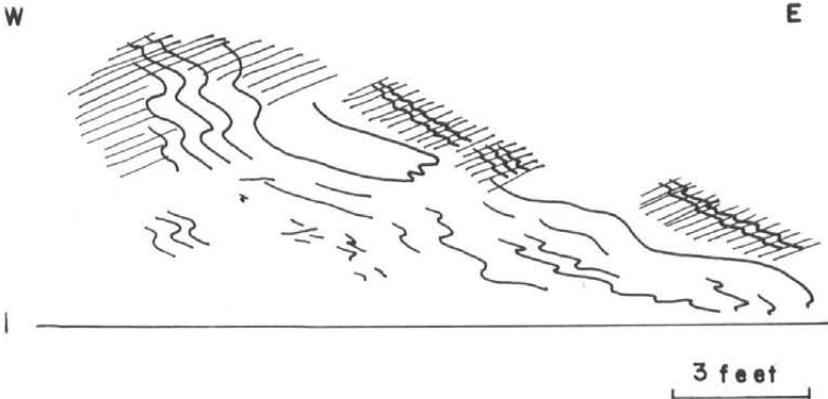
Figure 5.   Minor folds and slip cleavage in black phyllite and thin-bedded impure limestone of the Northfield formation. Exposure on south end of Long Hill, 1 mile northwest of Brown School.

(1950). An eastward-closing fold that has an axial trace lying just east of the above can also be followed to the Strafford area and identified as their Grannyhand "anticline". Both of these axial surfaces dip west into younger synclines (Plate 2), and presumably re-emerge somewhere in the Waits River formation on the southwest and west limbs of these later folds. Many outcrops in the Waits River formation in the central and western portions of the Reading-Pomfret fold belt show fairly large isoclinal recumbent folds (Plate 9, Fig. 1), suggesting that major recumbent folds are indeed present in this area. They have not been located there, however, because of the lack of good marker beds.

Possibly the westward-closing fold delineated in Reading and West Windsor by the Standing Pond and a non-calcareous zone in the Waits River is to be correlated with the Old City fold. A more plausible, reconstruction, however, is indicated by the cross sections accompanying the state geologic map (Doll *et al*, 1961). This would identify the West Windsor fold with a still lower axial surface, that of the Strafford Village "syncline" of White and Jahns.

White and Jahns (1950) designate the westward and eastward closing folds as "synclines" and "anticlines" (the quotation marks here and above are ours) respectively. This is correct if the Standing Pond *overlies* stratigraphically, the adjacent parts of the Waits River, but the terms would have to be reversed if the opposite proved to be true. This is still a moot point as we have noted earlier. It has, however, been well substantiated that these intricate folds are in fact large recumbent

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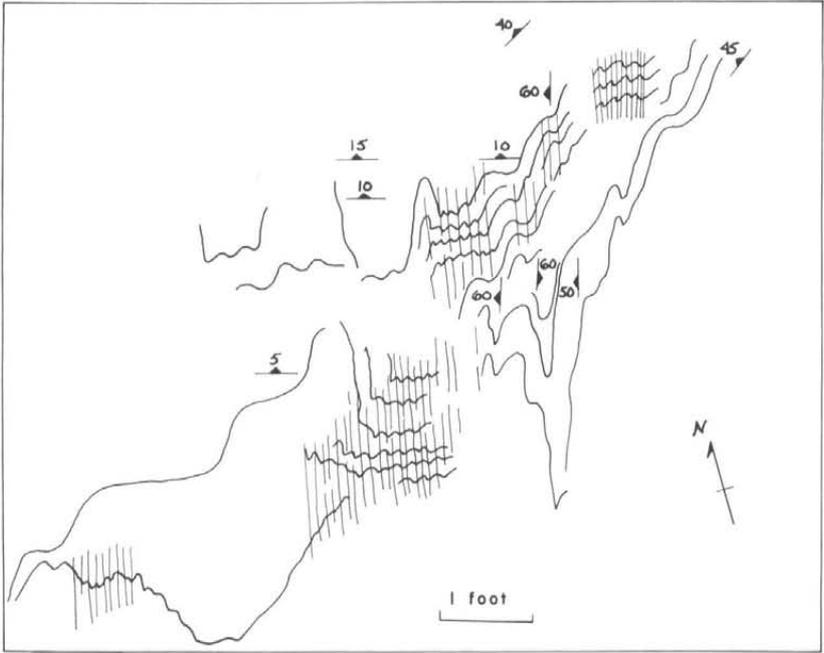
Figure 6. North-south trending minor folds and cleavage in a gently northward dipping impure quartzite bed of the Waits River formation, 1¼ miles northwest of Bailey Mills.
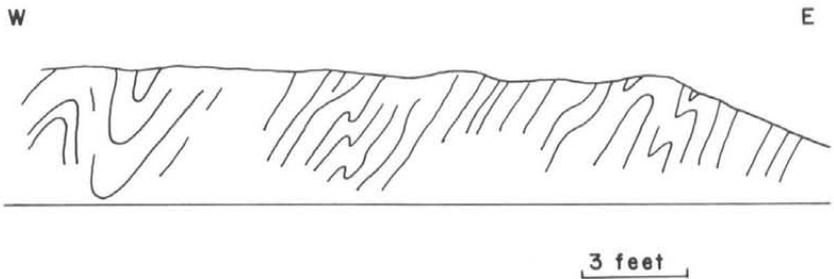
Figure 7. Folds in thick-bedded metamorphosed impure limestone of the Waits River formation. Outcrop 1¼ miles due north of Pelton School.

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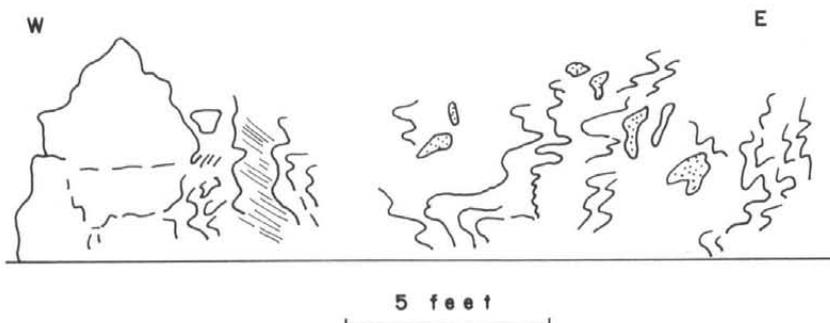
Figure 8.   Flexure folds in Waits River formation in a road cut 1 mile north of Reading Hill.

isoclinal structures (White and Jahns, 1950, Billings, *et al*, 1952, Lyons 1955, Doll *et al*, 1961). Further, it has been established that these recumbent folds are older than the domal structures with which they are associated, the later doming having refolded the axial surfaces of the isoclinal folds (White and Jahns, 1950, Lyons 1955). Rosenfeld (1960) in a regional study on spirally arranged inclusions in rotated garnets confirms this interpretation. The early rotation of the garnets can be related to the recumbent folds and the later rotation to the doming. According to Rosenfeld (personal communication) the two stages of deformation are shown in some garnets by a reorientation of the rotation axes, and in some by an actual reversal of the sense of rotation. Many thin sections provide further evidence (Plate 9, Fig. 2) of late stages of deformation.

A synclinal axis trends northwest near South Woodstock, and a central syncline can be traced from the north part of Reading to East Barnard. Both of these folds are late and related to the doming. The South Woodstock fold is the synclinal trough separating the Pomfret and Chester domes. The central syncline, the Proctorsville syncline of Thompson (1950), separates the domes from the Green Mountain anticlinorium (see Plate 1, cross-sections B-B' and D-D').

The westernmost area of the Gile Mountain formation terminates by structural closure west of East Barnard village. It has been interpreted by some (Ern 1963, Goodwin 1953) as part of a huge recumbent anticline once connected over the Strafford-Willoughby arch and the Pomfret and Strafford domes to a root zone farther east. The plunge of the

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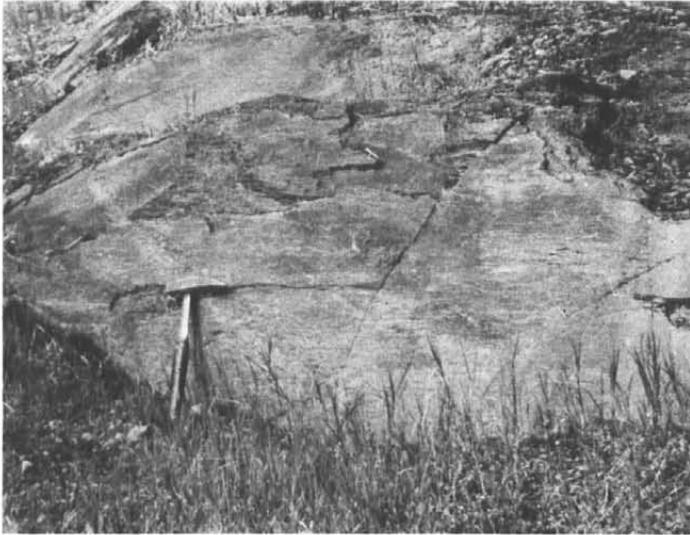
PLATE 8

Figure 1. Bedding-cleavage relations, Waits River formation: Locality 1.2 miles east of Barnard village, quartz-biotite schist and garnetiferous phyllite. View is to south, beds dip 70° west, cleavage dips 30° west. (E.H.E.)

Figure 2. Sinistral fold, Waits River formation: Locality 1.3 miles northwest of Hewetts Corners, Pomfret. Sequence of carbonate rocks folded in sinistral pattern. (E.H.E.)

PLATE 9

Figure 1.   Fold, Waits River formation: Vermont highway #12, 1.6 miles south of Barnard village. Nose and lower limb of sinistral fold; surface has slickensides parallel to fold axis. (E.H.E.)

Figure 2.   Garnetiferous phyllite, Waits River formation: Shows warping of axial surfaces of crinkles around garnets, and rim of chlorite around garnet; g, garnet; mu, muscovite; chl, chlorite; q, quartz. (P.H.C.)

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hinge-line or arch bend of such a recumbent structure would be to the north. In northern Vermont the western band of Gile Mountain joins the main area of that formation at the termination of the Strafford-Willoughby arch in the southwest part of the Island Pond quadrangle (Goodwin, 1963).

The interpretation of Ern and Goodwin would follow if the Standing Pond were shown to be older than the adjacent parts of the Waits River. Most investigators, however, have believed stratigraphic tops to lie the other way across the Standing Pond. With this latter interpretation the western band of the Gile Mountain formation occupies the axial zone of a large syncline, the Townshend-Brownington syncline of Doll *et al*, (1961), overturned toward the east. Either interpretation is consistent with the known relations in the Woodstock quadrangle, and even with the cross sections (Plate 1). The differences would appear only in more extended reconstructions at depth and above ground. We prefer in this report to leave the question open.

ECONOMIC GEOLOGY

There are no active mines or quarries in the Woodstock quadrangle at the present time. Granite (granodiorite) was once quarried on a small scale in Plymouth (Dale, 1923), and talc and soapstone were quarried in Plymouth and Bridgewater (Perry, 1929; Morrill and Chaffee, 1957). A talc quarry is now active, however, at Hammondsville, just south of the quadrangle boundary in Reading township.

The dolomites of the Tyson and Hoosac formations were worked for many years in Plymouth (Dale, 1915; Perry, 1929) but could not compete with the more extensive deposits in the western part of the state. Two quarries, however, were recently operated in the Tyson dolomite in the eastern part of the Rutland quadrangle south of West Bridgewater. The stone was crushed and used for road metal in the reconstruction of U.S. Highway #4. The hematite and magnetite associated with the Tyson dolomites in Plymouth were worked on a small scale for iron ore in the nineteenth century (Hitchcock, *et al*, 1861), as were some bog ores in the same general area.

Plymouth and the nearby towns experienced a minor gold rush many years ago (Hitchcock, *et al*, 1861; Perry, 1929) but the deposits proved to be of little value. The gold occurred in quartz veins, mainly in the Pinney Hollow and Stowe formations, and was also obtained from stream gravels.

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THE GEOLOGY OF THE
BURKE QUADRANGLE, VERMONT

By
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ABSTRACT

The Burke quadrangle is located in northeastern Vermont, 18 miles south of the International Boundary and about 10 miles west of the New Hampshire boundary. It is largely a mountainous region lying east of the Green Mountain anticlinorium. The considerable amounts of granitic rocks link the Burke area with the New Hampshire plutonic belt.

Three metamorphosed sedimentary formations are found in the quadrangle: the Albee formation, of Ordovician age, which is part of the New Hampshire sequence of metamorphic rocks, and the Waits River and Gile Mountain formations, of presumed Silurian and Devonian age, which belong to the Vermont sequence of rocks and which have a wide distribution in the central and eastern parts of the state. The two sequences are separated by the Monroe fault, but an important unconformity may also occur between them. The Albee formation consists of schist, quartzite, and amphibolite. The Waits River formation is made up of impure limestone, phyllite, schist, quartzitic beds, amphibolite (representing volcanic tuff, lava, and intrusives), and acid volcanics and intrusives. The Gile Mountain formation is comprised of alternating bands of phyllite, schist, quartzose phyllite, and amphibolite. Granite has intruded the sedimentary rocks and in places the two form mixed zones of hornfels and granite.

The meta-sedimentary rocks are strongly deformed and show in their minor structures evidence of two phases of movement: (1) an earlier folding accompanied by a well-developed schistosity, which is generally parallel to the bedding, and (2) a later one which folded the schistosity and produced a slip cleavage. The two movements are probably successive phases of the Acadian orogeny. The strata have a steep dip mainly to the east-southeast, but they dip more gently in the

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western part of the quadrangle and dip steeply to the east-northeast in the northern part. The rocks are part of a major north-northeasterly-plunging structure which extends westwards for some 10 to 15 miles and the central zone of which is the Willoughby arch, an anticlinal structure in the Waits River formation which may mark the culmination of a large-scale recumbent fold; the axial plane of the fold parallels the arch.

Metamorphism has occurred in two distinct phases: (1) a lower grade phase, which accompanied the formation of the earlier folds and the schistosity, and (2) a later higher grade phase, which was post-deformational in the Burke area. It was caused by an increase in temperature of the rocks which were at moderate depth in the earth's crust, and resulted in the growth of porphyroblasts of actinolite-tremolite, biotite, garnet, and staurolite. In the final stage granitic magma was intruded and the adjacent rocks were thermally metamorphosed to a high grade, with the appearance of andalusite and sillimanite.

INTRODUCTION

Location

The Burke quadrangle is located in northeastern Vermont between 44° 30' and 44° 45' North Latitude and 71° 45' and 72° 00' West Longitude (Fig. 2). It has an area of approximately 216 square miles and includes parts of Caledonia, Essex, and Orleans counties.

Methods of Study

Thirty-eight weeks were spent mapping the Burke quadrangle during the summers of 1955, 1956, and 1957, and an additional two weeks were spent in the field in June of 1958. Field mapping was carried out by means of traverses spaced as frequently as time and terrain permitted. Streams in particular were followed as much of the area is mantled with glacial deposits. Often, however, the streams had not eroded down to bedrock so the hillslopes had to be searched for outcrops. Locations were determined by sighting or by pace and compass. In the forested country compass traverses were made; locations were found by the use of a pocket altimeter and by frequent stereoscopic examination of air photographs along the route. Outcrops were located on three inches equal one mile photographic enlargements of the one inch equals one mile U. S. Geological Survey topographic map of the Burke quadrangle, published in 1951.
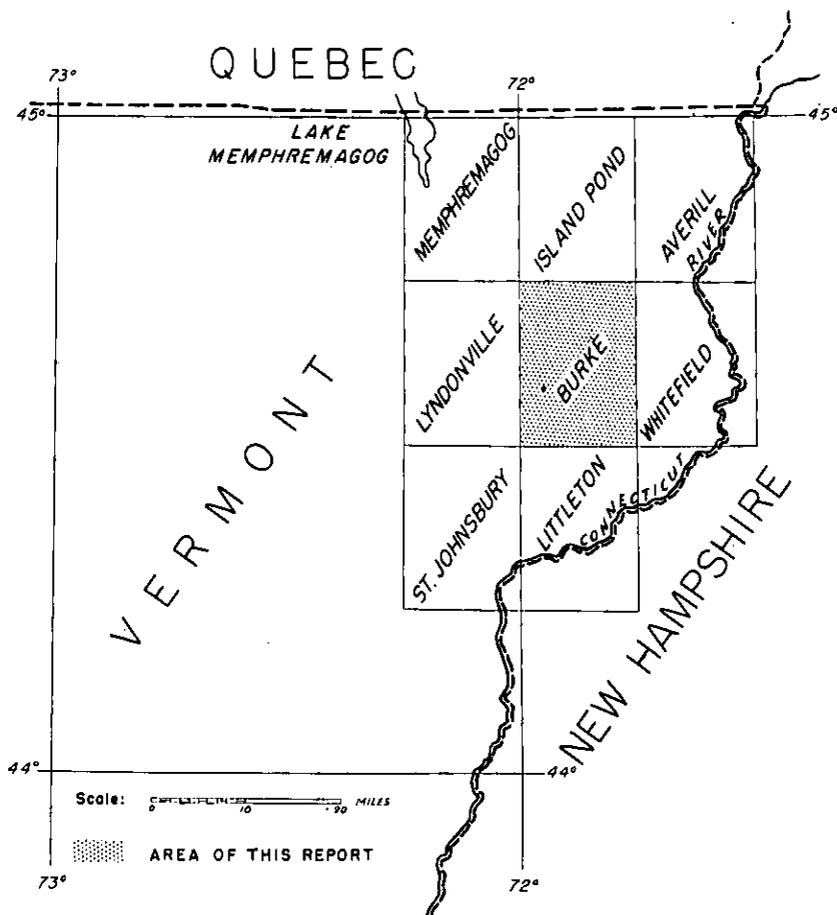
Figure 2. Location of Burke quadrangle.

A total of 3856 measurements of structural features were made in the field, including the strike and dip of bedding, schistosity, cleavage, and joints and the trend and plunge of fold axes and lineations. In addition, 31 trends of glacial striae were recorded. Hand specimens totalling 1819 were collected for detailed lithologic study; of these, 1029 were marked with their field orientation so that they could be re-oriented in the laboratory for further structural determinations.

In the laboratory all of the hand specimens were examined under a binocular microscope. Many planar and linear structures were measured

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on the oriented specimens. Re-orientation was done by placing an oriented specimen in a clamp attached to a universal ball joint which permitted the setting of the specimen in its correct geographic position. A Brunton compass was used to position the specimen and to measure additional planes and fold axes; for some measurements an adaptation of a contact goniometer was used. When planes were inaccessible or only showed as lineations on two or more non-parallel surfaces, their orientation was determined by sighting along a plate, mounted on a ball joint, and placing the plate parallel to the plane to be measured. The attitude of the plate could then be determined readily by Brunton compass. The pitch of linear features was measured on oriented (or re-oriented) surfaces and their plunge determined by construction on a Wulff stereonet. The structural data were plotted on maps (see Plates 2, 3, and 4); they were also plotted on the lower hemisphere of a Schmidt stereonet for the purpose of preparing stereograms for the quadrangle as a whole and for twelve subareas, whose arbitrary boundaries are shown on Plate 4. Nearly all of the stereograms were contoured and used to obtain additional structural data, the details of which are given in the section on Structure.

A total of 201 thin sections were examined under the polarizing microscope.

All directions of strike of planar surfaces are stated in this report from 0° to 180° east of north.

Reference to localities is made by a letter and four-number grid system. The letter refers to the particular ninth of the quadrangle (Fig. 3), the first two numbers are inches and tenths of an inch from the western edge of the ninth, and the last two numbers are inches and tenths of an inch from the southern edge of the ninth.

Acknowledgments

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Figure 3.   Index letters to Burke quadrangle ninths.

John Dennis, who completed mapping of the Lyndonville quadrangle during the author's first field season and was later revising the Littleton quadrangle mapping.

The writer is pleased to acknowledge discussions and advice given freely by Dr. John C. Haff of Mount Holyoke College and by Dr. Walter H. Newhouse of the University of Chicago. He has also benefitted from the assistance of his colleagues at the Chicago Natural History Museum, Dr. Eugene S. Richardson, Jr., Dr. Rainer Zangerl, Dr. Albert Forslev, who made a number of X-ray diffraction determinations, and Miss Maidi Wiebe, who helped in the preparation of the illustrations.

Mr. Malcolm Hibbard in the summer of 1956 and Mr. Christos Alex in the summer of 1957 provided valuable field assistance.

The residents of the Burke area were always extremely helpful; in particular the writer would like to mention Mr. and Mrs. Thomas Gould of East Burke, Mr. Edward Lund, Fire Warden of Granby, and Mr. and Mrs. Kenneth Hoffman of Kirby.

Last, but not least, the writer wishes to express his great debt to his wife, Dr. Mary Vogt Woodland, who was a constant source of encouragement and provided much help in all aspects of the work.

Physiography

The Burke quadrangle lies in the northern portion of the New England Division of the Appalachian Highlands. Approximately the western third of the area consists of partially forested hills and valleys trending north-south; almost all of the farm land of the quadrangle is restricted to this section. Its relief is subdued, with the lowest point 660 feet, in the Passumpsic riverbed in the southwestern corner of the quadrangle, and with the land surface rising gradually northwards to over 1900 feet around Newark Street (Fig. 4). This area is underlain by the calcareous and associated rocks of the Waits River formation, and the trend of the hills and valleys follows in general the regional strike of the strata. Some of the lower hills are composed of glacial deposits, but most have a rock core and are not unlike rock drumlins, e. g., Graves Hill.

The eastern two-thirds of the quadrangle together with the northwestern part is mountainous and virtually under continuous forest cover. Burke, Umpire, East Haven, East, and Seneca mountains all rise to over 3000 feet, East Mountain being the highest with an elevation of 3420 feet. An extensive low-lying boggy area, largely below 1200 feet and ringed with mountains, occupies much of the Township of Victory in the southeastern part of the quadrangle. Burke and Umpire mountains (Fig. 1) stand above the surrounding country like monadnocks and both have a characteristic radial drainage pattern. The rugged nature of this terrain can be correlated with its underlying rock types. Thus, Kirby, Burke, Umpire, and East mountains are composed of a resistant hornfels-granite complex; East Haven and Seneca mountains are composed of rocks of the Gile Mountain formation, in part hornfelsed, and biotite granite. The aforementioned low-lying area in the southeast is believed to be mainly underlain by amphibole granite, an earlier and presumably more easily eroded granite than that in Burke, Kirby, and the other mountains.

The drainage of the area is mainly to the south via the East and West branches of the Passumpsic River which join just north of Lyndonville and via the Moose River. These are all tributaries of the Connecticut River. The northern flank of Bull and Seneca mountains in the north-northeastern part of the quadrangle drains north into the Nulhegan River in the Island Pond quadrangle and thence into the Connecticut River. In the east-northeastern part of the Burke quad-

Figure 4. Subdued topography developed on the Waits River formation in the northwestern portion of the Burke quadrangle. View from the road to the summit of Burke Mountain. Mount Pisgah (in the Lyndonville quadrangle) in left background; Bald Mountain (in the Island Pond quadrangle) in right-center background.

rangle the North Branch of Paul Stream, Madison Brook, and Stony Brook flow eastwards into Paul Stream in the Guildhall quadrangle and thence into the Connecticut River.

There are two dams and accompanying hydro-electric plants on the Passumpsic River, one just west of Red Village and the other in the extreme southwestern corner of the quadrangle.

There are several ponds in the area, Newark, Bald Hill, and Center ponds being the largest. The ponds are chiefly the result of poor drainage impeded by thick glacial deposits. Some of the small ponds are kettle lakes, such as Duck Pond, northwest of East Burke, and Lily Pond, southeast of Lyndonville.

The whole of the Burke quadrangle was glaciated during the Pleistocene. The evidences of glaciation are many; they include the thick glacial deposits which mantle most of the ground and the erratics, amongst which one of the most interesting is the large anvil stone of amphibolite now standing at the west side of the road north of Bundy School (D3847). Areas extensively covered with glacial deposits are shown on Plate 1. The presence of glacial till on the highest peaks such as Burke Mountain proves that the area was completely submerged beneath a thick cover of ice. Glacial striae and grooving can be seen in places, and their trends are recorded on Plate 1. These show that the ice moved generally in a south-southeasterly direction. Roche moutonnées are found in several areas; some of the best examples are on the summit of Spur Ridge (G2907) and southeast of Gallup Mills (I2749). Their long axes are oriented in a north-northwest-south-southeast direction, which is in keeping with the striae trends.

Late glacial deposits of sand and gravel abound in the valleys. Sinuous eskers parallel much of the length of the East and West branches of the Passumpsic, e. g., northeast of White School (E0628) and southeast of West Burke (D1336). Kames are also found along the sides of the valleys and give rise to characteristic hummocky topography, such as along Route 114 north of East Burke (D3409) and northwest of Ward School (D1214). They are composed of crudely stratified sand and gravel, often with many large boulders mainly of local origin. Large quantities of stratified sand with minor gravel occur in the lower reaches of the East and West branches of the Passumpsic and east of Lyndonville; about 50 feet are exposed in the bank east of the Canadian Pacific Railroad just north of Lyndonville (G0231). These deposits which choke the Passumpsic valley represent the remnants of outwash, the upper elevation of which lies at about 860 feet (probably rising to the

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north up the valleys), and which were deposited when the ice paused in its recession at some point to the north, possibly just south of the south end of Lake Willoughby (Lyndonville quadrangle) at the divide between the St. Lawrence and Connecticut drainage systems. The Passumpsic River has now cut deeply into these deposits; a meander scarp in the latter rises above the present floodplain of the East Branch north-northeast of Lyndonville (G0837). In the Egypt area (e. g., G0553) the sand has been and still is subject to wind action and small active sand dunes are present. Varved clay deposits occur at widely scattered points in the southwestern and west-central parts of the Burke quadrangle (e. g., G1606, D1616) and at one locality in the southeast (I0412). The varved clay passes upwards into sandier deposits and finally into sand and gravel. It represents deposits laid down in temporary lakes and perhaps underlies the outwash sands, which were the final deposits laid down. Varved clay, however, is found at higher elevations up the brooks draining the west slopes of Kirby Mountain; the highest elevation at which the writer found it is 1620 feet, in Hawkins Brook (H0213). These higher occurrences were probably laid down in small lakes impounded between the ice that lay in the valley and the mountain side. The one example in the southeastern part of the quadrangle suggests that the low-lying area of Victory also was the site of a late glacial lake.

The "biscuit-board" pattern of the contours on Seneca, East, East Haven, Umpire, and Kirby mountains is remarkably similar to that formed by cirque development in areas of mountain glaciation. Cirque forms are well shown in the headwaters of North Branch, Paul Stream (C2028), the headwaters of the tributaries on the northern side of Madison valley (particularly C2318), the headwaters of the west branch of Greer Brook (E3930), the headwaters of Dish Mill Brook (H2654), and the headwaters of Barnes and Simpson brooks on the western slopes of Kirby mountain (H0628; H0333). They were probably eroded by mountain glaciers after the main continental glaciation. The glacial Madison valley is a hanging valley which plunges steeply eastwards via a series of waterfalls near the eastern boundary of the Burke quadrangle and extending into the Guildhall quadrangle to the over-deepened valley of Paul Stream.

Settlement

Settlement is essentially restricted to the western third of the Burke quadrangle coinciding roughly with the subdued topography developed

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on the Waits River formation. It includes the eastern outskirts of Lyndonville and the villages of East Burke and West Burke which are the principal areas of population and the communities of Burke Hollow, Newark Street, East Haven, East Lyndon, and Red Village.

Dairy farming, lumbering, and tourism are the major industries of this settled area. Near East Burke Darling State Forest Park, embracing an area of more than 1745 acres of Burke Mountain and the northern part of Kirby Mountain, is a scenic attraction during the summer and a skiing center in the winter. Extensive panoramic views are obtained from the summit of Burke Mountain (Fig. 4).

The remainder of the Burke quadrangle is heavily forested and much more inaccessible. The small population is restricted to a few isolated communities such as Gallup Mills, Granby, and Victory. The cutting of lumber and pulp wood is the only significant industry; practically this whole area has been cut over at least once. Gallup Mills was a thriving community at the turn of the century when lumbering in the Moose valley was an important operation and a railroad extended northward up the Moose valley from North Concord.

Evidences of more extensive farming are found throughout the quadrangle in the pastures and orchards encroached upon by forest and by the presence of cellar holes, stone walls, and corduroy roads in the woods, e. g., north of Newark Hollow, north of Granby, around Lost Nation, and around Kirby Pond.

Previous Work

Early geological work in the region was carried out by Hitchcock (1861) and S. R. Hall (1861). Richardson (1906) made a number of references to the area; he described the pink granite of Newark and referred to fibrolite on Burke Mountain and to diabase dikes in Newark. Dale (1909, 1923) described the granite quarries in Newark and Kirby. The small part of the Township of Westmore that falls in the Burke quadrangle was included in the geological map of Jacobs (1923). Detailed geological accounts of the neighboring quadrangles (Memphremagog, Doll, 1951; Lyndonville, Dennis, 1956; Littleton, Eric and Dennis, 1958; and St. Johnsbury, L. M. Hall, 1959) have been published by the Vermont Geological Survey. Geological mapping of the Guildhall, Island Pond, and Averill quadrangles was in progress during the preparation of this report and reports have now been published (Johansson, 1963; Goodwin, 1963; Myers, 1964).

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STRATIGRAPHY

General Statement

Three formations have been mapped in the Burke quadrangle—the Albee, Waits River, and Gile Mountain—all of whose contacts have in general a north-northeasterly strike. The latter two occur within the "Vermont sequence" while the Albee formation is part of the "New Hampshire sequence" (White and Jahns, 1950). The two sequences are separated by the Monroe fault, which was recognized to the south (Eric, White, and Hadley, 1941) and has been mapped, with some gaps, for over 80 miles from the Hanover quadrangle (Lyons, 1955) northwards into the Littleton quadrangle (Eric and Dennis, 1958), and is believed to occur at North Stratford, New Hampshire (Billings, 1956). In the Burke quadrangle the line of the fault is largely interrupted by granite intrusion and high-grade metamorphism; the contact is recognized only for a distance of a little over a mile on the eastern border in the vicinity of Granby. No actual contact was discovered as the exposures are not plentiful along its assumed line, which appears to follow a small valley and strikes northeastward into the Guildhall quadrangle. In the latter the Meetinghouse slate is found immediately west of the Albee formation (Johansson, 1963) as it is in the Littleton quadrangle (Eric and Dennis, 1958). It has been suggested that the Monroe contact between the Vermont and New Hampshire sequences may be an unconformity; the arguments for and against a fault or an unconformity and the structural and stratigraphic implications have been fully discussed by Eric and Dennis (1958), L. M. Hall (1959), Cady (1960), and Johansson (1963). No new information bearing on this problem has been discovered by the writer in the Burke quadrangle.

Thicknesses: The only one of the three formations exposed in the Burke quadrangle that has both an eastern and a western contact is the Gile Mountain. There is but a small portion of the Albee present along the eastern border of the quadrangle; the western contact of the Waits River formation lies in the Lyndonville quadrangle to the west (Dennis, 1956, Plate 1). White and Billings (1951) give a figure of 10,000 feet for the thickness of the Waits River in the Woodsville quadrangle.

The Gile Mountain formation has a maximum width of outcrop of about eight miles measured normal to the strike and along a line from a point between East Burke and Hartwellville in the west and the

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Monroe fault near Granby in the east. The average dip of this section is between 60° and 70° southeastwards. Thus the apparent thickness of the formation at this latitude is 36,000 to 39,000 feet. Estimates of true thickness are exceedingly difficult to make in highly deformed metamorphic rocks. This is true in the Burke area and a study of the literature shows it to be so for eastern and central Vermont in general. The prevalence of minor folding suggests that repetition of beds by folding is probably important. Geologists working in other areas of eastern and central Vermont have applied a rough factor to correct for this thickening of beds. The writer is certain that the great apparent thickness of the Gile Mountain formation in the Burke quadrangle is the result of considerable repetition, but there are insufficient data to estimate its extent. Doll (1944) gives a figure of 6,500 feet for the Gile Mountain (excluding the Meetinghouse slate) in the Strafford quadrangle and White and Billings (1951) estimate a thickness of 6,000 to 7,000 feet in the Woodsville quadrangle. Thickening of the order indicated suggests repetition by major folding. Although there is no direct evidence for this, it is possible that it exists; this is discussed in the section on Major Structural Features.

Age of formations: No data bearing on the age of the three formations were found in the Burke quadrangle. The Albee formation was considered Ordovician in age by Billings (1934, 1937) based on its relations in New Hampshire. The determination of the age of the Waits River and Gile Mountain formations has been fraught with difficulties. Two possibliities have been discussed and supported by various geologists. An Ordovician age was suggested by Richardson (1902) and in more recent years by Cady (1950), White and Billings (1951), Lyons (1955), and Billings (1956). The rocks of southern Quebec along the line of strike of the Waits River and Gile Mountain formations are now called the St. Francis formation and assigned to the Ordovician (Cooke, 1950, 1957). A Silurian or Devonian age has been given to the Waits River and Gile Mountain formations by Doll (1943a, 1943b, 1944, 1951), Hadley (1950), Cady (1956), Dennis (1956), Eric and Dennis (1958), and L. M. Hall (1959). Billings (1956), although he placed the formations in the Ordovician, considers a Silurian or Devonian age a distinct possibility. In keeping with most of the recent views, the Waits River and Gile Mountain formations are assigned a Silurian and/or Devonian age in this report.

Full discussion of the age problem is given in Billings (1956), Dennis (1956), and Cady (1960).

Sequence: The correlation and order of the formations of the Vermont

sequence have been given a number of interpretations (see L. M. Hall, 1959, pp. 37–41 and Fig. 15, and Cady, 1960, for a complete summary of views). The most favored order is that the rocks are progressively younger eastwards from the Green Mountains so that the Waits River is followed upwards by the younger Gile Mountain and by Meetinghouse slate. The reverse order has also been suggested by White and Jahns (1950), who reject it, by Billings (1956), who considers it unlikely, and by Murthy (1958, 1959a, 1959b), who favors it. A further complication is the suggestion that the Northfield slate, a formation that underlies the Waits River to the west in the Hardwick quadrangle, is the equivalent of the Meetinghouse slate. This involves difficulty in explaining the eastern contact of the Gile Mountain formation and Meetinghouse slate or the western contact of the Northfield slate and the Waits River formation depending upon whether the Waits River is regarded as being older or younger respectively than the Gile Mountain. In the first case the Waits River would be apparently missing and in the latter case the Gile Mountain; these omissions may be explained tectonically or by facies change, which would mean equivalence in part of the two formations. No evidence has yet been forthcoming to support either of these possible explanations or the equivalence of the Northfield and Meetinghouse slates. The relationship between the Waits River and Gile Mountain formations in the Burke quadrangle will be referred to again in the section on Major Structures.

Albee Formation

The Albee formation was named for Albee Hill, New Hampshire, in the Littleton quadrangle (Billings, 1934). In the Burke quadrangle it has very restricted outcrops, all of high metamorphic grade. The exposures northeast of the village of Granby present the more typical lithologies, as described by Billings (1937) and by Eric and Dennis (1958). Here the rocks include quartzose and pelitic beds in the staurolite grade—not unlike parts of the Gile Mountain formation. They are fine-grained, gray to green rocks, composed of quartz, abundant biotite, in places sericite, garnet (up to 1.15 mm.), staurolite (up to 1 mm. long), sparse to abundant chlorite, and rare plagioclase (Fig. 5).

