THE GEOLOGY OF THE LYNDONVILLE AREA, VERMONT

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Lake Willoughby, seen from its north shore.

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ABSTRACT

The Lyndonville quadrangle is situated in northeastern Vermont, between the Green Mountain anticlinorium and the New Hampshire plutonic belt.

Two sedimentary formations crop out within the quadrangle, the Waits River formation (Barton River member) and the Gile Mountain formation. Both belong to the St. Francis group which extends from Quebec into Massachusetts and which is here tentatively assigned Devonian age. The rocks consist of an alternation of graywackes, quartzites, siliceous limestones and volcanics, in which are emplaced a number of cross-cutting "granite" plutons.

The Gile Mountain formation lies at the core of a syncline, the Brownington syncline, which is flanked on its east by an arch-like structure, the Willoughby arch. Two phases of deformation are recorded in the rocks: an early one marked by a sericite schistosity essentially parallel to the bedding and isoclinal drag folds; and a later one dipping away from the crest of the arch, marked by slip cleavage in some rocks and schistosity in others, and by minor folds which face away from the crest of the arch (with certain exceptions). There is evidence for thinning of the limestones at the crest of the arch, and thickening on its eastern flank. Considered in conjunction with the presence of gneiss domes to the south along the same trend, these structural features lead the writer to interpret the Willoughby arch as a domal structure caused by the diapir-like rise of a subjacent mass. The arch-like pattern of the slip-cleavage and schistosity is not interpreted as deformation of a pre-existing horizontal structure but as a flowage phenomenon initiated by the doming.

The cleavage and schistosity are a structural "reference grid" used to distinguish at least two phases of metamorphism: an early sericite phase and a later garnet-staurolite phase. The loss of water in the early low-grade phase caused unusual textures in the later, higher grade phase.

The "granites" have the composition of granodiorites. They show intrusive relationships toward the country rock, but some of the evidence can be interpreted in favor of replacement.

The two phases of structural and metamorphic evolution recorded in the rocks may represent two distinct cycles, or stages of an essentially continuous history at the time of the Shickshockian orogeny.

INTRODUCTION

Location

The Lyndonville quadrangle is situated between latitudes 44°30′ and 44°45′ and longitudes 72°00′ and 72°15′. Its northern limit is less than 20 miles south of the Canadian border, and its southeastern corner is some 13 miles northwest of the New Hampshire state line at Littleton dam. U.S. Route 5, Vermont Routes 5A and 12, and the Canadian Pacific Railway's main line from White River Junction to Montreal traverse the area.

Geologic Setting

The Lyndonville quadrangle lies in a belt of Lower Paleozoic metamorphic rocks occupying the Connecticut River depression between the Green Mountains and the White Mountains. Only two stratigraphic units occur within the quadrangle: the Barton River member of the Waits River formation, and the lower part of the Gile Mountain formation. The Waits River formation includes a lens of Standing Pond lava. Two granodiorite plutons and numerous related dike rocks occur within the area; these are part of a plutonic belt extending from Massachusetts into the Eastern Townships of Quebec. Structurally this belt is characterized by an alignment of domes and arches (cf. Fig. 3) flanked on its west by a syncline.

Regional geology is outlined on Plate 2. Correlations in various neighboring areas, by different authors, are shown on Pl. 4.

Previous Work

Early work in the region, on a reconnaissance scale, was carried out by Adams (1845), Hitchcock (1861), and Richardson (1902 and 1906). Detailed work in adjacent quadrangles is confined to the Memphremagog quadrangle and parts of the Irasburg quadrangle (Doll, 1951), and the Littleton quadrangle (Billings, 1937, and Eric, 1942).

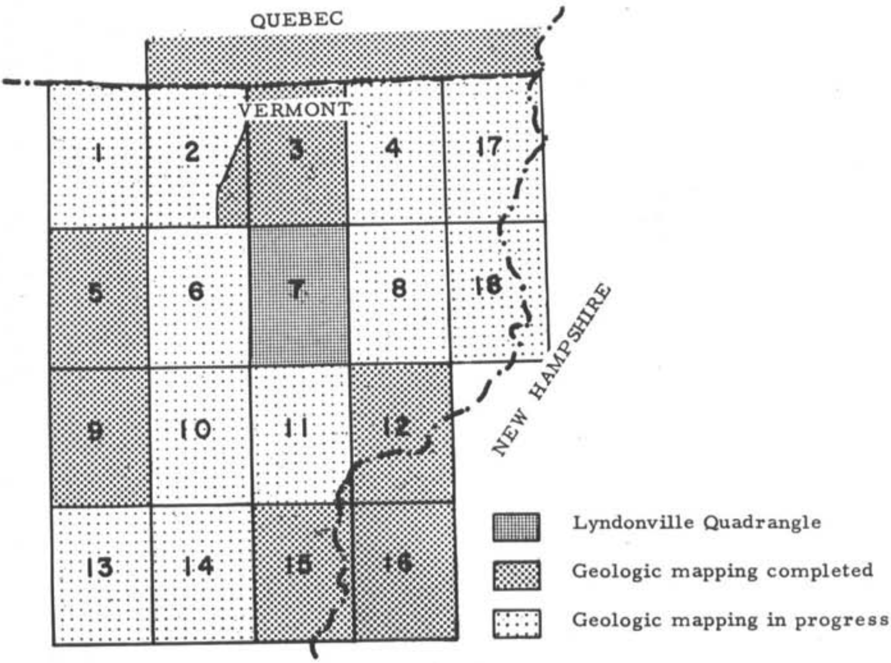

Figure 1. Status of detailed mapping in northeastern Vermont. Numbers denote the following quadrangles:

1. Jay Peak
2. Irasburg
3. Memphremagog
4. Island Pond
5. Hyde Park
6. Hardwick
7. Lyndonville
8. Burke
9. Montpelier
10. Plainfield
11. St. Johnsbury
12. Littleton
13. Barre
14. East Barre
15. Woodsville
16. Moosilauke
17. Averill
18. Guildhall

A detailed review of previous work that bears on the stratigraphy of the region is given in the section on Stratigraphy.

Purpose of Study

The original purpose had been the study of the structural relations of the granitic plutons. Field work soon showed that exposure conditions in the area would not permit adequate observations to be made to this end. On the discovery of the interesting structural features of the

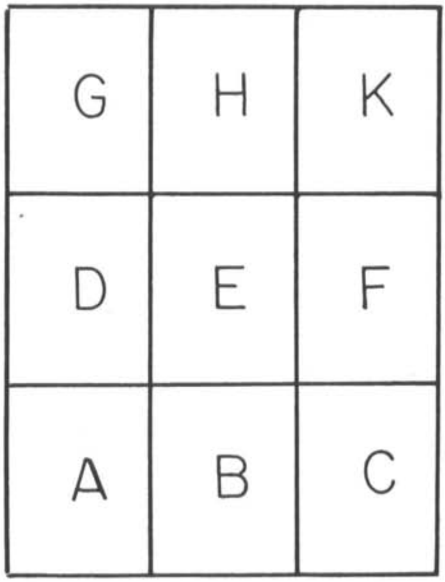

Figure 2. Reference letters of quadrangle divisions.

Willoughby arch, the main object became further elucidation of this structure, and the relation of metamorphic history to structural history. An attempt was also made to bring some light to bear on the age problem.

Method of Study

Field mapping was carried out in the summers of 1954 and 1955, a total of 7 months. Most of the fall and winter of 1955/1956 was devoted to laboratory and library studies, and to the preparation of the manuscript, including illustrative material.

As topographic bases, the writer used a photostatic enlargement of the U.S.G.S. one-inch map mounted and cut into cards, and aerial photographs covering the quadrangle area. Location on the base map was by pace and compass traverses, and direct location from topography in suitable areas. In mountainous, forest covered terrane, location was aided by the use of an aneroid barometer. A modification of Rich's (1921) reference system was found useful, and is used throughout this report: the quadrangle map is divided into its nine sections designated by consecutive letters of the alphabet, from left to right and from the bottom up. Thus the SW corner of the sheet is "zero" point and the SW ninth is A. Within each ninth, the SW corner is again zero, and decimal coordinates are used to fix any point; the first two figures of the reference matrix designate inches and tenths going east; the last two similarly designate inches and tenths going north. For further precision, a third figure may be introduced to designate estimated hundreths of an inch, but this has not been found necessary in the present report.

Acknowledgments

The writer is primarily indebted to Professor Walter H. Bucher of Columbia University who directed the work, and who gave constant support and constuctive criticism. Professors Marshall Kay and Arie Poldervaart, also of Columbia University, gave valuable help in the course of the project.

The field work was carried out under the auspices of the Vermont Geological Survey. The advice and guidance of Dr. Charles G. Doll, State Geologist, are acknowledged with gratitude. Dr. Wallace M. Cady, of the United States Geological Survey, gave a great deal of valuable help and advice, both in the field and with the manuscript.