Outcrops on Lees Hill and Miles Mountain are of high-grade sillimanite hornfels associated with granite. It is possible that these outcrops are part of the Albee formation, but at their high grade of metamorphism there is no way to distinguish them from the Gile Mountain formation, at least not those examined in detail by the writer. Possibly one dis-

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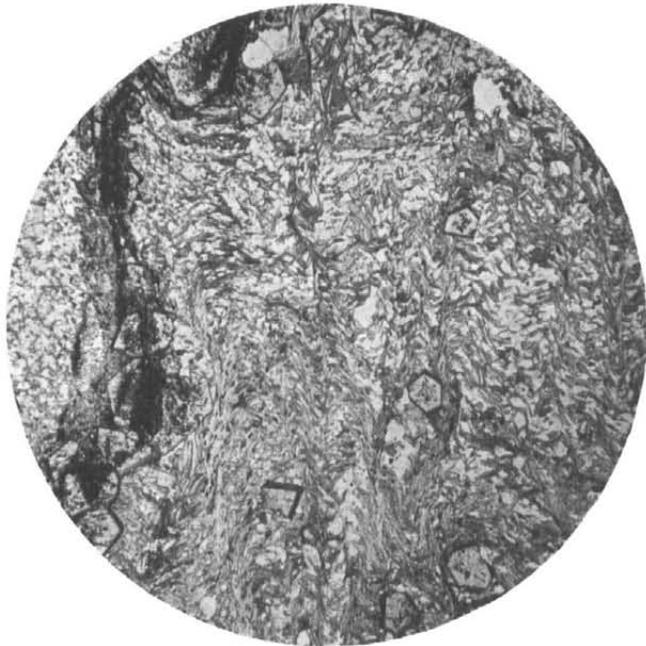
Figure 5. Photomicrograph (×20) of staurolite-garnet-mica-quartz schist from the Albee formation near Granby (I4251). Biotite and sericite form a wavy schistosity (S₁) and are in part oriented into a cross-cutting cleavage (S₂). Staurolite, garnet, and biotite occur as porphyroblasts. Note linear trend of garnets with opaque inclusions subparallel to S₂.

tinction, which may reflect the metamorphic history rather than the composition of the original strata, is that these rocks contain in general more prominent sillimanite and fewer or no sericitic aggregates after andalusite.

Amphibolite: Northeast of Granby amphibolitic bands are common in the limited exposures of the staurolite zone. Quite distinctive from the amphibolites of the Waits River and Gile Mountain formations, these are green, fine-grained, foliated rocks, often with small white or pink elongate spots or streaks up to 7 mm. long, which may represent deformed amygdules. The amphibolite is composed of abundant blue-green hornblende 0.1 to 0.9 mm. long, some altered to chlorite, and of very fine-grained interstitial areas (grain size 0.01 to 0.04 mm., seldom up to 0.2 mm.) containing rare, poorly twinned plagioclase (var. labradorite?), untwinned plagioclase, and quartz. Some of the interstitial quartz is

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rather larger, up to 0.36 mm. Poikiloblastic garnet is common (up to 0.9 mm.), apatite is sparse; there is some rare sericite, and iron ores are common. Thin quartz veins cross-cut the foliation.

These amphibolites represent metamorphosed diabase sills, dikes, or flows.

Waits River Formation

Richardson (1906) named the thick sequence of limestone and schist in eastern Vermont the Waits River limestone, re-named the Waits River formation by Currier and Jahns (1941). Doll (1951) mapped calcareous rocks in the Memphremagog quadrangle as the Ayers Cliff and Barton River formations and Dennis (1956) mapped the southward continuation of the Barton River formation in the Lyndonville quadrangle as Waits River. Eric and Dennis (1958) show this formation to outcrop in the northwestern corner of the Littleton quadrangle. Lithologically similar rocks outcrop, as continuations of the Lyndonville and Littleton occurrences, throughout the western part of the Burke quadrangle, essentially west of Route 114, but also south of East Burke in the structurally complex area of East Lyndon and Simpson Brook (Plate 1).

Lithologic types include impure recrystallized limestone, calcareous phyllite, calcareous quartzose phyllite, calcareous quartzite, phyllite, quartzose phyllite, amphibolite, meta-quartz keratophyre, calc-silicate hornfels, and staurolite-garnet-biotite schist. These occur interbedded in varying proportions and thicknesses, in bands from a few inches to several feet thick, more rarely as much as nine to ten feet thick. Some outcrops contain considerable impure limestone with subordinate phyllite or schist and quartzose rocks, while others contain little or no limestone. Outcrops believed to be typical of the general occurrence in the quadrangle are well-exposed in Mountain Brook (G2446), in Dish Mill Brook (D3104), in Calendar Brook (D0116), and in Roundy Brook, north of Burke Hollow (D3033). North of East Sutton Ridge and Packer Cemetery, granitic sheets, sills, and tongues are prevalent, intruded mainly parallel with the limestone bedding, with the consequent development of calc-silicate minerals and occasionally of calc-silicate rocks and skarn. Amphibolite and meta-volcanics are confined to the more easterly outcrops.

Impure limestone: This is a light gray to blue-gray and dark gray rock which weathers to a buff-brown color. It is compact and fine-to medium-grained, with a poor foliation formed by sericite, phlogopite, or biotite. It frequently contains much quartz and mica, thus grading into cal-

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careous quartzose phyllite. Occasionally it has a fissile structure and weathers to a loose brown sand. Thin, recrystallized segregation veins of quartz, with some calcite, less than one-eighth inch thick parallel the foliation. Quartz veins, a few millimeters to a few feet thick, commonly occur, but in general are not so prevalent as in the Gile Mountain formation.

Between West Burke and the northern outskirts of Lyndonville the limestone outcrops are frequently well-bedded and regular, e.g., in Quimby Brook, at the western edge of the Burke map (G0054). Farther to the east and southeast the limestone bands are usually strongly deformed internally and are discordant at their contacts with phyllite, sometimes in a very intricate, complex fashion. Indeed, in some occurrences 2 to 3 inches thick, tightly folded limestone bands or lenses have external contacts which are tectonic in origin rather than original bedding, the calcareous material having flowed into its present position. These occurrences are typical of calcareous rocks and marble flow-folded under the influence of strong deformation. Flow folding is well exhibited in Mountain Brook (G2446), in Simpson Brook (G2727), and west of Burke Hollow (D2031). (See Figs. 6, 7, 8, and 9).

Microscopically the limestone contains calcite and quartz in varying proportions, the latter often about equal in amount to the calcite, and ranging in grain size from about 0.06 mm. to 0.2 mm. and up to 0.35 mm. Sericite is common, particularly in the more phyllitic limestone; phlogopite or biotite is common (up to 0.3 mm.). Chlorite flakes, up to 0.87 mm. and randomly oriented, are present in some thin sections. In one slide cut from a flow-folded specimen the flakes are bent and deformed, with undulose extinction. Tremolite porphyroblasts (sometimes actinolitic) are commonly developed in radial, brush-like aggregates, in places up to 2 cm. in diameter (Fig. 10). Opaque minerals are ubiquitous; fine graphite dust and sulfide grains are common. In the higher metamorphic grades, e. g., on Burke Mountain and adjacent to the granite masses in the Newark area, the impure limestone has been converted into calc-silicate hornfels with diopside and grossularite. The rocks range in texture from those with grains less than 0.5 mm. to those with garnet dodecahedrons up to 5 mm., amphibole nearly 5 mm. long, and diopside up to 3 mm. long.

Pelitic rocks: Interbedded with the calcareous rocks, the pelitic rocks are usually well-foliated, biotite-rich, and dark-colored, with an abundance of fine opaques and much pyrite, which gives rise commonly to rusty-colored outcrops upon weathering. In general, the mineralogy of

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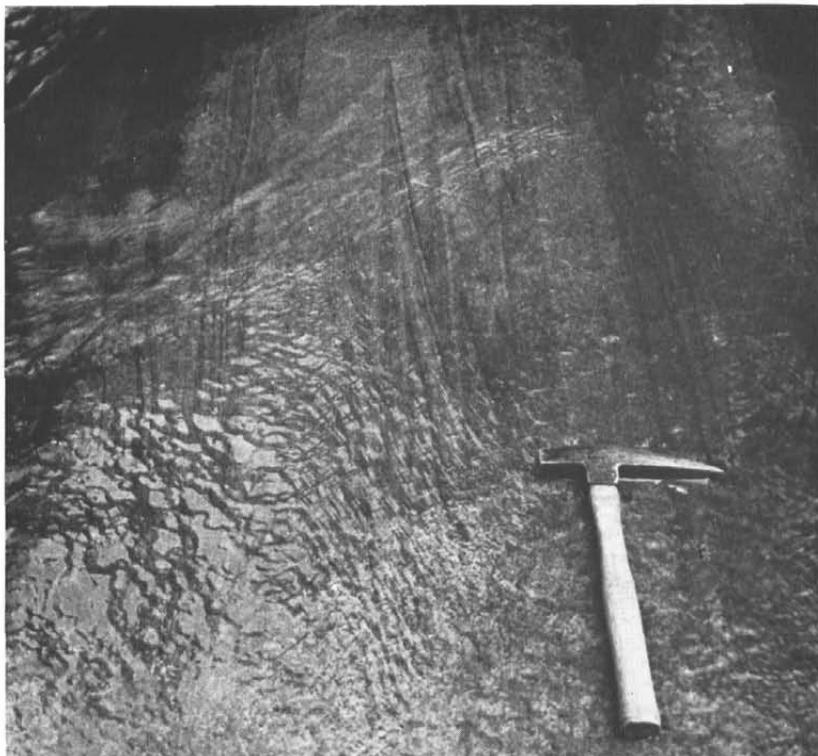
Figure 6. Sharp minor folds in Waits River impure limestone in Mountain Brook (G2446), looking south.

the pelitic rocks is similar to that of the pelite described below in the Gile Mountain formation. Garnet and staurolite occur in the higher metamorphic grades. Andalusite and kyanite are very rare; kyanite occurs at B0347 and both are found together in an unusual occurrence on the east side of Route 114 (B1848) associated with abundant biotite and pegmatitic material (this is further described under Metamorphism). Tourmaline schist occurs at D4053 near pegmatitic veins.

The pelite bands in the southwestern part of the Burke quadrangle have been intensely folded and often appear as discontinuous outcrops surrounded by flow-folded limestone. They are also strongly cleaved, usually possessing one or more cleavages cutting the main schistosity.

Psammitic beds: Quartzose rocks are commonly interbedded with the limestone and pelitic strata and are similar to the psammitic beds in the

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Figure 7. Flow folding in Waits River impure limestone in Simpson Brook (G2728), looking west.

Gile Mountain formation. The quartzose types are more plentiful in the zone transitional to the Gile Mountain than in the Waits River proper. They are fine-grained (the quartz grains are 0.03 to 0.3 mm.), usually with a well-developed sericite foliation, and have biotite both sub-parallel to the foliation and as porphyroblasts up to 0.45 mm.; opaque dust streaked in the foliation is sometimes present, and sulfides are ubiquitous, often elongated in the foliation. Chlorite occurs occasionally in the foliation in flakes 0.06 mm. across, but more commonly as porphyroblasts up to 0.65 mm. Small garnets are occasionally present. At one locality (G2552) in the more pelitic bands transitional to phyllite staurolite has been observed. Plagioclase (albite-oligoclase) is a minor constituent in

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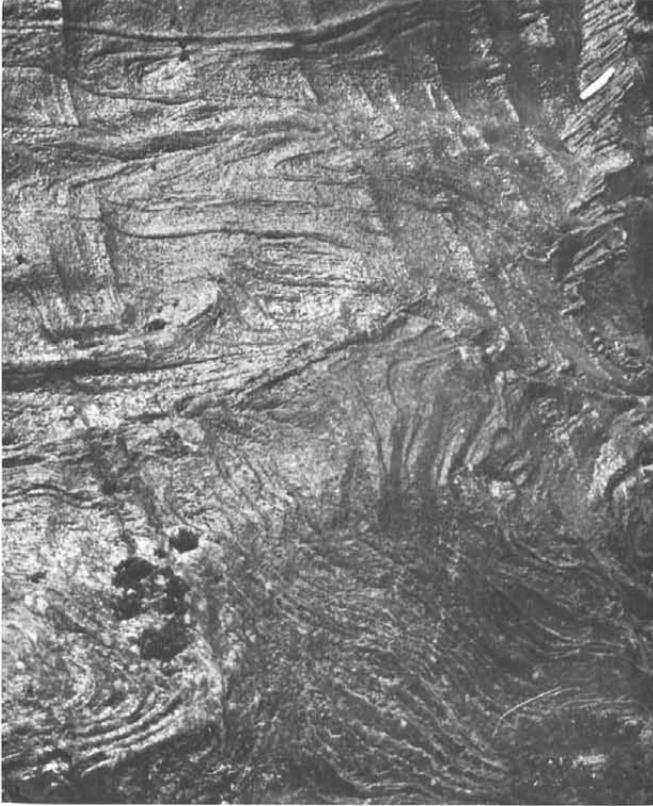
Figure 8. Flow-folded Waits River impure limestone cut by joints, which are healed by quartz. Note quartz rodding, caused by deformation of a quartz vein; rodding trends 152° and plunges 40° to the southeast (parallel to penknife, three and one-half inches long). In Simpson Brook (G2727), looking west.

some of the thin sections. Calcite may occur in small quantities in the bands transitional to the impure limestone. Poikiloblastic muscovite is present in some psammitic rocks, notably in the East Passumpsic valley between Hartwellville and White School. Accessory minerals are minute greenish tourmaline, zircon, sphene, and apatite. Quartz veins, from 0.5 mm. to a few centimeters thick, are frequent, usually paralleling the foliation and in places containing calcite; they are often folded and some-times are cut by a cleavage.

Amphibolite: Amphibole-rich rocks occur in general in the belt of country comprising the transition between the Waits River and Gile

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Figure 9. Flow-folded Waits River impure limestone with veins and masses of quartz. On hillslope west of Burke Hollow (D2031), looking east.

Mountain formations, e. g., along the East Passumpsic valley. The easternmost bands closest to the Gile Mountain formation probably can be correlated with the Standing Pond amphibolite described by Doll (1944) in the Strafford quadrangle and mapped by L. M. Hall (1959) in the St. Johnsbury quadrangle. The rocks are greenish in color and normally have a rather crude foliation parallel to that of the adjacent phyllite or limestone. The outcrop on Spur Ridge (G2908), however, is a massive sill-like exposure at least 12 feet thick, with only one boundary visible. Other similar outcrops nearby indicate a comparatively thick occurrence of amphibolite in Spur Ridge. The weathered surfaces show a marked reticulate structure, produced by the more resistant material of thin cross-cutting veins (Fig. 11). These veins probably represent metamorphic differentiation by secretionary growth of the central quartzose-feldspathic material and enrichment of hornblende along the borders (Eskola, 1932; Ramberg, 1952, pp. 216–218 and Fig. 105).

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Figure 10. Photomicrograph (×20) of Waits River impure limestone with calcite, quartz, and radiating "spongy" porphyroblasts of actinolitic tremolite. West slope of Graves Hill (G1326).

Usually the amphibolite is poorly exposed in scattered outcrops a few feet across which sometimes may be parts of a band 10 to 20 feet thick but which may also be bands 3 to 10 feet thick interbedded with phyllite and limestone. The weathered surfaces of amphibolite are frequently pitted as a result of solution of carbonates.

Amphibole is always a major constituent, medium to coarse grained and up to 19 mm. in length; it is usually a green, poikiloblastic, xeno-blastic hornblende, although in some cases it may be actinolitic (Z varying from blue-green to dark green). Sometimes the amphibole has altered in part to chlorite. The groundmass of the amphibolite is fine-grained (0.02 to 0.1 mm.), generally of untwinned plagioclase and quartz (Fig. 12). Definite determination of the plagioclase is usually not possible because of the small grain size and diffuse grain boundaries. Twinned plagioclase (up to 0.5 mm.) is sometimes common, but more usually is rare, and the twinning is often indistinct. The twinned plagioclase seems

Figure 11. Reticulate weathered surface of amphibolite on the northern slope of Spur Ridge (G3008).

to vary in composition from a fairly sodic type to labradorite. Biotite (0.02 to 0.36 mm.) is usually present in small amounts, but in some occurrences it is common, giving a slightly purplish hue to the fresh rock surface. Chlorite (0.1 to 0.36 mm.), often penninite, is rare, but in some thin-sections it is common. A small amount of clinozoisite and epidote occurs very seldom. Opaque minerals include sulfides and ilmenite, altered in part to leucoxene. Accessory minerals include apatite (in grains up to 0.2 mm.) and sphene (up to 0.78 mm.), partly altered to leucoxene. Calcite is often present in small quantities, particularly when the amphibolite is interbedded with limestone. Tourmaline (pleochroic colorless to smoky gray, blue, or brown) occurs in a few thin sections; in one coarse-grained calcareous "inclusion" in amphibolite tourmaline is abundant in a stellate pattern and is associated with axinite and quartz.

Two unusual occurrences at G1010 and at G1015 merit special mention. These are green rocks with a well-developed schistosity, which has a

Figure 12.   Photomicrograph (×18) of Waits River amphibolite showing poikilo-blastic hornblende in a groundmass of plagioclase, quartz, calcite, apatite, chlorite, and iron ores. Northwest of Graves Hill (G0930).

silky, greasy feel. They are composed of abundant blue-green or colorless amphibole, which is granulated and much reduced; abundant minute flaky talc and chlorite (some chlorite in aggregates and lenticles); abundant dusty opaques and grains of iron ores; a little fine-grained felsic material; and occasional garnet and apatite. A thin section (Fig. 13) of this rock shows it to be made up of exceedingly fine-grained, ground-up minerals, streaked out in places into curved shear planes, sometimes parallelled by lenticular areas of chlorite and amphibole. These two occurrences are of amphibolite highly sheared to the point of mylonitization. In the exposure at G1015 normal amphibolite outcrops within 100 yards of the sheared rock.

Light-colored hornblende quartzose bands: These are characteristically developed as bands a few inches thick intercalated among phyllite and limestone beds in the transition zone between the Waits River and Gile Mountain formations. As they are more common in the latter, a description is deferred to the section dealing with the Gile Mountain.

Meta-acid igneous rocks: Metamorphosed acid lavas and perhaps tuffs

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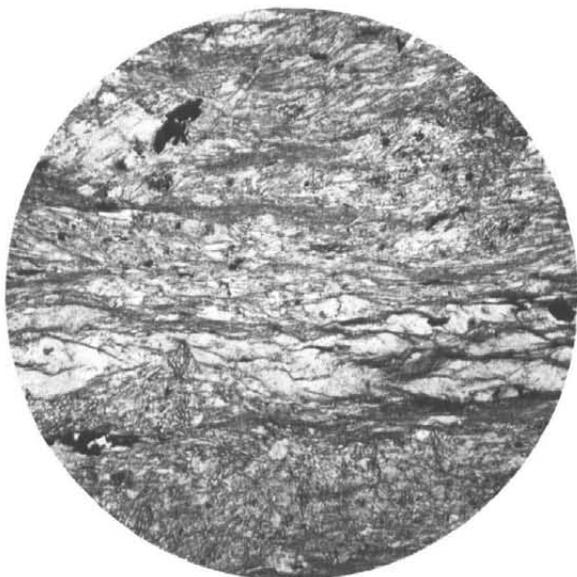
Figure 13. Photomicrograph (×18) of sheared amphibolite showing granulated amphibole, chlorite, garnet, opaques, ? talc. From the southern slope of Shonya Hill (G1015).

are exposed at a number of localities (Plate 1), and appear to represent a single volcanic episode. They occur in the transition zone between the Waits River and Gile Mountain formations. Limestone lies near most of the meta-acid outcrops; just south of one occurrence, at G3050, there is amphibolite. These meta-acid types are placed, therefore, within the Waits River formation.

Eric and Dennis (1958) refer to meta-rhyolite tuff associated with amphibolite just within the Gile Mountain formation of the Littleton quadrangle. It is probable that this occurrence should be correlated with the acid lavas and tuffs of the Burke quadrangle.

These meta-acid igneous rocks are gray, fine-grained, crudely foliated, and composed mainly of sodic plagioclase, quartz, and biotite. The plagioclase is albite or oligoclase, which in some thin sections is in microlites, often bent, 0.2 to 0.3 mm. long, in a felted or almost bostonitic texture (Fig. 14). In other sections the plagioclase is xenomorphic granular, often less than 0.05 mm., both twinned and untwinned. Some euhedral blastophenocrysts occur up to 0.5 mm., with inclusions of biotite. Quartz may be sparse to common; it is usually fine grained although some

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Figure 14. Photomicrograph (×18) of a crudely-foliated meta-quartz keratophyre containing oligoclase in a felted texture, quartz, and biotite. From a small stream northeast of Mount Hunger (G2950).

coarser grained pods also occur. Biotite (or lepidomelane) is either scattered throughout, in flakes 0.02 to 0.25 mm., or is aggregated into trains producing a crude foliation, or appears as abundant porphyroblasts up to 4 mm. Muscovite is sparse, but in one thin section it is concentrated in linear aggregates. Calcite occurs sparsely in the groundmass and also in amygdules, sometimes with quartz. A little sphene and apatite are present in some specimens. Opaque dust may be plentiful; iron ores, particularly sulfides, are common in some instances. Some microcline and cryptoperthite occur in one thin section. This same slide also has an area of a coarser grained aggregate of microcline, cryptoperthite, and quartz which may represent a cognate xenolith. A small sheared lenticle in Mountain Brook (G2645) may also be representative of these meta-acid flows. It is highly altered, with many opaques and abundant small tourmaline; the feldspar is nearly all altered to muscovite.

These meta-acid occurrences represent lava flows with an affinity towards quartz-keratophyre. The sodic nature of the feldspars may be secondary, resulting from the metamorphism, in which case they would have originally been more calcic.

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A series of outcrops in the East Passumpsic River at D3918 presents varied metamorphosed igneous rock types. The exposures are poor and the relations to the country rock and of the types one to another are not known. They are gray, medium- to fine-grained rocks with a crude biotite foliation, and are composed predominantly of feldspar, biotite, and quartz, with garnet and accessory minerals. The kind of feldspar (up to 6 mm.) varies from specimen to specimen. In some it is mainly plagioclase, in laths and sometimes in larger blastophenocrysts of albite or albite-oligoclase; the laths are frequently bent. The twin lamellae are often very fine and indistinct, and the grains are usually full of indeterminate granular alteration products, including calcite and possibly zoisite. A little myrmekite is sometimes found as well as microantiperthite. There are also patches of fine granular material which is probably untwinned plagioclase (albite?) replacing feldspar. Cryptoperthite is present, usually in small quantities, but in one thin-section it is common. Quartz occurs sparsely to abundantly. Biotite is a common constituent, often in reticulate aggregates. Garnets (up to 3 mm.) are sparse to abundant, poikiloblastic and xenoblastic to idioblastic, altered slightly to chlorite and epidote. Chlorite is present in some slides; calcite is often sparse, but is common in a few thin sections. Accessories are apatite and sulfides. One specimen from a varied outcrop contains abundant ragged poikiloblasts of blue-green hornblende.

It is difficult to interpret these occurrences, particularly as the outcrops are small and their relationships obscure. They may represent metamorphosed intrusives of a quartz diorite type, or they may represent varied acid flows (now sodic) to be correlated with those described above, or there may be both flows and intrusives present.

Two other obscure exposures may also represent correlatives of the flows. At G1112 there is a light-colored, banded, felsitic rock composed of fine-grained feldspar (0.01 to 0.06 mm.), including rare twinned plagioclase, and of quartz, with biotite (partly altered to chlorite) in aggregates, chlorite streaks, and sparse apatite and opaques. There is a suggestion of a folded, sheared foliation. This is possibly a sheared metamorphosed rhyolite or tuff. The other exposure is part of an outcrop at G2717 composed of abundant biotite, with mesh-like inclusions of rutile, anhedral feldspar, abundant quartz, calcite, apatite, pyrite, and a little muscovite and sphene.

GILE MOUNTAIN FORMATION

Gile Mountain schists was the name given by Doll (1944, p. 18) to rocks well-exposed on a mountain of that name in the Strafford

quadrangle. In the Burke quadrangle the Gile Mountain formation occupies a wide belt of country approximately east of Route 114 and extending from the southern border to the northern border of the quadrangle except where interrupted by areas of granite (Plate 1). The eastern boundary is mappable for only a very short distance in the vicinity of Granby village, where the lithologic trend is northeast-southwest in the small stream valley about three-quarters of a mile northeast of the Granby postoffice. The trend turns north-south before disappearing into the probable granite area just southwest of the point where the Granby road crosses the Granby-Victory town line. The western contact of the Gile Mountain with the Waits River formation is arbitrarily mapped as extending in a general northeast-southwest direction, apart from the complication of the boundary in the Shonya Hill—Graves Hill area produced by severe folding and shearing. The contact is interrupted by the granite intrusion north of East Haven. Where it reappears to the north of the granite mass it has a northwest trend (Plate 1).

The rock types comprising the Gile Mountain formation include fine-grained phyllitic quartzite, quartzose phyllite, dark gray phyllite, quartz-sericite-biotite schist, garnet-sericite-biotite schist, staurolite-garnet-mica schist, andalusite- and sillimanite-bearing hornfels, amphibole quartzite, and amphibolite. The characteristic lithology of much of the formation is a banded alternation of light gray quartzose phyllite and dark gray pelitic or slaty phyllite (Fig. 15). This lithology is well-exposed in accessible outcrops east of the road in the Kirby Pond area (e. g., H 0311) and in the Moose River and adjacent road cuts between Moccasin Mill and the confluence of the East and West branches.

The individual bands range generally from a few millimeters to several centimeters in thickness but occasionally as much as 3 to 4 feet; very thinly banded or pin-striped outcrops occur at I4157. Banding of this type is not always developed. Quartzose phyllite may predominate in the outcrops, with the pelitic layers subordinate; this appears to be more noticeable in the outcrops close to the boundary with the Waits River formation.

Highly metamorphosed, bare exposures are to be seen on top of Burke Mountain around the foot of the fire tower and particularly in the vicinity of West Peak.

Quartzose phyllite: The quartzose rocks are, in general, light-colored, foliated, and never composed purely of quartz. They occur from the lowest metamorphic grade of the area (the biotite zone) to the highest grade (the sillimanite zone), maintaining their characteristics except

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Figure 15. Typically banded quartzose and pelitic phyllite of the Gile Mountain formation northeast of Kirby Pond (H0211).

when metasomatised and granitized. Close inspection reveals a foliation parallel or closely parallel to bedding contacts, where the latter are observed by lithologic contrast. This schistosity may be due to sericite or to microlaminae of biotite which may represent original clay partings in some cases but in others are due to crystallization segregation along closely spaced shear planes. These alternative origins produce very similar results and care is needed to discriminate between them; sometimes it is impossible to decide in a restricted outcrop. Often the rock possesses microlaminations about a millimeter thick composed of alternating lighter and darker quartzose layers, obviously an original sedimentary feature.

Microscopically these rocks are composed predominantly of a quartz mosaic (individual grains vary in size from 0.01 to 0.35 mm., averaging 0.06 to 0.15 mm.), usually, but not always, with sericite in flakes (0.02 to 0.2 mm.) which are oriented to form a poor to fairly well developed schistosity and which are accompanied occasionally by chlorite flakes (0.02 to 0.06 mm.). The sericite schistosity is microfolded and cross-cut by a cleavage related to the limbs (usually the short limbs) of the

Figure 16. Photomicrograph (×18) of a quartz schist from the Gile Mountain formation showing a microfolded sericite schistosity and cross-cutting slip cleavage with sericite; biotite and chlorite porphyroblasts. South of Spur Ridge (G2703).

microfolds and with sericite aligned parallel to it. This structure is not so well developed as in the more pelitic bands but in very narrow bands it may be dominant (Fig. 16). Often the false cleavage cutting poorly developed schistosity in the sericite-poor beds is a fracture cleavage (close joints) (Fig. 38).

Red-brown, pleochroic biotite, in flakes 0.03 to 1.1 mm., is ubiquitous. Sometimes it lies parallel with the main schistosity, forming the latter, particularly in the absence of sericite. Sometimes it is aggregated into streaks within which the flakes are not well oriented. Occasionally it is also in part parallel with the cleavage. Frequently it is porphyroblastic without obvious orientation and cutting both schistosity and cleavage. Inclusions of opaque minerals in the biotite parallel the main schistosity. Chlorite porphyroblasts up to 0.63 mm. are common. Some are sub-parallel to the foliation and occasionally to the slip cleavage, but frequently they show no visible orientation. Opaque minerals are common and include fine dust and grains of sulfides up to 0.8 mm. across. The latter are sometimes elongated in the schistosity, while the dust is frequently in streaks parallel to both the main schistosity

Figure 17. Photomicrograph (×18) of a folded quartz vein. The schistosity (S₁) parallels the vein; both were folded together. Slip cleavage (S₂) is well developed and has ruptured the quartz vein in places. Biotite porphyroblasts. West of Shonya Hill (G0821).

and the slip cleavage. Accessory minerals are tourmaline (0.03 mm.), zircon (up to 0.04 mm.), rare plagioclase (albite-oligoclase, 0.08 to 0.24 mm.), rare perthite (0.18 mm.). Calcite (0.08 to 1.5 mm.) occurs sparsely near the Waits River contact.

In thin section quartz veins are common, with their thickness ranging from 0.1 mm. to a few millimeters and with grains up to 1.4 mm. In the field veins up to 5 feet thick occur. The veins often parallel the foliation, but also crosscut it; they are intricately folded and sheared by the movements that caused the slip cleavage. In Figure 17 the sericite foliation can be seen to cut the quartz vein. Some calcite also occurs with the quartz.

Higher grades of metamorphism are shown by the rare development of garnet (up to 0.9 mm.) and of staurolite poikiloblasts. In the highest grade, the sillimanite zone, the quartzose rocks still remain remarkably recognizable; the grain size has coarsened a little and the texture is hornfelsic and "sugary." The foliation is often but by no means always

destroyed. Microscopically varying amounts of metasomatism may be seen; this is described in the section on Metamorphism and in Woodland, 1963.

Before metamorphism the quartzose phyllites were coarse siltstones and fine sandstones mainly of a quartz wacke type.

Pelitic rocks: Intercalated with more quartzose rocks the pelitic rocks occur in bands from less than 1 mm. to several inches thick, more rarely several feet thick. Typically, they are gray to dark gray and fine-grained, with a well-developed schistosity, frequently cut by a later cleavage and sometimes strongly lineated. The development of porphyroblasts in the higher grades of metamorphism coarsens the texture, but the groundmass remains surprisingly fine-grained even into the highest grades (sillimanite). The pelitic rocks thus include slaty phyllite, phyllite, and schist. Much of their occurrence is of a grade higher than that normally associated with phyllite, yet their field appearance is not that of typical schist.

Sericite-muscovite is an abundant constituent together with increasing amounts of fine-grained quartz (0.03 to 0.08 mm.) as the rock grades into more quartzose types. Chlorite laths 0.06 to 0.18 mm. long occur in some pelite of the lowest grade. Red-brown biotite is ubiquitous, usually porphyroblastic, with cleavage flakes up to 1 mm. apparently not oriented, but cutting across the schistosity and cleavage where the latter are developed (Fig. 31). Sometimes, however, it is oriented in and is responsible for the schistosity, and sometimes it is partially oriented in the cleavage planes, showing the structural control of post-tectonic crystallization. Rarely some thin pelitic bands (up to 2 to 3 cm. thick) are exceedingly rich in biotite, being composed almost exclusively of this mineral.

Minute carbonaceous specks frequently occur; in some thin-sections they are abundant and streaked out in the schistosity and cleavage. Within porphyroblasts of biotite, chlorite, garnet, and staurolite the trace of the schistosity is often clearly marked by the opaque streaks. Sulfide grains are common (up to 0.2 mm.), as well as some ilmenite altering to leucoxene. The pyrite is sometimes elongated and oriented within the schistosity. Minute tourmaline occurs sparsely.

Increasing grade of metamorphism is indicated by the appearance of idioblastic garnet (usually up to 1 mm., but more rarely 4 to 8 mm.) and staurolite (up to 12 mm. long) which is often idioblastic (Fig. 18), with inclusions of quartz and opaque minerals, but is sometimes xenoblastic

Figure 18. Photomicrograph (×44) of staurolite-biotite schist from the Gile Mountain formation showing euhedral staurolite and biotite with many opaque dusty inclusions in a groundmass of sericite, quartz, and iron ores; the sericite forms a sharply microfolded schistosity; shadowy chlorite. From the Moose valley (F1523).

and "spongy" with inclusions. The staurolite may have a "corona" of sericite or may even be almost or completely replaced by sericite-muscovite. Occasionally it is partly replaced by chlorite. Some outcrops on the lower western slopes of Kirby Mountain have bands of spotted phyllite. The spots are ovoid (up to 9 mm. long) and consist of aggregates of biotite, chlorite, muscovite, and quartz. These may represent altered cordierite, although this mineral has not been identified in any thin-section. Higher grade pelite contains andalusite (Fig. 19), often replaced completely by sericite, and sillimanite. These occur close to the granite contacts and in rocks that are clearly distinguishable in the field as having undergone a "contact" type of metamorphism. They are more fully described later in the section on Metamorphism and in Woodland, 1963.

Chlorite porphyroblasts (up to 6 mm.) occur plentifully in some areas in rocks of the garnet-staurolite grade (Fig. 65). In the more quartzose pelites accessory minerals are minute zircon and olive-green tourmaline. Quartz veins paralleling and cutting the schistosity are common.

Figure 19. Photomicrograph (×18) of Gile Mountain andalusite hornfels showing idioblastic andalusite (var. chiastolite) with inclusions of opaques, quartz, biotite, and muscovite in a fine groundmass of quartz, biotite, muscovite, and opaques. From East Mountain (F2947).

Tourmalinization of phyllite adjacent to a granitic dike has been noted at G3627. The phyllite has been converted to a tourmaline-quartz rock; the tourmaline is oriented lengthwise in bands and accompanied by considerable magnetite.

Amphibolite: Bands of amphibolite parallel with the bedding foliation outcrop at several places in the Burke quadrangle, as shown on Plate 1. Total thicknesses are not exposed, but some bands must be over 10 feet thick. They are crudely foliated and composed of abundant blue-green or green hornblende; plagioclase (andesine to labradorite), sometimes poorly twinned but more commonly untwinned and fine-grained; quartz; some biotite (usually rare); occasionally chlorite; and, in two thin sections, sparse garnet. Accessory minerals are iron ores, sphene, and apatite.

One exposure at I0941 contains a band of dark greenish-gray rock which has abundant chlorite aggregates, common biotite aggregates, much iron ore, common quartz and epidote, and a fine-grained ground-

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mass (0.02 to 0.04 mm.), which probably is mainly untwinned plagioclase and iron ores. It is possibly a hornfelsed amphibolite which has suffered diaphthoresis.

The amphibolite in general is interpreted as representing metamorphosed basaltic tuff and lava.

In the northern part of the area there are closely similar occurrences of a gray-green, massive, well-jointed but not foliated amphibolite. The exposures are all relatively small and the contacts are poorly exposed, so that their actual dimensions and relationship to the country rock are not known, but they appear to represent local cross-cutting intrusions. They are composed of abundant green or blue-green hornblende up to 4 mm. long, altered in part to chlorite; much minute acicular amphibole; badly altered plagioclase, poorly twinned and often full of needle inclusions of amphibole; and fine, granular, diffuse areas presumably of untwinned plagioclase and quartz. The plagioclase is exceedingly difficult to determine, but some of the albite twins in one thin section are labradorite and in another thin section appear to be about the composition of oligoclase-andesine. A small amount of myrmekite is present in one thin section. A little biotite and accessory sulfides, ilmenite, sphene (altering to leucoxene), and apatite also occur. In one specimen a little calcite is present in the more felsic portions and there are calcite amygdules. This nonfoliated amphibolite is either metamorphosed diabase or metamorphosed gabbro.

Amphibole-bearing quartzite: A characteristic rock type which occurs in thin beds about 1 to 3 inches thick, rarely a few feet thick, is a fine-grained, light-colored quartzose rock with prominent porphyroblasts of amphibole. It is interbedded with phyllite and quartzose rocks, particularly in the lower western slopes of Kirby Mountain and to the west in the zone transitional to the Waits River formation. It is also interbedded with impure limestone and amphibolite of the Waits River formation as mentioned above (p. 31).