The writer had the benefit of many stimulating discussions with Professors Marland P. Billings and James B. Thompson, of Harvard University, who have a long and intimate knowledge of the region. Dr. S. Sen of Calcutta University introduced the writer to petrofabric analysis, and has been very helpful in many aspects of the work.

The writer also owes much to discussions with Prof. and Mrs. Bertram Woodland, who have begun work on the Burke quadrangle, and has had the benefit of a day spent in the field with Mr. Leo Hall, who has started work on the St. Johnsbury quadrangle.

Mr. Arno Weilenmann helped with the drafting, and Mr. Rae L. Harris with the thin section photo-micrography.

The help of many other persons who have furthered the work in one way or another is gratefully acknowledged.

Throughout the field work, the writer has had occasion to appreciate the helpful attitude of the local residents toward his work. Mr. Maurice Brouha has given valuable help as voluntary field assistant fcr a number of weeks.

Physiography

The area is very hilly, with differences in relief of up to 1500 feet (Lake Willoughby). The lowest elevation is in the Passumpsic River at

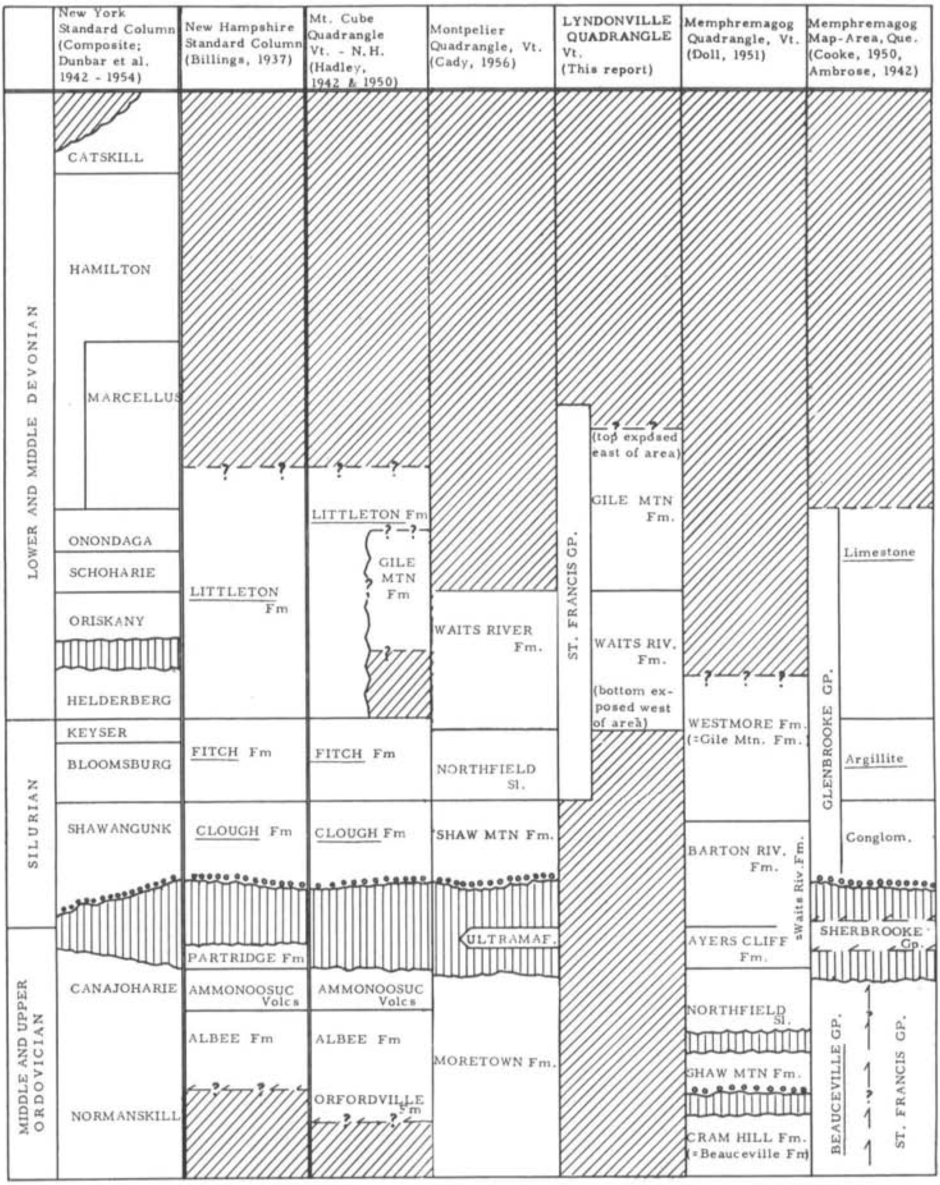

PLATE 4. Correlation chart, northeastern Vermont and adjoining region.

PLATE 5. Topographic expression of the Perry fault. Negative No. 8612 26–7, Fairchild Aerial Surveys, Inc., Courtesy Brown Co., Berlin, N. H., and Fairchild Aerial Surveys, Inc.—(⅔ original size)

North of the fault note the topographic expression of granite sheets; much of that area is, however, underlain by sediments. A few N30°W joints are clearly expressed.

On the south side of the fault, the curving outcrops of Gile Mountain beds dominate the picture. The smoother ground in the southeast corner marks the beginning of calcareous Barton River lithology. A number of N60°W (*ac*) joints show up plainly as straight lines. Close to the Perry fault, southwest of Bean Pond, there are a number of southwest striking linears, presumably tension joints associated with the faulting.

Lyndonville (just below the 700 ft. contour), the highest at the summit of Wheelock Mountain (2783 ft.). Geologic structure is well expressed in the topography, brought out by stream erosion and glaciation.

The eastern Gile Mountain/Waits River contact is along comparatively gently dipping beds, and thus forms a prominent scarp by lithologic contrast: the more resistant Gile Mountain quartzites rest on the more easily weathered Waits River formation. Wheelock Mountain, the highest peak in the Lyndonville area, and many other prominent mountains are aligned along this contact (see Pl. 1). The western contact is steep to vertical and is not brought out in the topography; nevertheless, the Lamoille and Barton rivers follow the strike close to the contact.

Doll (1951, p. 13) has interpreted the topographic profile shown on Pl. 6 as roches moutonnees. The present writer would prefer to regard them as scarps made by "capped" mountains. This interpretation is fairly well established for Mt. Hor, whose southern and eastern slopes consist mainly of the relatively soft Waits River formation, while it is capped by Gile Mountain siliceous schist and by granite on its back slope. The granite mountains to the northwest of Mt. Hor fit into the same pattern if interpreted as thick conformable granite sheets.

Granites often, though not necessarily, form elevations. The Black Hills, Grays Mountain, May Hill, Wheeler Mountain, Hedgehog Mountain and Haystack Mountain are the only prominent peaks which (as far as exposure conditions reveal) are entirely in granite. Mount Hor and Mount Pisgah are chiefly carved in sediments, reinforced by granitic vein rocks and adjoining granite sheets or massifs.

Air photographs reveal that many stream valleys follow fractures in the bedrock; certain straight lines on air photographs can be interpreted as fractures; joint measurements in the field confirm this interpretation.

Geologic considerations offer an explanation for the Lake Willoughby gap (see Frontispiece). Observations on joints in the face of Mount Pisgah show that a vertical joint system striking at N 30° W is pronounced in the granite, but is very much subdued in the adjoining sedimentary rocks. It would seem that any adjustment along joint planes would at first tend to be taken up by plastic yield in the sediments, rather than by rupture. But the granite will tend to adjust by shearing. Once sheared, the granite offers greatly reduced erosion resistance, and the shear zone would provide a ready channel for preferential erosion; the effects were undoubtedly accentuated by the circumstance that the direction of glacier movement, as seen in striations elsewhere, was at a relatively small angle to the shear zone. The lake bottom opposite the summit of Mt. Pisgah is at a depth of 210 feet, and so total relief here reaches 1800 feet (Doll, 1951, p. 14). Similar considerations might explain the steep cliffs on the east shore of Crystal Lake.