This amphibole-bearing quartzite is composed of a quartz mosaic (grain diameters 0.01 to 0.15 mm.) with large porphyroblasts, up to several centimeters long, of green or dark green amphibole which may lie in the crude foliation, but frequently do not, and which are often plumose or feathery in habit. The amphibole is usually strongly pleochroic, with Z = dark green, i. e., hornblende; sometimes Z = bright blue-green or occasionally yellow-green, perhaps representing an actinolitic hornblende. Sometimes the amphibole is partly altered to a chloritic material.

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Twinned plagioclase is a rare constituent. Untwinned plagioclase is also believed to be present but the fineness of grain and commonly diffuse nature of the grain boundaries make it extremely difficult to identify, so that its quantity and composition have not been determined. Chlorite sometimes occurs as minute flakes (0.03 to 0.21 mm.) in the foliation and frequently as porphyroblasts (var. penninite) up to 0.63 mm., in places aggregated in streaks. Biotite is present in variable quantity, sometimes sub-parallel with the foliation and poorly oriented within streaks; it is occasionally associated with hornblende and shows alteration to chlorite. Xenoblastic poikiloblasts of garnet often occur, at times notably "flattened" in the foliation. Calcite is a constituent of some bands and is accompanied by "dusty" opaque minerals marking the foliation. Sulfides are always present, and sphene and apatite may be accessory.

These rocks may be interpreted as altered tuffaceous sediments or dolomitic quartzose sediments. Their frequent association with amphibolite points to a contamination with volcanic ash in some occurrences. In one outcrop similar-appearing material occurs as an incomplete shell around a pod-shaped, calc-silicate core, indicating a reaction between aluminous material and the dolomitic content of the original impure carbonate core, which may have been a concretion.

Dark hornblende-rich bands: Dark hornblende-rich bands are sparsely distributed throughout the Gile Mountain formation. Characteristically they are about 1 inch thick, but may vary from $\frac{3}{4}$ to 2 inches thick. They contain abundant well-developed poikilitic hornblende (up to 6 mm. long) sometimes length-parallel to the boundary of the bands, but frequently plumose and bent (Fig. 20); X = yellow to yellow-green, Y = yellow-green to green, Z = green to blue-green. It is often partly altered to chlorite. Pink idioblastic garnets up to 2.5 mm. are a common constituent. These are poikilitic, with inclusions of many opaque minerals and of some quartz, but rarely of amphibole and of chlorite (altered amphibole), and are sometimes outlined with opaques. Chlorite, var. pennite (in flakes 0.8 mm.) may be abundant. In one thin-section a little clinozoisite occurs as well as a little biotite, associated with the hornblende. The groundmass is made up of very fine, diffuse grains 0.01 to 0.04 mm., and is very "dirty" with opaque minerals and inclusions. In part some twinned plagioclase (0.12 to 0.36 mm.) can be identified. The remainder may be untwinned plagioclase and quartz. Opaque minerals are common and include pyrite and much fine dust, sometimes in aggregated streaks. Sphene, in part altered to leucoxene, is common in one

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Figure 20. Photomicrograph (×18) of a dark hornblende-rich band from the Gile Mountain formation showing plumose, porphyroblastic hornblende and idioblastic garnet in a very fine groundmass of plagioclase, quartz, and iron ores. From northeast of Kirby Pond (H0211; see Fig. 15).

case, and sparse apatite (in grains 0.12 to 0.15 mm.) in another.

These thin hornblende-rich bands are interpreted as originally basic tuff or tuffaceous sediment.

MEETINGHOUSE SLATE

Meetinghouse slate was the name given by Doll (1944) to a band of slate in the Strafford quadrangle as a member of the Gile Mountain formation. The name was raised to formation rank by White and Jahns (1950).

No exposures of this formation were found during the field work in the Burke quadrangle. Eric and Dennis (1958, Plate 1) on their geologic map of the Littleton quadrangle show Meetinghouse slate extending north-eastwards to the southern border of the Burke quadrangle at I0200. The terrain to the north of their map is obscured by glacial deposits. The nearest exposures, on Mitchells Knoll and on the northwestern slopes of Miles Mountain, are part of a granite-hornfels complex of high meta-morphic grade (sillimanite zone). Whether the slate formation is present

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could not be determined. Moreover, the low ground of Cold Brook below about 1200 feet and the area to the north are considered to be underlain by granite (p. 52).

In the vicinity of Granby exposures of the Albee formation occur within 100 feet, in some cases, of typical outcrops of the Gile Mountain formation which have a marked pin-striped lithology (I4154). There is no evidence that the Meetinghouse slate is present between the two.

INTRUSIVE ROCKS

Granitic Rocks and Hornfels-Granite Complexes

Granitic rocks are found extensively throughout the northern, northeastern, and south-central parts of the Burke quadrangle. They all cut and intrude the metamorphic rocks, i. e., the Albee, Waits River, and Gile Mountain formations, and are post-tectonic. Although referred to in general as "granite", the rock types include adamellite, granodiorite, and quartz diorite as well as true granite. As their modes of occurrence differ, each general area will be described separately.

Kirby-Burke-Umpire-East Haven Mountains and Hobart Ridge: Kirby, Burke and Umpire mountains owe their elevation and ruggedness directly to the resistance of their rocks, a complex of granite and hornfels. Because of the nature of this complex no boundaries between granite and country rock can be drawn. The continuous forest cover and presence of glacial drift on all of the lower slopes also mitigate against any attempt to map in detail even the larger areas of mainly hornfels or mainly granite within the complex. The complex is bounded on the north, west, and south by the Gile Mountain formation; the eastern boundary is more complicated and will be considered below. The approach to the complex is marked by the increasing metamorphic grade of the Gile Mountain formation—the appearance of staurolite, then andalusite, almost invariably retrogressed to muscovite "shimmer aggregates", and sillimanite, also muscovitized, followed by the appearance of granitic stringers and dikes.

Traverses within the complex reveal a very mixed suite of rocks. Sometimes the granite exposures are relatively large, e. g., several hundred yards across, as in the vicinity of I0952, or the hornfels may extend over broad areas, the hornfels always being of high metamorphic grade and usually with some granitic stringers, although occasionally with none. More often, however, the granite and hornfels occur together in the same exposures frequently in alternating zones or as agmatite

Figure 21. Mixed granite and hornfels (agmatite) on eastern side of the parking area, summit of Burke Mountain (H1149).

(Fig. 21). The contact between the two although intricate is normally sharp (Fig. 22), often following bedding foliation, and there is no difficulty in discriminating between the granite and the metamorphic rock. On Burke Mountain there are excellent exposures on the road to the summit, on the summit, in the area around West Peak, and along the fire road, which extends along the western flank at an elevation of 1900 to 2300 feet. The structures are generally well preserved in these sillimanite-grade rocks; the bedding, schistosity, and cleavage are still evident and antedate the high-grade metamorphism and intrusion. In places there are, however, zones of dark, fine-grained, equigranular, biotite-rich rock

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Figure 22.   Contact between Gile Mountain hornfels and granite. Note the original sedimentary banding with the schistosity parallel to it still preserved, although the hornfels is in a high grade of metamorphism (sillimanite zone). Western slope of Kirby Mountain (G4028).

with no internal structure and a somewhat granitic appearance, but still distinct from the normal granite. The contact between it and the granite is usually very intricate and not sharp as is the case with the normal hornfels. Apparently it is a granitized sediment and represents a mixed rock; its total volume appears to be very subordinate.

Another type of mixed rock which is more prevalent is migmatite (Fig. 23), an intimate mixture of granitic stringers and patches with intervening hornfels and the previously described mixed rock. Sometimes the hornfels has been thoroughly shattered and broken up and cut by granitic stringers. A more unusual occurrence is for the hornfels, including some granitized rocks, to be brecciated and contorted wildly and to contain few granite stringers—all in all, giving the appearance of having been rendered to a semi-plastic condition and stirred up. The

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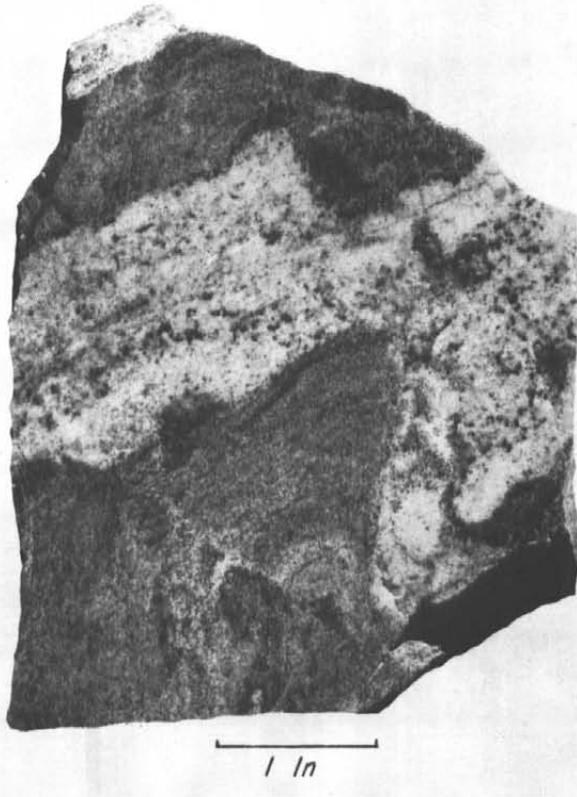
Figure 23.   Photograph of a hand specimen of migmatite from outcrop on road to the summit of Burke Mountain (H1156).

granitic hornfels breccia is presumably what S. R. Hall (1861) and Richardson (1906, p. 84) describe as occurring in Granby, Victory, and Concord.

Xenoliths are abundant in the granite proper and are commonly ellipsoidal, biotite-rich pods from a few millimeters up to several inches in length. (The larger masses of hornfels alternating with the granite and together comprising the complex are not considered xenoliths in the strict sense of the term). Other xenoliths are angular masses of calc-silicate hornfels up to several feet long and 1 to 6 inches wide, representing sections torn off original bands of limestone. The biotite-rich pods may be sharply delineated or fade into the granite, but the calc-

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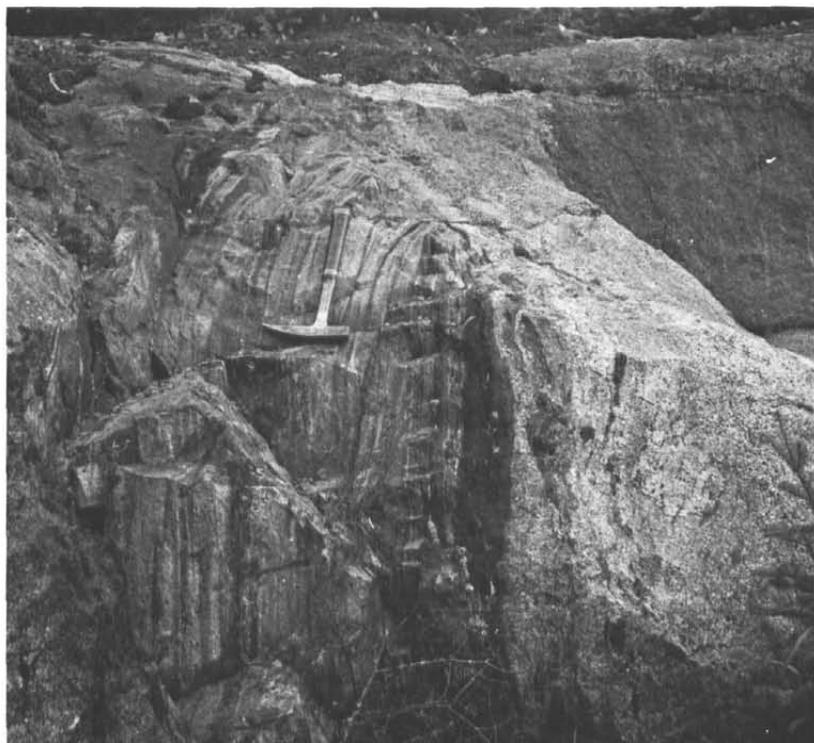
Figure 24. Contact between hornfels and granite showing the regular banding of calc-silicate hornfels. Note biotite-rich pods in granite. Eastern side of the parking area, summit of Burke Mountain (H1149).

silicate xenoliths invariably have sharp contacts and show no macroscopic evidence of assimilation (Fig. 24).

East of Umpire Mountain granite crops out over an extensive area of Hobart Ridge and the southern slopes of East Haven Mountain. On the latter the contact between the granite and the Gile Mountain formation is irregular and can be mapped only approximately. There is little of the granite-hornfels complex as developed on Burke and Umpire mountains. South of Umpire Mountain there is apparently a large continuous granite mass, but farther south the hornfels and granite recur in the Stanley Brook area and represent here the southern boundary of the granite.

The granite in the quarries on the south slope of Kirby Mountain, just south of the Burke quadrangle boundary, have been described by

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Richardson (1906) and by Dale (1909, 1923). It is possible that the Grout quarry, described by Dale (1909, p. 29), is the one found during the field work approximately at H2302. Dale describes the rock in this quarry as a gray biotite granite, whereas in the quarries just south of the quadrangle and to the west it is quartz monzonite. Eric and Dennis (1958) also describe the rock from the northern boundary of the adjacent Littleton quadrangle as Kirby quartz monzonite and state that ". . . in a few places where microcline is abundant (more than 67% of the feldspar) it is a true granite" (p. 28).

A detailed investigation of the petrography of the granite of the area extending from Kirby to East Haven Mountains has not been carried out, but the rock is mainly a fine- to rarely medium-grained, sometimes porphyritic, hypidiomorphic granular, biotite or biotite-muscovite quartz monzonite. It contains oligoclase, kaolinized and sericitized, occasionally zoned, sometimes overgrown with orthoclase and with some myrmekite borders against microcline; turbid orthoclase, kaolinized and sericitized; slightly clouded microcline, occasionally cryptoperthitic (?) and enclosing a little plagioclase; quartz; reddish-brown biotite, sometimes bleached and containing rutile needles and ore specks; sometimes muscovite; and accessory zircon and apatite.

In places a medium-grained melanocratic phase is present, e.g., at H1050 at the top of the ski lift on Burke Mountain. This rock is composed of oligoclase or oligoclase-andesine, often zoned, cloudy with kaolin, and with inclusions of apatite, microcline, and amphibole; microcline; very subordinate orthoclase; reddish-brown biotite; a little green biotite; hornblende with patchy pleochroism (X = yellow green, Y and Z = olive green), and in some cases altering slightly to biotite; quartz; and accessory tourmaline, sphene, apatite, and zircon. A little carbonate is present as a secondary product. The relationship of these granodioritic phases to the normal leucocratic granite and quartz monzonite is not known; they may represent contaminated zones, but are considered more likely to represent separate intrusive material.

The East Haven Mountain medium-grained biotite granite contains microcline, orthoclase, antiperthite, oligoclase, biotite (chloritized in part), quartz, myrmekite, and accessory apatite.

Southeast Victory: The low-lying Victory area is covered with moundy sand and gravel and numerous boulders of granite, as well as extensive swamps. The higher ground northwest and southwest of Damon Crossing and on Mitchells Knoll has exposures of high-grade banded hornfels and granite, with granite conspicuous in some places on the lower slopes.

These occurrences are similar to the granite-hornfels complex of Burke Mountain. The outcrops northwest and southwest of Damon Crossing may be the complex zone forming part of the roof of the pluton. The hornfels is identical to that on Burke Mountain and is believed to be of the Gile Mountain formation.

The northern slopes of Miles Mountain are heavily covered with glacial deposits and definite outcrops are located only on the higher slopes. These exposures are of sillimanite-bearing hornfels and of granitic rocks similar in occurrence to the Burke Mountain complex. Eric and Dennis (1958) describe about half of the outcrops on Miles Mountain in the Littleton quadrangle as granitoid dikes and sills in silimanite-grade rocks of the Albee formation. The granitic rocks range from light gray granite to dark gray tonalite. A thin section from one specimen of quartz diorite collected by the writer is composed of acid andesine, zoned and antiperthitic, a little cloudy and sericitized; biotite, partly chloritized; quartz; rare orthoclase; rare muscovite; sphene; zircon; and apatite.

The low hill at I3205 is covered with large blocks of banded hornfels-granite, none of which are *in situ*, but are considered locally derived. Not a single outcrop within the Burke quadrangle was found on Temple Mountain; the entire slopes are littered with granite boulders. An outcrop on the northernmost peak of the mountain about one-quarter mile east of the Burke quadrangle boundary (i. e., in the Guildhall quadrangle) is a fine-grained quartzose phyllite assigned to the Albee formation (W. I. Johansson, personal communication, 1957). It is not possible to determine metamorphic grade from such a small occurrence, but its appearance is of a considerably lower grade rock than the sillimanite hornfels of Miles Mountain.

Lees Hill has outcrops of sillimanite hornfels and biotite granitic rock which are very similar to those on Miles Mountain, and may represent a small area of injected rock of the Albee formation.

In the vicinity of Little Roundtop and southwest of Gallup Mills there is an area of hornfels-granite complex where no contact between the Gile Mountain formation and the granite could be mapped, with the exception of one small knob of granite northwest of Little Roundtop.

Many of the boulders on the lower slopes of Miles Mountain and along the crude oil pipeline are of a melanocratic rock type. Some along the pipeline east of the Concord road and in the Moose River probably came from bedrock cut into during the laying of the pipeline. One thin section from a representative specimen is composed of microcline (? some cryptoperthitic); oligoclase, some zoned, some cloudy with kaolin and

antiperthitic; scarce orthoclase; myrmekite; dark brownish-green biotite with inclusions of sphene, apatite, and zircon; hornblende (X = yellow-green, Y = olive green, Z = dark green); and accessory epidote, sphene, and apatite. The relationship of this monzonitic rock to the leucocratic granite and quartz monzonite is not known.

A similar rock type is present in the low-lying ground north and northeast of Seneca Mountain, e. g., at C4059. Where the Grand Trunk Railroad crosses the Nulhegan River in the Averill quadrangle this same rock type is seen to be cut by biotite granite. Doll (1951, p. 44) states that dark, biotite-hornblende granite is cut by later, light-colored, coarser granite in the Memphremagog quadrangle. It is possible that much of the Victory basin-like area may be underlain by this melano-cratic phase of the intrusion and may account for the erosion of the area to a low basin in contrast to the surrounding mountains of hornfels, leucocratic biotite granite, and quartz monzonite. It may be a situation analogous to that of the basin-like area north of Seneca Mountain and extending into the Island Pond and Averill quadrangles.

Newark: Considerable outcrop and, indeed, cliffs of granite are present on Packer Mountain, Hawk Rock, Walker Mountain, Deer Hill, and in La Pawac and Jack brooks. The western and northern granite boundary is in contact with the Waits River formation, and the eastern and southern one with the Gile Mountain formation. In Mill Brook (B1703) there is in contact with the Waits River a microgranite, which may be a fine-grained offshoot from the border of the main mass. Unfortunately the boundaries of the main mass are not well exposed, but they are probably relatively well defined. A mixed granite-hornfels zone is absent except along the northern contact, where there is a confused alternation of granite, hornfels, schist, phyllitic limestone, and quartzose rocks. In the East Passumpsic River north of B2045 and in the two nearby streams flowing into the Passumpsic, one from the east and the other from the west, thermal metamorphism of the impure limestone is scarcely notice-able, even in outcrops close to the granite. Calc-silicate hornfels is present, however, at a number of localities along the northern contact zone. Schist and pelite show characteristics of the higher grades of meta-morphism and include the sillimanite zone, particularly along the northern contact and in a narrow zone against the granite's eastern boundary, e. g., on the mountain east of Lost Nation.

The Newark granite has been described by Dale (1909, 1923) from one quarry east of Center Pond at approximately B0329. This was not def-initely located during the field work, but there are numerous large ledges

in the vicinity, and one of them may have been the site of a small quarry. Dale (1923, pp. 112–113) describes the rock as a light pinkish biotite-granite with very sparse porphyritic feldspar up to 1.5 by 0.5 inch in a groundmass of coarse-grained (up to 0.8 inch) orthoclase and microcline (with some perthitic intergrowths), medium-grained quartz, kaolinised albite to oligoclase-albite, biotite, and accessory magnetite, pyrite, sphene, and allanite.

There is a pink granite at a number of places west of the East Passumpsic River; the best locality is an old quarry behind Shepard's farm (B2040), where many large quarried blocks remain. Not all of the rock is pink in color, for it grades into a normal light-colored biotite granite. The distribution of the pink-colored rock is very difficult to determine in the field because the weathered granite outcrops tend to have a pinkish color even when the fresh rock is white or creamy-colored. The pink color is possibly caused by finely divided hematite in the microcline. Some outcrops on the eastern shore of Center Pond show surfaces weathered to a deep pink or red. The pink rock at Shepard's quarry is composed predominantly of slightly cloudy microcline microperthite phenocrysts up to 13 mm. long, with plagioclase inclusions which are kaolinized and sericitized. The microcline is also often cryptoperthitic with irregular vein-like intergrowths and with disturbed indistinct twinning. Plagioclase (up to 3.0 mm.) is common; it is much kaolinized and sericitized, with some zoning. There is little orthoclase. Myrmekite occurs in grains up to 0.65 mm. and as borders to the oligoclase, which in turn borders the microcline. The sparse biotite is green; some is partly chloritized. Quartz (up to 4.75 mm.) is common and has slightly undulose extinction. Accessory minerals are apatite, sphene, iron ores (associated particularly with biotite), epidote, and very rare allanite and zircon.

Thin sections from other localities show a very similar composition: abundant microcline microperthite and cryptoperthite, and cloudy oligoclase, occasionally antiperthitic. The biotite is either green or a dirty brownish-green and partly chloritized. A small amount of tiny garnet is present in a specimen from Deer Hill and rare fluorite in specimens from near the top of Packer Mountain (B0524) and from a granite ledge on the east side of Route 114 near Benchmark 1162.

The Newark granite is strikingly different from the granitic rocks of Kirby, Burke, and East Haven mountains in the presence of abundant perthitic microcline and green, partially chloritized biotite.

East Mountain: Granitic rocks are exposed extensively on East Moun-

tain and on the spur which extends southward. Their contacts with the Gile Mountain formation are difficult to determine because of the extensive glacial deposits which mantle all of the lower slopes. However, on the south and west the contact is relatively sharp, with little or no "mixed zone." For example, in an excavation just off the new summit road at F2947 granite overlies andalusite schist (Fig. 19). On the lower eastern slopes of the southern spur hornfelsed Gile Mountain strata occur. The northeastern spur has many large outcrops of hornfels as well as of granite and of mixed rock, and comprises a hornfels-granite complex similar to that of Burke Mountain. However, the easternmost knob of the spur is entirely granite, as is the adjacent area of the Guildhall quadrangle. Hornfels also occurs on the top of East Mountain and on the lower slopes of the northwestern spur, with granite on the upper slopes. The northern boundary of the granite is not known, for glacial drift obscures the lower slopes into Madison Brook; the inaccessibility of the northern slope of East Mountain precluded a more detailed search there for outcrops in the time available. The granite may be continuous across Madison valley with the granite of the rugged southern spur of Seneca Mountain (at C3014), but no outcrops were found in the valley in this vicinity. It is possible that much of the northern flank of East Mountain is a Gile Mountain hornfels-granite complex. The relationship between the granite and hornfels of the northeastern spur to the granite, gneiss, and hybrid-like rocks in Madison Brook around C4112 is also not known.

The main igneous mass of East Mountain appears to be a medium- to fine-grained, often porphyritic, biotite or biotite-muscovite granitic rock similar to the normal Burke and Kirby quartz monzonite or granite. Aplite dikes cut it sporadically. A coarse leucocratic phase is present and is well exposed on the new road to the summit at F3047. It is composed of abundant, slightly cloudy microcline (3.2 mm.), some of which is cryptoperthitic and microperthitic with cloudy plagioclase inclusions (up to 0.4 mm.); abundant quartz; sericitized and kaolinized oligoclase, some antiperthitic and with myrmekite borders; a little orthoclase; sparse green biotite, in part bleached and altered to chlorite; rare muscovite with symplectic borders against microcline; and accessory sphene and garnet. Some of the granite has a slightly pinkish color.

Melanocratic diorite is exposed on the road at approximately F3356. Its constituents are oligoclase or calcic oligoclase with a little zoning, highly altered to kaolin, sericite and carbonate; hornblende (X = yellow-green, Y = dark green, Z = green), much altered to biotite and chlorite

and with sphene inclusions common; reddish-brown biotite (apparently mainly after amphibole) often bent and strained and with inclusions of sphene and, in some, of reticulated rutile; some cloudy orthoclase (it is difficult to discriminate its quantity as the feldspars are so altered, but it appears to be very subordinate in amount); a little myrmekite; sparse quartz; and accessory sphene, altering to leucoxene, iron ores, and apatite. Hornblende may exceed feldspar; the color index of the rock then rises to about 50. The diorite is associated with dark hybrid rock, hornfels and granitized hornfels, contaminated granite, and some pegmatitic masses composed of apatite, biotite, tourmaline, amphibole, feldspar, and quartz.

Seneca Mountain: Granitic rock comprises the whole of the southeastern spur of Seneca Mountain, and it also occurs in a hornfels-granite complex on the extreme northeastern spur. The granite of the southeastern spur is in contact with hornfelsed Gile Mountain formation on the northern flanks and also in the North Branch of Paul Stream just east of the Burke quadrangle boundary. To the south, as mentioned above, it is not known whether the granite is continuous under the broad Madison valley to East Mountain. The rock is fine to medium in grain and similar in hand-specimen to the granite and granodiorite of the Burke-Kirby area. A thin section from C4129 is composed of abundant microcline (up to 4 mm.) with some inclusions of cloudy plagioclase and perthitic intergrowths; kaolinized and sericitized oligoclase, some with myrmekite borders against microcline; a little orthoclase; abundant quartz with undulose extinction; reddish-brown biotite; and a little secondary muscovite. In contrast, a specimen from C4217 is composed of abundant plagioclase (oligoclase to acid andesine), finely twinned and often zoned, slightly altered to kaolin and sericite, and occasionally antiperthitic; abundant reddish-brown biotite, slightly altered to chlorite along cleavages and some bleached; quartz with fine needle inclusions; sparse symplectic muscovite; sericite aggregates in streaks up to 7.3 mm. long; and accessory apatite, iron ores, and zircon. This rock is a quartz diorite of the trondhjemite variety.

A series of exposures occurs in Madison Brook near the eastern boundary of the Burke quadrangle. The rocks are quite varied and, as they are restricted in outcrop to the brook, their relationships to the granite of Seneca Mountain on the north and to that of East Mountain on the south are not known. They include biotite "granite," gneiss, meladiorite, basic intrusives, and very dark basic material with considerable pyrite. Downstream in the Guildhall quadrangle biotite gra-

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nitic rocks crop out together with masses of Gile Mountain hornfels, although the ridges both to the north and south are apparently entirely of "granite."

The melanocratic dioritic rocks are composed of euhedral to sub-hedral hornblende (X = colorless, Y = yellow, Z = light yellow green or X = yellow green, Y = green, Z = green), partly altered to biotite and chlorite, and with inclusions of sphene and apatite; oligoclase to acid andesine, slightly kaolinized and sericitized, rarely zoned, with occa-sional overgrowths of orthoclase and rare myrmekite rims; a little ortho-clase and quartz; and accessory sphene, apatite, epidote, and tourmaline (as inclusions in the plagioclase). Reddish-brown biotite, with inclusions of apatite and sphene, may be abundant. Lighter colored quartz diorite contains hornblende (X = yellow green, Y = green, Z = bluish-green), altering to chlorite and containing rutile needles; andesine; quartz; and accessory sphene, apatite, and epidote. One hand-specimen is cut by veins (up to 2.5 mm. thick) of a gray-colored glass which appears under the microscope as light brown isotropic material containing rounded quartz crystals, many opaque granules, and, in some zones, incipient spots of crystallization and minute crystallites. These veins presumably represent chilled offshoots from an adjacent basic dike. The host rock shows quartz with undulose extinction, cracks, and dusty lines of inclusions; much cloudy and sericitized plagioclase (oligoclase-andesine) with some carbonate and with the twins ruptured, bent, and poorly defined; some cloudy orthoclase; brownish-green bent biotite; colorless amphibole; a little chlorite; common sphene altering to leuco-xene; apatite; and pyrite. Near the veins the alteration increases and aggregates of biotite and sphene occur at a high angle to the contact. One dark green basic zone contains abundant masses of pyrite (up to 9.5 mm.); abundant reddish-brown biotite, bleached and chloritized; much chlorite and light green amphibole; rare plagioclase; apatite; quartz; and common colorless mica, particularly adjacent to and penetrating the pyrite masses.

The gneiss is composed of a regular alternation of dark schistose bands 8 to 17 mm. thick and white granulose bands 3 to 7 mm. thick dipping to the southwest. The dark bands are composed of euhedral and subhedral green hornblende, biotite (mainly green but a little dark brown), plagio-clase, quartz, and accessory sphene, apatite, epidote, and sulfides. The light-colored bands have minor biotite and hornblende, abundant acid andesine (2.6 mm.), which has perthitic intergrowths of microcline (0.15 mm.) and cryptoperthitic patchy intergrowths, occasional oscil-

latory zoning and patchy kaolinization, rare microcline (up to 0.36 mm.), and rare myrmekite. This occurrence of quartz diorite-gneiss is unique in the Burke quadrangle; its origin is obscure, but it may represent meta-somatised banded Gile Mountain strata.

McSherry Mountain-Newark Street: The northwestern section of the Burke quadrangle, around Bald Hill Pond and Newark Pond and extending southward to East Sutton Ridge and eastward to Walker Pond and Newark Street, has many exposures of granitic rock, particularly in the high ground such as McSherry Mountain. Unfortunately large areas are also covered with glacial drift, and the contacts of granite with the Waits River formation are, in general, impossible to map. However, a characteristic feature is that many of the granite exposures overlie Waits River strata and the latter dip at low angles beneath the granite. Examples of this relationship occur on the lower eastern slope of McSherry Mountain (A0751), north of Newark Pond, (A0842), and northeast of Bald Hill Pond (A1558). In addition, granitic exposures alternate with metamorphosed calcareous rocks, for example, along the road north of Bean School and on the slopes northwest of Newark Street. Elsewhere sills and dikes frequently cut the Waits River strata, as in the East Sutton Ridge area. The granitic rocks are considered to have essentially a sheeted form following the structure of the Waits River formation in the northwestern area; some of the sheets are thin offshoots while others must be several hundred feet thick, such as that on McSherry Mountain. The intrusions extend from the granitic areas of Bald Mountain to the north (Island Pond quadrangle), Haystack Mountain (Lyndonville quadrangle), and the area south of Lake Willoughby (Lyndonville quadrangle) mapped by Dennis (1956, Plate 1) as containing an abundance of granitic dikes. There is, in general, little evidence of advanced thermal metamorphism in the exposed rocks of the Waits River formation except locally where there are bands of calc-silicate hornfels.

The granitic rocks are very variable in texture; the larger masses are normal, fine- to medium-grained biotite granite or granodiorite. They pass frequently into a very coarse pegmatitic phase, which may often comprise the whole of many exposures. Fine-grained, very leucocratic white phases also occur as well as dikes and sills up to one or two feet thick.

A leucocratic specimen from A2316 is composed of slightly clouded microcline (up to 2.1 mm.), microperthitic and cryptoperthitic, with muscovite inclusions; kaolinized and sericitized oligoclase, some anti-

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perthitic; quartz (up to 1.2 mm.), a little orthoclase; sparse muscovite; rare myrmekite; and iron ores. A sheeted rock overlying the Waits River formation at A1622 has abundant quartz; oligoclase (up to 1.8 mm. occasionally), normally zoned, with some antiperthitic; orthoclase; a little microcline; sparse brown biotite; and sphene. Quartz also occurs in rounded grains dotted sparsely throughout the feldspars; it appears to be of secondary metasomatic origin.

At A2455 an exposure of "granite" has a platy flow structure trending 154° and dipping 15° northeast.

Contact Relations and Mode of Emplacement

All of the granitic rocks appear to be related and perhaps co-magmatic. The melanocratic amphibole-bearing granodiorite to monzonite varieties were intruded earlier than the leucocratic quartz monzonite and granite. They are all post-tectonic intrusions and probably are all members of the New Hampshire magma series (Billings, 1934), one of the seven series recognized in New Hampshire (Billings, 1956). The New Hampshire magma series is believed to be Middle Devonian in age (Lyons, et al., 1957), although according to the new time scale (Kulp, 1961) the radioactive age falls within the Mississippian or Pennsylvanian time span.

The contact relations of the granitic rocks are not known in detail because of the poor exposures. Over large areas there is not a sharp contact, but rather a broad zone of mixed granite and hornfels. In general the granite cross-cuts the regional structure, but in local exposures the contacts are seen to both parallel and cut across the bedding schistosity. In the northern part of the Burke quadrangle the metamorphosed sediments dip beneath the granite; this suggests that at least in this area the intrusion is in the form of sheets and tongues more or less parallel to the bedding.

There is little, if any, evidence of large-scale deflections by the granite of the strike of the bedding of the metamorphic rocks. The east-west strike in Ferdinand, in the northeastern part of the quadrangle, may be related to the intrusion of the granite to the north; the bedding of the Gile Mountain strata dips steeply northwards into the granite. East-west strikes are also prevalent in the Gile Mountain near the granite contacts on East Haven and Hobart Ridge, and there is a slight suggestion of a bending of the strike in the area northwest of Burke Mountain. The hornfels-granite complex zones provide evidence of much disruption of the intruded rocks and rotation of the resultant blocks. On the

northern flank of Burke Mountain east-west strikes are common in the hornfels of the complex. These relations imply much fracturing and much penetration of the metamorphosed sediments by the granite and an intrusion mechanism by stoping. The granite is not, however, surrounded everywhere by complex zones.

One point of interest is that the hornfels on Burke Mountain includes a noticeable number of thin calc-silicate bands. Limestone bands are rarely observed in the Gile Mountain formation; perhaps the metamorphism of the calcareous bands to calc-silicate rocks makes them more apparent than in their lower metamorphic grade outside of the complex. However, it could be that these Burke Mountain hornfelsed rocks belong to the transition zone between the Gile Mountain and the Waits River formations and have been brought to their present elevation by the granite or by a previous fold structure which has been disrupted by the granite.

The rocks of the complex also show some evidence of assimilation and granitization; these are believed to be on a small scale. The mechanism of intrusion appears to have been by a combination of forcible intrusion, utilizing essentially the structural weaknesses of the intruded rocks, mainly the bedding schistosity, by intimate penetration of the rocks and stoping, and to a smaller extent by assimilation and granitization.

Granitic Dikes

Granitic dikes cut the country rock usually near the main "granite" masses or near the granite-hornfels complexes. They are prevalent in the area north and south of East Sutton Ridge. The majority of the dikes are thin, usually less than one foot thick, but very occasionally up to 25 feet thick. Some are simple pegmatites, while others are fine-grained, sometimes porphyritic, representatives of the larger granitic intrusions ranging in type from microgranite to microdiorite.

One occurrence at G3627 contains rare brown tourmaline, and the phyllite in contact with it has very abundant brown tourmaline. Sparse green hornblende together with brown biotite is present in a dike of microgranodiorite at A0313, and abundant hornblende altering to biotite and chlorite occurs in a porphyrite dike six feet thick at I2147. On East Mountain (F3356) there is a more unusual dike. This is a microsyenite: greenish, mottled with pink; one foot thick, with slickensided walls and joints. It is composed of abundant blue-green hornblende; common green biotite, slightly altered to chlorite; microcline (crypto-

perthitic); a little orthoclase; some kaolinized and sericitized feldspar (probably plagioclase); epidote; scarce interstitial quartz; and accessory sphene, apatite, and iron ores.

Mafic Dikes

Thirty-seven post-metamorphic mafic dikes have been encountered, mainly in the northern half of the quadrangle; they are described in detail in Woodland, 1962. The exposures are generally poor and it was often impossible to determine either the thickness or the attitude of the intrusions. Richardson (1906, p. 111) mentions briefly the occurrence of diabase dikes in Newark.

The dikes have a gray to greenish-gray or dark gray color; they range in thickness from less than one inch to several feet and appear to dip very steeply. They are aphanitic to fine- to medium-grained rocks, occasionally with a narrow light-colored chilled border, often with amygdules of calcite and chlorite (rarely analcite), and generally with sparse to abundant phenocrysts. Microscopically they comprise a varied suite which possesses one common characteristic—they have all suffered considerable deuteric alteration. Carbonates and chlorite are ubiquitous. The rock texture is intersertal with the latter minerals, together with other groundmass minerals, occupying the interstices between the feldspars.

The fine grain and altered nature of the rocks usually make it very difficult or impossible to identify all of the constituents, particularly in the mesostasis. Plagioclase often is one of the major constituents, but in a rare instance it is present in only small quantities. It may be in small laths 0.04 to 0.06 mm. long or in larger laths up to 1.25 mm. long, felted, rarely pilotaxitic, but commonly in ocelli which are crudely variolitic. The plagioclase is much altered to sericite, kaolin, and calcite; its determination normally is not possible but in some cases the feldspar appears to be andesine and perhaps occasionally labradorite. Phenocrysts of plagioclase sometimes occur up to 4.5 mm. long, also much altered, and containing inclusions of apatite. Alkali feldspar is present in some dikes in discrete crystals as well as mantling the plagioclase, and may be present in the groundmass in other dikes. Phenocrysts of olivine (up to 1.5 mm.) occur in about half of the thin sections examined; sometimes they are very abundant and almost invariably completely altered to serpentine-carbonate or talc-carbonate-serpentine aggregates. Pyroxene (augite) is of frequent occurrence both as phenocrysts (up to 1 mm. or even 3.5 mm.) and in the groundmass; it is also often altered to carbonate-

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serpentine aggregates. Brown amphibole occurs in about half of the sections, sometimes as phenocrysts (up to 3.7 mm. long) but more often in the groundmass; rarely it may be after pyroxene. It is a brown hornblende or barkevikite. Biotite is present in over half the sections but usually sparsely. In some it is common (olivine is then absent and pyroxene may or may not be present) and is pleochroic to a dark red-brown. It is chloritized in two occurrences. Analcite occurs in a few slides usually interstitially but rarely as larger patches (up to 1.5 mm.) with many minute inclusions; it is also in amygdules with calcite. Quartz, in small quantities, is very rare. The interstitial material is abundant, sometimes preponderantly so, and often difficult to resolve, but includes chlorite (often in spherulites), carbonate, antigorite, and chlorophaeite. Apatite is sometimes present and is occasionally very abundant as needles throughout the thin section. Sphene occurs sparsely. Opaques are abundant, and in some cases are so prevalent as to make the thin section very dense; included are magnetite, ilmenite (altering to leucoxene), leucoxene, and sulfides. The latter are sometimes very noticeable even in hand specimens. In two thin sections the opaques occur in a mesh-work of fine rods; in one the mesh is very irregular throughout the abundant groundmass; in the other the mesh forms a very regular network, with angles of approximately 87° and 93° in small patches, super-imposed on the groundmass throughout the slide. Granular opaque minerals also occur abundantly.

The metamorphic effects on the intruded country rock are not notice-able in the field. In thin section a limestone and a schist contact have been examined. In one case (A3742) the immediate limestone contact (about 0.35 mm. thick) is composed of fine-grained quartz (0.04 to 0.36 mm.), phlogopite, calcite, and many opaque grains. The individual grains are larger away from the contact and contain masses and streams of opaques; microcline, apatite, chlorite, and tremolite also occur. In the other case (B1948) the schist against the dike is composed of quartz grains with fine needles and felted masses of sillimanite; abundant biotite, much of which is altered considerably (in part to sillimanite); abundant sericitic aggregates; and opaques. The fibrolitic sillimanite, however, was probably present before the intrusion of the dike.

The mafic dikes are much altered but appear to range in type from amygdaloidal diabase to olivine diabase, analcite diabase, and lampro-phyric types. The latter predominate and include intrusions of camptoni-tic affinities, although none possess the typical panidiomorphic granular texture. They may be related to the White Mountain plutonic-volcanic

series of New Hampshire or to the intrusive necks of the Monteregian Hills of Quebec (see Woodland, 1962, p. 1107–1109).

Metamorphosed Mafic Dikes: A few metamorphosed dikes cut the Gile Mountain formation in a number of localities, e. g., at F2802 and H0513. These dikes contain plagioclase (often untwinned), hornblende, biotite, a little quartz, and iron ores; chlorite is common in some and rare epidote occurs in one specimen. They were intruded before the metamorphism of the area.

STRUCTURE

Minor Structural Features

The structural features described herein include all planar and linear structures and mesoscopic (Weiss and McIntyre, 1957, p. 577) and microscopic folds. Mesoscopic or minor fold structures are those that occur within the limits of an average hand specimen or small outcrop in the field, and range in size from about one-quarter inch up to several yards across; generally they are between one-half inch and three inches across. In contrast, major fold structures cannot be viewed directly in the field in the Burke area, and their presence and form can only be determined by indirect means; they may range in size up to regional structures miles across.

The minor structural features of the Burke quadrangle have been made the subject of intensive study so that they could be grouped by type and relative size, their geometry described, regional preferred orientations, if any, defined, and any areal change of preferred orientations discovered. It was hoped this would lead to and support the description of major structures and the interpretation of their origins.

To this end, many planar, linear, and fold features were measured in the field and a total of 1029 geographically oriented specimens were collected. The latter have all been studied in the laboratory with a binocular microscope to determine the types of planar and linear features present, and whenever necessary the specimens have been re-oriented in a simple apparatus which enabled the field orientation to be reproduced and the orientation of additional structural surfaces and linear elements to be measured, as described under Methods of Study. In this way many more data have been obtained to supplement the field readings. One important result of the laboratory examination has been a more detailed appreciation of the geometry, origin, and mutual relationships of the structural features. The accumulation of such orientation

data in many cases would have been difficult or impossible to obtain in the field, particularly in the relatively poor light in the forests and in the stream courses completely overshadowed by trees.

Most of the measurements have been plotted in the conventional way on Plate 2 (planar features map), Plate 3 (linear features map), and Plate 4 (way-up of bedding and shear sense map), but in some areas there were too many readings to plot on a map of the scale of one inch equals one mile. All of the measurements have been plotted on the lower hemisphere of the Schmidt stereonet (Lambert equal-area projection) or on the lower hemisphere of the Wulff stereonet. (See Plate 5 and Fig. 56).

PLANAR FEATURES

These include all parallel planes, regardless of origin, which can be discerned in the rock because of a compositional or mechanical inhomogeneity, for example, repeated planes along which a rock tends to part. Planar features of the rock fabric have been designated "s"-surfaces by Sander (1930). In the Burke area "s"-surfaces include bedding, schistosity, false cleavage, and joints.

PRIMARY PLANAR FEATURE

Bedding (S_0): Compositional banding which represents original depositional layers is easily recognizable throughout the Burke quadrangle. It is particularly well developed in the Gile Mountain formation, where the alternating layers vary from a fraction of an inch to several feet in thickness. Characteristically it provides a striped or banded appearance of alternating light gray quartzose phyllite and dark gray slate, phyllite, or schist. A well-developed schistosity is nearly always present, and in most exposures this is sensibly parallel to the bedding layers and is termed a bedding schistosity (S_1). In the quartzose rocks the schistosity is marked not only by the alignment of sericite but often by a fine striping a millimeter or two apart caused by small biotite flakes also aligned parallel to the main compositional banding and appearing to represent original silt or clay partings. Although the biotite laminae may well be derived from such partings, they are essentially of mechanical origin. Comparatively rarely the bedding is seen to be tightly folded, with the schistosity and fine striping paralleling the axial plane and the limbs, but cutting across the bedding layers at the nose of the folds, proving the mechanical nature of the schistosity and of the fine biotite laminae. These folds are designated as earlier or B_1 folds; by far the more abundant folds in the area deform the schis-

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tosity, and these are called later or B_2 folds (see the section on Minor Folds). It can not be assumed, therefore, that schistosity is always parallel to original bedding in outcrops that show no distinctive bedding layers, although more often than not it is likely to be parallel or sub-parallel. The great majority of bedding orientations recorded in the field were actually measured on foliation planes sensibly parallel to the primary compositional banding and are referred to in this paper as bedding schistosity or S_1. These, together with the relatively small number of bedding (S_0) readings where the latter are not parallel to S_1, are plotted on the planar features map, Plate 2, which thus shows the present attitude of the bedding in the Burke quadrangle.

The bedding contacts maintain remarkable constancy of attitude over large areas and give the impression of a simple structure with a steep east-southeast dip, becoming gentle and even horizontal in the western part of the quadrangle and dipping steeply to the east-northeast in the northern part. The monotonous nature of the strata within the Waits River and Gile Mountain formations obscures any repetition of beds caused by isoclinal folding or strike faulting. However, in the south-western portion of the quadrangle there is evidence of considerable repetition. This is indicated by the successive occurrence across the strike of outcrops of Waits River lithology alternating with outcrops of the Gile Mountain type. Unfortunately the evidence is provided, not by repetition of a sharp, clearly defined boundary between the two forma-tions, but by repetition in the transitional zone between the two. Struc-tural complexity is very evident in this area; minor folds abound, and often outcrops show the rocks to have been strongly deformed and contorted, sometimes to a point that defies geometric reconstruction. Limestone bands are infolded sharply with phyllite and die out repeatedly along the strike as do bands of phyllite interbedded in more quartzose rocks; the intricate nature of the limestone folds, particularly, and their disharmonic relation to their envelopes imply a large measure of rock flowage. Shears are also evident at some localities, particularly at G1010 and G1015. It is likely that other tectonic breaks also occur, perhaps representing sheared-out limbs of folds. This complexity of structure in the southwest, the gradational lithology, and the scattered nature of the outcrops (due to glacial and post-glacial cover) render the mapping highly conjectural in places; the pattern represented on the geologic map (Plate 1) is thus a generalized interpretation of the field data.

Elsewhere in the Burke quadrangle the evidence for repetition is not available. Minor folds occur throughout the area, but in general outcrops

do not show the structural complexity of those in the southwest, with the exception of the outcrops in the granite-hornfels complexes. North of East Burke the evidence for any complexity of the boundary between the Waits River and Gile Mountain formations is obscured by thick, widespread drift deposits.

Evidence bearing on the original order of superposition of the strata is exceedingly scanty and is rendered difficult to interpret because of the strong deformation and recrystallization of the rocks. Repeated isoclinal folding is believed to affect the beds, with both reversed and normal limbs present. However, it is reasonable to suggest that, if the formations as a whole are in their normal order, that is, if they dip and face to the east, indications of a normal order should be more prevalent than those of a reversed order in a random occurrence of appropriate outcrops. A complicating factor is the presence of later folds which may result in local reversals of the original attitudes of limbs of the earlier folds. However, the later folds appear to be mainly of a minor character and are not accompanied by large-scale structures so that inversion effects are restricted to a very local scale, which can be recognized; the way-up data in such cases are rejected.

The features that were examined in the field include cross-lamination, graded bedding, attitude of drag folds, and the relationship of bedding (S_0) to schistosity (S_1), and are shown diagrammatically on Plate 4. Structures similar to cross-lamination are rare throughout the area. Secondary cleavages often simulate this primary structure and usually it is not possible to be sure whether it is cross-lamination or cleavage which is present. Of the ten cases of accepted cross-lamination in the area, seven indicate that the bedding is right-way-up and faces eastwards, and three indicate that the bedding is upside-down and faces westwards. Graded bedding is present in the banded pelitic and psammitic beds of the Gile Mountain formation. Rarely a psammitic bed grades upwards through a semi-pelitic zone into a succeeding pelitic band within the space of two or three inches. Normally the alternating psammitic and pelitic bands have sharply differentiated boundaries on both sides so that it is impossible to determine top and bottom, although the banding probably represents a type of rhythmically graded bedding. Only in five instances was the way-up determined; in three the beds dip and face eastwards and in the other two the beds dip eastwards and face westwards.

The attitude of minor drag folds has long been used in determining the correct order of deposition. White and Jahns (1950, p. 197) suggest

the use of the terms sinistral and dextral to describe the form of folds with one limb shorter than the other as exposed on a horizontal (or vertical) surface. L. M. Hall (1959, pp. 61–63) shows that, if for a horizontal exposure the plunges of the folds are also known, the shear sense that produced the folds can be deduced and thus the position of the minor folds in the major structure and the way-up of the beds can be determined. However, the minor folds so used must be congruous folds related to the major structure. In the Burke quadrangle only the earlier folds (B_1) can be used safely, as the later folds (B_2) are super-imposed on a folded structure (see section on Minor Folds below). A further complication is that early folds may be rotated bodily around the B_2 axis or the plunge may be inverted by rotation around an axis at a high angle to the B_1 axis. However, a total of thirteen early minor folds not deformed or rotated by later folds (so far as could be ascertained) have been analyzed; nine indicate that the strata face and dip eastwards and four that they dip eastwards and are upside-down (see Plate 4).

The relationship of the attitude of bedding to that of axial plane cleavage, slaty cleavage, and fracture cleavage in folds is well-known (e. g., Billings, 1942; Wilson, 1947). In the Burke area the mutual relations of bedding (S_0) and schistosity (S_1) have been noted wherever an angular relation could be discerned and where later folds do not cause complications—a total of thirty-one observations. In seventeen cases the beds are right-way-up (they dip eastwards in twelve instances and westwards in five), and in fourteen they are upside-down (they dip eastwards in thirteen instances and westwards in one). (See Plate 4).

Thus, out of a total of fifty-nine determinations the strata are right-way-up in thirty-six localities (thirty-one with an easterly dip) and upside-down in twenty-three (twenty-two with an easterly dip).

Secondary Planar Features

Schistosity (S_1): The dominant planar feature of the metamorphic rocks of the area is a well-developed schistosity or foliation, that is, a structure caused by the parallel orientation of platy or prismatic minerals throughout the rock and producing an infinite number of parallel planes along which the rock tends to part. This agrees with the definition of what Leith (1905, p. 105) describes as flow cleavage, which is developed parallel to the axial plane of folds and which is parallel to the plane containing the greatest and intermediate axes (AB) in the strain ellipsoid. Fairbairn (1935, p. 592) refers to this

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structure as "axial plane" foliation and Mead (1940, p. 1010) calls it flow cleavage developed perpendicular to the least axis of strain. As noted above, in the great majority of outcrops the schistosity is sensibly parallel to the primary bedding, and is referred to as a bedding foliation or schistosity. The schistosity is seen to be parallel to the axial plane of the rare earlier minor folds. It is accordingly not to be interpreted in the sense of bedding foliation as defined by Mead (1940, p. 1009).

Schistosity is plotted on Plate 2; poles to schistosity are plotted on the lower hemisphere of a Schmidt stereonet in Figure 56A and on Plate 5 (Figs. 1 to 11).

The schistosity is defined mainly by the subparallel orientation of mica (sericite, muscovite, and biotite) and is best developed in the pelitic and semi-pelitic rocks. Quartz and calcite often appear "flattened" and show a dimensional orientation with their longer direction parallel to the schistosity. The quartzose rocks usually have sericite in subparallel orientation throughout and frequently it is also somewhat concentrated in narrow zones; the quartzose phyllite thus usually has a well-developed foliation. It is not possible to determine the origin of the schistosity (S_1) surface, but the fact that it plainly cuts the bedding layers at the noses of minor isoclinal folds and is axial plane to these folds indicates that it was formed during the deformation producing the tight isoclinal folds (thus explaining why it is nearly always subparallel to bedding) and is accordingly an axial plane schistosity. The rocks have been elongated parallel to the schistosity and normal to the B_1 or earlier fold axis direction, as is indicated by the usually considerable "thickening" of the isoclinal minor fold hinges and by rare boudinage of original bedding parallel to B_1 caused by movement on the schistosity. It is probable that movement parallel to the schistosity occurred throughout the area and resulted in the obliteration of original bedding and destruction of fold hinges; it may explain the scantiness of the observations of the latter. The development of the schistosity essentially parallel to the bedding may have been partly caused by translatory movements being initiated and continued on the original bedding surfaces. The latter must have been prominent, particularly in the banded shales and siltstones of the Gile Mountain formation.

False Cleavage (S_2): Later cleavages transect the schistosity very commonly in the Burke quadrangle. Collectively these are called false cleavage and include slip cleavage and fracture cleavage. They are plotted on Plate 2; poles to cleavage are plotted on the lower hemisphere of a Schmidt stereonet in Figure 56B and on Plate 5 (Figs. 12 to 19).

Figure 25. Drawing of a hand specimen of Gile Mountain banded phyllite, showing bedding (S_0), striking 40°, dipping 63° to the southeast; schistosity (S_1), striking 39°, dipping 72° to the southeast; and cleavage (S_2), striking 41°, dipping 50° to the southeast. From the southwestern slope of Sugar Hill (G3306).

Cleavage occurs throughout the area, but is not everywhere equally developed as structural surfaces, and even in an area such as the southwestern section (subareas I and II—see Plate 4), where it is, in general, prominent, it is not present in some exposures while in others it is the dominant structure. Cleavage is best developed in the pelitic and semipelitic rocks where it sometimes nearly obliterates the schistosity (Fig. 25). An outcrop or even a hand specimen frequently shows two or more later cleavages cutting the schistosity. Often these may be caused by differences in lithology so that the attitude of the cleavage surfaces in quartzose bands varies from that in pelitic layers. This refraction of the cleavage is clearly seen in Figure 26. In other cases the relationship

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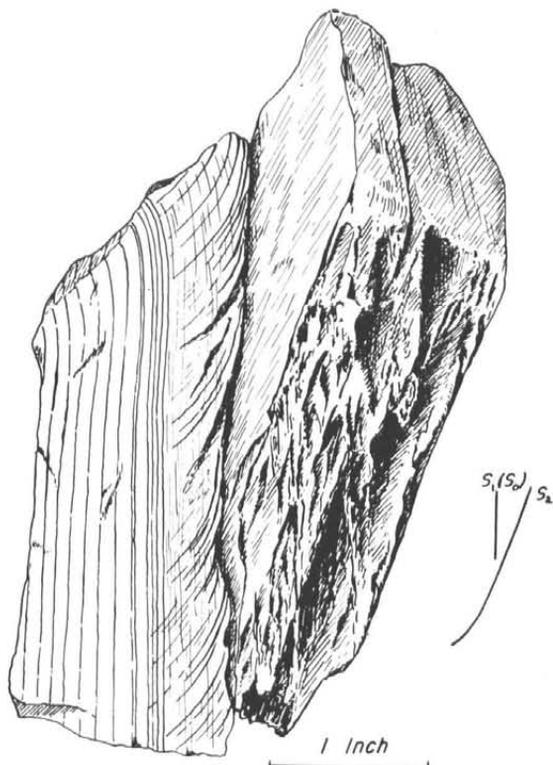
Figure 26.   Drawing of quartzose and pelitic phyllite showing "refraction" of cleavage (S₂) in the quartzose band. From western foot of Sugar Hill (G3107).

between two or more cleavages is not clear, although one is usually dominant and much better developed than the others. A few cleavages (S$_3$) apparently later than S$_2$, are plotted on Plate 2. Where later B$_2$ folds deform the schistosity (S$_1$) surfaces the dominant S$_2$ surfaces are parallel to the axial planes (Figs. 27 and 37). Displacement on and parallel to the cleavages is often evident; it is more apparent in the quartzose bands and also in the finely banded quartzose and pelitic or semi-pelitic beds (Fig. 30, a and b). The amount of movement rarely exceeds two or three millimeters. Quartz veins, often parallel to the schistosity, are disrupted by the cleavage and, when displaced (Fig. 28), give rise to cleavage mullions (Wilson, 1953).

The characteristics of the cleavage and its development vary with the lithology. The pelitic and quartzose beds offer contrasting situations and

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Figure 27. Typically banded Gile Mountain strata deformed by later (B_2) folding. Schistosity (S_1) is parallel to the bedding (S_0); slip cleavage (S_2), parallel to penknife, is axial plane to the B_2 fold. Note drag folds in the quartzose band and folded quartz veins (bottom center and on right). On east side of road up the Moose valley (F1232; see Fig. 42).

are described separately below; intermediate types are formed in semi-pelitic rocks.

Cleavage in pelitic beds: The stages in the development of cleavage can be established by study of hand specimens from different outcrops. Examination of rough fractured surfaces approximately normal to the schistosity and lineation, as well as examination of polished surfaces cut with this orientation, frequently shows the deformation of the schistosity and the formation of the later cleavage. Examination of thin sections, also prepared normal to the schistosity and the lineation, provides detailed evidence of mineral orientations and of the nature of the structural surfaces.

The earliest stages of deformation show the schistosity (S_1) with regular zones of "crinkling," which produce a rippling of the surfaces. The "crinkle" zones, S_2, are spaced about one to two millimeters apart

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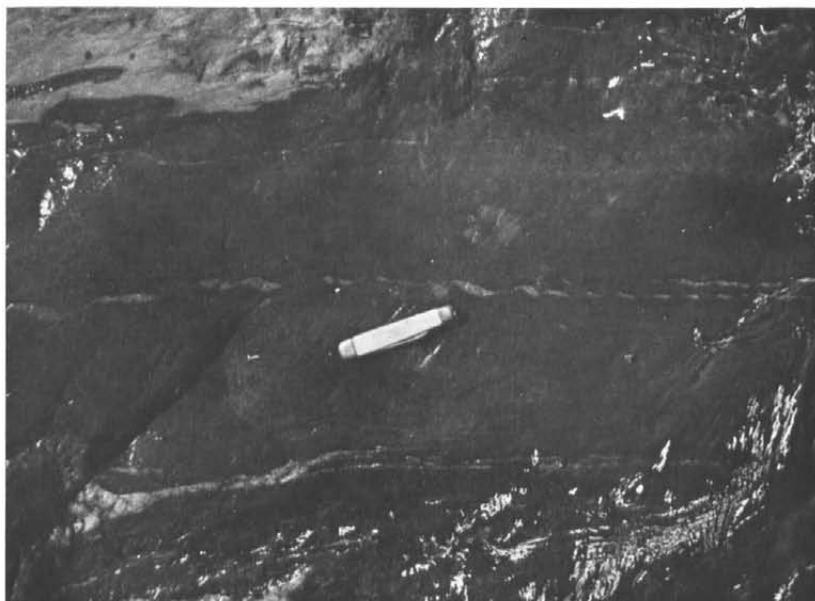
Figure 28. Quartz vein disrupted into cleavage mullions by movement on slip cleavage (S₂), which parallels penknife. In Gile Mountain formation, Sheldon Brook (G2302).

(Fig. 29a), and the mica is oriented oblique to the trend of the zones. At this stage S_2 is not a well-developed structural surface and does not give rise to a cleavage. Further deformation results in the development of microfolds in S_1 (Fig. 29, b and c), the short limbs of which line up and form the S_2 surfaces. As the micas of the short limbs lie within and parallel or sub-parallel to the trends of the limbs, the S_2 is a well-developed cleavage surface. The sense of movement on both S_1 and S_2 surfaces is evident at this stage and is in the same direction on both (Fig. 29b), although no slip is discernible on S_2. Apparently further deformation and slip on S_1 result in rotation of S_2 as the short limbs of the microfolds become more overfolded. A stage is then reached when the rotation of S_2 surfaces brings them into a favorable position for actual slip to take place, and displacement on S_2 surfaces occurs. Shears, however, may arise before the overturned limbs have produced "through-going" surfaces; the shears (S_2) transect the limbs at a small angle and movement along them drags the mica into parallelism with S_2 (Fig. 30, a, b, and c, and Fig. 31). Meanwhile, the S_2 surfaces have become closer together

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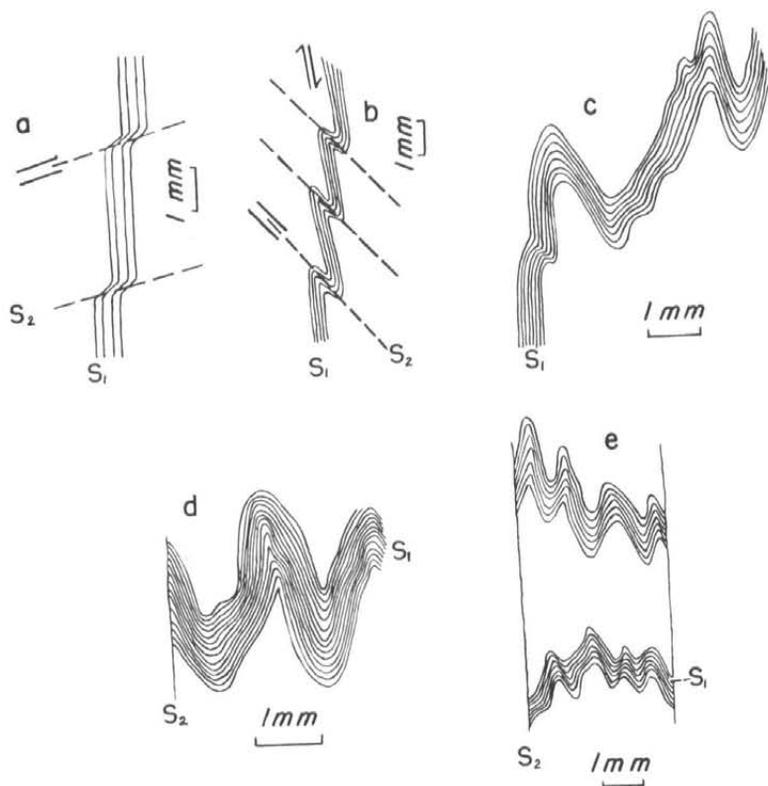
Figure 29. Series of drawings from one specimen showing the deformation of schistosity (S₁) and the development of cleavage (S₂) in the Gile Mountain formation. From the western slope of Shonya Hill (G1020).

    a. Crinkling of S_1 and incipient development of S_2.
    b. Microfolding of S_1 and development of S_2 on aligned short limbs.
    c. Isoclinally folded S_1.
    d. Tight isoclinal folds, sheared in places by S_2.
    e. Minutely flexured S_1 between parallel S_2 planes.

due to the microfolding of the S_1 surfaces (Fig. 29, d and e). Initiation of displacement on the S_2 surfaces limits further slip on S_1. As a result of the stresses between adjacent S_2 surfaces, the S_1 surfaces are thrown into further microfolds, the limbs of which are sub-parallel to S_2 (Fig. 29e and Fig. 32). Continued movement results in more and more mica becoming oriented parallel to the S_2 surfaces so that S_1 is almost obliterated. Its presence at this stage is shown by relict noses of microfolds

Figure 30a. Photograph of polished surface of finely banded Gile Mountain phyllite; dark spots are biotite porphyroblasts. From the summit of Shonya Hill (G1120).
a. Microfolds in S_1, which is parallel to S_0, and development of S_2 cleavage. Note thinning of long limb (on the right).

lying between the S_2 surfaces, which are now extremely well developed throughout and form a true schistosity (Fig. 33). The sense of movement on S_2 is now not possible to determine. This apparent development of S_2 surfaces is very similar to that described by Sorby (1880), White (1949), Brace (1953), De Sitter (1956), Weiss and McIntyre (1957), and L. M. Hall (1959).

Heim (1878; quoted in Harker, 1886) describes cleavage related to minute faults resulting from microfolding as "Ausweichungsclivage." Sorby (1858, 1880) calls it "close-joints cleavage," which, he says, could

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Figure 30b. Enlargement of a portion of (a) to show development of S_2 along the short limbs of the microfolds.

develop into a true slaty cleavage. Bonney (1886) calls a similar structure "strain-slip cleavage," and Harker (1886) uses the term "false cleavage." Dale (1892, 1894) uses the term "slip cleavage," which he equates with "ausweichungsclivage" and strain-slip cleavage. Leith (1905, p. 120), under the general context of "fracture cleavage," describes "false cleavage" as ". . . usually the result of closely spaced parallel overthrust folds grading into minute faults. . ." and equates it with "ausweichungscliv-age," slip cleavage, and strain-slip cleavage. Mead (1940, p. 1010) describes shear cleavage as consisting of ". . . roughly spaced surfaces of shear displacements on which platy minerals have developed and into which they may have been dragged." He contrasts it with fracture cleavage

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Figure 30c. Enlargement of a portion to the left of (a) to show development of S_2 in more homogeneous pelitic rock.

by suggesting that the former is particularly ". . . a phenomenon of rock flowage rather than of fracture" (p. 1011). He further states, however, that ". . . there is little importance in distinguishing between these two varieties of cleavage (that is, shear cleavage and fracture cleavage) because they have the same relationship to causal stresses and accomplish the same deformational results" (p. 1017). White (1949, p. 593) describes the deformation of rock during the formation of slip cleavage as comprising two components, a flattening of thin plates bounded by slip planes and a distortion by simple shear along the slip planes. De Sitter's ideas (1956, p. 97) are essentially similar.

White (1949, pp. 592–593) prefers not to use Leith's term "fracture cleavage" because:

". . . it is questionable that slip cleavage is entirely independent of

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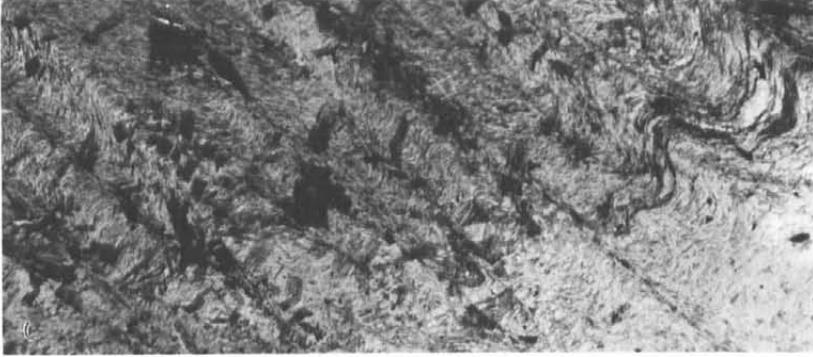
Figure 31. Photomicrograph (×13) of Gile Mountain banded phyllite showing microfolded schistosity (S_1) well developed in the pelitic layer and less well developed in the quartzose layer (bottom right-hand corner). S_2 cleavage is developed and transects the limbs of the microfolds in S_1. Biotite porphyroblasts (some altered to chlorite) are common. In Barnes Brook (G2817).

mineral orientation. . . . Fracturing on the theoretical shear planes of the stressed rock as a whole does not seem to be the primary cause of the localization and orientation of the slip cleavage . . . (it) places undue emphasis on one aspect of this planar element, whereas actually there is clear evidence of appreciable flowage in the process leading to its formation."

He thus calls the structure "slip cleavage" after Dale, and this is the term widely used for a similar structure in all of the recent reports on Vermont geology.

A cross-cutting structure which may give rise to S_2 surfaces has also been observed to occur in some cases when the schistosity is deformed into minute chevron folds, the alignment of whose sharp angular hinges produces a parallel structure (S_2) crossing the general trend of the schistosity. There are no platy minerals oriented parallel to these structures, which are called "Knickungsebene" by Brögger and referred to by Harker (1886).

S_2 surfaces are also developed in the Burke area in a kinematically different way to that outlined above. In Figure 34 the S_2 surfaces are prominent but relatively widely spaced. The S_1 schistosity is folded (note the sharp fold on the left) and is displaced along the S_2 shear surfaces. The sense of slip is opposite to that on S_1—which is the reverse of the position previously described (see Fig. 30b). Figure 35 also shows the same relationship, although this occurrence is complicated by the pres-

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Figure 32. Photomicrograph (×18) of banded phyllite, with minor calcite, from the transition zone between the Waits River and Gile Mountain formations, showing folded sericite schistosity (S_1) cut by S_2 cleavage, biotite porphyroblasts (some lying parallel to S_2), and opaques in pelitic band (lower area). Note that S_2 surfaces are also developed in the quartzose band (upper area). From south of East Haven village (E1651).

ence of a prominent S_2 surface, the sense of slip on which is opposite to that on the other S_2 surfaces and appears to be related to a minor fold visible in the quartzose band. This fold structure and the related slip surface are thus analogous to those described above and shown in Figure 30b. The opposite slip sense on S_1 and S_2 is well shown, too, in Figure 36; the development of a new flexure with a new incipient S_2 surface, the sense of slip on which is opposite to that on the other S_2 surfaces, can also be seen.

Hoeppener (1956) outlines the development of S_2 structures whose orientation and movement sense are similar to those of Figure 34. He states that S_2 surfaces first form at angles of 115° to 145° to S_1 surfaces, and the slip sense on S_2 is opposite to that on S_1, in each case towards the obtuse angle between them. Rotation of S_2 is accompanied by a flattening and flexure of the S_1 surfaces into microfolds between the S_2 planes, and new S_2 surfaces form, the movement direction on which is opposite to

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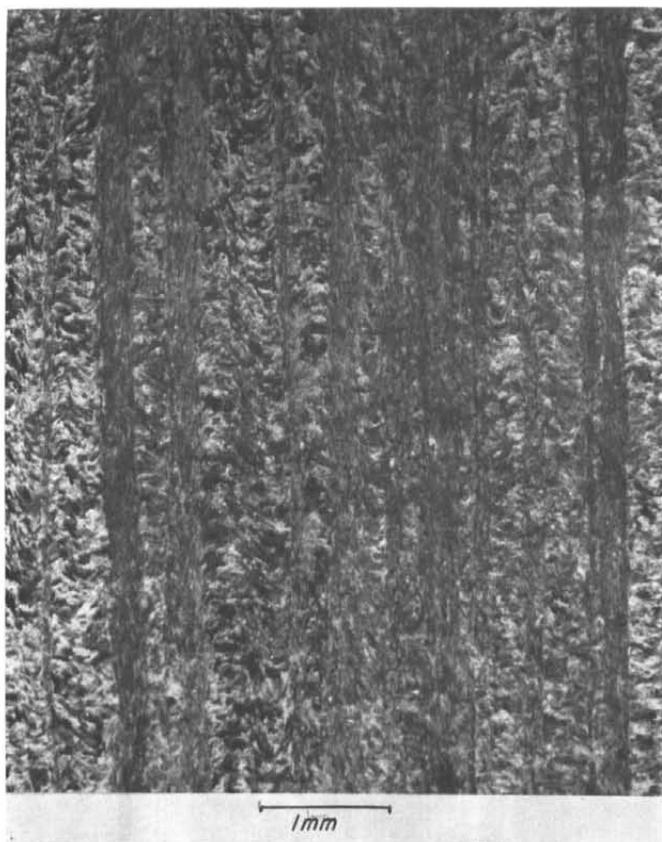
Figure 33. Photograph of polished surface, etched with hydrofluoric acid, showing well-developed S_2 cleavage and relict schistosity (S_1), the latter appearing as isolated fold noses. From the southern slope of Bemis Hill (G1440).

that on the old S_2 surfaces; then it becomes no longer possible to determine the movement sense on the S_2 surfaces. The S_2 surfaces are finally rotated into planes perpendicular to the least strain direction. The development of S_2 surfaces in the pelitic portion of Figure 35 and also in Figure 36 appears to be in agreement with Hoeppener's description.

It is not clear why these two kinematically different S_2 surfaces should arise together, but it is probable that the lithology of the rocks and the initial attitude of the S_1 surfaces in the stress field causing the new structures are both responsible. The lithology of the specimen in Figure

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Figure 34. Drawing of a hand specimen of Waits River banded phyllite showing folded bedding schistosity (S_1) cut by S_2 cleavage. Note small isoclinal fold in S_1 on the left. From Mountain Brook (G2446).

30a is a finely banded quartzose and pelitic phyllite, and it is possible that the micro-overfolds and attendant shears on the overturned limbs are controlled by flexural slip folding of the thin competent bands.

Brace (1953) describes different patterns in the development of slip cleavage:

". . . one if west limbs (of microfolds) are thinned and brought into sub-parallel arrangement, another if the east limbs are so affected.
. . . Both limbs may become aligned, forming "Totfalten" (Ampferer, 1938)."