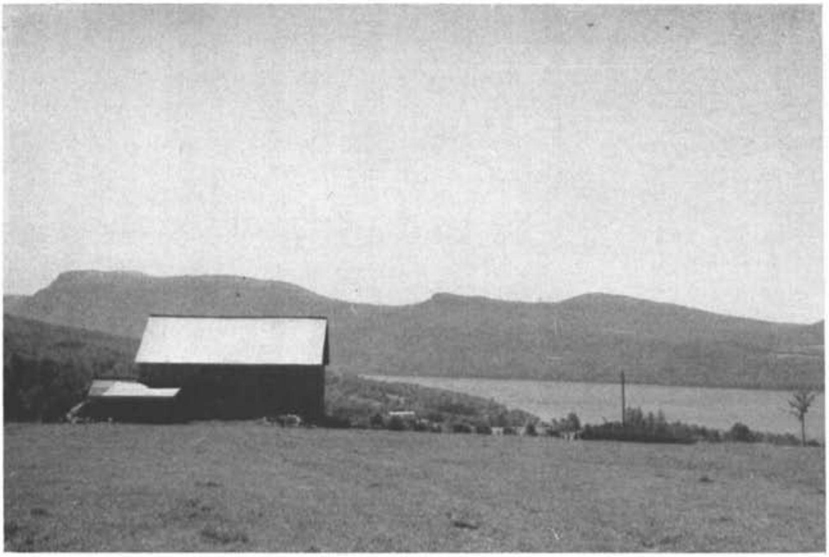

PLATE 6. Saw-tooth scarp pattern of granite sheets; Mt. Hor at extreme left. Looking southwest from above Westmore village.

There is considerable evidence of former glaciation in the area. Scattered observations of glacial striations indicate a somewhat variable direction of ice advance within a few degrees of due south. The last stage of glaciation produced valley trains now forming terrace deposits along Miller Run and Sutton River. These often carry large boulders of locally derived Barton River limestone up to 7 feet and possibly more in diameter. There are a number of small kames and eskers along Miller run (see Pl. 15).

Evidence of the great thickness of the continental glacier in the region is provided by one large boulder of granite (some 20 feet in diameter) on the very top of the northern ridge of Big Rock Hill, at H 0615.

Pl. 16 brings out the contrast between the resistant granite and quartzite mountains in the background, and a part of a peneplain which is expressed by areas of fairly flat plateau-like country at an average elevation of 1500–1600 ft. Drainage southward is toward the Connecticut River and northward toward the St. Lawrence. The divide is mostly carried by the band of Gile Mountain formation, the chief exception being near Lake Willoughby where the glacially carved gap has brought the divide farther south.

The drainage basin of the Barton River was extended artificially in 1810. The swamp at Runaway Pond had originally been a filled pond, draining into the Lamoille River. In 1810 it was decided to drain some of it toward Barton. Soon after excavations started, the water in the pond breached the divide, now weakened by digging, and the whole pond drained into the Barton River. Since then, the divide has been just south of the Glover/Greensboro townline.

The flow of the Passumpsic River is controlled by two dams south of Lyndonville.

STRATIGRAPHY

Lithologic Descriptions

WAITS RIVER FORMATION

GENERAL STATEMENT

The name Waits River limestone was introduced by Richardson (1906) for a thick sequence of limestones and schists in eastern Vermont. The evolution of the nomenclature is traced under Age. The Lyndonville quadrangle is underlain by the upper sedimentary member of the Waits River formation, first named Barton River formation by Doll (1951) in the adjoining Memphremagog quadrangle (cf. correlation table, Pl. 4).

Doll did not use the names Waits River and Gile Mountain in his 1951 report because, at that time, no correlative link to the type localities had been established. The present writer retains the names Barton River and Ayers Cliff as members of the Waits River formation. The Ayers Cliff member does not crop out in the Lyndonville quadrangle.

The type locality of the Barton River member is in Newport City in a low cut on U. S. Highway 5 leading to Derby. According to Doll, the Barton River formation consists of "... intercalated impure calcareous rocks ranging from limy quartzites to limestones, amphibolitic layers, slates, phyllites, quartzites and schists. . ." The bottom of the Barton River member does not crop out within the Lyndonville quadrangle and the immediate stratigraphic problem was to find an adequate definition of its contact with the overlying Gile Mountain formation.

DISTRIBUTION

The Barton River crops out in the Memphremagog quadrangle in a north-south trending belt averaging 6 miles in width. This belt follows the western edge of the Lyndonville quadrangle, its western (lower) contact remaining outside the quadrangle throughout. A second, more easterly belt, crops out east of the Brownington Syncline (Doll, 1951), and underlies most of the eastern part of the quadrangle. This belt is a structural arch with a gentle northward plunge; it disappears under the Gile Mountain formation north of Newark, and joins the western belt around the Brownington syncline north of Joe's Pond. South of Danville and north of the Canadian border the Barton River has not yet been mapped as such, but it is represented by parts of what has variously been mapped as Waits River, Memphremagog, Tomifobia and St. Francis (see under Age). The Barton River/Gile Mountain contact, which is of importance in the Lyndonville area is, by definition, the same as the Waits River/Gile Mountain, Memphremagog/Gile Mountain and Barton River/Westmore contacts of the earlier literature. A definition of the contact as recognized in the Lyndonville area, can only be given after each formation has been described in detail.

As the bottom of the Barton River is not exposed in the area under investigation, an independent estimate of its thickness by the present writer is not justified. Doll calculated the thickness as 8800 feet using Currier and Jahns' factor of 0.4.

Age

The question of age relationships is taken up separately below. As for the sequence within the quadrangle, pillows in the Standing Pond lava near the Barton River/Gile Mountain contact indicate that the Barton River is stratigraphically older than the Gile Mountain formation (see below, under Standing Pond lava).

Lithological Detail

Definition of formations within the St. Francis group (cf. correlation table) hinges on the relative proportion of three key lithologies found throughout the group. There are, of course, variations within each lithology from the bottom to the top of the group, and some of these may be significant.

1. Silicious limestones (Doll's impure calcareous rocks and amphibolitic layers).
2. Argillaceous schists and phyllites (Doll's slates, phyllites and schists).
3. Quartz schists and siltstones (Doll's quartzites).

In the Barton River the limestones should typically occupy more than 25 per cent of a given outcrop area, while quartzites should be very subordinate. The practical aspects of fixing the boundary are discussed below.

1. Siliceous limestone. Typically, this is seemingly coarse grained, bluish gray when fresh and weathering gray to rusty, according to iron content (which normally is very low). It consists of re-crystallized calcite and quartz grains, averaging 0.02–0.05 mm in diameter and variable amounts of disseminated carbon. It should properly be called a marble. Normally this rock is remarkably free of argillaceous and magnesian impurities; where present these impurities have been recrystallized to sericite, tremolite and phlogopite. Argillaceous varieties of the carbonate lithology member do occur recrystallized to amphibolites. As exposed many of these amphibolites appear to be almost monomineralic, but closer examination invariably shows the presence of a considerable amount of calcite matrix. White and Jahns (1950) state that they have found gradations between pure calcarenites and argillaceous rocks, but in the Lyndonville area the writer has found no gradation: the calcareous bands may be more or less impure; the argillaceous bands may be locally calcareous; but contacts between the two are always sharp.

The limestone bands vary in width from little more than an inch, forming septa between argillites, to cliffs exposing thicknesses of more than 50 feet; in the western part of the sheet, in Lyndon township particularly, great areas seem to be underlain by Barton River containing over 90 per cent of limestone.

The limestone is often traversed by thin veins (about 2–3 mm in width) of fine-grained siliceous and/or calcareous matter; these veins are parallel to sub-parallel to the bedding and they sometimes outline intense internal deformation. Near the contact with the argillites, the limestone is often dotted with amphibole porphyroblasts.

2. Quartzite. Quartzites are exceedingly rare in the Waits River formation. Essentially, there appear to be two types: a tough bluish silty rock which often weathers like siliceous varieties of the limestones (in some exposures it can be mistaken for the latter in the absence of further tests), and a whitish rock which can be observed to grade into a more argillaceous lithology. Both are more typical of the Gile Mountain formation, and will be described under that heading in greater detail.

3. Argillites. The argillaceous schists and phyllites vary in composi-

tion, and their variety is accentuated by metamorphism. The following are characteristic:

a) a dark, almost black dense graphitic schist containing much pyrrhotite aligned in grains parallel to the bedding; the rusty weathering of this rock is very characteristic.

b) a gray brittle quartz-sericite-biotite schist containing varying amounts of carbon and, sometimes, sulphides. Because of the latter the rock usually weathers a rusty brown. This type is normally better foliated than the graphitic type, but gradational types between the two are common. Petrographic study shows that the argillite is a fine-grained graywacke. In general, rusty weathering of the argillaceous lithology is typical of the Waits River formation, while it is rare in the Gile Mountain. However, this difference is merely indicative, not diagnostic. It points up a change in iron content, particularly iron sulphide.

Detailed petrographic descriptions of all lithologies, particularly the argillites, are given in the section on petrology.

GILE MOUNTAIN FORMATION

General Statement

The name Gile Mountain formation was introduced by Doll (1944), in the Strafford quadrangle. Its type locality is on Gile Mountain, in the southern part of the Strafford quadrangle. According to Doll it there consists "principally of quartz-mica schist . . . rocks in subordinate amount are thin beds of massive and sheared quartzite, occasional coarse feldspathic schists, calcareous beds and some graphitic layers." Doll includes the schists east of the Monroe line, an extension of Hadley's Orfordville, in his Gile Mountain. As mentioned under Age, Kruger (1946) and Lyons (1955) show that there is uninterrupted transition from the Gile Mountain into the Orfordville, vindicating Doll's interpretation.