His first two cases are identical with the two varieties described above,

Figure 35. Photograph of polished surface, etched with hydrofluoric acid, of Gile Mountain phyllite showing microfolded bedding schistosity (S_1) cut by a well-developed S_2 cleavage. Note that the cleavage (S_2) associated with the fold in the quartzose band in the right-center has a reverse sense of slip to that on the other S_2 surfaces. From the hill north of Sheldon Brook(G2204).

that is, S_2 surfaces related to micro-overfolds, with movement sense on S_1 and S_2 the same (Fig. 30) and S_2 surfaces accompanied by micro-folding, with the movement sense on S_1 and S_2 opposed and directed into the obtuse angle between them (Fig. 34). Alignment of both limbs of microfolds will arise in both cases when flattening between the S_2 surfaces causes flexuring of the S_1 surfaces and continued slip on S_2 shears out the limbs, leaving the detached noses of S_1 as relicts between the S_2 planes (Fig. 33) analogous to the "Totfalten" of Ampferer. As the S_2 surfaces

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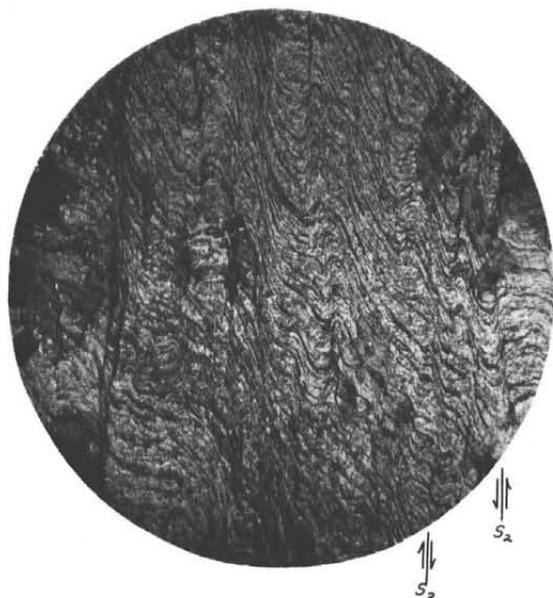
Figure 36. Photomicrograph ($\times$17) of microfolded schistosity (S_1) in phyllite, with the development of slip cleavage (S_2). Note the sharp new flexures, the limbs of which parallel incipient S_2 cleavage and the slip movement on one of which is in an opposite direction to that on the other S_2 surfaces. From the northern slope of Spur Ridge (G3008).

are frequently subparallel to the axial planes of minor folds in the schistosity, their ultimate attitude would appear to be in the AB plane of the strain ellipsoid and normal to the direction of greatest stress. When this stage is attained, no further slip on S_2 is possible. Rotation of S_2 planes into an axial plane attitude is clearly demonstrated in Figure 37. Recrystallization of mica parallel to the S_2 surfaces has often accentuated them. However, although post-deformational growth of biotite porphyroblasts is to be seen many times as preferential along S_2 planes, their (001) cleavages are frequently oriented at a high angle to S_2, thereby producing "edgewise" biotite as seen on the cleavage surfaces.

Cleavage in quartzose rocks: S_2 cleavage surfaces are not often present in the quartzose rocks, but some outcrops do have a cleavage cross-cutting the schistosity (S_1). These cross-cutting surfaces are generally relatively widely spaced (one-eighth to one-half inch apart) and usually show as shear surfaces with visible displacement of the bedding schistosity (Fig.

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Figure 37.   Photomicrograph (×18) of a minor fold in Gile Mountain banded phyllite showing microfolded bedding schistosity (S_1) and the development of S_2 cleavage, which has been rotated into an axial plane position in the core of the fold. From northeast of Granby (I4157).

38). Sometimes the planes contain biotite porphyroblast concentrations, which have formed after the deformation and which make the cross-cutting S_2 structures very prominent (Fig. 39) and similar to Dale's (1896) "cleavage banding".

The S_2 cleavage of quartzose rocks has the characteristics of Leith's (1905) "fracture cleavage". It is, however, also axial plane to later folds (Figs. 38, 40, and 41). In Figure 41 it is clearly seen as a shear cleavage with displacement of the bedding schistosity, the fold being essentially a shear fold.

Cleavage in calcareous rocks: S_2 cleavage structures are rarely observed in the impure limestone. Intergranular and intragranular adjustments were apparently continuous during the movements so that the material deformed plastically (Fig. 48) without the development of shear surfaces. The paucity of mica and the relatively poor development of S_1 surfaces did not provide the conditions for inter-layer slip and microfolding as described under the pelitic rocks. Calcareous phyllites deformed much like the latter, however.
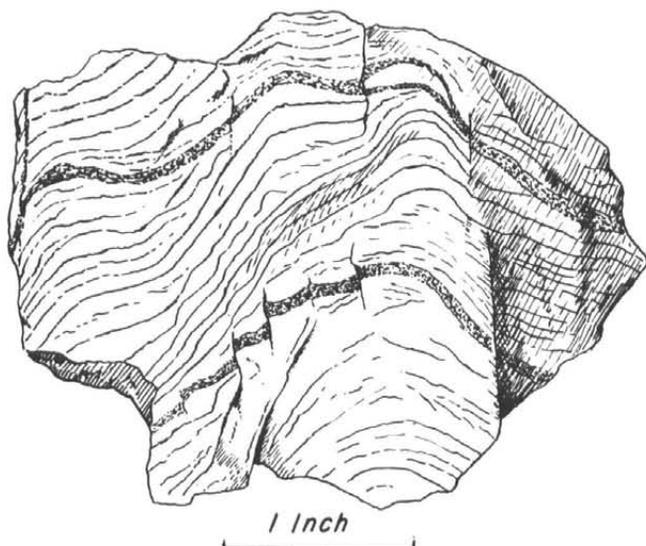
Figure 38. Drawing of a hand specimen of Gile Mountain quartzose phyllite, showing an open fold in the bedding schistosity and quartz veins (dotted) cut by fracture cleavage, movement on which has caused displacement of the quartz veins. From west of Sugar Hill (G2708).

Analysis of attitudes of S_2 cleavage: Sometimes the divergent attitudes of the S_2 cleavage cannot be explained by differences of lithology, as described in the preceding section, but may be due to: (1) the development of microfolds in S_1, the limbs of which are not parallel but are more or less equally developed and transect the general trend of S_1, or (2) rotation of the surfaces during the course of deformation or external rotation of fold limbs which would bodily rotate the slip surfaces relative to the stress field. It is difficult to explain why there are two kinematic variations of the S_2 surfaces, particularly as the actual distribution of each is not known. The writer is inclined to believe that the micro-overfold type is the more common in the Burke quadrangle because it occurs more often in the hand specimens.

It is probable that this micro-overfold type of S_2 cleavage is the result of initial deformation of S_1 by microfolding with concomitant slip on S_1 and flexural slip folding of the more competent rock layers. Evidently the initial attitude of the well-developed S_1 surfaces in relation to the applied stress favored slip on them (simple shear in a homogeneous field). As the microfolds develop, the micas on the over-

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Figure 39.   Drawing of a hand specimen of Gile Mountain quartzose phyllite showing slightly flexed bedding schistosity (S_1) cut by fracture cleavage (S_2). From the western slope of Shonya Hill (G0822).

folded limbs line up and produce the S_2 surfaces, which, in the earliest stages, are oriented so that slip (also oblique or simple shear relative to the greatest stress direction) takes place on them in the same direction as on the S_1 surfaces. In Figure 30a it can be seen that the long limbs of the very asymmetric folds have been thinned and the short overturned limbs have been preserved; slip on the S_1 surfaces of the long limbs and on the S_2 surfaces was in the same direction, but it was apparently opposite in direction on the short limbs. Early slip on S_2 surfaces, however, would lead to shearing-out of the overturned limbs. As the S_2 surfaces are rotated towards the AB plane, slip on S_2 diminishes and further deformation takes place by new microfolding of S_1 and the rotation of more mica into parallelism with the S_2 planes. The sense of slip is opposite on opposite limbs of the microfolds.

The other main type of S_2 surfaces is developed as shears, the attitude and slip movement of which are complementary to those of the S_1 surfaces; the initial mutual relation of the S_1 and S_2 surfaces conforms to that predicted by the strain theory of rock deformation (Swanson, 1927). The S_2 surfaces are gradually rotated into a sub-parallelism with the axial planes of minor folds and thus into the AB plane, and further deformation takes place by flexuring of the S_1 surfaces.

The microfolding of the schistosity in the pelitic bands and the development of the new slip surfaces (S_2) result in an extension of the

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Figure 40. Photograph of polished surface of Gile Mountain quartzose phyllite showing schistosity (S₁) cut by shear cleavage (S₂). Biotite is aligned in S₂, while the biotite in the S₁ laminae is oriented parallel to the S₂ direction. From west of Brookside School (G2923).

rock mass initially parallel to a direction bisecting the acute angle between the S₁ and S₂ surfaces, but, as the S₂ is rotated into parallelism with the AB plane, the extension direction lies parallel to it.

Slip sense on S₁: The relationship between the direction of slip sense on S₁ and on S₂ surfaces can sometimes be determined by examination of hand specimens and particularly by the form of small, even microscopic, folds in the S₁ surfaces. The geometry of sixty-nine of these folds in the Burke quadrangle is shown on Plate 4; an analysis of them indicates that in thirty-seven the slip movement was upwards on the

Figure 41. Photograph of polished surface of Waits River quartzose phyllite showing shear fold in bedding schistosity (S_1) caused by slippage on S_2 cleavage. Biotite is aligned in both S_1 and S_2. From west of Sugar Hill (G3009).

west side of the predominantly easterly-dipping S_1 surfaces and in the remaining thirty-two it was upwards on the east side.

The slip sense on S_1 in any specimen may also be determined from the attitude of S_2, if it is assumed that the angular relation between S_1 and S_2 surfaces conforms in general to the strain theory of deformation (Swanson, 1927), i. e., when the slip sense on S_1 is towards the obtuse angle formed by the intersection of the S_1 and S_2 surfaces. This is the relationship exhibited by all of the Burke hand specimens where slip sense on S_1 could be determined; it is also the relationship widely accepted by geologists and used for a long time to determine the po-

sition of an outcrop in a fold and to determine the way-up of bedding (see, for example, Leith, 1913; Billings, 1942; Wilson, 1947; Shrock, 1948). The direction of apparent slip sense may also be ascertained if it is taken as being approximately normal to the line of intersection of S_2 with S_1. Using these criteria, a number of oriented specimens have been analyzed and the slip sense determined by construction on the lower hemisphere of a Wulff stereonet; the plots of the slip sense are shown in the eight stereograms of Plate 4, together with the geographic area covered by each plot. The stereograms indicate that in the majority of cases the slip sense is downwards on the east side of the S_1 surfaces, but upward slip on the east is common in areas A, C, and E. This is in agreement with the data derived from the folds in S_1, although the diagrams show slip sense downwards on the east more preponderantly than do the folds. Downward slip on the east has mainly a southeastward plunge in areas B, C, and D, a southwestward plunge in area F, a southward plunge in areas E and G, and a southeastward to southwestward plunge in area A. Upward slip directions (on the east side of S_1) from the east and southeast have gentler plunges in area A; likewise, those from the southeast in area C and those from the southeast, east, and northeast in area E have gentler plunges. Slip movement is thus mainly downwards on the east side of the schistosity (S_1) in a general southerly direction; upward slip on the east side may represent movement on the gentler, less common, opposite limbs of isoclinal folds in the schistosity (S_1).

The spread of the directions shown on the plots may indicate differences in the main stress vectors; for example, the diagrams of areas D and F suggest this. However, some spread of points would be expected even in a homogeneous stress field, because the initial attitude of the S_1 surfaces would have varied somewhat around the preferred orientation. Slip directions plotted in the northeastern quadrant of the stereogram occur in areas B and E and to a lesser extent in A; these are to be correlated with southeastward- and southward-plunging lineations (B_2) and with S_2 surfaces differently oriented to those associated with northeastward-plunging lineations (B_2) and southward-plunging slip vectors. The former S_2 surfaces either strike east of and dip more gently than the S_1 surfaces or they strike west of and dip more steeply than S_1, while in the latter case the attitude of the S_2 surfaces relative to S_1 is the reverse.

Relationship of S_1 and S_2: The relationship between the schistosity and cleavage surfaces is shown in Figure 57A for the Burke quadrangle as

a whole and in Figures 57B to K for the subareas. Over the entire quadrangle the S_2 maximum lies west of and dips more gently to the southeast than the S_1 maximum. This is true also for subareas I, II, III, V, and VI. Insufficient data are available for preparing a stereogram of S_2 in subarea VII; however, the measurements available show that S_2 trends in part west of and dips more steeply than S_1, and in part trends east of and has about the same dip as the S_1 surface. In subarea VIII the S_2 cleavage tends to parallel S_1 or else to strike east and dip more gently. S_2 surfaces strike west of and dip more steeply to the east than S_1 in subareas IX and XI, while in subarea X they have a more easterly strike and a gentler dip. The relationships in subareas VII, VIII (in part), IX, X, and XI are correlated with B_2 lineation plunging in an easterly or southerly direction, while in the other subareas the lineation maximum plunges to the northeast.

Joints: A detailed study of all of the joints in the Burke quadrangle has not been carried out, but the attitudes of the joints in three areas have been analyzed. Contoured stereograms of poles to all of the joints measured in the metamorphic rocks of subarea II, in the hornfels of subarea IV, in the granite of subarea IV, and in the Newark granite are shown on Plate 5, Figures 31 to 34, respectively. The metamorphic rocks of subarea II have a joint plane maximum trending 127° and nearly vertical and a sub-maximum trending 58° and dipping 70° to the northwest. The hornfels of subarea IV has two maxima of nearly vertical joint planes trending 78° and 143°; this relationship strongly suggests a conjugate shear joint system. The granite of the same area has a different and rather scattered joint pattern. The majority of joint planes either strike 53° and are vertical or strike 32° and dip 16° to the northwest; the latter represents sheeting. The origin of the joints in the granite thus appears to be quite different from that of the joints in the hornfels. The steep cliffs developed on the west-southwestern flank of Burke Mountain and associated with an extensive rock fall appear to be controlled by jointing in the hornfels trending 143°. The vertical cliffs on the northeastern slope of the northern spur of Umpire Mountain also may be due to northwest-southeast-trending joints in the hornfels and phyllite. The data are insufficient to explain as due to jointing the precipitous eastern flank of Umpire Mountain which trends north-south.

The Newark granite has a maximum joint plane attitude trending 48° and dipping 75° to the northwest, a second maximum trending 166° with a near-vertical dip, a third maximum trending 156° and dipping 78° northeast, a sub-maximum trending 106° and dipping 80° to the

southwest, and a second sub-maximum of horizontal joints; the latter represents sheeting. The steep cliffs on Packer Mountain and Hawk Rock are probably due, in part, to the joints striking 166°, while the south-facing cliffs on Walker Mountain may have been developed as a result of the joints striking 106°.

Acid dike trends: The trends of 87 acid dikes, which lie mainly in the northern half of the Burke quadrangle, have been measured. Of these 56% lie between N.18°W. and N.10°E.

Poor exposures almost invariably prevented measurement of their dips.

Faults: Only a few faults have been observed in the Burke quadrangle (e. g., at A0503), and all have small displacements of the bedding (not exceeding three inches). It is possible that the phyllonite that occurs in the southwest (G1010 and G1015), south of Shonya Hill, represents a moderate dislocation, but the extent of displacement, if any, cannot be estimated. It is also possible that large-scale strike-faulting may occur, yet be undetected, because of the absence of marker horizons in the Waits River and Gile Mountain formations.

MINOR FOLDS

Folds are common throughout most of the quadrangle, but are apparently particularly numerous in the southwestern part where the contact between the Waits River and Gile Mountain formations is complex, as a result of folding on a relatively large scale. The Albee formation has so few exposures in the quadrangle that little opportunity is offered for observing fold structures; no fold axes have been measured in this formation. All of the folds which have been observed in the field are small-scale, normally less than three to four inches across, but some a few feet across do occur (Figs. 42 and 43). Two generations of folds can be distinguished, and they are described below separately as earlier and later folds. The time interval between the deformations causing the two series of folds is not known; the series probably belongs to the same general orogeny and represents successive phases.

Earlier folds (B_1): Very few folds of the earlier deformation have been identified. The considerable minor folding, particularly in the southwest, nearly all belongs to the later deformation. No early folds have been discovered in the Albee formation.

Early folds are recognized by the bending around of the beds of differing composition that are cut by the extremely well-developed schistosity (S_1), which is axial plane to the folds. Slip cleavage (S_2),

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Figure 42. A large, relatively open, later (B₂) fold in banded Gile Mountain strata. The schistosity (S₁) parallels the bedding (S₀); the slip cleavage (S₂) is axial plane to the fold and is parallel to the penknife. The fold axis plunges 45° northeastwards and parallel to the hammer handle. On the east side of the road up the Moose valley (F1232; see Fig. 27).

if present, cuts the schistosity often at a small angle, and is not axial plane to the earlier fold. The folds are isoclinal and often very tight, with the result that the axial plane schistosity is practically parallel to the limbs. The schistosity is so well developed that it is the dominant structural surface of the rock; the bedding is not present as structural surfaces. The bedding around the nose of the fold may be flayed out along the schistosity because of slippage along the latter (Fig. 44). Some of the folds are so tight as to be represented by very narrow inter-digitations of bands, as seen in plan view (Fig. 45). The bedding bands may be rendered inconspicuous by the pervading schistosity and show up merely as shadowy lines (Fig. 46). In some cases translation along the schistosity has left noses of folds as relicts suspended between the new mechanically-produced surfaces (Fig. 47). It is probable that translation on the schistosity has obliterated much of the evidence of early folds, particularly in the layers which are more homogeneous in composition, where they would be obscure at best.

Figure 43. A steeply-plunging, later (B_2) fold in banded Gile Mountain strata. The schistosity (S_1) parallels the bedding (S_0). The fold axis plunges 70° northeastwards and parallel to the hammer handle. On the east side of the road up the Moose valley (F1234).

Undoubtedly the later deformation has also served to obscure the earlier folds as the later folds have refolded the earlier ones. Examples of refolded folds have been rarely identified in the Burke quadrangle, presumably because the relationships in the outcrops would be complex. Highly deformed rocks crop out fairly often in the southwestern area, and these usually defy geometric analysis. Figure 48 is an example of a refolded fold in banded limestone of the Waits River formation. The earlier fold is a small, tight, isoclinal drag fold which is now folded over the crest of the later fold. There is no visible cleavage parallel to the axial plane of the later fold. Another example of a refolded fold is shown in Figure 49. Here the earlier fold is tightly isoclinal with a quartzose phyllite core and a well-developed axial plane schistosity (S_1), which is parallel with the limbs but cuts the nose and flays out the bedding contact. The later fold is an open fold which has a cleavage, sub-parallel
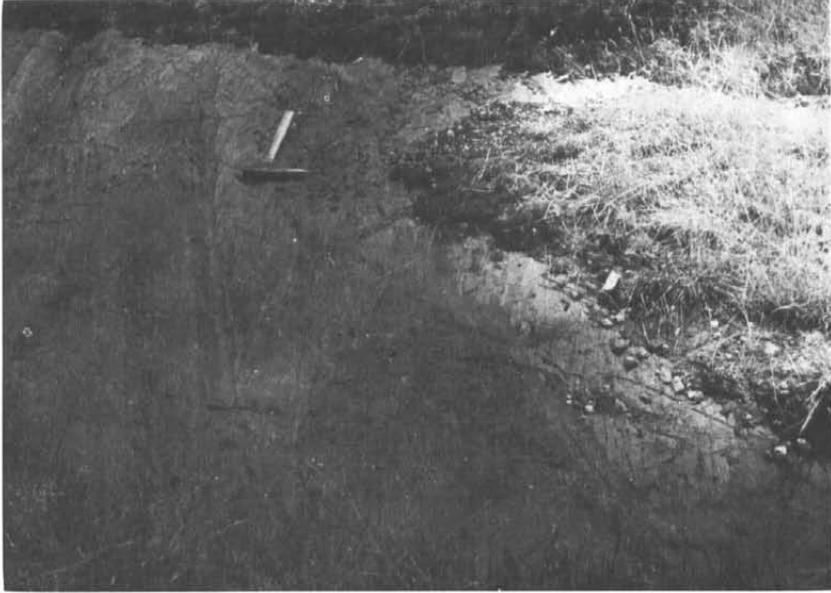
Figure 44. Plan view of plunging earlier (B₁) folds in the Gile Mountain formation. The schistosity (S₁) is parallel to the axial planes of the folds (parallel with hammer handle). Note how the pelite of the cores has flayed out along the schistosity. Northeast of Kirby Pond (H0512).

to its axial plane, developed in the quartzose core and also developed in the pelitic envelope, where it has a different attitude.

Of the thirteen earlier folds which could be analyzed, six have a sinistral pattern in plan view (see White and Jahns, 1950; L. M. Hall, 1959), five plunging northwards and one plunging southwards; seven have a dextral pattern, three plunging northwards and four plunging southwards. Most of these earlier folds occur in the southwestern ninth (G) of the quadrangle.

Later folds (B_2): Abundant in the southwestern part of the Burke quadrangle (Figs. 6, 7, and 8), later folds are also prevalent to the north, around Burke Hollow (Fig. 9), Hartwellville, and in the Moose valley (Figs. 42 and 43), and along the eastern border at I4258. At the latter locality the schistosity of the Albee formation is deformed into small folds; this deformation is equated with the later deformational episode which affects the Waits River and Gile Mountain formations.

Later folds are identified by the criterion that the schistosity (S_1)

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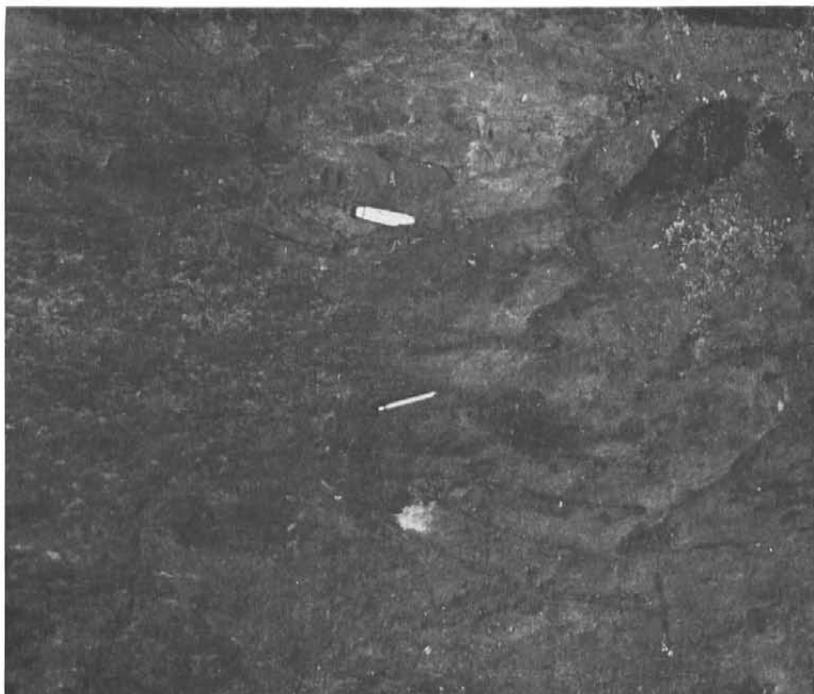
Figure 45.   Interdigitations of pelite in quartzose phyllite which represent very tight, isoclinal, earlier (B_1) folds. The schistosity (S_1), parallel to the penknife, is axial plane to these folds. The earlier (B_1) folds are flexed into an open later (B_2) fold, the axial plane of which, parallel to the pencil, marks the slip cleavage (S_2). West of Sugar Hill (G2907).

is folded. It is actually the S_1 surfaces which define the form of the fold. As the S_1 surfaces are parallel to the sedimentary banding, the latter is usually folded congruently with the former. (However, Figure 48 and Figure 49 are contrary examples).

The later folds vary from relatively open ones (Fig. 38 and Fig. 42), usually in the more quartzose bands, to relatively tight, isoclinal ones (Fig. 48 and Fig. 50). Figure 50 shows clearly the mechanism of folding in a series of bands of differing competence. The quartzose bands are folded by concentric shear along the well-developed S_1 surfaces, while the pelite band is microfolded and has developed a slip cleavage which is parallel to the axial plane of the minor fold. This has resulted in the thickening of the hinge by slippage on the S_2 surfaces. The pelite folds

Figure 46. Photograph of a shadowy relict of an earlier (B_1) fold cut by an extremely well-developed schistosity (S_1), which is axial plane to the folds. Note the rippling of the schistosity at the bottom. From Barnes Brook, northeast of East Lyndon (G2816).

are, therefore, similar folds, which have been formed partly by shear folding. The same relationship is shown in Figure 37, where the S_2 cleavage is more pronouncedly developed in the core of the fold and is parallel to the axial plane, while out on the limbs it is poorly developed and has not been rotated into parallelism with the axial plane.

Figure 51 shows a characteristic drag-like fold form in a more quartzose band and intercalated quartz vein, while the pelitic rock above and below has been microfolded and has developed dominant S_2 surfaces, although the relicts of the S_1 surfaces are clearly evident.

Some of the later folds are entirely shear folds. These are observed in

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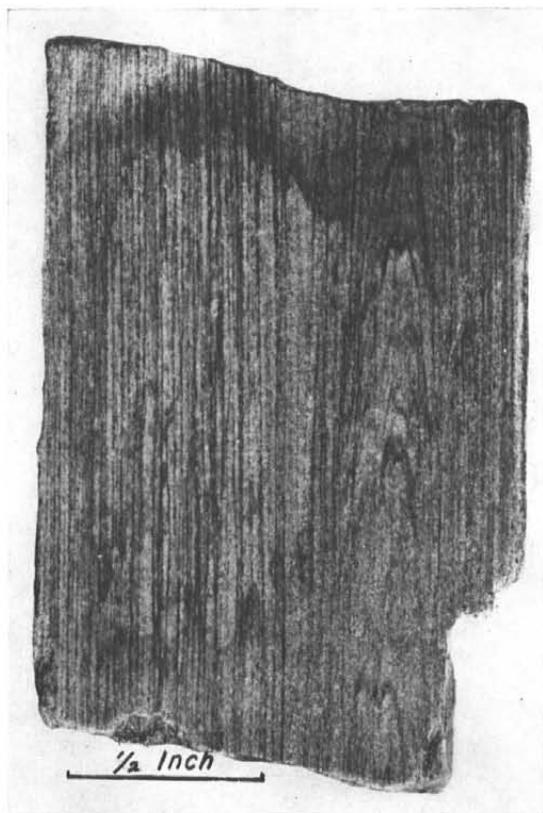
Figure 47.   Photograph of a relict earlier (B₁) fold now seen only as an isolated fold core. A perfect schistosity (S₁) has developed axial plane to the fold. From north of Flower Brook (E1627).

quartzose rocks which have a well-developed S_2 cleavage. In Figure 41 the flexuring of the S_1 surfaces between the S_2 surfaces is slight, but it is more pronounced in Figure 40 where a considerable amount of biotite has recrystallized along the S_2 surfaces. Quartz veins are deformed congruently with the folding upon shearing of the S_1 surfaces (Figs. 17, 51, 52, and 53), proving that the veins belong to an earlier phase of regional metamorphism. Earlier folds have been refolded by the later folds; only rarely are examples actually seen, although they must be common in the southwestern part of the quadrangle. Figure 45 shows a broad flexuring of the very tight earlier isoclinal folds marked by the

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Figure 48. Photograph of folds in Waits River impure limestone and phyllite, with quartz veins. Note the discontinuous beds over the arch of the fold; these represent tight isoclinal earlier (B_1) folds which have been refolded by the more obvious later (B_2) fold. Note, also, the absence of axial plane cleavage in the limestone. From Simpson Brook (G2728).

interdigitating pelitic bands. Figure 48 illustrates a refolded fold in limestone where no axial plane S_2 cleavage has developed. Figure 49 shows an isoclinal earlier fold deformed by an open later fold; the fold axes of the two folds are coincident, but the S_2 cleavage, sub-parallel to the axial plane of the later fold, cuts across the S_1 axial plane schistosity of the earlier fold.

LINEAR FEATURES

Fold Axes: The trend and plunge of fold axes have been measured wherever possible in the field; additional axes have been measured on oriented specimens in the laboratory. Still more fold axes have been constructed on the Wulff stereonet by the method described by Wegmann
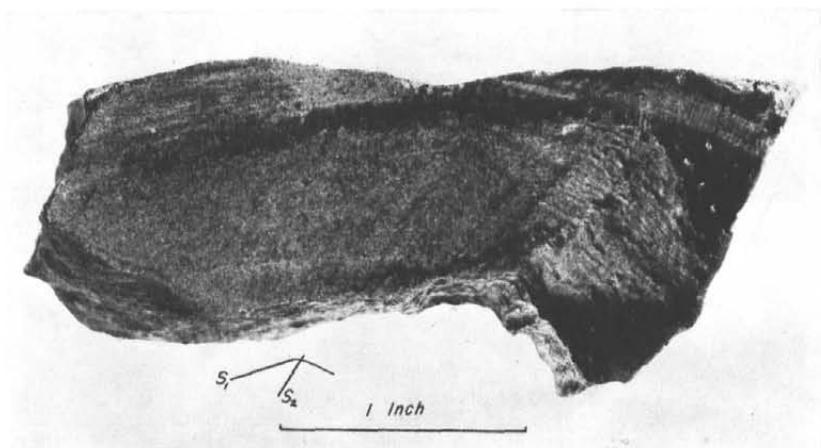
Figure 49.   Photograph of an isoclinal earlier (B_1) fold with a quartzose core and a pelitic envelope. The schistosity (S_1) is axial plane to the fold. This earlier fold has been deformed by a later open fold with the development of axial plane cleavage (S_2). S_2 has a different orientation in the quartzose and in the pelitic layers. Note how the biotite porphyroblasts outline the lithologic boundary. From west of Sugar Hill (G2908).

(1929): A series of readings of strike and dip are made in the field or on a hand specimen at intervals across a folded surface. Poles to these planes are plotted on a Wulff stereonet. If the fold is cylindroidal, these points will lie along a zone corresponding to a great circle. The pole to this great circle defines the fold axis.

B_1 *folds:* The trend and plunge of the 13 earlier (B_1) fold axes measured are shown on Plates 3 and 4 and have been plotted on the lower hemisphere of a Schmidt stereonet together with B_1 lineations, which represent the intersections of bedding (S_0) and schistosity (S_1), and are mainly in the southwestern area. The resultant dot plot has been contoured, as shown in Figure 56C. The maximum trends 24° and plunges 42° to the northeast. This indicates a northeasterly plunge for the earlier fold structures, and probably represents the attitude for the quadrangle as a whole.

B_2 *folds:* The trend and plunge of the 168 B_2 folds measured are shown on Plate 3. The pattern, axial plunge, and attitude of the axial planes of 63 of these later folds are shown on Plate 4. A total of 38 folds have a dextral pattern in plan view, 28 plunging northwards and 10 plunging southwards; 25 folds have a sinistral pattern in plan view, 21 plunging northwards and four plunging southwards.

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Figure 50. Photograph of a sharp later (B₂) fold in Gile Mountain banded phyllite. Note the slip cleavage (S₂) developed in the pelite band, which is "thickened" in the core of the fold parallel to S₂. The quartzose phyllite retains the schistosity (S₁), which is folded by concentric shear. From west of Sugar Hill (G3107).

The trend and plunge of the B_2 folds have also been plotted on the lower hemisphere of a Schmidt stereonet for the quadrangle as a whole and for subareas I and II. These dot plots have been contoured, and the resultant stereograms are shown in Figure 56D and on Plate 5, Figures 20 and 21, respectively. The composite diagram shows that the maximum of the B_2 fold axes trends 21° and plunges 23° to the northeast. The axes lie in a girdle striking 14° and dipping 70° to the southeast; this orientation is similar to the attitude of the maximum of the S_1 surfaces (see Fig. 57A). The B_2 fold axes in subarea I have a very similar attitude to that shown on the composite diagram, but in subarea II the maximum plunges 39° due south, although there is a sub-maximum trending 27°
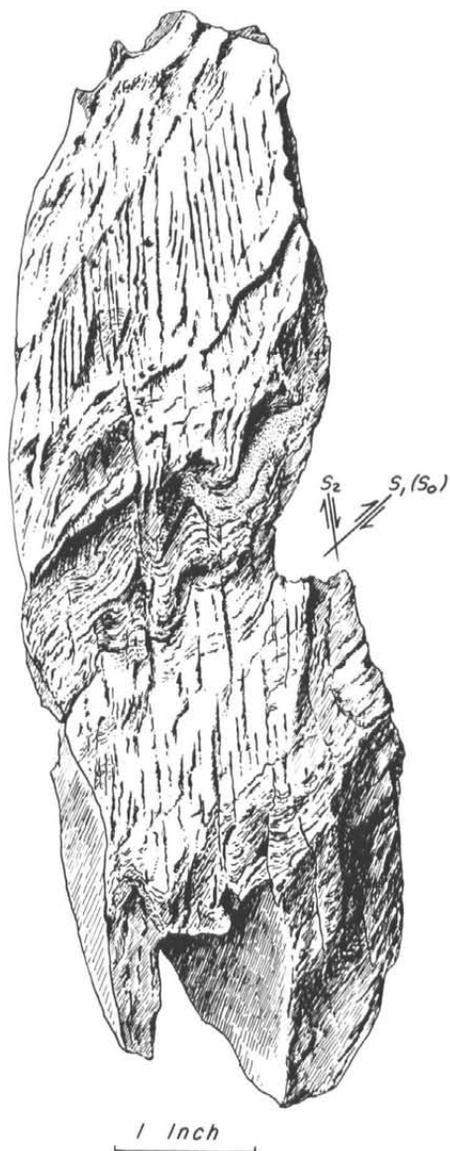
Figure 51.　Drawing of a hand specimen of phyllite showing drag folding of inter-
calated quartzose band and quartz vein (later folds) and development of slip cleavage
(S_2) in the pelitic bands. The cleavage is sub-parallel to the axial planes of the folds and
almost obliterates the schistosity (S_1), which is parallel to the bedding (S_0). From the
southern slope of Bemis Hill (G1440).

99

½ Inch

Figure 52. Photograph of later (B₂) folds in Gile Mountain banded phyllite. Axial plane cleavage (S₂) is weakly developed and masked by biotite porphyroblasts. Note the folded quartz vein, which is showing incipient boudinage. From Sheldon Brook (G0703).

and plunging 20° to the northeast. The axes lie in a girdle striking 12° and dipping 76° to the southeast, which is very similar to that of sub-area I.

Lineation: Lineation is frequently observed on schistosity (S₁) and cleavage (S₂) surfaces. It is usually prominent on these surfaces in pelitic rocks, but fainter in quartzose rocks, and absent in the impure limestone. The great majority of lineations noted in the Burke area are caused by the intersection of planar features and by the corrugation of surfaces due to microfolding. Mineral alignment, other than that produced by intersecting surfaces, is quite subordinate; amphibole is

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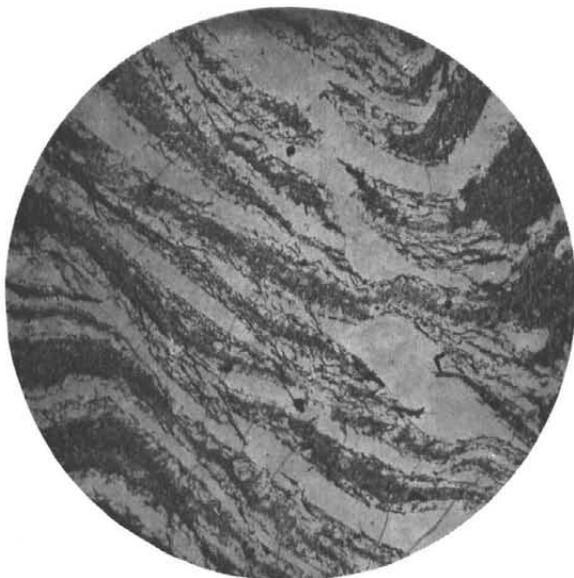
Figure 53.   Photomicrograph (×18) of folded quartz veins in phyllite. The schistosity (S_1) of the phyllite has been microfolded and is cut by slip cleavage. From south of Red Village (G1501).

sometimes aligned in the foliation of amphibolite and biotite sometimes occurs in streaks.