The rocks mapped as Gile Mountain by the writer are in direct strike continuity with Doll's (1951, p. 33–37) Westmore formation of the Memphremagog quadrangle. The name Westmore was dropped when it became clear that the Westmore was the equivalent of the Gile Mountain, which latter name had precedence. The evidence for this correlation will be summarized at the end of this section.

Distribution

The Gile Mountain formation of the Lyndonville quadrangle is confined to the Brownington syncline. Both its contacts within the quadrangle are lower contacts with the underlying Barton River formation. This contact will be discussed below.

Immediately southeast of the Lyndonville quadrangle another belt of Gile Mountain comes to overlie the Barton River. It is this belt which Eric and White mapped as Ticklenaked, and which White later (1951) correlated with Doll's Gile Mountain type section. The writer, in a number of brief reconnaissance traverses, was able to follow this belt into the type area of the Westmore, thus correlating the Gile Mountain and the Westmore. A similar interpretation had been suggested by Richardson (1906) for his Vershire schist which, on the whole, can be taken as equivalent to the Gile Mountain.

Regionally, therefore, the Gile Mountain formation can be followed along the Connecticut River as far south as Ascutney, where it becomes indistinguishable from the Orfordville, and as far north as the St. Francis group has been mapped in Quebec, where the Gile Mountain constitutes the upper member of the St. Francis "series."

The western belt, previously recognized by Richardson (1898) and White and Jahns (1950), branches southward from Barton on the west side of the Wheelock arch. It is removed by erosion over a culmination in Walden and Cabot, and reappears in Barre (White & Jahns, 1950); it has not been recognized south of the White River. There are indications that outliers of this belt occur at the Cabot culmination (Richardson, 1906, p. 93), but the area between Cabot and Barre has not yet been mapped in detail.

Lithologic Detail

Most of the Gile Mountain within the Lyndonville quadrangle is of garnet and staurolite grade; consequently much of the detailed petrography will be more fully discussed in the section on metamorphism. Doll (1951, p. 33) states that the rocks of the Westmore formation ". . . are largely phyllites and schists, with smaller amounts of limestones and quartzites, all interbedded. . . . The light-colored rocks are proportionately greater in amount than in the formations to the west, and sulphides appear to be greatly reduced in comparison."

In the Lyndonville quadrangle, the Gile Mountain formation consists mainly of gray quartz-sericite-biotite schists, and impure silty biotite-quartzites. Limestones are rare and more often impure than in the Barton River formation. South of Crystal Lake, however, there is a small area in which the lithology of the Gile Mountain resembles that of the Barton River in its limestone content. It is difficult to decide whether this is due to faulting along the Perry fault, or whether it represents a calcareous facies within the Gile Mountain.

1. *Quartz-sericite-biotite schists*. All of these contain porphyroblasts of one or more of the following: biotite, chloritoid, garnet, staurolite. However, these porphyroblasts are generally very hard to identify in exposures and hand specimens; only the conspicuous minerals are incorporated in the nomenclature. These schists are dark gray when fresh and exhibit typical sericite luster, with biotite distributed in flakes and clots. The schists are well foliated; the biotite porphyroblasts (as well as the garnets and staurolites) are post-deformation, as will be shown in the section on petrology. Frequently a lineation can be seen, sometimes more than one. The sericite is often associated with very fine-grained quartz, in very variable proportions; macroscopically, the relative amounts of quartz and sericite can be roughly estimated by toughness and lack of foliation (quartz-rich) and sheen (sericite-rich). The more quartz-rich types occur near the bottom of the formation, and partly give way upwards to silty quartz-schists and quartzites which are abundant near the center of the syncline.

Close to the Barton River contact there are transitional types to the Barton River schists. The typical Gile Mountain schist is poor in sulphides and carbon, and does not weather a rusty red nearly so often as the Barton River schist.

2. *Quartzites*. These are micaceous (sericite and biotite) throughout. They are fine grained (of the order of 0.01 mm grain size) even in the higher metamorphic grades, where they are interbedded with staurolite schists. Often they are finely banded, a banding which is not unlike varving; rhythmic interbedding with sericite-biotite schists, in bands averaging 6″–9″ thick was observed in the eastern part of the town of Sheffield.

3. *Limestones*. The limestones of the Gile Mountain are siliceous, as are nearly all the limestones of the St. Francis group. They are completely recrystallized as in the case of the Barton River and should, properly, be called marbles. Greater amounts of impurities account for a higher proportion of amphibolites than in the Barton River formation.

There are also occasional lenses of calc-silicate rock in the terrigenous lithologies of the formation, suggesting the presence of calcareous lenses before metamorphism.

In the Littleton quadrangle the Gile Mountain occurs in a far less metamorphosed form than in the Lyndonville area, the gray schists being represented by biotite-porphyroblast slates. When comparing the two belts it is important to remember that grade of metamorphism is not the only difference, and that this is a region of comparatively rapid sedimentary facies change eastward; also, unlike the western belt, the eastern belt contains volcanics.

THE WAITS RIVER/GILE MOUNTAIN CONTACT

Previous workers (Richardson, 1898; Eric, 1942; Doll, 1944 and 1951) have defined the difference between the Waits River and Gile Mountain formations by the relative abundance (Waits River) and lack (Gile Mountain) of calcareous rocks in a gradational sequence of inter-bedded argillites, quartzites and siliceous limestones. Although the disappearance of the limestones, when going from the Waits River into the Gile Mountain is gradual, it occurs within a comparatively narrow zone, usually within 1000 to 1500 feet. There is no one marker horizon to use as a reference, and the bands of the various inter-bedded lithologies seem to vary in thickness and composition along the strike. There are insufficient exposures for an attempt to follow any one horizon, or even to establish comparative sections. The chief criterion used in mapping was the relative abundance of limestone bands: less than 25 per cent of limestone in a large exposure, or a group of closely spaced exposures, was mapped as Gile Mountain if occurring near the estimated contact; (but there are areas *within* the Barton River—such as west of Parker Pond—which have less than 5 per cent of limestone exposed). Supporting criteria were prominence of siliceous schists, absence of rusty weathering, absence of sulphides. The stratigraphic significance of the Standing Pond volcanics[1] was not realized until late in the second season of mapping, when Dr. Cady drew the writer's attention to the similarly placed (though lithologically different) Standing Pond volcanics farther south. It was gratifying to find that the Waits River/Gile Mountain contact had been consistently mapped with reference to the volcanics wherever they occur, in spite of the very rough and empirical criteria which had to be used. The revision of the contact in the Littleton quadrangle was also confirmed by the discovery of the Standing Pond volcanics.

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[1] See Pl. 3.

Age

PRELIMINARY REMARKS

Only two sedimentary units crop out within the Lyndonville quadrangle: the Waits River and the Gile Mountain formations. If no time concept were attached to stratigraphic units, this chapter would be short. But, as in most metamorphic terranes, the assignment of units to a position in the relative time scale of geology is difficult. In the Lyndonville area the problem is particularly intriguing: depending on the weight given to apparently conflicting evidence, the formations have been described elsewhere as Ordovician, Silurian or Devonian. Each stratigraphic interpretation leads to a different concept of the structure. The problem is, therefore, of sufficient importance to deserve a critical survey of the literature.

The review that follows is by no means exhaustive. The writer has merely attempted to bring together important evidence and views of the past, and has omitted mention of work which sheds no significant light on the specific problem to be considered.

All references to localities and quadrangles are summarized on Pl. 3, the regional key map.

EARLY WORK

Adams (1845, p. 49 and 62) was one of the first to describe the rocks underlying the Lyndonville quadrangle. He named them "Calcareo-Mica Slate," and placed them in the "Primary System," but did not attempt to establish a succession.

Hitchcock (1861, p. 475–488) gave a detailed lithologic description and changed the name to "Calciferous Mica Schist." His section XI passes through the Lyndonville quadrangle (p. 660–661).

Logan (1848 and 1863, p. 406–437) traced his Gaspé series (to which he assigned Siluro-Devonian age on paleontologic evidence), from the Gaspé peninsula type section to the Vermont border at Stanstead. He was able to distinguish two formations: one, more calcareous, to the northwest and the other, mainly terrigenous, to the southeast. In the Eastern Townships the calcareous formation (correctly recognized as the older) was found to contain "Helderberg" fossils in three, now classic, localities. ("Helderberg" at that time was not defined as now, and included some late Silurian.) These localities are at Famine River, Dudswell and Lake Memphremagog. In each case the fossils are in the lowest part of the calcareous formation, and in each case this overlies a conglomerate.

Hunt (1854) traced Siluro-Devonian rocks from Gaspe to Lake Memphremagog and on into Massachusetts; he also suggested (p. 199) that the metamorphic rocks of Massachusetts (the continuation of the New Hampshire metamorphics) might be highly metamorphosed Devonian. Hitchcock (1841, p. 586–587) had already remarked that many of the granitic rocks of the area were in all probability transformed sediments.

Hitchcock (1861, p. 447–451) recognized "Upper Helderberg" occurring at Bernardston, Massachusetts, and in a belt of limestone running south into Vermont from the Lake Memphremagog locality. Hitchcock, however, believed the conglomerate to be younger than the fossiliferous limestone with which it is associated. The "Calciferous Mica Schist," therefore, had to be older than the Upper Helderberg, but no definite age was assigned to it. Hitchcock (1870) published a provisional geological map of New Hampshire. His Coos formation included very largely those rocks which are now variously named Gile Mountain, Orfordville and Littleton (low grade). The high-grade metamorphic Littleton of today he named the White Mountain series (adjective: Montalban). The Albee-Ammonoosuc-Partridge sequence of today, he correlated with Logan's Quebec group.

Dana in his textbook (1875) quoted Logan as his authority for listing formations between northern Vermont and Gaspé as ranging from Niagaran to possible Lower Devonian.

Selwyn (1882, p. 2–3) concurred in Logan's age assignment, but observed that the separation of Silurian and Devonian fossils presents a problem.

Ells (1887, p. 7–14) reviewed previous work and gave a brief account of his own reconnaissance work in the Eastern Townships. He was impressed by the great contrast between the fossiliferous rocks and the volcanic assemblage associated with them. He also described (p. 16–17) a "Llandeilan" graptolite assemblage found in Castle Brook, Magog. Ells was evidently of the opinion that all the terrigenous rocks of the Memphremagog area were related, belonged to one sequence, and should be classified as Llandeilan in accordance with the age of the Castle Brook graptolites, as then understood. Ruedemann (1947) classified this fauna as late Normanskill.

RICHARDSON'S WORK IN EASTERN VERMONT

Richardson (1898, p. 296) assigned a Lower Trenton age to his Washington limestone and Bradford schist (now Waits River and Gile Mountain formations) and based his opinion on the fossil evidence at Castle Brook; he saw in the Northfield slates the equivalent of the fossiliferous Magog slates; in his structural interpretation the Waits River was older than the Northfield slate, and hence (in his reasoning) Lower Trenton.

In 1902 Richardson gave the first more or less detailed description of an area in eastern Vermont, with his report (1902) on the "Terranes of Orange County, Vermont." This report contains a reconnaissance map, cross-sections, and drawings of the Castle Brook graptolite fauna. Arguments for an Ordovician age of the Waits River/Gile Mountain are reinforced by these graptolites. Richardson here correlates the present Northfield slate with the Magog graptolite slate and observes (p. 97) that there is no unconformity to the west of it in the area covered by the report.

In his report on northeastern Vermont (1906, p. 115) Richardson changed the name to Waits River limestone, because the name Washington limestone had been pre-occupied. In 1898 he had given the name "Bradford schist" to the non-calcareous members of Hitchcock's Calciferous Mica Schist. Again, this name was changed in his 1906 report (p. 115) to "Vershire schist," the name "Bradford" having been preoccupied. Richardson's (1906, p. 63) Pl. XVII gives the first structural section in the vicinity of the Lyndonville quadrangle which bears some resemblance to present-day interpretation. It is drawn from west to east, a few miles south of the southern edge of the Lyndonville sheet. Richardson now (1906, p. 84) cast doubt on Hitchcock's (*loc. cit.*) correlation of the Coventry crinoidal limestone with the Lower Helderberg of Lake Memphremagog. He cited (1902 and 1906, p. 112) several fossil finds in the Waits River and Gile Mountain, but admitted that none of them were diagnostic. In 1919 Richardson correlated the Irasburg conglomerate with various conglomerates in Vermont and Quebec, some of which are now known to be part of the Shaw Mountain formation and its correlatives, and some to be even older. This conglomerate horizon Richardson defined as the erosional unconformity at the base of the Ordovician. His view was in part based on Ruedemann's dating of the "graptolites" in this formation as Deepkill.

Convinced that graptolites had to be found in the formations immediately overlying his conglomerate, Richardson discovered several "graptolite" localities, and submitted specimens to Ruedemann. Richardson's reports published in 1919 and 1928 review the paleontological evidence for Lower Trentonian age of the Northfield slate-Waits River. However, in 1931 Foyles (p. 252) showed that one of the "proven" graptolite assemblages represented mica streaks. Thus doubt was cast on all the graptolite finds in Vermont.

RECENT DETAILED MAPPING IN THE WAITS RIVER FORMATION

Detailed work on the geology of northeastern Vermont started in 1937 when White (1951, p. 646) began mapping the Vermont portion of the Woodsville quadrangle. One of his main problems was to relate the rocks of eastern Vermont (Richardson's Waits River limestone and Vershire schist) to those of western New Hampshire, which had just been described by Billings (1937) in his classic paper on the Littleton-Moosilauke area. Eric (1942) mapped the Vermont portion of the Littleton quadrangle as part of the same program. The Waits River limestone was re-named Waits River formation, after Currier and Jahns (1941); and the Vershire schist became the Ticklenaked formation (White, quoted by Eric, 1942). Evidence for a fault contact, the Monroe fault, between the "Vermont Sequence" and the "New Hampshire Sequence" was presented by Eric, White and Hadley (1941).

Eric (1942) defined his Ticklenaked-Waits River contact essentially by the characteristic presence of limestone in the Waits River formation.

Currier and Jahns (1941) described a crinoidal limestone (possibly related to Hitchcock's "Helderberg") in their Shaw Mountain formation (p. 1496). This places the formation above the Deepkill, and is further evidence of the misidentification of Richardson's "graptolites" ("Deepkill"). They also noted (p. 1498) the lithologic similarity of the Shaw Mountain and the Clough. They seemed to be of the opinion that Richardson had placed the Waits River below the Northfield slate throughout his reports. Richardson was never too explicit on this point, and it seems to this writer that he quietly and unobtrusively reversed his succession between 1906 and 1919. It is certain that in 1898 his Waits River was below the Northfield slate; however, later reports are ambiguous—evidently Richardson was puzzled as to the true relationships; his 1928 report implicitly places the Waits River above the Northfield slate.

Currier and Jahns (*loc. cit.*) were also the first to accurately define the Northfield slate, a term which had been used rather loosely so far. As defined by Currier and Jahns, the Northfield slate ". . . lies unconformably above the Shaw Mountain formation and conformably beneath the . . . Waits River limestone of Richardson. . . ." In the same paper the Northfield slate is tentatively correlated with the slaty basal beds of the Tomifobia formation, removing the correlative link with the Castle Brook graptolite slate, and the name Waits River formation (p. 1491) was substituted for "Waits River limestone."

Currier and Jahns also correlated the Magog graptolite slate with their Cram Hill formation below the Shaw Mountain; the Cram Hill thus becomes, in part, the correlative of the Beauceville in Canada, dated by means of the Castle Brook graptolites as Normanskill and older (Ruedemann, 1947). They also closely examined, and rejected, all the available "graptolite" specimens found by Richardson.

This work places a lower age limit (post-Normanskill) on the Waits River/Gile Mountain, and removes the upper age limit.

Doll (1943) found large crinoid calyces in the Gile Mountain at Westmore, which led him to date it tentatively as Lower Devonian. Later (1943b) he found the impression of a large spirifer in the Gile Mountain of the Strafford quadrangle.

In the meantime Doll (1944), working in the Strafford quadrangle, had discovered a marker horizon between the Waits River and Ticklenaked formations. Doll was able to show that this horizon, the Standing Pond amphibolite, transgressed lithologic boundaries and was, probably, a time-stratigraphic horizon of volcanic origin. He applied the name Memphremagog formation to the Waits River, the Standing Pond amphibolite and a discontinuous garnet schist associated with the Standing Pond. The correlative of the Ticklenaked, Richardson's Vershire schist, he named Gile Mountain schist. The correlative of Richardson's Waterford slate was mapped as a separate lithologic unit, the Meetinghouse slate.

In the extreme southeastern corner of Doll's map (1944) there appears a strip shaded as Gile Mountain. This is in the same belt that Hadley (1942) had mapped as Orfordville, one more indication of the interesting resemblance between the two formations.

Working in the Memphremagog area Doll (1951) divided the equivalents of the old Tomifobia formation into three new formations. From west to east (upwards) these are: the Ayers Cliff, Barton River and Westmore formations. Doll recognized the equivalence of his Ayers Cliff and Barton River formations with the Waits River, and dropped the latter name as too comprehensive. He also suggested that his Westmore and Gile Mountain might be equivalent.

DETAILED WORK IN CANADA

Kerr (1923) quoted by Clark (1934, p. 11) applied the name Tomifobia slates to the equivalents of the Waits River limestone in Canada and included the slates below it. Clark (1934) defined the Tomifobia slates as "calcareous slates east of the Bunker thrust;" hence the lower limit is "faulted;" the upper limit, being east of the Memphremagog quadrangle, is not described.