The pitch of some lineations has been measured in the field, but the great majority of determinations have been made on oriented specimens in the laboratory, where the nature of the lineation can be ascertained more readily and where fine lineations, which might be undetected in the field, are more visible. The pitch of the lineations has been converted to plunge by construction on a Wulff stereonet, and the plunges are plotted on Plate 3. The plunges have also been plotted on a Schmidt stereonet and the dot plots contoured; the resultant stereograms are shown in Figures 56E and F and on Plate 5, Figures 22 to 30.

B_1 *lineation:* Frequently individual hand specimens possess more than one lineation. Rarely one of the lineations is caused by the intersection of bedding (S_0) and schistosity (S_1); these instances are designated B_1 lineations, for they are the "b" lineations of an earlier deformation. As they are few in number, they are included with the earlier (B_1) folds on the aforementioned stereogram, Figure 56C.

B_2 *lineation:* The most prominent lineations are caused by the inter-

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section of S_2 cleavage surfaces and schistosity; these are designated B_2 lineations, the "b" lineations of a later deformation. They are shown for the quadrangle as a whole in the stereogram, Figure 56E, and for sub-area I in Figure 22 on Plate 5. The "b" lineations are often very coarse and mark the axes of microfolds in S_1. Frequently, however, it has not been possible to determine the type of lineation. The undetermined ones have been called undifferentiated lineations on the maps and stereograms; in many cases they are probably B_2 lineations. The stereogram, Figure 56F, includes all of the 330 B_2 lineations and all of the 618 undifferentiated lineations measured. The two diagrams, Figures 56E and F, are very similar; there is no evidence that the undifferentiated lineations are statistically different from the B_2 lineations. This is also true in subarea I, where only the B_2 lineations are included in its stereogram (Plate 5, Fig. 22), whereas in all of the other subarea lineation stereograms (Plate 5, Figs. 23 to 30) undifferentiated lineations are included with the B_2 lineations because of the scarcity of data. It is possible that some "a" lineations, i. e., fine striations parallel to the direction of transport of the deformed rock, are included among the undifferentiated lineations. An analysis of the stereograms shows that in subarea V there are two concentrations, one of which is mainly of B_2 lineations and the other mainly of undifferentiated lineations; they lie 60° apart on the great circle defining a girdle of lineation. In subarea XI a similar relationship is found, but in the remaining subareas there is no mutually exclusive preferment of B_2 or undifferentiated lineations. The stereograms on Plate 5, Figures 23 to 30, thus probably define the B_2 lineation directions.

When specimens show two or more lineations, it is sometimes possible to determine their order of development. Thus, in Figure 54 a coarse lineation marks the intersection of the prominent S_2 surface with the schistosity (S_1); a later lineation crosses both the S_1 and S_2 surfaces, but no measurable surface coincides with it. It appears to be caused by an incipient cleavage formed later than the S_2 cleavage and presumably in response to a changed stress field. Figure 55 exhibits three lineations. Firstly, there is a coarse one caused by S_2 intersecting S_1. This is cut by a prominent ripple-type lineation which appears to be due to the intersection of an incipient cleavage, but which is not seen as a structural surface. The third is a finer but more generally developed ripple-type lineation caused by another incipient cleavage. The two ripple-type lineations do not appear to cut one another and they may be con-temporaneous.
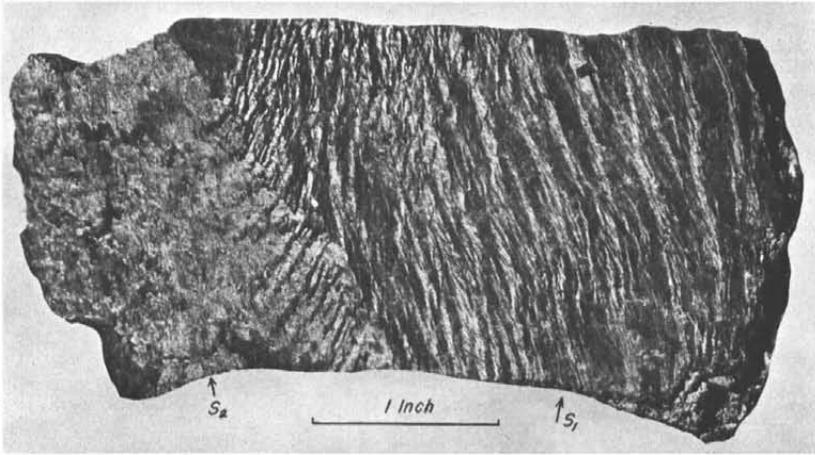
Figure 54. Photograph of phyllite showing two lineations. One lineation (B_2) is caused by the intersection of schistosity (S_1) and S_2 cleavage and produces a strong rippling of S_1. The other lineation cuts both S_1 and S_2, and is caused by a later incipient cleavage (S_3) intersecting them. From Red Village (G1005).

All multiple lineations have been measured and plotted on the lineation stereograms.

The B_2 lineations (Fig. 56E) have a maximum concentration trending 40° and plunging 45° to the northeast, which is somewhat different from the maximum concentration of the B_2 folds (Fig. 56D). The lineations fall on a girdle striking 13° and dipping 65° to the southeast, which, as with the girdle of B_2 folds, is similar to the attitude of the maximum of the S_1 planes. The subarea diagrams (Plate 5, Figs. 22 to 30) show, however, a diversity in the attitude of the lineation maxima. In subareas I and II the maxima are similar to that for the composite plot; in III the maximum is to the north-northeast and much gentler; in V there are two maxima, one plunging gently to the northeast and the other plunging gently to the east-southeast. In subareas VI to X the lineation maxima lie much farther to the east, even to the east-southeast in some cases.

Lineation girdles: Although the lineation stereograms have a distinct maximum in all cases, a well-developed girdle is also often present. Weiss and McIntyre (1957) give an excellent account of the geometry of superimposed fold systems and, in particular, point out that an early rectilinear lineation, if folded in a later deformation, will produce a girdle of points around the later fold axis direction, whereas lineations

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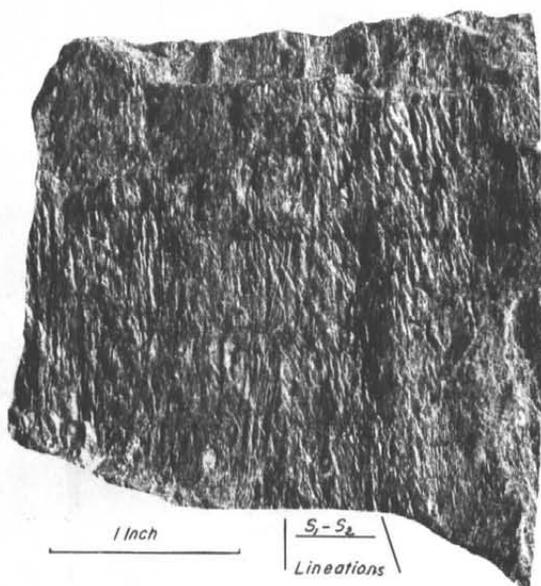
Figure 55. Photograph of schistosity (S_1) of Gile Mountain phyllite showing multiple lineations. The intersection of S_1 and S_2 surfaces produces a coarse horizontal lineation. A rippling of the S_1 surface in two other directions may be caused by incipient cleavages. From north of Granby (F3701).

produced by the intersections of a later cleavage across earlier folded surfaces will fall on a great circle defined by the later cleavage surface. The lineation girdles in the Burke quadrangle do not conform to the former of these geometric relations, as the poles to the lineation girdles do not coincide with the B_1, the B_2, or the β_2 directions (Fig. 57A). The girdles do, however, coincide approximately with the attitude of the S_2 plane maximum in subareas III, V, VI, VII, VIII (in part), IX, and XI, although the girdles lie nearer to the plane of the S_1 maximum for the quadrangle as a whole and for subareas II and VIII (in part) and between the maxima of S_1 and S_2 planes for subarea I. The girdles may thus represent the intersections of an S_2 cleavage across previously folded S_1 surfaces. However, the axes of folds in S_1 also define a girdle with a similar attitude to the corresponding lineation girdle (Figs. 57A, B, and C), and S_2 is often seen to be axial plane to folds in S_1. The B_2

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lineations thus are mainly contemporaneous with the B_2 folds in the S_1 surface, not later.

It is suggested that the explanation for the lineation girdle is that the generation of S_2 cleavages at varying attitudes around β'_2 would produce different intersections of S_2 on S_1. Likewise, varying attitudes of S_1 would produce a similar spreading of the lineation plunges. Why some subarea girdles parallel the S_1 surface maximum and others parallel the S_2 surface maximum is not completely understood. It may be that in the former instance the attitudes of the S_1 surfaces vary less than the S_2 and in the latter instance the converse is true.

Axial directions: Regional axial directions may be constructed (Wegmann, 1929) from a stereonet plot of the poles to a form surface by determining the normal to the great circle along which these poles lie (i. e. a π-diagram). The result is similar to a β-diagram (Sander, 1942). Axial directions have been constructed in this way from the strikes and dips of the S_1 surfaces (β_2 directions) and of the S_2 surfaces (β'_2 directions) for the whole of the Burke quadrangle (Fig. 57A) and for the subareas (Figs. 57B to K). β_2 directions plunge a moderate or small amount to the northeast in the quadrangle as a whole and in subareas I, II, III, V, VI, and VIIIA; they plunge steeply eastwards in subareas VIII (and VIIIB) and gently southwards in subareas VII, IX, X, and XI.

The S_2 planes do not form a folded structure; the β'_2 directions directions simply indicate the plunge of the line of intersection of all S_2 surfaces in the quadrangle. On the assumption that all of the S_2 surfaces are genetically related, their intersections define the intermediate strain direction for the field concerned. β'_2 directions plunge a moderate to small amount to the northeast in the quadrangle as a whole and in subareas I, II, III, V, and X; moderately to the east in subareas VI and VIII, and gently to the southeast in subarea IX.

Regional "b" directions can also be constructed statistically from the plunge of the intersections of the maxima of the S_1 and S_2 surfaces, which are plotted in Figures 57A and K. Northeasterly plunges are recorded for the quadrangle as a whole and in subareas I, II, V, and VI; southeasterly plunges in subareas VIII and IX; southerly plunges in subarea III, and southwesterly plunges in subareas X and XI.

Rodding: Occasionally quartz rodding is present in some outcrops (e. g., Fig. 8). It is caused by the close folding and shearing of quartz veins during the later phase of deformation. The rods represent the cores of nearly detached folds.

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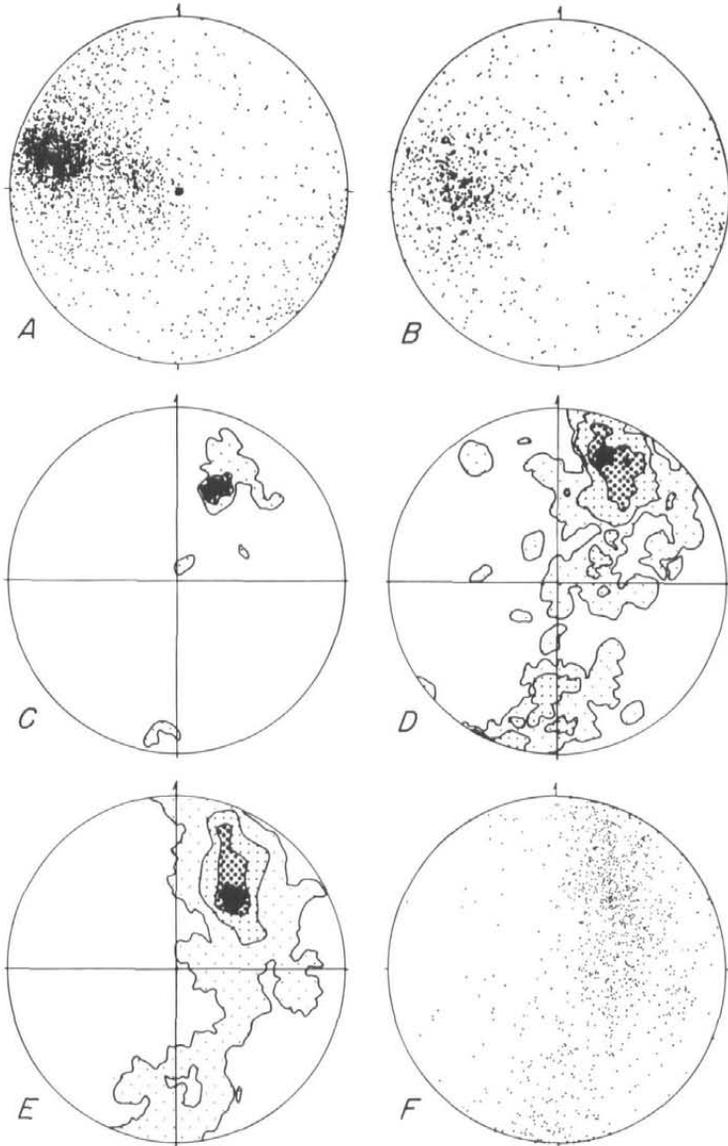
Figure 56.   Composite stereograms of Burke quadrangle structural data.
    A. Plot on the lower hemisphere of a Schmidt stereonet of poles to 1811
        schistosity (S₁) planes in the Waits River and Gile Mountain formations.
    B. Similar plot of poles to 766 cleavage (S₂) planes.

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Boudinage: Boudins have been observed at a small number of localities in the Burke quadrangle. The most common variety is produced from quartz veins by cross-cutting S_2 cleavage surfaces, along which some movement has occurred (Fig. 28). These boudins are similar to what Wilson (1953) describes as cleavage mullions. Comparable structures, but oval in cross-section, are well seen in a disrupted band of calc-silicate hornfels, intercalated in a fold and cut by cleavage, on Burke Mountain (H1254).

More interesting boudins are found in the large outcrops on the east side of Route 114 near the northern boundary of the quadrangle (B1848 and B1751). These have the characteristic elliptical cross-sections, the long axes of which vary in length from five inches to five feet. These boudins are composed of calcareous amphibole-quartz rock. A thin section of one specimen contains, in addition, clinozoisite, sphene, and plagioclase and is cut by narrow, folded quartz veins. Garnet occurs in some of the boudins. The outer surface of the boudins has a sheen of mica. One series of boudins, two and one-half inches thick, is intercalated in impure limestone. The plunge of the line between the boudins at these localities seems to be approximately down the dip of the strata; the boudins thus trend between 69° and 79° and plunge 56° to 59° to the northeast. This orientation is parallel to one of the two B_2 lineation maxima of subarea VIII and is also parallel to β_2; the boudins thus lie parallel to the "b" direction of the subarea.

RELATIONSHIPS OF BURKE STRUCTURAL DATA

The relationships of the various structural data are shown on the composite stereograms in Figures 56A to F and are shown synoptically on Schmidt stereonet plots in Figures 57A to K and in plan on the outline map in Figure 58. The simplicity of the data for the quadrangle as a whole (Fig. 57A) is misleading, as is evident from a study of the subarea data (Figs. 57B to K). The synoptic plot shows β_2 closely coincident with the maximum of B_2 fold axes and with the maxima due

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C. Similar plot of 33 earlier (B_1) fold axes and lineations; contoured at 6–12–15% per 1% area.

D. Similar plot of 168 later (B_2) fold axes; contoured at 1–3–6–9% per 1% area.

E. Similar plot of 330 B_2 lineations, due to the intersection of schistosity (S_1) and cleavage (S_2); contoured at 1–4–7–10% per 1% area.

F. Similar plot of 948 lineations, including the 330 B_2 lineations in (E) and 618 undifferentiated lineations.

KEY TO FIGURE 57

	Plane normal to pole maximum	Pole maximum
S₁ schistosity		
S₂ cleavage		
	Girdle	Maximum
B₁ folds & lineations		
B₂ folds		
B₂ lineations		
β₂ pole to πS₁ girdle	■	
β₂′ pole to πS₂ girdle	◆	
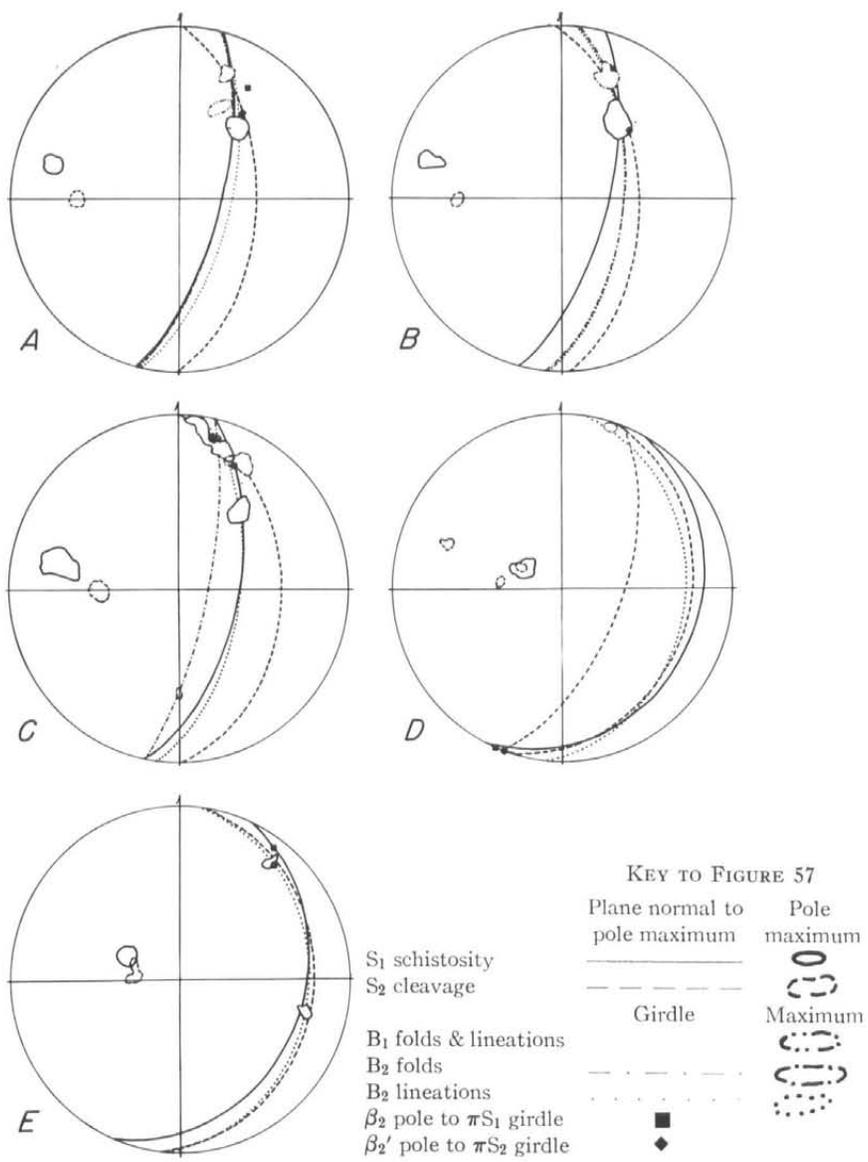
Figure 57.   Continued on page 109.

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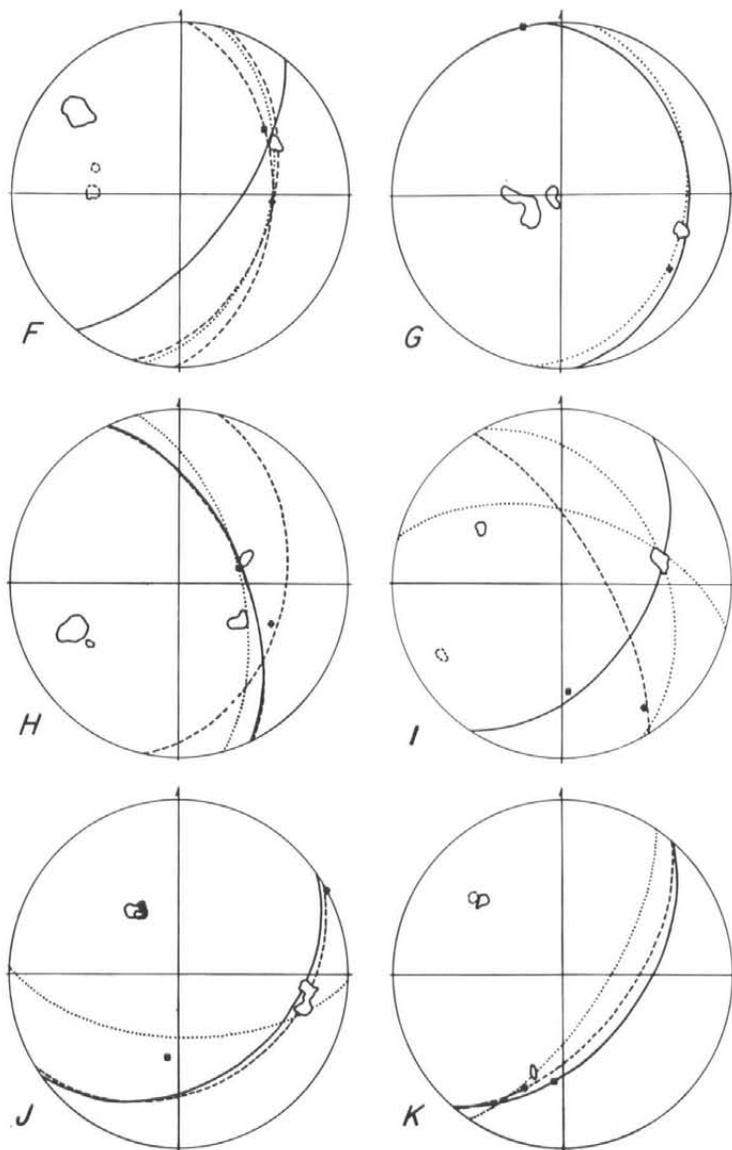
Figure 57. Synoptic stereograms of Burke quadrangle structural data.
A, entire area; B, subarea I; C, subarea II; D, subarea III; E, subarea V;
F, subarea VI; G, subarea VII; H, subarea VIII; I, subarea IX; J, subarea X;
K, subarea XI.

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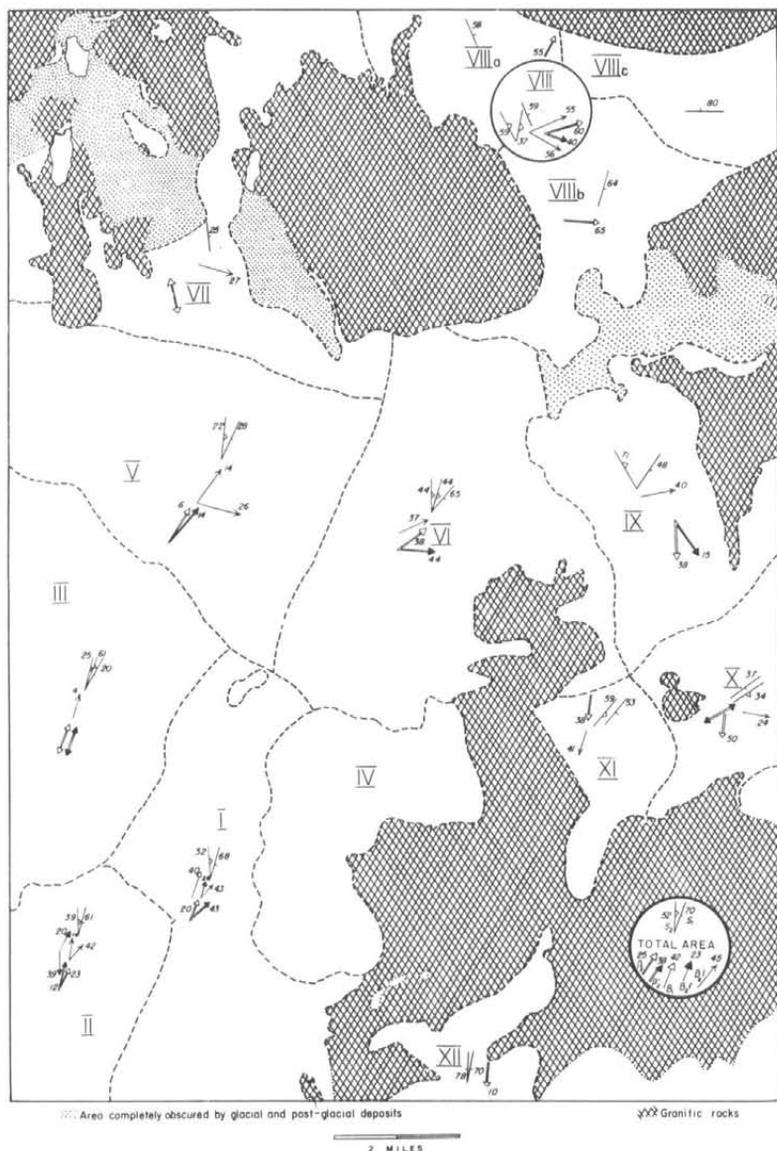
Figure 58. Map of generalized, statistically-derived structural data of the Burke quadrangle and its subareas. Roman numerals refer to subareas. All symbols as used for total area diagram: S_1, schistosity; S_2, cleavage; β_2, pole to πS_1 girdle; β'_2, pole to πS_2 girdle; B_1 earlier fold axes and lineations; B_2f, later fold axes; B_2l, later lineations.

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to the intersection of S_1 and S_2 planes. β'_2 lies close to the maximum of the B_2 lineation, and the girdle of B_2 folds and B_2 lineations is coincident with the S_1 plane maximum.

The schistosity (S_1) has a strong preferred orientation throughout; its strike, however, shows a distinct swing to the northwest in the northern part of the quadrangle and a more easterly swing in the eastern part. The S_2 cleavage also shows a strong preferred orientation. Over much of the area it has a constant relation to S_1, namely, it trends west of and has a gentler dip than the latter; in three subareas (III, IX, and XI) it has a steeper dip, while in one subarea (X) it trends slightly east of S_1. The β_2 and β'_2 directions do not coincide. The north-northeast plunge of β_2 steepens from 20° to 23° in the southwestern part of the quadrangle to 55° in the northeastern part, and the β_2 trend varies from northeasterly in the western part of the quadrangle to easterly in the center and northern parts and to southerly in the eastern part. β'_2 plunges northeasterly in the southwestern subareas (I and II), but plunges easterly in subareas VI and VIII and southeasterly in subarea IX. These variations indicate an inhomogeneity in the structure of the quadrangle as a whole.

The small amount of detail available on B_1 folds and lineations indicates that they plunge 42° to the northeast. This is the regional plunge of the structures formed during the earlier phase of deformation. Although these structures have been affected by the later deformation, it is considered that the later movements superimposed mainly smaller scale structures on the earlier structures, so that the present regional B_1 plunge is closely similar to its original attitude and represents, therefore, the axis of intermediate strain of the earlier deformation. The axis of greatest strain may lie in the schistosity (S_1) maximum and normal to the B_1 direction; this axis would trend about 172° and plunge 48° to the southeast. It may lie at a small angle to the S_1 plane if the deformation that produced the schistosity was substantially one of simple shear.

The regional B_2 fold maximum trends 21° and plunges 23° northeast; plunge to the south is important only in subarea II. The regional B_2 lineation trends 40° and plunges 45° northeast, but the trend actually varies throughout the area, as noted above. In particular, the northern and eastern subareas show an easterly or southerly plunge, compared to the northeasterly plunge in the western subareas. The regional β_2 axis is close to the B_2 fold maximum, while the β'_2 axis is close to the B_2 lineation maximum. The subareas, however, again show greater varia-

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tions. The northern and eastern subareas have β_2 axes plunging in an easterly or southerly direction.

The girdle of B_2 lineations may be explained by the varying attitude of S_2 surfaces superimposed on strongly preferred, but also varying, S_1 surfaces. Later S_2 surfaces may perhaps also have developed in part after the folding of S_1 surfaces. The lack of coincidence between the β_2 and β'_2 axial directions may be explained by the inclined and somewhat varying attitude of the S_1 surfaces to the axis of greatest stress; β'_2 would tend to develop parallel to the axis of intermediate strain, but the β_2 axis (and the B_2 fold axis) would develop at varying trends and plunges depending upon the attitude of the S_1 surfaces. For the area as a whole the attitude of the maximum of S_2 planes may approximate the AB plane of the strain ellipsoid, and may contain the greatest and intermediate axes of strain. The intermediate axis would be defined by the β'_2 axis, which trends 37° and plunges 38° northeast. The least strain axis would then coincide with the S_2 pole maximum, which trends 271° and plunges 38° northwest.

Major Structures

The constancy of strike and dip of the bedding schistosity (S_1) suggests that the overall structure in the Burke quadrangle is a simple one. The bedding schistosity strikes north-northeast and dips steeply to the southeast over much of the quadrangle. The dip is more gentle to the west and in the northwestern area it is horizontal; in the northern part the strike swings to the northwest and the dip is steep to the northeast. (The deflection of strike to a more easterly direction in subareas VIIIC, X, and XI may be explained as the effect of nearby granitic intrusions). The rocks may thus be interpreted as homoclinal and, as the dip flattens to the west, as being on the eastern limb of a very large anticlinal structure, the axis of which lies mainly to the west of the Burke quadrangle but trends north-northeastwards through the northwestern part of the quadrangle, where horizontal or gentle dips to the northeast or north are characteristic. This large structure would plunge north-northeast parallel to the B_1 folds and B_1 lineations. The abundant B_2 folds and the S_2 cleavage were formed during a later deformation; the plunge of these later folds is mainly to the northeast, but southward plunges also occur. The slip movement appears to have been mainly upwards to the northwest on the west side of the bedding schistosity.

The plunge of the minor folds in the southwestern part of the quadrangle, particularly the infold of Gile Mountain strata surrounded by

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Waits River strata in the Shonya and Graves hills area, appears to be abnormal. If the Gile Mountain formation is younger than the Waits River, the Gile Mountain must lie in a stratigraphic syncline, but structurally it seems to behave as an anticline, plunging northwards at its northern end and southwards at its southern end. Possibly, however, the plunge of the later folds may obscure the true plunge of the major structure, which is an earlier structure, as shown by the trace of its axial surface paralleling the schistosity (S_1). Alternatively, it may be that the folds and their plunges have been inverted, as L. M. Hall (1959) suggests for the isoclinal folds involving the Waits River-Gile Mountain contact in the eastern part of the St. Johnsbury quadrangle. Reversal of plunges by oversteepening, that is, by rotation around an axis normal to the axial plane is, however, unlikely, as no evidence to support such rotation has been found in the Shonya-Graves area. If the Gile Mountain formation is older than the Waits River, the folds have a normal form for the inverted limb of a major structure.

Turning to the surrounding regions, the northwestern trend of the northern part of the Burke quadrangle continues into the Island Pond quadrangle (Goodwin, 1963); the Gile Mountain formation strikes into the Westmore formation mapped by Doll (1951) in the Memphremagog quadrangle, indicating that the two formations are to be correlated with one another. Westwards, in the Lyndonville quadrangle, the dip of the Waits River formation is gentle over the crest of the major structure called the Willoughby arch (Dennis, 1956), and then steepens to the west and is overlain by a band of Gile Mountain strata which is continuous with the Westmore formation of Doll. The Gile Mountain strata are in a syncline overturned to the east; this has been called the Brownington syncline by Doll (1951) and is recognized by Dennis (1956). To the south, the northeasterly trends of the bedding schistosity of the Burke quadrangle continue into the Littleton quadrangle (Eric and Dennis, 1958) and into the St. Johnsbury quadrangle (L. M. Hall, 1959) where the Willoughby arch and the Brownington syncline are also recognized.

The Willoughby arch has been interpreted by Dennis (1956) and by L. M. Hall (1959) as a later structure and to be a cleavage arch. Hall considers the schistosity over much of the arch to be a later schistosity, of the same age as the slip cleavage to the east of the arch. This is in agreement with the structural interpretations made in the Woodsville quadrangle by White (1949) and by White and Billings (1951). Farther south, in the Strafford area, White and Jahns (1950) describe a broad

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cleavage arch (which includes the Strafford cleavage dome) in the Waits River formation. They ascribe its origin to a later upward flowage of the calcareous rocks superimposed on a homoclinal sequence which produced the later reverse drag folds on the flanks and also large folds which were subsequently rolled over the arch into a recumbent position. White and Jahns tentatively correlate the band of phyllite to the west of the cleavage arch with the Gile Mountain formation and they consider the band to be a homocline, the sequence being repeated by a large strike fault in the calcareous rocks just east of the band of phyllite. They also consider the evidence for whether the western band of Gile Mountain is a syncline or the western limb of a larger syncline; in the latter case the Waits River formation of the cleavage arch would lie stratigraphically above the Gile Mountain band and the cleavage arch would be superimposed on the syncline. White and Jahns reject both of these alternatives. Murthy (1957), however, in his description of the geology of the East Barre quadrangle, favors a major synclinal structure. He also suggests that facies changes are important and account for the southern termination of the western band of Gile Mountain in the Randolph quadrangle and for the northern termination of the Waits River of the cleavage arch in the Island Pond quadrangle. (See the discussion of Murthy's explanation by White, 1959, and by Dennis, 1959).

The interpretation of the major structure of an area like the Burke quadrangle is fraught with difficulty because Burke comprises only a relatively small part of the major structure. An interpretation must be consistent with the detailed minor structural data described under Minor Structural Features, and it should also be reasonably in accord with the information available from the aforementioned neighboring quadrangles. The northwestward swing in strike of the bedding schistosity (S_1) proves that the Waits River formation of the Willoughby arch indeed terminates in the Island Pond quadrangle by structural closure. The arch, according to White and Jahns (1950) and to Dennis (1956), is formed by the later S_2 cleavage produced by upward and outward rock flowage; however, the bedding schistosity (S_1) is also deformed into the north-north-eastward-plunging arch structure. This deformation of the earlier-formed schistosity was accompanied by the formation of the B_2 folds and S_2 cleavage surfaces. If these structures are to be related to upward flowage caused by the rise of subjacent granitic rocks (Lyons, 1955; Dennis, 1956; L. M. Hall, 1959), then the regional structure of northeastern Vermont would be similar to one of the two structures shown diagrammatically by Eric and Dennis (1958, Fig. 13, p. 58). Because of the great thickness

of Gile Mountain strata exposed in the Burke quadrangle, the presence of a syncline (Ibid, Fig. 13B) immediately west of the Monroe contact is attractive. No direct evidence for such a syncline has been detected, but, as pointed out above, the minor structures of the eastern subareas, as exemplified in the B_2, β_2, and β'_2 directions, are different from those of the western subareas. The absence of the Waits River formation on the eastern limb of the syncline could be explained by facies change, for rapid facies changes may be expected normal to the length of an ortho-geosyncline such as the one in which the sediments of the Waits River and Gile Mountain formations were deposited. If the contact between the Gile Mountain formation and the Albee formation in this area is a fault, the presence of a major unconformity between the Vermont and New Hampshire sequences is still not excluded (see L. M. Hall, 1959, p. 86).

The writer is, however, attracted to another interpretation of the major structure. This considers large-scale recumbent folding to be present, with the Brownington syncline representing the downward-facing nose of such a fold, the axial plane of which lies across the Willoughby arch. Eric and Dennis (1958, p. 61) briefly mention this as a possibility; see also Goodwin (1963) and Ern (1963). The following data from the Burke quadrangle support the presence of this type of structure:

(1) The schistosity (S_1) is nearly everywhere sensibly parallel to the bedding (S_0), but is seen to be axial plane to the few earlier (B_1) folds recognized in the area. This relationship suggests that translatory movements parallel to the bedding were important in the development of the schistosity and that the folds that were formed in the bedding were to a considerable extent obliterated by continued movement on the mechanically generated schistosity, thereby explaining why few such folds are recognized in a sequence that must be greatly thickened.

(2) The direction of the maximum of the B_1 folds and lineations is similar, but not identical, to that of the B_2 folds and B_2 lineations (Fig. 57A). This suggests that the deforming forces that produced both the B_1 and B_2 folds were operating in general from the same direction.

(3) If it is assumed that the S_2 surfaces represent the AB plane of the strain ellipsoid, then the S_2 pole maximum corresponds to the axis of least strain and corresponds approximately with the direction of greatest stress.

(4) Likewise, the pole maximum of the S_1 surfaces may approximate the least strain axis of the rotational strain that accompanied the development of the S_1 surfaces and that resulted in rotation of the latter

almost into the AB plane. Such a relation between the S_1 maximum and the least strain axis would be expected if considerable translation had occurred.

(5) The statistical axes, i. e., the pole maxima of the S_1 and S_2 surfaces, for the Burke area as a whole (Fig. 57A) lie relatively closely together. The attitudes of the pole maxima vary from subarea to subarea (Figs. 57B to K), but they all lie in a general westerly direction with variable plunges. This may be indicative of the rotational nature of recumbent folding.

(6) The attitudes of the B_2, β_2, and β'_2 axial directions also vary from subarea to subarea. In general, however, there is a relatively consistent variation of each of the three axial directions between the western subareas and the northern and eastern subareas. This might be due to difference in tectonic level, for example, the eastern subareas might be on the opposite limb of a large-scale recumbent fold.

(7) If a large-scale recumbent fold is present, then the contact between the Waits River and Gile Mountain formations is an inverted one, for the latter would be in the core of an anticline. The apparently inverted northward-plunging isoclinal folds that affect the contact in the south-western area of the quadrangle (see section CD on Plate 1) would then be normal for an inverted limb. L. M. Hall (1959) reports abnormal fold structures in the Waits River-Gile Mountain contact in the St. Johns-bury quadrangle and interprets them as folds, the plunge of which has been inverted.

(8) The shear sense, which is mainly up on the west side of the S_1 surfaces, would also be normal, as would dextral minor folds plunging northwards and sinistral minor folds plunging southwards.

(9) In subarea VIII the relationships (Fig. 57H) are interesting in that the β_2 direction trends 75° and plunges 60° northeastwards. This suggests that a major fold plunging in this direction may be present in the area rather than a north-northeastward-plunging one. The β'_2 direction is even more to the east, trending 112° and plunging 40° southeastwards.

Large-scale recumbent folds across the Strafford dome have been mapped by White and Jahns (1950), who interpret them as later stage folds which were rolled over the arch by continued upward flowage of the calcareous Waits River formation. Lyons (1955) has also mapped recumbent folds in the Hanover quadrangle. Murthy (1957, pp. 62-63) describes recumbent minor folds on the top of the arch in the East Barre quadrangle; he regards these as rotated earlier folds (but see White, 1959, and Murthy, 1959a and b). L. M. Hall (1959, Plate 2) also maps recum-

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bent minor folds, on the crest of the Willoughby arch in the St. Johnsbury quadrangle.

Unfortunately the data on way-up of bedding in the Burke quadrangle (as described under Bedding, S_0) do not provide evidence of large-scale repetition in the Gile Mountain formation. On balance, the data favor the interpretation that a small majority of the outcrops concerned are the right-way up. However, the author considers these data too sparse to be conclusive one way or the other. On Plate 1, the Gile Mountain is shown as being younger than the Waits River formation in keeping with the generally accepted interpretation of the sequence; however, as is described above, it is possible that the Waits River is the younger of the two.

In summary, large-scale recumbent folding is a distinct possibility. The earlier and later deformation phases recognized in northeastern Vermont would then be successive phases in a continuously developed tectonic framework and the major structure would be related to both. The Strafford-Willoughby arch would represent a culmination, and its steeply-plunging northern end in the Island Pond quadrangle would be a major crossing structure producing a plunge depression transverse to the major recumbent fold.

The relationships at the Monroe contact remain obscure, but perhaps major faulting is present. Further difficulty in building up the overall structural and stratigraphic picture is because of probable facies changes, both normal to and along the strike of the structures. If the Brownington syncline is the downward-facing nose of a recumbent anticline, the structural relations west of the syncline would be more complex than present interpretations indicate. The writer is not able to comment on this problem as he is not well enough familiar with the ground in the neighboring areas.

If the hypothesis of major recumbent folding has to be ruled out, then the interpretation applied to the structures by Eric and Dennis (1958), and indicated in their diagrammatic section in Figure 13B, is a distinct possibility.

METAMORPHISM

The sedimentary rocks and associated extrusive and intrusive rocks of the Burke quadrangle have experienced a complex metamorphic history. There is evidence for two distinct phases of progressive metamorphism and perhaps for a final retrogressive stage. Differences in metamorphic environments are expressed in the resultant mineral assemblages,

and these have been described in the geologic literature and grouped into a series of metamorphic facies broadly related to the kinds and intensity of the metamorphism (Eskola, 1915, 1921; Fyfe, Turner, and Verhoogen, 1958). Metamorphic grade is also indicated by the appearance of index minerals in sediments which are essentially isochemical (Barrow, 1893; Tilley, 1924). The appearance of new minerals corresponds to changes in the conditions of metamorphism, particularly temperature, so that zones of metamorphism can be recognized and traced in the field. Thus, a line drawn on a map marking the first appearance of a particular index mineral, representing the passage from a lower to a higher grade of metamorphism, defines the outer limit of the higher grade zone. Such lines of equal grade are called isograds (Tilley, 1924). Careful recording of the mineralogic variation of a particular widespread rock type enables the zone of progressive metamorphism to be mapped. Pelitic rocks commonly have been chosen for this purpose as they are sensitive to changes in metamorphic conditions.

The isograds of pelitic sediments throughout the Burke quadrangle are shown on Plate 1. The index minerals of progressive metamorphism utilized are biotite, almandine garnet, staurolite, andalusite, and sillimanite. These isograds are very generalized, particularly in some areas, because of the glacial drift cover, the lack of outcrops of the appropriate chemical composition, and the impossibility of covering every bit of the ground, especially in the more inaccessible areas, in sufficient detail in the time available. Furthermore, although the index minerals are usually developed as porphyroblasts (except sillimanite, which is often only present microscopically), it is not always possible to recognize them in the field. Even though many specimens were collected for later laboratory examination it is probable that occurrences of index minerals were missed. Much of the area of the Waits River formation west of the East Branch of the Passumpsic River presents difficulties because of the paucity of outcrops, particularly of rocks of the appropriate composition. It is possible that the garnet zone is more wide-spread than shown on Plate 1. However, although the lines may be amended by further detailed work, they show the generalized distribution of zones of progressive metamorphism.

Mineralogy of the Metamorphic Rocks

Biotite: Deep brown biotite is ubiquitous in the pelitic and quartzose rocks of the area. It occurs as small flakes aligned in the schistosity and, less commonly, in the false cleavage. Occasionally it occurs in narrow,

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biotite-rich layers parallel to the false cleavage where frequently the highly perfect basal cleavage of the biotite is oriented at high angles to the rock cleavage. Occasionally layers one to two centimeters thick and paralleling the main schistosity are composed almost entirely of biotite. Porphyroblasts of biotite occur plentifully throughout the area; they cut both the schistosity and false cleavage and are undeformed. Characteristically they are pleochroic straw-yellow to red-brown, they possess prominent pleochroic halos and frequently lines of dust inclusions oriented parallel to the surrounding schistosity, and they sometimes show alteration to chlorite. In the hornfels zone the biotite often has the characteristic hornfels pleochroism to a very dark brown in contrast to the red-brown biotite of the lower grades. It is sometimes bleached and partly altered to chlorite; in some thin sections it is seen to be replaced by felted fibrous sillimanite.

Feldspar: Feldspar occurs in most of the thin sections. In the lower grades of the pelitic rocks it is inconspicuous; in the psammitic rocks plagioclase (usually oligoclase, sometimes albite), untwinned or very rarely with albite twinning, is very sparse. Plagioclase is present in the amphibole quartzite bands and is an important constituent, usually fine grained and untwinned, of the amphibolite (see below, under amphibolite assemblages). Twinned and untwinned plagioclase, fine grained and difficult to determine, but identified as oligoclase in some slides, is present, usually in small amounts in the hornfels; rare myrmekite is present in a few slides. Microcline is present in only two sections out of thirty. A number of thin sections of what in hand specimen appears to be granitized sediment contain notable amounts of feldspars which have very irregular, poorly defined boundaries against the groundmass, and contain abundant bleb-like inclusions, particularly of quartz. In some sections they attain a size up to 1.58 mm., and are untwinned or simply twinned. Optical determination is very difficult, but in several specimens the maximum refractive index indicates oligoclase. It is possible, however, that some orthoclase is also present. These poikiloblastic feldspars appear to have grown metasomatically in the groundmass incorporating much of the latter within the structure, and point to the action of soda- and possibly potash-bearing solutions. Microcline and plagioclase occur also in some thin sections of calcsilicate hornfels.

White mica: White mica occurs as very small flakes (sericite) forming the schistosity in all zones of metamorphism. It is most abundant in the pelitic layers but occurs also in the quartzose rocks where it also forms

the schistosity. Individual grains may be bent when the schistosity is thrown into microfolds. The microfolding results in the sericite forming the secondary false or slip cleavage; whether any new recrystallization of sericite takes place in the latter is difficult to determine but in the extreme examples when the cleavage pervades the rock mass and is indistinguishable from a schistosity some recrystallization is likely to have taken place.

White mica is a very common constituent in hornfels occurring in the great majority of specimens examined in amounts varying from rare to abundant. It occurs in two characteristic ways—as aggregates replacing porphyroblasts of other minerals and in some rocks as very common large porphyroblastic plates up to 12 mm. across. The latter have a peculiar metallic luster and dominate a type of hornfels in which there is little or no relict foliate structure and biotite is much reduced in amount. This hornfels rarely has porphyroblasts of aluminum silicates, although sillimanite is present as microscopic needles and fibers in the quartz and mica and in some cases as partially muscovitized fibrolite. This rock type thus seems to have been extremely muscovitized. The pseudomorphous replacement aggregates of muscovite are discussed below under "shimmer aggregates". White mica appears also to replace biotite in some specimens in what were apparently biotite-rich laminae and is accompanied by granular iron ores, fibrolite, and tourmaline.

Garnet: Idioblastic, wine-colored to deep red almandine garnets (up to 6 mm.) occur frequently, and their appearance marks the garnet zone. They occur also in the staurolite zone (Fig. 65), but in the high grade zone are usually small, xenoblastic, and rarely are conspicuous. Some are altering to chlorite and in one instance to green biotite. The refractive index of some idioblasts is 1.795 ($\pm$ 0.004) and the specific gravity is 4.05; these data indicate 70% almandine and 30% pyrope (Winchell and Winchell, 1951, pp. 485–488).

The garnets occur in the pelites and semi-pelites, also in amphibole quartzite bands and rarely in amphibolite. They usually have inclusions of quartz, which are sometimes arranged helicitically. They are sometimes fractured and occasionally granulated. Grossularite occurs in calc-silicate hornfels bands, particularly on Burke Mountain, in some cases up to 33 mm. in diameter.

Staurolite: Staurolite has a wide occurrence in the pelitic rocks of the area. It is frequently idioblastic (Figs. 18 and 59); cruciform twins and twins crossing at 60° both occur, but one type seems to characterize a particular area. Large cruciform twins (up to 22 mm. long) are very

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Figure 59. Photomicrograph (×18) of idioblastic staurolite in Gile Mountain horn-fels. Note the arrangement of the inclusions of quartz and opaques and the small, lath-like, inclusion-free areas, which represent previous biotite flakes. The groundmass contains quartz, abundant biotite, muscovite, and opaques. From the western slope of East Mountain (F2355).

abundant on Bull Mountain and its northern slopes. The porphyroblasts are usually full of inclusions, sometimes arranged helicitically and parallel to the surrounding schistosity and cleavage. They often show complete or partial alteration to sericite and chlorite.

Idioblastic staurolite occurs rarely within the sillimanite zone: xenoblastic grains are often present, however, and over 25% of the thin sections contain them. These occur sparsely throughout the groundmass and are also included within "shimmer aggregates." The former apparently are relicts, but the latter occurrence is more difficult to explain. Some aggregates are undoubtedly pseudomorphous after staurolite and the included grains are relict; others are not after staurolite, and this occurrence is discussed below under "shimmer aggregates."

Andalusite: Porphyroblasts of andalusite have been identified in four thin sections of hornfels and in a number of other specimens it was found by microscopic examination of crushed "shimmer aggregates." At locality F2948 perfect crystals up to 17 mm. long are abundant in a

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large exposure of Gile Mountain schist underlying granite. In thin section (Fig. 19) the porphyroblasts are pleochroic colorless to rose and contain opaque inclusions. In the other sections the andalusite is less well formed or is xenoblastic and associated with much muscovite (Fig. 60) and in some cases with sillimanite needles and fibers.

Sillimanite: Sillimanite is of frequent occurrence in the thin sections of the high-grade hornfels. It is presumed to have a general distribution throughout these more highly altered rocks and it seems justified placing them in the sillimanite zone. It occurs both as porphyroblasts up to 22 mm. long, as needles, and as matted fibrous aggregates. The porphyroblasts are well developed on Burke Mountain and are often associated with muscovite; they may be seen lying within muscovite aggregates. Some also occur as parallel aggregates of fibers (fibrolite). Large porphyroblasts are also common on Lees Hill (Fig. 61) and Miles Mountain in rocks believed to be in the Albee formation. Minute needles are frequently observed in the quartz grains; these are assumed to be sillimanite as they occur only in rocks bearing sillimanite or within the sillimanite zone. Matted fibrous aggregates of sillimanite (fibrolite) also occur within muscovite (Fig. 62) and also as sinuous streaks and patches associated with the penetrating dark brown biotite and obviously replacing the latter (Fig. 63). Needle and fibrous sillimanite (fibrolite) inclusions have also been noted in andalusite. In some staurolite xenoblasts, particularly those from "shimmer aggregates," there are curious matted anastomosing fibrous inclusions which may also be sillimanite.

Kyanite: Large brilliant blue kyanite porphyroblasts up to 87 mm. long have been found at one locality, in the roadside outcrop at B1848 of schists of the Waits River formation in the zone transitional to the Gile Mountain formation. The kyanite (Fig. 64) occurs embedded in coarse quartz or pegmatitic vein material which is associated with contorted masses of biotite-rich schist. Coarse pink andalusite crystals also occur with the kyanite, and sillimanite needles (fibrolite) are present in the associated quartz and plagioclase and as fibrolite masses after biotite.

The only other occurrence of kyanite found is at B0347 where it is in small plates visible only with a hand lens.

Cordierite: No cordierite has been definitely identified in any thin section, but some "spotted" rocks are present in a few localities, the spots of which may represent pseudomorphs after this mineral. At locality B1848, near the aforementioned kyanite occurrence, are irregular spots composed of a highly birefringent aggregate which may

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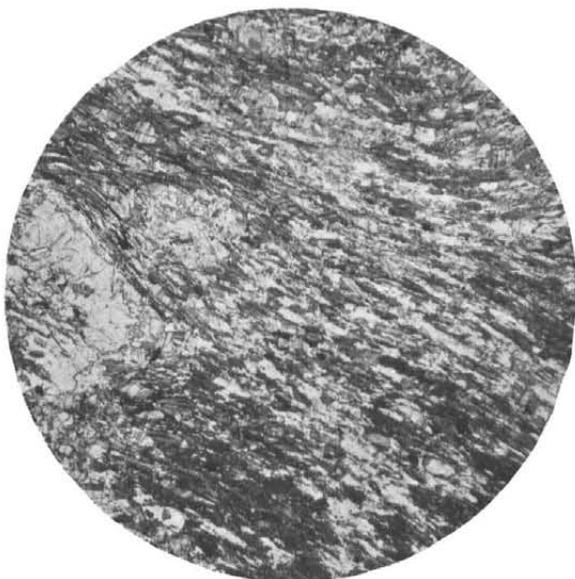
Figure 60. Photomicrograph (×18) of andalusite-sillimanite schist. The andalusite (on the left) has inclusions of quartz, biotite, and opaques and is surrounded by a corona of muscovite. Sillimanite (fibrolite) occurs in quartz and after biotite (near top and right-hand border). Staurolite xenoblasts are also present. Groundmass contains quartz, biotite, and muscovite. From the eastern side of Route 114 (B1751).

be after cordierite. On the western slope of Kirby Mountain phyllite contains prominent ovoid spots up to 9 mm. long which are composed of aggregates of biotite, chlorite, muscovite, and quartz; these may also be after cordierite but more likely after garnet.

Chlorite: Chlorite is of frequent occurrence throughout the area. Much of it occurs as an alteration product of biotite, amphibole, staurolite, and garnet in all stages of retrogression in all of the zones of metamorphism and in the granitic rocks. In addition, it occurs as porphyroblasts (Fig. 65), sometimes ghost-like, but usually well-defined and large (up to 6 mm. across). This occurrence is also widespread, but it is particularly noticeable in the garnet and staurolite zones of the Gile Mountain formation in the Moose valley east of East Haven Mountain. It does not appear to be replacing any particular mineral. Optically it is slightly pleochroic colorless to light green and full of opaque inclusions. Most of the specimens examined have a refractive index of $N\beta = 1.638$ (± 0.002) and very low anomalous birefringence, and are

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Figure 61. Photomicrograph (×18) of sillimanite-garnet hornfels, showing idioblastic sillimanite (one surrounded by garnet) containing inclusions of quartz, biotite, and muscovite; some fibrolite. The other minerals are biotite, muscovite, a little microcline, oligoclase, and iron ores. From Lees Hill (I2725).

nearly uniaxial positive. According to Winchell and Winchell (1951, p. 383) this would be ripidolite, an alumina-iron-rich chlorite; one specimen is probably diabantite. Penninite, with ultra-blue polarization colors also often occurs, apparently as an alteration of biotite or amphibole.

It is difficult to determine the relationship of the chlorite porphyroblasts to the biotite, garnet, and staurolite porphyroblasts. They are of late growth and are later than the development of the slip cleavage. It is possible that they represent a retrograde development during falling temperature contemporaneous with the chlorite formed by the alteration of biotite, hornblende, staurolite, etc., or else they were formed contemporaneously with the other porphyroblasts during progressive metamorphism. Dennis (1956, p. 74) describes euhedral porphyroblasts of chlorite occurring ". . . randomly distributed over the whole (Lyndonville) quadrangle, but comparatively rare." Doll (1951, p. 71) also records large chlorite porphyroblasts in the Memphremagog

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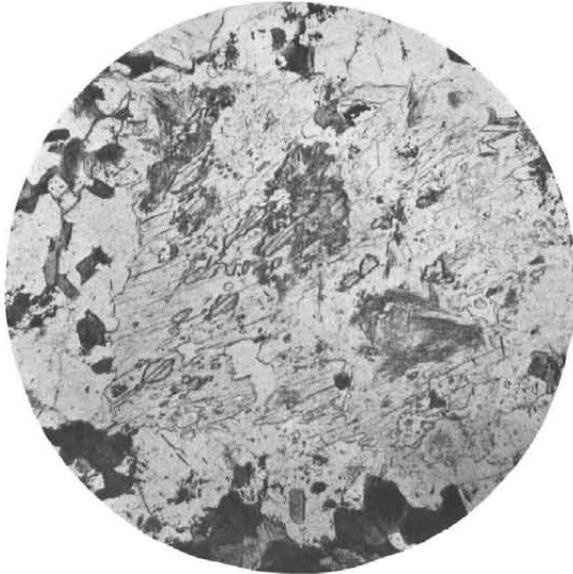
Figure 62. Photomicrograph (×44) of hornfels showing the large square cross-section of a "shimmer aggregate," composed of muscovite, with abundant felted fibrolite inclusions, and prisms of sillimanite; some biotite and quartz occur within the pseudomorph. The groundmass contains quartz, biotite, plagioclase, and iron ores. From Burke Mountain (H0653).

quadrangle. Porphyroblastic chlorite thus has a wide regional development in northeastern Vermont.

Tourmaline: Tourmaline is an occasional accessory mineral, some of which may be recrystallized detrital material. However, in some cases larger prisms occur up to 5 mm. long. Tourmaline has also been noted in the "shimmer aggregates," where it may indicate metasomatic introduction of boron. It is pleochroic brownish (e) to olive-green or bluish-green (o). Tourmalinized phyllite occurs adjacent to a granitic dike on the lower western slope of Kirby Mountain (G3627) and near pegmatitic material in the roadside outcrop north of Bundy School at D4053. Tourmaline is conspicuous with axinite in amphibolite at D3016; the tourmaline is pleochroic colorless (e) to blue and smoky gray (o).

"Shimmer aggregates": Perhaps the most characteristic macroscopic feature of many of the outcrops of the sillimanite zone, as well as part of the staurolite zone, is the presence of abundant large aggregates of

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Figure 63. Photomicrograph (×44) of hornfels showing abundant biotite, in part altered to flamboyant masses of fibrolite; abundant muscovite, with inclusions of fibrolite, biotite, and iron ores; xenoblastic staurolite (right-hand border); quartz; and sparse oligoclase. Garnet and coarser sillimanite occur in an unphotographed portion of the slide. From the col between Burke and Kirby mountains (H0843).

muscovite, called "shimmer aggregates" by Barrow (1893). In the Burke area the mica making up these is often quite large, certainly larger in size than many typical "shimmer aggregates" described by others. These aggregates weather out prominently on relict bedding or schistosity surfaces and have diverse orientations; although many lie with their long direction parallel with the surfaces, many others do not. The aggregates retain the shape of the replaced mineral, and range in size from 8 mm. to 40 mm. in length and 2 mm. to 7 mm. square in cross-section. Occasionally some stubbier masses occur, with pairs arranged in orientation similar to that of staurolite twins; such aggregates are replacing staurolite and are present more usually in the staurolite zone (e. g., at B1353), and rarely in the high-grade hornfels zone.

The form of the characteristic aggregates indicates that they are pseudomorphous after andalusite or sillimanite. In cross-section they often possess a dark core, caused by inclusions of opaques or biotite,

Figure 64. Photomicrograph (×18) showing an andalusite porphyroblast with inclusions of kyanite (top left), a kyanite porphyroblast (top right), and biotite, in part altered to fibrolite (right). The large area in the center is partially kaolinized oligoclase with inclusions of fibrolite; quartz with fibrolite inclusions is also present. Xenoblastic staurolite occurs elsewhere in the slide away from the quartz-oligoclase vein. Muscovite is abundant in the vein and elsewhere. From east of Route 114 (B1848).

which is strikingly similar in appearance to chiastolite. In thin section the aggregates are seen to be composed of randomly oriented muscovite flakes up to 0.7 mm. across. In one thin section (Fig. 62) from Burke Mountain (H0653) the aggregates also contain a little quartz, biotite, and sillimanite. The latter occurs as needles, felted fibers, and larger prisms up to 1.26 mm. long and 0.18 mm. square. In cross-sections of the pseudomorphs the sillimanite grains lie normal to the "c"-axis and the several separate grains of a pseudomorph extinguish together under crossed nicols; they are never seen to occupy the cores, but in some cases are in a discontinuous ring between the center and the periphery. In longitudinal sections of the pseudomorphs the sillimanite prisms are parallel to the "c"-axis, their ends splaying out into fibers penetrating the muscovite. The needles and fibrous masses occurring within the muscovite plates have a random orientation, and give the appearance of replacing the muscovite. Crushed "shimmer aggregates" from other

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Figure 65. Photomicrograph (×18) of staurolite-garnet schist, showing idioblastic staurolite and garnet with traces of relict wavy schistosity. Biotite porphyroblasts are abundant; chlorite porphyroblasts also occur (center). The groundmass is composed of quartz, sericite, and iron ores. From the Moose valley (F1519).

specimens reveal the presence of grains of tourmaline, staurolite, and rare andalusite, as well as sillimanite. The staurolite fragments contain matted anastomosing fibers believed to be sillimanite, which appear to be replacing the staurolite.

The pseudomorphs present some difficulties in determining their original identity. The section described above indicates that they were sillimanite initially; it is perhaps strange, however, for the felted fibrous variety to occur as well as the coarser prisms as relicts of a single porphyroblast. Furthermore, no unaltered sillimanite porphyroblasts as large as the usual pseudomorphs are found on Burke or Kirby Mountain, although some as long as 22 mm. are prominent in places. The presence of andalusite grains in some of the pseudomorphs suggests that they were originally that mineral. Andalusite, altering to sericite, occurs in a number of thin sections of hornfels, e. g., in the Moose River at J2147 and near the northern boundary of the Burke quadrangle at B1349 and B1850 (Fig. 60); the andalusite sometimes contains sillimanite needles. No fresh andalusite porphyroblasts are found on Burke or Kirby Moun-

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tain; fresh andalusite is prominent on the southern spur of East Mountain. The evidence suggests that the "shimmer aggregates" were originally in many cases andalusite crystals which were in the process of inversion to sillimanite, and the former were altered to muscovite by potash metasomatism which the sillimanite mainly resisted. Some of the pseudomorphs, however, are probably after sillimanite, as partially muscovitized sillimanite is present.

Muscovitization of andalusite and sillimanite is a common phenomenon. Billings (1938) describes such an alteration of sillimanite in western New Hampshire and suggests that it was caused by metasomatic introduction of potash derived from intrusions of the New Hampshire magma series. Heald (1950), too, describes replacement of sillimanite by muscovite in western New Hampshire. Billings (1937) also records the retrograde alteration of sillimanite by reaction with biotite (and water) to produce staurolite and muscovite. It is possible that the staurolite grains in the "shimmer aggregates" are the result of such a reaction, although the presence of included fibrous masses suggests that the staurolite is more likely relict in the process of altering to sillimanite.

Hornblende: Hornblende occurs abundantly mainly in amphibolite but also in thin amphibole quartzite bands. It is often poikiloblastic and xenoblastic, but sometimes develops large prismatic forms, and has a dark green to black color in hand specimen. Another characteristic occurrence is as feathery sprays on foliation planes. The pleochroism varies from $X =$ yellow, light green, or nearly colorless to $Z =$ blue-green, dark green, or light green. No detailed study has been made of the hornblende to determine whether its properties vary systematically with metamorphic grade. Intensely blue-green hornblende is found throughout the area in thin sections of amphibolite from the Waits River, Gile Mountain, and Albee formations, ranging from the upper biotite zone to the staurolite zone (see Eric and Dennis, 1958, p. 50). Hornblende often shows patchy alteration to chlorite and occasionally to biotite.

Tremolite: Porphyroblasts of tremolite are found sporadically in the impure calcareous rocks of the Waits River formation. In some areas it is abundant and characteristically weathers out on the rock surfaces in large, radial, brush-like aggregates. Light green actinolite takes its place in some exposures.

Diopside: Light green, granular diopside, colorless in thin section, is abundant in calc-silicate hornfels bands on Burke Mountain and adjacent to granite in the northwestern area of the Burke quadrangle. It is accompanied by tremolite or actinolite, grossularite, an epidote

mineral (xenoblastic, with ultra-blue interference colors, apparently mainly clinozoisite), and rarely vesuvianite.

Sphene: Sphene occurs sparsely in calc-silicate rocks and often abundantly in amphibolite. It is xenoblastic, "drop-like" in form, and frequently has rims of leucoxene. In a rare instance in an amphibolite sphene surrounds ilmenite.

Characteristic Mineral Assemblages

In pelitic and semi-pelitic rocks (see, also, Woodland, 1963):

Biotite zone	quartz-sericite-biotite-plagioclase ± groundmass chlorite (± chlorite porphyroblasts)
Garnet zone	quartz-sericite-biotite-garnet-plagioclase ± chlorite
Staurolite zone	quartz-sericite-biotite-garnet-staurolite ± chlorite
	quartz-sericite-biotite-staurolite ± chlorite
	quartz-chlorite-sericite-garnet-staurolite
Sillimanite zone	quartz-biotite-plagioclase-muscovite-garnet ± sillimanite
	quartz-biotite-muscovite-plagioclase-staurolite-sillimanite-garnet
	quartz-biotite-plagioclase-staurolite-sillimanite-garnet
	quartz-biotite-muscovite-plagioclase-staurolite-andalusite-sillimanite
	quartz-biotite-muscovite-plagioclase-andalusite-sillimanite ± garnet
	quartz-biotite-plagioclase-staurolite-andalusite-sillimanite
	quartz-biotite-muscovite-andalusite

Kyanite is present in a few localities; in one specimen it occurs with andalusite and fibrolite. Accessory minerals include tourmaline, apatite, sphene, and opaques.

In quartzose rocks: (i. e., those mainly composed of quartz in all zones and which grade into semi-pelitic rocks upon an increase of mica):

Biotite zone	quartz-biotite-sericite-plagioclase (albite) ± chlorite ± calcite
	quartz-biotite-sericite
Garnet zone	quartz-biotite-sericite-garnet-plagioclase (albite-oligoclase or oligoclase)
	quartz-biotite-sericite-oligoclase

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	quartz-biotite-chlorite-garnet
	quartz-biotite-sericite-chlorite-garnet-calcite
Staurolite zone	quartz-biotite-sericite-plagioclase ± garnet ± staurolite
Sillimanite zone	quartz-biotite-muscovite-sillimanite
	quartz-biotite-plagioclase-microcline
	quartz-biotite-plagioclase-garnet
	quartz-biotite-muscovite-plagioclase-sillimanite

In amphibole quartzite (light-colored variety):

Garnet zone	quartz-hornblende-plagioclase (oligoclase-andesine)-garnet-biotite-chlorite ± calcite
	quartz-hornblende-plagioclase (oligoclase-andesine)-biotite-chlorite ± calcite
	quartz-hornblende-plagioclase-garnet-calcite

The amphibole varies from very pale green (Z) to green and to blue-green in different specimens, and is often partly altered to chlorite and calcite. It is believed to be aluminous hornblende, but may be actinolitic in some cases. The biotite is sometimes partly altered to chlorite. Chlorite (often var. penninite) is usually seen to be pseudomorphous after amphibole or biotite. The plagioclase is invariably very fine grained (less than 0.05 mm.) and almost entirely untwinned; its refractive index is higher than that of quartz. Accessory minerals include apatite, sphene, and iron ores.

In amphibole quartzite (dark-colored variety):

Garnet zone	quartz-hornblende-garnet-plagioclase-chlorite-clinozoisite
	quartz-hornblende-garnet-plagioclase-biotite-chlorite

The hornblende has Z = blue-green or green, but is often bleached with patchy polarization and prominent pleochroic halos around inclusions; the alteration is possibly metamictization by the inclusions. Accessory minerals include apatite, sphene, and abundant opaques, including sulfides and ilmenite.

In amphibolite:

Biotite zone	hornblende-plagioclase-biotite-quartz ± calcite ± chlorite
	hornblende-plagioclase-quartz-clinozoisite-axinite-tourmaline-calcite
Garnet zone	hornblende-plagioclase-biotite-quartz-clinozoisite-calcite ± chlorite
	hornblende-plagioclase-biotite-quartz ± chlorite

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	hornblende-plagioclase-garnet-quartz ± chlorite
	hornblende-plagioclase-biotite-garnet-quartz ± chlorite
Staurolite zone	hornblende-plagioclase-quartz ± chlorite
	hornblende-plagioclase-biotite ± chlorite
	hornblende-plagioclase-garnet-quartz
Sillimanite zone	hornblende-biotite-plagioclase-quartz

Hornblende forms the major part of the rock and ranges in color from green to bright blue-green (Z). The plagioclase varies in composition from oligoclase-andesine to labradorite; it is generally fine grained and untwinned. Biotite is rare to common in amount and is often seen to be forming from hornblende. The chlorite is apparently always retrograde, forming after hornblende particularly. Axinite and tourmaline occur in only one locality, at D3016, in the Waits River formation; they are abundant and represent metasomatic introduction of boron. Accessory minerals are apatite, sphene (altering to leucoxene), epidote, ilmenite, and sulfides.

In impure limestone:

Biotite and garnet zones	calcite (dolomite)-quartz-phlogopite-sericite-chlorite
	calcite-quartz-tremolite and/or actinolite-phlogopite-sericite-chlorite
	calcite-tremolite-quartz-chlorite-sphene
Staurolite and sillimanite zones	(interpreted as representing the hornblende hornfels facies of contact metamorphism in thin bands of the Gile Mountain formation on Burke Mountain and in the transition zone of the Waits River-Gile Mountain formation)
	diopside-grossularite-tremolite-labradorite-calcite-quartz
	tremolite-actinolite-diopside-microcline-clinozoisite-calcite-quartz
	actinolite-grossularite-vesuvianite-sphene-quartz
	diopside-grossularite-actinolite-hornblende-calcite-quartz-sphene
	tremolite-calcite-clinozoisite-microcline-plagioclase-diopside-quartz-sphene

In addition, the following assemblages occur adjacent to granite or in areas of granitic dikes in the Waits River formation in the northwestern part of the quadrangle:

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diopside-tremolite-calcite-clinozoisite-plagioclase-quartz-sphene
diopside-actinolite-oligoclase-quartz-calcite-clinozoisite-sphene
tremolite-microcline-andesine-quartz-phlogopite-calcite-clinozoisite
tremolite-quartz-microcline-plagioclase-clinozoisite-phlogopite-
    muscovite-sphene

Scapolite is present in a calc-silicate rock adjacent to granite in the north (B1850), and vesuvianite is found in the skarn rocks of this area.

Discussion

Textures and Metamorphic Grade

The texture and mineralogic composition of much of the pelitic, of the semi-pelitic, and, to a lesser extent, of the quartzose rocks of the area strongly suggest that the metamorphism took place in two main stages separated by a deformational episode. These events were probably not unrelated, but represented a continuous sequence and may have even overlapped in time in different areas. The groundmass of the pelites in the biotite, garnet, and staurolite zones is fine grained and shows a well-developed schistose structure, the mineral assemblage remaining essentially similar, e. g., quartz-sericite-biotite. The schistosity is disrupted by a slip cleavage, and in places nearly obliterated by its development. Porphyroblasts of biotite, garnet, staurolite, and chlorite have grown in the rock cutting both the schistosity and cleavage, which are frequently seen as relicts traced by inclusions in the porphyroblasts (Fig. 65). In some cases biotite, in addition to unoriented porphyroblasts, has formed parallel to the later slip or fracture cleavage presumably by post-cleavage crystallization controlled by structural surfaces. However, rarely, a biotite porphyroblast may be deformed by the slip cleavage. Some garnets show extensive fracturing, which occurs rarely in parallel planes at a high angle to the schistosity and which may also have been produced during the later movements. Doll (1951, p. 73) describes fractured and broken staurolite from the Memphremagog quadrangle. Normally the porphyroblasts show no evidence of deformation and are not granulated, which indicates that they were formed during post-cleavage crystallization.

The grade of the initial metamorphism is indicated by the groundmass of the porphyroblast-bearing pelitic and semi-pelitic rocks in all but the high-grade hornfels. (The later recrystallization of the latter has been so extensive as to obliterate or greatly change the early metamorphic assemblages.) Biotite appears to have been a constituent of the pelites

throughout the area, and it is possible garnet may have been present in places; generally, however, the rocks seem to have been in the biotite zone of progressive metamorphism, which places them in the greenschist facies of regional metamorphism (quartz-albite-epidote-biotite subfacies), and perhaps locally in the garnet zone (quartz-albite-epidote-almandine subfacies).

The impure calcareous rocks of the Waits River formation carry tremolite (actinolite) porphyroblasts sporadically throughout, as a result of the reaction between dolomite and quartz. The calcite-quartz-tremolite assemblage is stable over a wide range of conditions—the whole greenschist facies, including chlorite, biotite, and low to medium garnet zones (Turner, in Fyfe, Turner, and Verhoogen, 1958). The accompanying pelites appear to be mainly in the biotite zone, although the garnet zone is also present, and higher grades with garnet, staurolite, kyanite, and sillimanite occur locally near the large Newark granite mass when the impure calcareous rocks carry diopside. It is not possible to determine whether all or part of the tremolite appeared in the early metamorphism or at the time of the development of the porphyroblasts in the pelites of the Gile Mountain formation. Certainly the knob-like tremolite-quartz aggregates which characterize some of the limestone outcrops show no evidence of distortion, even though the rock has been strongly deformed into flowage folds with concurrent recrystallization. This indicates that at least these porphyroblast occurrences are post-deformational.