Ambrose (1943) had published a map of that part of the Memphremagog map-area (Canada) which adjoins the Memphremagog quadrangle (Vermont). It shows Kerr's (1923) and Clark's (1934) Tomifobia formation without subdivisions, and without a name. Clark (in Cooke 1937), had defined the St. Francis formation as including the rocks southeast of "a line drawn from Lac Rocheux in Adstock township to Elgin Lake in Stratford." He also stated that the St. Francis formation was made up of two members: a western member made up chiefly of volcanics, and an eastern member consisting of slates, quartzites and graywackes, and calcareous beds. Cooke (1950) detached the volcanic member from the St. Francis group and correlated the sediments with the old Tomifobia formation, dropping the latter name. Thus the St. Francis group, as now defined by Cooke, correlates with the Northfield, Waits River and Gile Mountain formations as far east as Cooke's (1950) Victoria River fault; this latter may represent the Monroe fault, and the rocks to its east may represent the Albee formation.

Clark (1934, p. 11) states that "graptolites" in the Tomifobia slates resemble those of Castle Brook. In 1937 Clark (in Cooke 1937, p. 42) states that Ruedemann identified graptolites found in the St. Francis series near lac Rocheux as Diplograptus (glyptograptus) euglyphus of Normanskill age. These may be the same that are briefly referred to in Clark's 1934 paper and which Morin (1954) later had re-examined and rejected.

Morin (1954) suggested Siluro-Devonian age for the St. Francis mainly on the evidence of good strike alignment with similar lithologies leading to proven Devonian in Gaspe. This is essentially Logan's argument. Boucot (1954), on an unpublished map, indicates similar relationships.

In the meantime good collections of fossils had been made from those localities in the Eastern Townships whose Siluro-Devonian age has never been in dispute.

Harvie (1912) collected an assemblage from what is now known as the Glenbrooke series, in Glenbrooke Creek. McKay (1921) studied the fauna of the Famine limestone in the Beauceville area. Burton (1930) wrote a report on the Lake Aylmer area, and his list of fossils was brought up to date by Cooke (1937, p. 50–51). It is interesting to note that McKay (*op. cit.*) has his Devonian in a syncline within the Beauceville, while Burton's Devonian is surrounded by his "Silurian" Disraeli series. There is this difficulty, in each case, to distinguish between the rocks on either side of the Devonian; yet evidence further south (Cooke and Clark, 1937) seems to suggest that they belong to two distinct sequences.

Laverdière (1936) mapped the Marbleton (Dudswell) area in detail, and described the fauna.

Cooke's Memoir 257 (1950) is a review and compilation of much of the preceding work by himself and others between Lake Memphremagog and Weedon. It introduces a new unit: the predominantly volcanic and coarsely clastic rocks of the Sherbrooke group (p. 57–62). Cooke's Memoir gives evidence that they overlie the Beauceville with what looks like a large angular unconformity but is, in all probability, a thrust plane. Cooke also shows that most of the Sherbrooke's contacts with adjoining formations are faulted: " . . . Where not faulted, the sprawling, irregular shapes of the Sherbrooke remnants, in the midst of an otherwise closely folded region, is excellent evidence of their unconformable character and rather flat-lying structure." (Op. cit., p. 60.) Cooke concludes that the Sherbrooke is Silurian.

In 1948b, Cooke had described the Bolton group as having very similar structural relations to the underlying rocks. However, as these latter rocks include Devonian, he felt compelled to make the Bolton group post-Devonian. This introduced a difficulty, for the Bolton group is metamorphosed and deformed much as the Sherbrooke group. Cooke explains this by means of a post-Devonian (Appalachian) period of deformation.

On his Dudswell map (1948a) Cooke proposes the extension southward of Burton's Weedon thrust, between the Devonian Lake Aylmer Group and the "Ordovician" St. Francis series. In Burton's area this thrust ran along a real lithologic boundary, the contact between the group of volcanics on the east and the Lake Aylmer group on the west. However, in Dudswell township the volcanics do not appear and Cooke was obliged to draw his thrust line through an area which Laverdière (1936) had mapped as consisting of one formation. He had to assume a thrust because the St. Francis series was supposed to be "Ordovician," and hence older than the Lake Aylmer which here underlies it directly (see also Cooke, 1950).

In the 1950 memoir (p. 71) Cooke also draws attention to the extraordinary resemblance between the Glenbrooke on the one hand and the Clough and Fitch (Billings, 1937) on the other.

RELATIONSHIPS IN THE CONNECTICUT RIVER VALLEY, VERMONT AND NEW HAMPSHIRE

Fossils had been found in the Fitch by Hitchcock in 1870 (p. 16). Hitchcock also remarked on a possible relationship to the Memphremagog Helderbergian. He admitted that fossil finds had been reported before, but his was the first authenticated discovery. It was considered of such importance at the time that news of the find was "immediately telegraphed to the Dartmouth Scientific Association, who happened to be holding a meeting the same evening" (p. 16). E. Billings examined the fossils and recognized Zaphrentis and Favosites.

A full description of the fauna of this locality and of others in this classical area is given by Billings and Cleaves (1934). The succession of the rocks of western New Hampshire was first established in its modern form by Billings (1937). Following this classical work, Billings' students have mapped key areas in the New Hampshire succession (cf. index map Pl. 3). In particular, Billings established the succession Albee, Ammonoosuc, Partridge, followed by the fossil-dated Clough (Middle Silurian), Fitch (Upper Silurian), Littleton (Lower Devonian). For ease of reference, the first three formations will in this report be referred to as the Albee-Ammonoosuc sequence.

Mapping in the Mount Cube quadrangle, Hadley (1938) encountered a new formation underneath the Albee, and named it the Orfordville. He recognized (1942, p. 159), that it resembles the Littleton formation on its east flank, but, owing to the age difference, felt obliged to separate it from the latter by assuming that the Northey Hill thrust extends this far south from the Moosilauke area. His correlation further obliged him (1942, p. 159) to postulate a stratigraphic displacement of about 13000 feet.

White and Billings (1951) state (p. 668): " . . . A large fault, with a stratigraphic throw of 12000 feet, must lie between the Orfordville and the Littleton formations. It is not easy to locate this fault, however, because the schists of the Orfordville and Littleton formations are lithologically similar."

A very similar problem is presented by the Monroe fault. It is mapped to the north of the Hanover quadrangle but is not shown to the south of it. There is no convincing evidence for the fault in the Hanover quadrangle, but Lyons (1955) maps it along the northern part of the Meetinghouse/Orfordville contact, and lets it vanish farther south. Lyons notes (p. 117) that the Littleton formation, where it crops out in the Hanover quadrangle, "closely resembles the near-by outcrops of the Orfordville formation, except that the Littleton possesses a distinctive 'pin-stripe' structure."

J. B. Thompson (personal communication, 1956) has observed a disconformable quartz conglomerate at the Orfordville/Gile Mountain contact in the Claremont area. He sees no evidence for a faulted contact, but he believes that tops of beds face west on the east side of the Monroe line. Thompson also finds that the Orfordville lithologies have closer affinities to pre-Clough lithologies (Albee, Cram Hill, Moretown) than to the Gile Mountain.

White (1946) gave 3 possible correlations for rocks of the St. Francis group, all of which would place them in the Devonian. His preferred correlation, on the basis of lithologic similarity, was Clough/Fitch = Shaw Mountain, Littleton = Northfield slate, and Waits River/Gile Mountain above Littleton in time. Later, however, (Billings and White, 1951) he felt that the evidence was in favor of Ordovician age.

Boucot *et al.* (1953) confirm the age of the Bernardston formation, tentatively correlate it with the Clough, and suggest that it might be equivalent to the similar units in Vermont (Shaw Mountain) and Quebec (Glenbrooke, Lake Aylmer and Famine).

Samples of some of the less metamorphosed dark shales were processed for microfossils by Foster D. Smith, Jr. Poorly preserved organic remains were encountered, including rare *Hystrichosphaerida* and *Chitinozoa* from sample No. M 15. These were submitted to Prof. L. R. Wilson, of New York University, for his opinion. He refers *Hystrichosphaerida* to the form genus *Leiosphaera* and states that this general type of *Leiosphaera* is encountered in rocks of Cambrian to Devonian age. This is a marine form, preserved most abundantly in shelf sediments. According to Prof. Wilson, the presence of *Chitinozoa* suggests that the beds are younger than Cambrian in age and that since no Silurian or Devonian forms were encountered in association with this sparse fauna, the formation could most easily be assigned to the Ordovician.

SUMMARY OF PRESENTLY HELD OPINIONS

In recent years, Siluro-Devonian age of the Gile-Mountain-Waits River has been accepted by Doll (1944 and 1951), White (1946), Hadley (1950), Boucot (1954), Morin (1954) and Cady (1956). These workers were predominantly concerned with the northern part of the belt.

Kruger (1946), Billings and White (1951), and Lyons (1955) all retain Ordovician age.

DISCUSSION

The evidence for a unified interpretation of northwest Appalachian geology is in part conflicting, in part suggestive. It is here proposed to start with a reasonable working hypothesis, and to discuss known facts and relationships in the region in the light of this hypothesis.

The most promising idea, in the opinion of the writer, is one put forward at various times, formally and informally, and by which the well-dated Clough would be equivalent to similar conglomerates in Vermont (Shaw Mountain) and Quebec (Glenbrooke, Lake Aylmer, Famine). Each of these conglomerates is associated with a fossiliferous limestone above it, and each (with the possible exception of the Shaw Mountain) is almost certainly the post-Taconic basal conglomerate, on the basis of fossil evidence previously cited. Making the Shaw Mountain equivalent to these dated conglomerates is the only major assumption in this discussion; the association of lithologies concerned is sufficiently uncommon to make such an assumption reasonable. Also, the map trends of the Glenbrooke and Shaw Mountain formations are so suggestive that Hitchcock (1861) had no hesitation in correlating the two.

To simplify the discussion, in view of the variegated nomenclature the following terminology will be used: *St. Francis group*—base of the Northfield slate up to and including top of the Meetinghouse slate, and correlatives in Vermont and Canada; *Albee-Ammonoosuc sequence*—Albee formation, Ammonoosuc volcanics and Partridge formation.

Following the Connecticut River from Brattleboro to Ascutney Mountain (cf. Billings *et al.*, 1952) it is found that the Standing Pond volcanics—an established horizon marker—are overlain first by the Orfordville formation, and then by the Gile Mountain (both of rather similar lithology!). To the west of this route, Gile Mountain appears in a number of synclines in the Standing Pond. The Orfordville in this area seems to lie in a syncline, as outlined by a quartzite. To the east, the next formation is the Littleton. As there is no need here for a thrust to separate the two formations, there is no immediate objection to placing the Littleton underneath the Orfordville stratigraphically; according to the hypothesis here used, both formations could be Lower Devonian. Going east, the next boundary of the Littleton is its *lower* contact (with the Clough, Fitch or Ammonoosuc, according to locality). This places the Orfordville in the trough of a synclinorium, with formations getting older going both east and west. The Northey Hill thrust is not considered here as a major dislocation of stratigraphic significance. There may well be overthrusting toward the west, but, if the hypothesis is granted that the Gile Mountain and Orfordville are both Lower Devonian, this thrusting would be largely intraformational, and therefore of minor importance. The disappearance of the calcareous facies of the Waits River in such a remarkably short distance as seen on Pl. 3, strongly suggests shortening by thrusting or recumbent folding.

North of Orford, the more characteristically volcanic lithology of the Albee-Ammonoosuc sequence which farther south underlies the Clough, Fitch and Littleton now comes to *overlie* the Orfordville. To keep strictly within the hypothesis adopted, the Ammonoosuc sequence here must be allochthonous. This is entirely compatible with its western boundary which, a little farther north, becomes the Monroe thrust. Still farther north, the Clough outlines a west-facing imbricate structure. Thus, wherever the Ammonoosuc group comes to overlie rocks of Gile Mountain/Orfordville/Littleton lithology, the contact must be tectonic, and related to the Monroe thrust; for in such cases there will be pre-Taconic rocks overlying Lower Devonian.

A further consequence of the present hypothesis is that the Shaw Mountain correlates with the Glenbrooke conglomerate; hence all rocks overlying the latter should be Siluro-Devonian. As Onondagan fossils are believed to occur in the Glenbrooke limestone (A. Wilson, quoted by Cooke, 1950), the St. Francis may well be Middle Devonian, in part.

CONCLUSIONS

The above deductions follow directly from strict application of the stated hypothesis to the whole of the region directly concerned. In the opinion of the writer, no major inconsistencies are encountered in this reasoning while a number of existing difficulties are removed. Further-more, in accepting Devonian age for the St. Francis, its correlatives and the Orfordville formation, it beomes possible to build a consistent framework for the structure and stratigraphy of the region.

The writer must apologize for discussing the relationships of a formation (the Orfordville) with which he has practically no field acquaintance. However, the position of the Orfordville is of importance in any attempt to assign an age to rocks of the St. Francis group. He has therefore felt justified in discussing the Orfordville formation in the light of the available evidence, and would like to acknowledge a debt of gratitude to Professors Billings and Thompson who are familiar with the problems concerned.

STRUCTURE

Introduction and Structural Setting

The major structural features of the Lyndonville quadrangle are the Brownington syncline (Doll, 1951, p. 51–52) and the Willoughby arch. The Brownington syncline extends from north to south throughout the length of the quadrangle (Pl. 1) and has as its core the lower part of the Gile Mountain formation. It is flanked on the west by a homoclinal zone of steeply-dipping Waits River beds and on the east by a broad, flat arch of Waits River sediments here called the Willoughby arch. The latter corresponds to the Danville "anticline" in the St. Johnsbury quadrangle (Eric, 1942). The eastern flank of this arch grades into a wide, steeply dipping homoclinal zone which extends into the Burke and Littleton quadrangles (Billings, 1937, Eric, 1942).

During his work the writer became aware that the above-mentioned arch might well be a structure corresponding to the domes of southeastern Vermont (Doll, 1944; White and Jahns, 1950; Lyons, 1955). Accordingly, the elucidation of the nature of the arch in the eastern part of the quadrangle assumes importance.

Terminology

There still is a great deal of ambiguity in the terminology of structural geology which has not always kept pace with conceptual advances. For this reason it is desirable to define the sense in which certain structural terms are here used.

Plunge of a linear feature: the angle between the line and the horizontal, measured in the vertical plane.

Trend of a linear feature: the bearing of its projection on to the horizontal plane, measured from the geographic meridian.

Attitude of a linear feature: its position in space, as defined by plunge and trend.

Attitude of a planar feature: its position in space, as defined by dip and strike.

It was found desirable to use a term which would convey the meaning of the German *Vergenz* (Stille, 1930). Accordingly, an asymmetric anticline is said to *face* in the direction towards which its apex is turned; similarly, an asymmetric syncline faces in the direction towards which it opens. The terms, anticline and syncline, are here used in their normal, stratigraphic sense, reserving the terms synform and antiform (Bailey) for purely structural forms. "Facing" of folds has been used before; Bain, (1931) used the term, but did not define it or use it consistently. Cummins and Shackleton (1955, p. 31) defined it in the sense here adopted, but the writer feels that their definition is too precise, while the usefulness of the term is enhanced by keeping the comparative freedom of Stille's original meaning of *Vergenz*.

Lithologic terms: In the sections on structure and metamorphism, rocks will usually be referred to by terms denoting grain size rather than composition. Thus, *pelites* are rocks with an original grain size of less than 0.005 mm. These correspond to the argillites, and a few of the impure quartzites of the stratigraphy section. *Siltstones* are rocks, with a grain size between 0.005 and 0.05 mm. These correspond to the quartzites of the stratigraphy section. The limestones are all wholly recrystallized, and are therefore referred to as *marbles*.

Major Structural Features

The *Brownington syncline* is a southward continuation of Doll's (1951) syncline of the same name. Its synclinal form, based on indirect evidence in the Memphremagog quadrangle, could be fully substantiated in the Lyndonville and adjacent quadrangles by two lines of evidence:

1. The band of Gile Mountain formation in the area has nearly vertical dips in the west and very shallow westerly ones in the east, suggesting a synclinal trough between.

2. The trough of the syncline is exposed and can be traced out in Danville and Walden townships (St. Johnsbury quadrangle).

Along the same trend, in the East Barre quadrangle, another band of Gile Mountain was found by White and Jahns (1950). There is no reason why this should not be a continuation of the Brownington syncline. This means that the regional plunge is northward in the St. Johnsbury quadrangle and southward where the Gile Mountain reappears in the East Barre quadrangle. The intervening area is evidently a dome or culmination; hence the outcrop pattern of the granites in this area is of some interest (Pl. 3).

The attitude of the beds (Doll, 1951, Pl. 1; and this report, Pl. 2) shows that the syncline is strongly overturned to the east. The bottom of the syncline in section CC' is apparently shallow because this is the lowest exposed part of the fold, while the axial line of the syncline lies farther west.

The Brownington syncline is one of the very rare east-facing major folds in a region of dominantly west-facing structures.