The growth of porphyroblasts during the later metamorphism resulted in the assemblages of the biotite, garnet, and staurolite zones, which are characteristic of the developments described from many normal regionally metamorphosed areas (Turner, in Fyfe, Turner, and Verhoogen, 1958). One notable feature is that the groundmass of the rocks in these three zones shows little evidence of recrystallization and the grain size remains small; the minerals are, however, the ones to be expected in equilibrium assemblages in these zones.

One exception is a chlorite schist which is exposed adjacent to amphibolite at G0931. This is the sole outcrop of this type found within the Burke quadrangle, but L. M. Hall (1959, p. 24) reports garnet-chlorite schist at the top of the Standing Pond amphibolite in the St. Johnsbury quadrangle. The schist (Fig. 66) is composed of fine laths of chlorite forming a perfect schistosity, with interstitial quartz and in places much sericite; iron ores occur throughout aligned in the schistosity. Rare green biotite also occurs. The chlorite flakes are bent into microfolds

Figure 66. Photomicrograph (×18) of staurolite-garnet-chlorite schist, showing poikiloblastic garnet (right) associated with poikiloblastic staurolite (below garnet), which is surrounded by muscovite aggregates and chlorite porphyroblasts. The ground-mass is composed of chlorite, quartz, and sericite aligned into a wavy schistosity; rare green biotite and tourmaline are also present. Note "halo" around garnet and staurolite poikiloblasts where groundmass chlorite is almost absent. From northeast of Lyndonville (G0930).

which produce a false cleavage parallel to the short limbs or to the limbs of sharp sigmoidal folds. This cleavage is developed imperfectly at irregularly spaced intervals and often fails to cross the width of the thin section. Sericite and opaques are, in part, aligned in the false cleavage. Interrupting both schistosity and false cleavage are aggregates of sericite and quartz with a rim of large twinned chlorite; the aggregates have no apparent orientation and lines of opaques parallel to the false cleavage cross them. Around the aggregates the groundmass is almost devoid of chlorite and is composed of quartz; this produces a halo-like effect. Large porphyroblasts of garnet (up to 6 mm. across) also interrupt the schistosity and cleavage; they have a very spongy peripheral zone with abundant inclusions of quartz and some chlorite. Staurolite is often associated with garnet. Sometimes staurolite twins penetrate to the center of garnet from its periphery; or an individual crystal may be

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enclosed in garnet. The staurolite always has a rim of sericite alteration; large chlorite flakes occur around the periphery of both staurolite and garnet. The garnet and staurolite also have a halo-like zone (0.2 to 1.4 mm. wide) poor in groundmass chlorite surrounding them. Some green tourmaline is present. The sericite aggregates are probably pseudomorphous after staurolite.

The paragenesis appears to have been: (1) chlorite schist formed during early deformation, with possibly the development of some garnets (now the inclusion-free cores of the porphyroblasts), (2) false cleavage produced, (3) further garnet growth, with staurolite porphyroblasts then growing in the garnet, and (4) partial retrograde alteration of staurolite to sericite and chlorite and slight alteration of garnet to chlorite. It is difficult to explain why the groundmass remained chloritic during stage (3) unless the rock was potash-deficient or the chlorite was originally alumina-rich. High alumina content stabilizes chlorite into higher metamorphic grades (Harker, 1939, p. 216; Ramberg, 1952, pp. 59, 141). The staurolite and garnet do not appear to be relict in a retrograde chlorite schist. The chlorite schist may be the result of differential shearing stress which affected the margins of the more resistant amphibolite mass more strongly than elsewhere during the early regional metamorphism and formation of schistosity. The porphyroblasts developed during the post-cleavage metamorphism, which was caused particularly by a rise in temperature.

The siliceous phyllitic marble of the Waits River formation, excluding the calc-silicate rocks, has a very similar texture throughout the Burke area except for the erratic appearance, both in quantity and size, of tremolite (actinolite), which is probably related more to varied original composition than to varied metamorphic conditions. Because of its mineral composition the impure marble probably suffered much greater recrystallization during the later movements, which caused considerable internal deformation, as is evidenced by the prevalence of flowage folds, the majority of which are believed to belong to this stage.

Amphibolite also has a similar texture throughout the quadrangle with the exceptions of: (1) the sheared and mylonitized schist in the southwestern part, (2) the granoblastic outcrops in the northern part, which probably represent small gabbroic intrusions that have retained their texture because of their size and shape, and (3) the coarser textured outcrop at B1855 which is adjacent to calc-silicate rock and close to granite. No petrographic evidence has been obtained to determine whether the amphibolite was altered from a lower grade during the later

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phase of metamorphism but it probably was. The amphibolite appears to have a broadly similar mineralogy throughout the metamorphic grades; even in the biotite zone there is an absence of chlorite (other than that pseudomorphous after hornblende), epidote minerals, and albite. (It is not possible to determine whether any of the amphiboles are aluminum-poor without chemical analysis). In this respect they may be compared with similar amphibolites of Banffshire, Scotland (Read, 1923, and Wiseman, 1934). Sutton and Watson (1951) suggest that the abnormalities of the Banffshire rocks, where the pelitic rocks also do not conform to the normal zones of regional metamorphism (see below), are the result of temperature control rather than of deficient stress. This may also apply to the Burke area, where temperatures relative to depth of burial were probably higher than normal because of the rise of large subjacent igneous masses. The mylonitized amphibolite with a chlorite-actinolite-? talc assemblage is characteristic of the greenschist facies, produced in this case by shearing stress. If this was developed during either the main regional metamorphism or the later slip cleavage phase, there is no evidence of recrystallization that might have been caused during the post-cleavage (porphyroblastic) metamorphism.

METAMORPHISM IN THE HORNFELS

The highest grade zone of andalusite and sillimanite has some distinct characteristics. The distribution of this zone is closely related to the occurrence of granitic rocks, and much of it is incorporated within a complex of granite and hornfels. The hornfels has a distinctive appearance, although varied, according to the original rock composition and to the extent of metamorphism and metasomatism. Banding caused by alternating beds of contrasting lithology is frequently well preserved, as is the schistosity parallel to it and a cross-cutting slip cleavage. The bedding foliation of hornfels masses may be regular but usually it is highly variable, folded, and contorted; where granitic patches and tongues occur throughout a mass the relict foliation of the hornfels in one part may bear no apparent relation to that in other parts. Occasionally an outcrop shows the rock to be highly contorted and broken up, with the fractured segments wildly disarranged (e. g., Fig. 21). The pre-existing structural surfaces within the complex are thus drastically altered, but in no definable geometric fashion. (The stereogram, Plate 5, Figure 4, of poles to bedding foliation for the Burke Mountain-Kirby Mountain-Umpire Mountain subarea is triclinic). The rocks are noticeably indurated and toughened compared to similar lithologies in the

lower metamorphic grades and, although the general grain size is still small (approximately 0.04 to 0.2 mm.), it is noticeably coarsened and in some cases it is very coarse grained, e. g., with very abundant white mica plates up to 12 mm. across. Adjacent to granitic occurrences the hornfels has often lost its schistose structure and has a "sugary" granulose appearance, particularly in the quartz-rich types. In part, these may fade gradationally into the normal hypidiomorphic granitic rock, although normally a sharp, but highly irregular, line can be drawn between the latter and the metamorphosed sediments. Hornfelsed quartzose phyllite has characteristically developed a purplish hue apparently due to the crystallization of numerous minute disseminated biotite flakes. The texture and appearance of the rock are typical of thermally metamorphosed rock within an aureole of an igneous intrusion.

As mentioned above, andalusite is present as fresh porphyroblasts in the vicinity of East Mountain and as relics in "shimmer aggregates"; it occurs along with staurolite and fibrous sillimanite. Both fibrous and coarser prismatic sillimanite is found with staurolite and garnet. Andalusite is normally associated with thermally metamorphosed aureoles in the hornblende hornfels facies (Turner, p. 207, in Fyfe, Turner, and Verhoogen, 1958), where it is often accompanied by sillimanite adjacent to granitic intrusions. Bosworth (1910) describes contact altered rocks of a granite impregnation zone around the Ross of Mull granite where sillimanite occurs as fibers and prisms, often in parallel intergrowths with andalusite. The sillimanite prisms within the andalusite behave almost as a single crystal. (See also Harker, 1939, Fig. 170B and 172B). This is identical with the interpretation given to some of the "shimmer aggregates" occurring in the granite-hornfels complex in the Burke area. Read (1927, p. 324) also describes sillimanite and andalusite in parallel intergrowths in pelitic schists of an injection complex in Cromar, Aberdeenshire, Scotland.

The presence of staurolite and garnet accompanying sillimanite is not uncommon; for example, Billings (1937, pp. 493, 551-552) refers to staurolite as occurring in the sillimanite zone as a retrograde mineral after sillimanite. In most cases in the Burke area the xenoblastic staurolite occurring in the same thin section (Fig. 67) as sillimanite appears to be in relict crystals, indicating incomplete reaction. It is possible that the staurolite grains that are present in the "shimmer aggregates" represent a retrograde product after sillimanite similar to the occurrence described

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Figure 67. Photomicrograph (×44) of staurolite-garnet-sillimanite hornfels, showing garnet (bottom left), staurolite (left and bottom right), and poorly-oriented biotite porphyroblasts in a groundmass of quartz, sericite, opaque dust, and scarce apatite and tourmaline. Fibrolite (right) occurs as tufted aggregates after biotite. From Victory (H3410).

by Billings; the evidence is not equivocal, however, and these, too, may be relict grains.

Not all of the staurolite of the aureole rocks is xenoblastic. In one thin section (Fig. 59) large porphyroblasts occur within a fine-grained hornfels-type groundmass, which is composed mainly of quartz, decussate biotite, and opaque minerals. The staurolite contains many inclusions, apparently of quartz and some opaques, arranged in a regular manner similar to that of chiastolite. The areas of the inclusions are interrupted by lath-shaped, inclusion-free areas which have a size and arrangement similar to the biotite outside of the porphyroblasts. It appears in this case that the staurolite developed late and grew in a rock already possessing a hornfelsic texture. In contrast with this, another thin section shows large staurolite porphyroblasts with inclusions of much finer grain size than the groundmass grains and arranged helicitically but with no relationship to the wavy schistosity of the groundmass. The

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staurolite thus grew before recrystallization of the groundmass and has been rotated so that its relict schistosity is no longer parallel with the existing schistosity. This rock contains poikiloblastic garnet, which may be of later origin than the staurolite, and fibrolite after biotite. The garnet of the hornfels is often quite granulated and sometimes shows evidence of rotation, as lines of inclusions are not parallel and do not lead into the surrounding schistosity.

The mutual occurrence of garnet, staurolite, andalusite, and sillimanite (fibrolite) is less common. It may represent polymetamorphism or a type of regional metamorphism described by Turner (in Fyfe, Turner, and Verhoogen, 1958, p. 211) as transitional between the hornblende hornfels facies and the almandine amphibolite facies or the Buchan type described by Read (1952, p. 278). It also fits into the low-pressure intermediate facies of Myashiro (1961). Read (1923, p. 59) describes the regional occurrence of andalusite schists in Aberdeenshire and Banffshire, Scotland, in which staurolite, garnet, and sillimanite also occur, together with cordierite. Regional thermal metamorphism is the explanation given by Read, 1952, p. 278) while Turner (in Fyfe, Turner, and Verhoogen, 1958, p. 206) suggests high temperatures, developed at higher than normal levels, as the cause. Harker (1939, p. 233), however, writes:

"If we suppose, after the production of staurolite, a decided falling off of shearing stress, while the temperature still remained high, the conditions would seem suitable for the formation of andalusite. . . This, it should be observed, is quite different from a second metamorphism, as, e. g., when a staurolite-schist is invaded by a later granite intrusion."

Doll (1951, p. 73) also records the occurrence together of staurolite and andalusite in the Memphremagog area and suggests that the latter formed at a later time than the staurolite, i. e., when the temperature was falling.

In the Burke area the assemblages of the sillimanite-andalusite zone have a close spatial relation to the intrusion of granite and are most widely developed in the mixed granite-hornfels complex. The texture and mineral assemblage of these rocks suggest that high temperature was the important factor in their metamorphism, together with metasomatism by fluids emanating from the granite. The country rocks were already in an advanced stage of metamorphism, caused by a rise in temperature, which was produced by a subjacent mass of granitic magma; it is believed that the development of porphyroblasts was initiated during this stage. Granitic magma was then injected into the

rocks, and the effect of heat and solutions on the immediately adjacent strata was marked by their alteration to a hornfels, bearing andalusite and sillimanite. Rocks with andalusite and no sillimanite may perhaps represent zones little affected by metasomatism, while those with extensive porphyroblastic development of sillimanite may represent the most altered and soaked rocks. This explanation thus does not involve a separate metamorphic episode nor yet is it Read's Buchan type of metamorphism nor Harker's interpretation, although stress effects would decline as the rocks became more plastic as the result of injection. With reference to the facies concept, the sequence may be described as a culminating metamorphic phase, producing the hornblende hornfels facies on rocks already at high temperatures and already in the almandine amphibolite facies of regional metamorphism (see Woodland, 1963, for further discussion of the metamorphism of pelitic rocks).

One further minor comment may be made about sillimanite. It has been noted that both fibrolite and coarser prismatic varieties occur within the same thin section. It might be that these represent two stages of development. Harker (1939, p. 328) holds that a similar occurrence of fibrolite and prismatic sillimanite in hornfelsed schist enclosures in the Ross of Mull granite, Scotland, represents two stages, older fibrolite and newer sillimanite, although Bosworth (1910) and Bailey and Thomas (1925) regard both as belonging to the thermal metamorphism caused by the granite. No positive evidence of age difference has been obtained in the Burke area; the two varieties are probably more or less contemporaneous. Much of the fibrolite occurs as sheafs within large muscovite plates and some replaces biotite. The large increase in muscovite in some of the rocks is probably of late metasomatic origin, produced by circulating solutions emanating from the granite, and it is possible that some or all of the fibrolite may be of similar origin, as is suggested by Watson (1948) for fibrolite in migmatites of Kildonan, Scotland. G. J. Williams (1934, p. 338) in describing roof pendants of biotite-muscovite schist on Stewart Island, New Zealand, explains muscovite containing knotted bunches of fine sillimanite (Ibid, Fig. 2A), apparently identical to occurrences on Burke Mountain, as representing muscovitization of sillimanite, the latter having been formed by the loss of silica by endogenous secretion. Late muscovitization may also explain the fibrolite-muscovite association in the Burke area, but the evidence from thin sections is not definite on this point; however, it appears more likely that the sillimanite is contemporaneous with or perhaps replacing the muscovite.

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The occurrence of kyanite in the northern part of the Burke quadrangle is quite unusual. It is found as rare flakes in a few specimens and as large, beautiful crystals up to 8.7 cm. long at B1848. The latter occurrence is in a coarse pegmatitic lens associated with large pink, slightly pleochroic andalusite and with muscovite (Fig. 64). Some of the oligoclase crystals have been deformed and show strained lamellae, and the quartz has undulose extinction. Some pyrite is present. The kyanite is, in part, altered to muscovite; one crystalloblast in thin section is strained and fractured. The associated contorted host rock is biotite-rich. Close to the margins of the veins the biotite is altering to flamboyant masses of fibrolite which also penetrate the associated quartz and plagioclase. The biotite is often bleached and appears to be altering to muscovite with the release of much iron ore; rarely it is also partly altered to chlorite. Ilmenite and sphene are present, both much altered to leucoxene. Fine opaques trace a helicitic structure in the biotite marking a relict wavy schistosity. Xenoblastic staurolite with inclusions of quartz and opaques is a common constituent. Tourmaline and zircon occur as accessories. Some small rare kyanite has also been found at this locality and at B0347.

Dennis (1956) records kyanite associated with granite plutons (all occurrences within 1,500 feet of a granite contact) and suggests that it formed as a contact aureole mineral because of low water pressure and concentration in the pelites. However, kyanite is generally regarded as the high pressure polymorph of Al_2SiO_5 (Fyfe, Turner, and Verhoogen, 1958, p. 165; Clark, et al, 1957), although the stability relations of the three polymorphs (kyanite, andalusite, and sillimanite) are not known for certain. The three have been recorded in the same rock but have not been regarded as contemporaneous (Barth, 1936, p. 790), and Hietanen (1956) suggests that their mutual occurrence in cordierite-mica schist is due to fluctuation of temperatures and stress. Bosworth (1910) describes kyanite on the Island of Mull, Scotland, occurring in pegmatite and passing into tourmaline-bearing kyanite gneiss and also, in another occurrence, kyanite associated with large garnet, staurolite, and fibrolite enveloped in muscovite and fringing kyanite. In the latter area Clough (1911) reports that Flett also detected andalusite besides sillimanite and staurolite, while MacKenzie (1949) states that this kyanite-bearing band extends into the aureole of the Ross of Mull granite where kyanite is transformed into andalusite, particularly in quartz-rich segregations. Read (1934) records kyanite with chloritoid produced from andalusite and staurolite by a later metamorphism, and

Tilley (1935) suggests that the kyanite which occurs in part as replacement of andalusite and cordierite in the hornfels zone of the Carn Chuinneag granite, Ross-shire, Scotland, has been formed during the initial stages of the dynamic metamorphism of the hornfels rocks. Barth (1936, p. 789) describes the occurrence of needle kyanite in shear zones in schists which carry porphyroblasts of staurolite and kyanite, and states that it ". . . crystallized directly from a solution, and apparently in this case did not depend on the action of shearing stress for its formation."

The occurrence of kyanite in the Burke quadrangle is very restricted and must represent specialized local conditions of formation. It appears, however, to be of late development, and the main large porphyroblastic example is associated with and presumably contemporaneous with pegmatitic material. The andalusite is slightly altering to muscovite and is in part intergrown with kyanite, but it is not possible to determine whether one mineral is replacing the other. The fibrolite is after biotite and seems to be directly related to the occurrence of the pegmatitic vein, although biotite altering to fibrolite is surrounded by andalusite in one instance. Clifford (1958) suggests that in the relatively low pressure field andalusite replaces kyanite with increasing temperature. It is possible that the andalusite and kyanite in this occurrence crystallized from aluminous-rich fluids under pressure and temperature conditions at or near the triple point for the three polymorphs, and the fibrolite may have formed contemporaneously from biotite while other biotite close by was muscovitized (see Woodland, 1963).

MUSCOVITIZATION

In many places in and adjacent to the granite there is evidence of late potash metasomatism of the hornfels. This is demonstrated by certain occurrences of abundant large muscovite plates (often containing fibrolite) and also by the extensive muscovitization of the andalusite, sillimanite, and staurolite porphyroblasts. The presence of very abundant biotite in an amphibolite of the hornfels zone may be due, too, to potash metasomatism; a rock from Hobart Ridge (I0941) composed of abundant masses of biotite, masses of chlorite presumably after biotite, plagioclase, and quartz may be another metasomatised amphibolite. Muscovitization of aluminum silicates is a common phenomenon; Billings (1938) describes such an alteration of sillimanite in western New Hampshire which he shows to have been caused by the action of potash-bearing solutions which he suggests were derived from nearby granite plutons.

Heald (1950) describes a similar effect in the Lovewell Mountain quadrangle, New Hampshire. In the Burke quadrangle the source of the potash-bearing solutions could have been the residual fluids of the adjacent granitic rocks.

SUMMARY OF THE GEOLOGIC HISTORY OF THE BURKE QUANDRANGLE

Sediments of Ordovician age, represented by the Albee formation in the eastern part of the Burke quadrangle, were laid down in what Kay (1951) has termed the Magog eugeosyncline. These were presumably affected by the Taconic orogeny. Further subsidence of the geosynclinal belt during Silurian and early Devonian time was accompanied by deposition of a considerable thickness of sediments, which, when indurated, included arenaceous and argillaceous limestone, shale, silty shale, quartz siltstone, and rocks grading from quartz wacke to quartz arenite. The calcareous rocks with interbeds of shale and arenite now comprise the Waits River formation, which crops out over the western portion of the quadrangle and extends westwards into the Lyndonville quadrangle. The Gile Mountain formation, which occurs in a wide belt east of the Waits River formation, was laid down as sediments which became indurated into a series of alternating thin bands of shale, siltstone, quartz wacke, and quartz arenite—a lithology characteristic of a rapidly sinking eugeosynclinal belt. The alternation of shale and arenite suggests a thick sequence of mainly graded beds. The calcareous rocks of the Waits River and the non-calcareous sequences in the Gile Mountain are probably equivalent, at least in part, with the former being deposited to the west at the edge of the eugeosyncline (perhaps representing a miogeosynclinal environment), and the latter to the east in the eugeosyncline proper. The exact relations between the two are open to doubt; the Waits River is believed to be older than the Gile Mountain by most workers in northeastern Vermont, while this author considers, on the basis of an alternative structural interpretation, that it may be, in part, younger.

Volcanism was sporadic, and is represented by lava and tuff, both acid and basic, which occur intermittently throughout the Gile Mountain strata and particularly in the transition zone between the Waits River and Gile Mountain formations. Some intrusive activity, resulting in mainly basic dikes and plugs, accompanied the volcanism.

In the Middle Devonian subsidence ceased and the area was subjected to considerable deformation during the Acadian orogeny. The rocks were

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strongly folded and underwent low-grade metamorphism (green-schist facies) during the early phase of the orogeny. The movements produced a well-developed schistosity in the rocks, and it is probable that recumbent folding took place as the rocks deformed under the influence of stresses which apparently operated in an east-southeast-west-northwest direction. Continuation of the movement then caused deformation of the earlier-formed schistosity, which was folded and cut by a newly-developed cleavage (slip cleavage) that almost or completely obliterated the schistosity in the more strongly deformed areas. The Willoughby arch structure resulted at this stage from a linear culmination zone.

Meanwhile, the temperature of the rocks of the area was increasing and towards the end of the movements the higher temperatures, accompanied probably by the rise of large subjacent masses of granitic magma, resulted in the thermal metamorphism of the rocks at what was probably a moderate depth in the earth's crust. Porphyroblasts of biotite, garnet, staurolite, tremolite, and actinolite grew at this stage, and cut both the schistosity and the cleavage. Granitic magma then broke into the structures, partly by following their structural surfaces and forcing the country rock aside and partly by wedging and intimate penetration of the rock. Adjacent to the granite the rocks were converted to hornfels, with the production of diopside, tremolite, and grossularite in the calc-silicate rocks developed from impure limestone and with the production of andalusite and sillimanite in the banded hornfels developed from the pelitic and semi-pelitic rocks. Sillimanite was abundantly produced in the intimately injected zones, now comprising the hornfels-granite complexes, where there is also evidence of the activity of fluids in the formation of migmatite and granitized rocks. Muscovitization was a late stage in events and led to the development of much muscovite in the hornfels and to the replacement of staurolite, andalusite, and, to a lesser extent, sillimanite by "shimmer aggregates" of mica.

As the rocks cooled, some retrograde metamorphic effects were produced; chlorite replaced hornblende, biotite, garnet, and staurolite and also appeared conspicuously as porphyroblasts in the pelitic and semi-pelitic rocks. At some later date, in the Mesozoic Era, basic dikes, mainly lamprophyres, were intruded into both the granitic and metamorphic rocks. Since that time, as far as it is known, the area has been subjected to continuous erosion, with periodic uplift enabling the removal of many thousands of feet of rock. The present configuration of the land surface broadly reflects the relative resistance of the various rocks to weathering and erosion. During the Pleistocene the area was completely

covered by ice, and since the withdrawal of the ice weathering and erosion have continued.

ECONOMIC GEOLOGY

The economic deposits of the Burke quadrangle are granite and sand and gravel. Granite has been quarried in the past on a small scale in Newark, e. g., at the northern end of Packer Mountain and northeast of Walker Mountain, and on Kirby Mountain, near the southern border of the quadrangle. There are many opportunities for granite quarrying; the most promising areas are in Newark, particularly east and west of Route 114. An attractive granite, some of it pink in color and within easy access of the highway, occurs west of Route 114 and north of East Newark School, in the slopes leading up to Walker Mountain and Hawk Rock.

Sand and gravel are obtained from eskers, kames, and late-glacial deposits in the valleys of the East and West branches of the Passumpsic River. Other sand and gravel deposits are found near Newark Hollow, Bean School, south of Umpire Mountain, and in the mounds throughout the low-lying area of Victory.

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PLATE I

VERMONT
WOODSTOCK QUADRANGLE
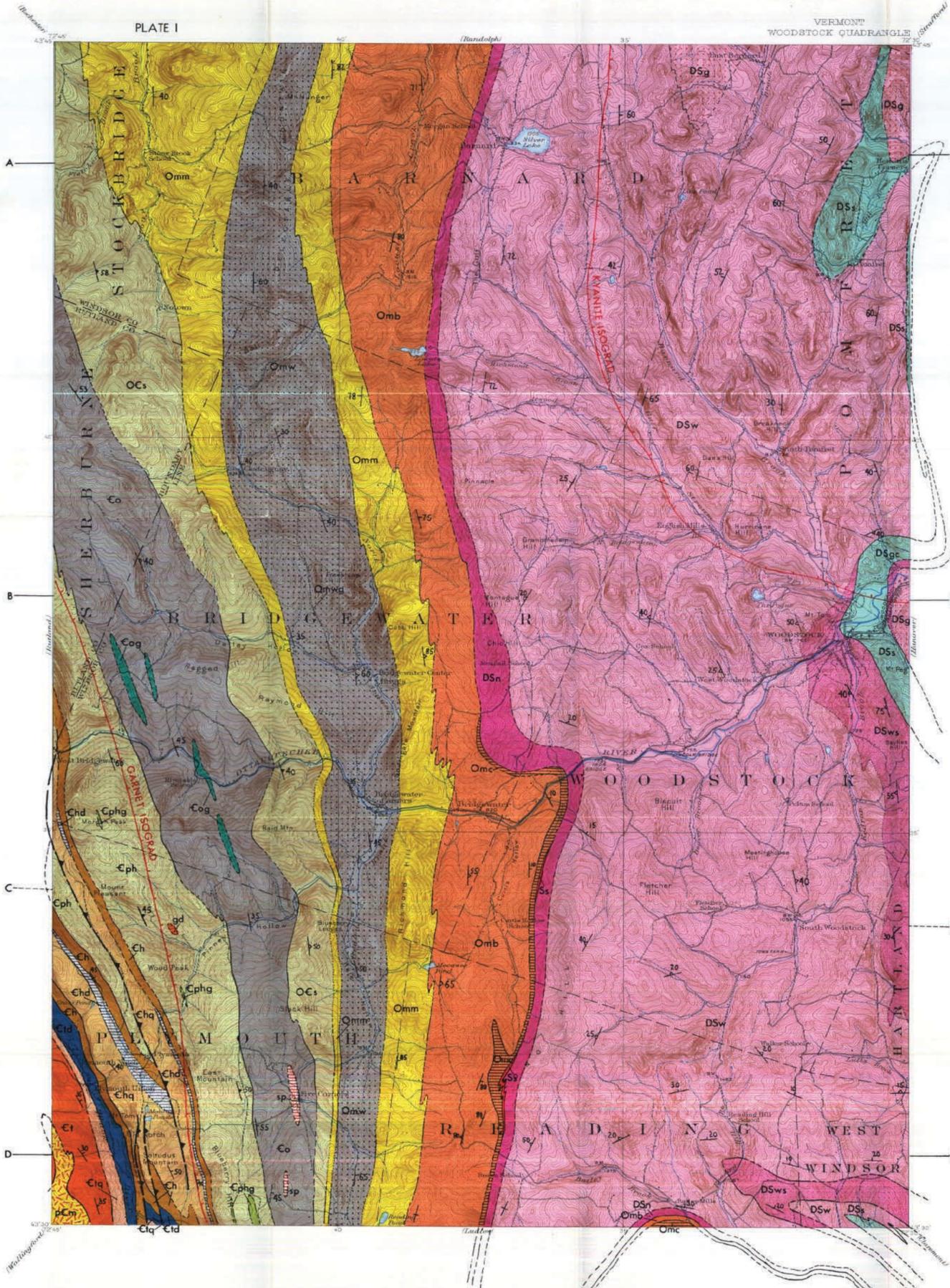
LEGEND

IGNEOUS ROCKS

DEVONIAN ? — gd — Granodiorite

ORDOVICIAN ? — sp — Serpentine, talc-carbonate rock

STRATIFIED ROCKS

SILURIAN OR DEVONIAN

DSg / DSgc — Gile Mountain formation: Gray mica schist with biotite and garnet, schistose quartzite, minor impure limestone; where limestones abundant, DSgc.

DSs — Standing Pond volcanics: Hornblende schist, amphibolite, hornblende garbenschiefer, locally with garnets two inches or more in diameter.

DSws / DSw — Waits River formation: Impure limestone or marble weathering to a brown punky crust. Minor mica schist with biotite and garnet, some quartzite. Where schist and quartzite abundant, DSws.

DSn — Northfield formation: Dark gray schist or phyllite with abundant garnet and biotite, minor impure limestone and quartzite.

SILURIAN

Ss — Shaw Mountain formation: Quartz conglomerate, quartzite, quartz-mica schist.

MIDDLE ORDOVICIAN

Omc / Omb / Omm / Omw / Omwg / Omm — Missisquoi formation: Cram Hill member, Omc, is rusty weathering carbonaceous schist. Barnard volcanic member, Omb, is biotite gneiss, hornblende gneiss, amphibolite. Whetstone Hill member, Omw, is mica schist or phyllite, locally carbonaceous, with amphibolite and volcanic conglomerate in upper part, and a thin garnetiferous quartzite, Omwg. Moretown member, Omm is mainly schistose quartzite in which granulitic layers averaging about half an inch thick are separated by paper-thin schistose laminae.

ORDOVICIAN OR CAMBRIAN

O€s — Stowe formation: Green quartz-sericite-chlorite schist; garnet and biotite abundant locally.

CAMBRIAN

€o / €og — Ottauquechee formation: Black, rusty-weathering phyllite interbedded with quartzite containing carbonaceous layers; minor green phyllite; biotite and garnet abundant locally. Greenstone schist, actinolitic greenstone, €og.

€ph / €phg — Pinney Hollow formation: Green quartz-sericite-chlorite phyllite, chloritoid abundant near base, garnet and biotite occur locally. Greenstone schist, €phg.

€h / €hd / €hq / €h — Hoosac formation: Quartz-albite-muscovite-biotite schist, carbonaceous phyllite near top. Quartzites abundant, €hq; dolomite and dolomite breccia, €hd.

€td / €tq / €t — Tyson formation: Schistose graywacke, minor conglomerate, €t; orthoquartzite with interbeds of carbonaceous pyritic phyllite, €tq; dolomite, €td.

PRECAMBRIAN

p€m — Mount Holly complex: Chloritic biotite gneiss, pegmatite.

METAMORPHIC ZONES

GARNET — Isograd, hachures on high grade side

CONTACTS

——— Good control.
– – – Fair control.
········ Inferred.
⊥⊥⊥ Fault, sawteeth on overthrust side.

SYMBOLS

40 — Strike and dip of bedding
+ — Horizontal bedding
35 — Strike and dip of schistosity
55 — Strike and dip of schistosity or bedding (where not differentiated)

KEY TO MAPPING
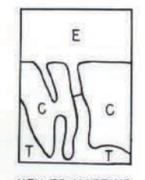
A – A′ (cross section)
B – B′ (cross section)
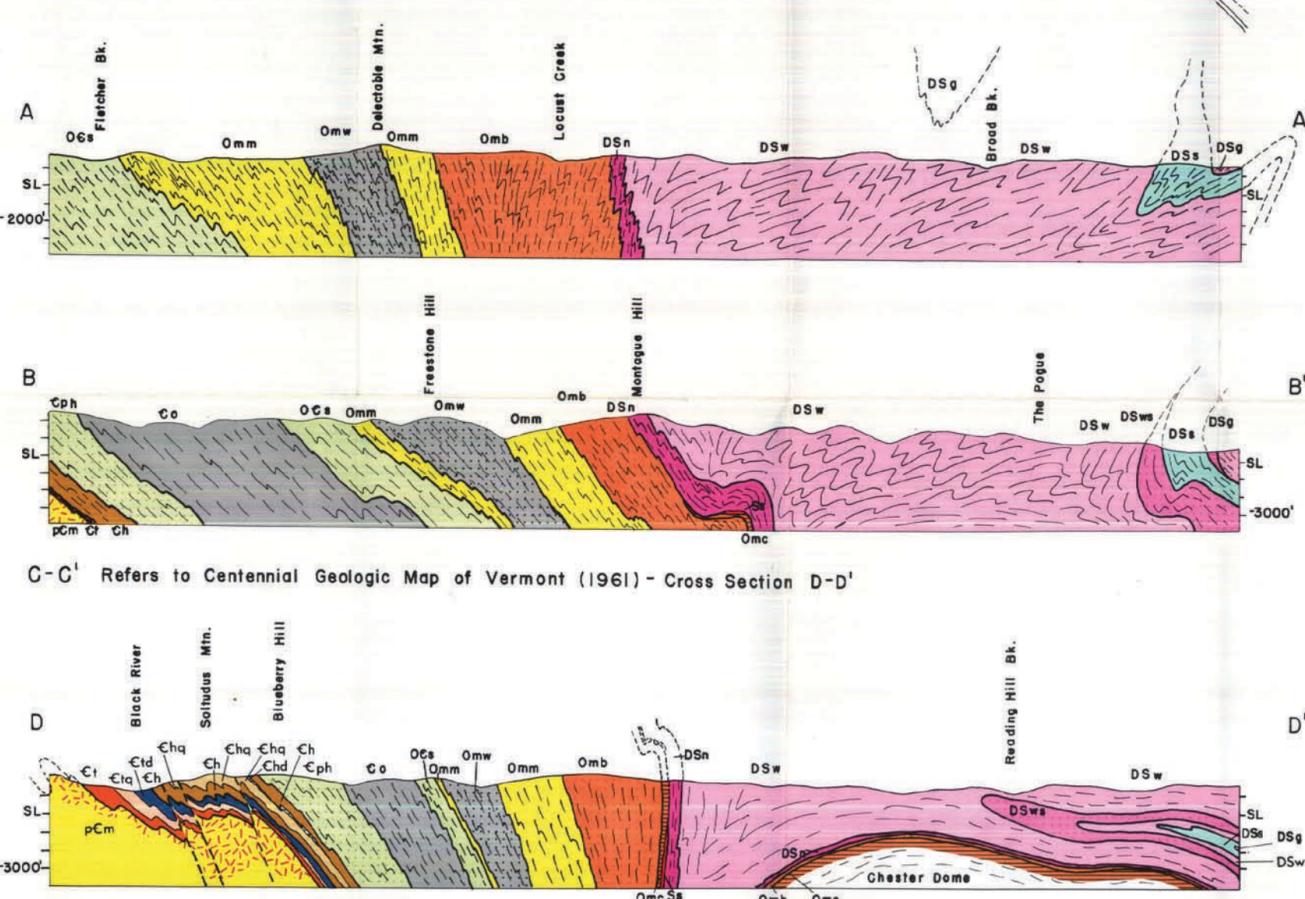
C–C′ Refers to Centennial Geologic Map of Vermont (1961) – Cross Section D–D′

D – D′ (cross section)

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE WOODSTOCK QUADRANGLE, VERMONT

PLATE 2

LEGEND

STRUCTURAL SYMBOLS

³¹⌐ Strike and dip of bedding

⁴²⊢ Strike and dip of bedding where steeper than cleavage

⁵⁶◀ Strike and dip of foliation or schistosity

⁸¹◀ Schistosity and / or bedding (undifferentiated by Chang)

⊥ Strike of vertical bedding

+ Schistosity horizontal

⁷⁰⌐ Slip Cleavage

²³↑ Bearing and plunge of linear element
  c – crinkle
  f – fold axis
  p – pebble elongation

↕ Lineation horizontal

ᡇ Pattern of minor folds

Fold axes

—+— Early (synclinal)

—+—⊢—+— Late (anticlinal)

⌂ Isoclinal (∪ points in direction of closure)

⊢ Thrust fault (dashed where inferred)

----- Geologic boundary

Formation symbols as shown on geologic map

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 29)

0   1   2   3   4   5 miles

TECTONIC MAP OF THE WOODSTOCK QUADRANGLE,
VERMONT