The *Willoughby arch* is the most interesting structure in the area. It has shallow dips adjoining the Brownington syncline, a broad, flat top (Pl. 1), and a steep eastern flank, mainly outside the Lyndonville quadrangle. The construction of Section CC′ has brought out thickening of the Barton River on the east flank of the arch. Field observations show that this thickening is due mainly to an increase in calcareous lithology. Thickening of the Barton River is even more conspicuous in the Burke quadrangle, where it has been confirmed by B. Woodland (personal communication, 1955). Thus, the beds on the crest of the arch have thinned, while on at least the eastern flank their thickness has increased. This is compatible with vertical arching, not with tangential shortening. The arch can be followed northeastward into the Island Pond quadrangle, where the Gile Mountain contact closes across its top. Extensive areas of granite here partially obscure the relationships. West of Island Pond the arch ends abruptly with a thick sequence of Gile Mountain, dipping vertically and striking almost at right angles to the trend of the arch. The strike of these vertical beds is parallel to what appears to be an important trend across the Island pond quadrangle at N 55° W, in straight alignment with the Monteregian intrusives, the Pilot Range Complex, New Hampshire, the Litchfield syenites, Maine, as well as an unusual amphibolite in Charleston, Vermont. Geologic mapping in the Island Pond quadrangle may reveal alkaline igneous rocks there along the same trend. This important regional alignment is here called the Brighton line, after the town of Brighton in Essex County.

To the south the Willoughby arch broadens and merges into a dome or culmination with westerly trend near Danville and Cabot. The *Perry fault*, along Willoughby brook, cannot be seen in outcrop. However, transcurrent displacement along the fault is readily inferred from the outcrop pattern near Perry station, Canadian Pacific Railway. The

Waits River/Gile Mountain contact is appreciably displaced in a left-lateral sense (northside displaced west).

On the map, the lateral displacement by the Perry fault is obscured, because the Waits River/Gile Mountain contact swings back to very nearly its original alignment northeast of Mt. Hor (K 1532). This is due to the steep, high southern slope of Mt. Hor, which brings the contact to a much higher level on that side of the fault and, hence, farther to the east (note prevailing shallow westerly dips). Moreover, near Lake Willoughby the contact begins to close eastward over the plunging nose of the Willoughby arch (maps, Pl. 1 and Pl. 3).

The air photograph (Pl. 5) gives an excellent picture of the curving of the bedding along the Perry fault. There is a strong suggestion that much of the displacement is taken up by a zone of flowage with a shear zone rather than by a single fracture. The Perry fault parallels the major regional *ac* joints rather closely; several of these joints show clearly in the photo. The tectonic map, Pl. 2, shows many air photograph linears interpreted as fractures; but apart from the Perry fault there is no evidence that any of these are associated with appreciable displacements.

The gap at Lake Willoughby appears to be due to a shear zone along a joint system which affected the granites more than the sediments. Its origin is discussed under Physiography.

Related Major Structures in Adjoining Areas

As mentioned above, the Brownington syncline continues into the Memphremagog quadrangle to the north, and a similar syncline of Gile Mountain appears to correspond to it in the area investigated by White and Jahns (1950), farther south. In the St. Johnsbury quadrangle, the Brownington syncline becomes very shallow, and the Waits River/Gile Mountain contact outlining it closes north of West Danville.

West of the Lyndonville quadrangle, the St. Francis group is essentially steeply homoclinal. Farther west, in the Beauceville formation, an anticline appears which can be followed north beyond Bunker Hill, Que. (W. M. Cady, personal communication, 1956) (regional map, Pl. 3). This anticline disappears toward the south, and in central Vermont there is no repetition from the western zone of Gile Mountain to the Green Mountain axis (Currier and Jahns, 1941).

The Willoughby arch has been recognized as an anticlinal structure by Richardson (1906) and Eric (1942) in the St. Johnsbury quadrangle. White and Jahns (1950) evidently picked up the same structure in the

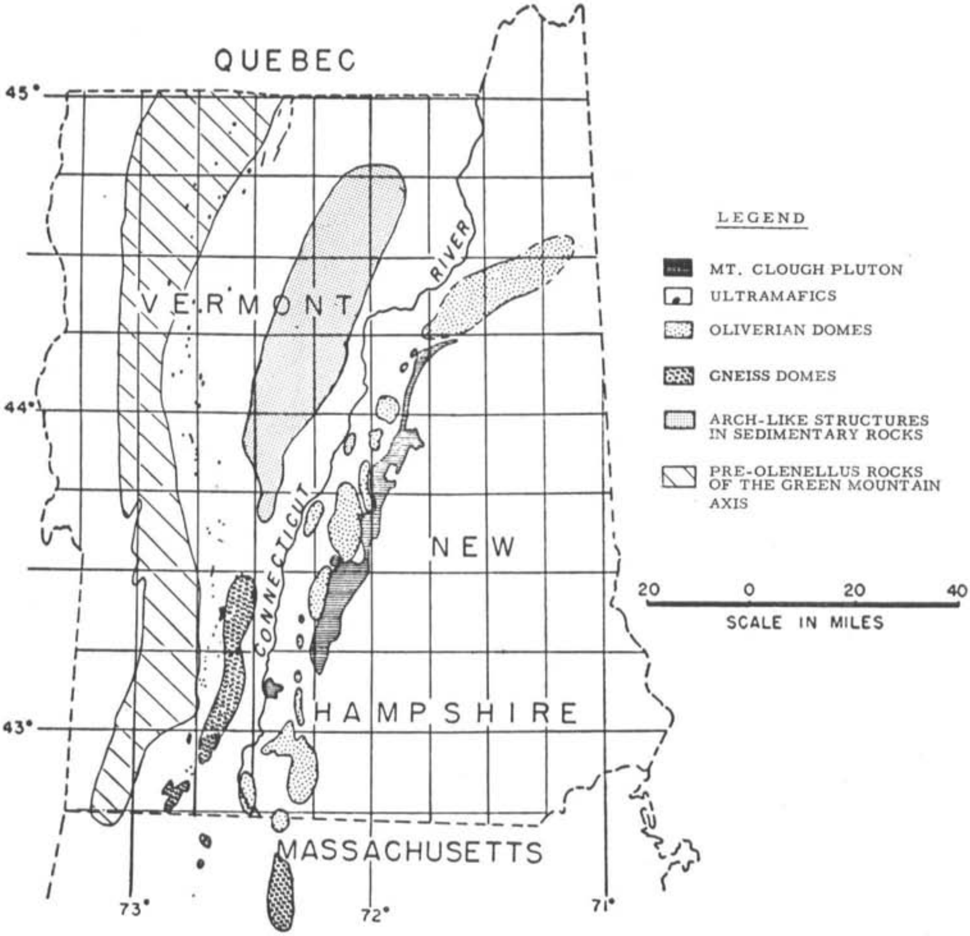


Figure 3. Domal Structures in Northwestern New England. Modified from Lyons (1955), Fig. 4.

Woodsville quadrangle to the south, and they and Doll (1944) also described it in the Strafford quadrangle. Lyons (1955) described the continuation of the arch in the Hanover quadrangle. A modified version of his Fig. 4, a sketch map of domal structures in New England, is reproduced as Fig. 3.

Thompson (1952) observed marked thinning in the mantling formations of the Chester dome.

East of the Lyndonville quadrangle there is another homoclinal zone, as far as the Monroe line which separates the St. Francis group from the New Hampshire sequence according to Eric, White and Hadley (1941). These authors interpret the Monroe line as a fault. Evidence for a faulted contact seems good in the Littleton and adjoining areas. But farther south the relationship is controversial; this writer, therefore, prefers to refer more non-committally to the Monroe *line*, agreeing in this usage with J. B. Thompson (personal communication, 1955).

In southern Vermont the zone of doming is considerably narrower than in northern Vermont, for here the Green Mountain anticlinorium and the White Mountain plutonic terrane approach to within 25 miles of one another (cf. Pl. 3; and Billings, Rodgers and Thompson, 1952, Pl. 2). It is no coincidence that the domal structures of southern Vermont are far more pronounced and more intensely deformed than corresponding structures in the Lyndonville and adjoining areas, where doming was less confined.

Minor Structural Features

Minor structures are of great importance for the interpretation of the movement pattern. The geometry of the elements concerned will be considered first, and will be followed by a kinematic analysis. Finally, an attempt will be made to draw conclusions as to the dynamics of rock deformation in the area.

The following coordinates and fabric designations will be used, following Sander (1948) and Cloos (1946, p. 6.)

b is the axis of major folding.

$b^{\prime}$ is any local fold axis.

B is any axis of internal or external rotation.

a is perpendicular to b within the *plane* of movement, ab.

c is normal to ab. The ac plane is the plane of maximum deformation, and often a plane of symmetry.

s-planes are mechanically preferred parallel surfaces of a fabric.

The relationship of the minor structures among themselves and toward the major structures is summarized in Fig. 4.

Geometry

Planar elements: (*a*) *Foliation*—Foliation in a rock is here defined as due to the presence of parallel planar elements following Fairbairn (1949, p. 5). Foliation may be due to sedimentation (Sander's Anlagerungs-

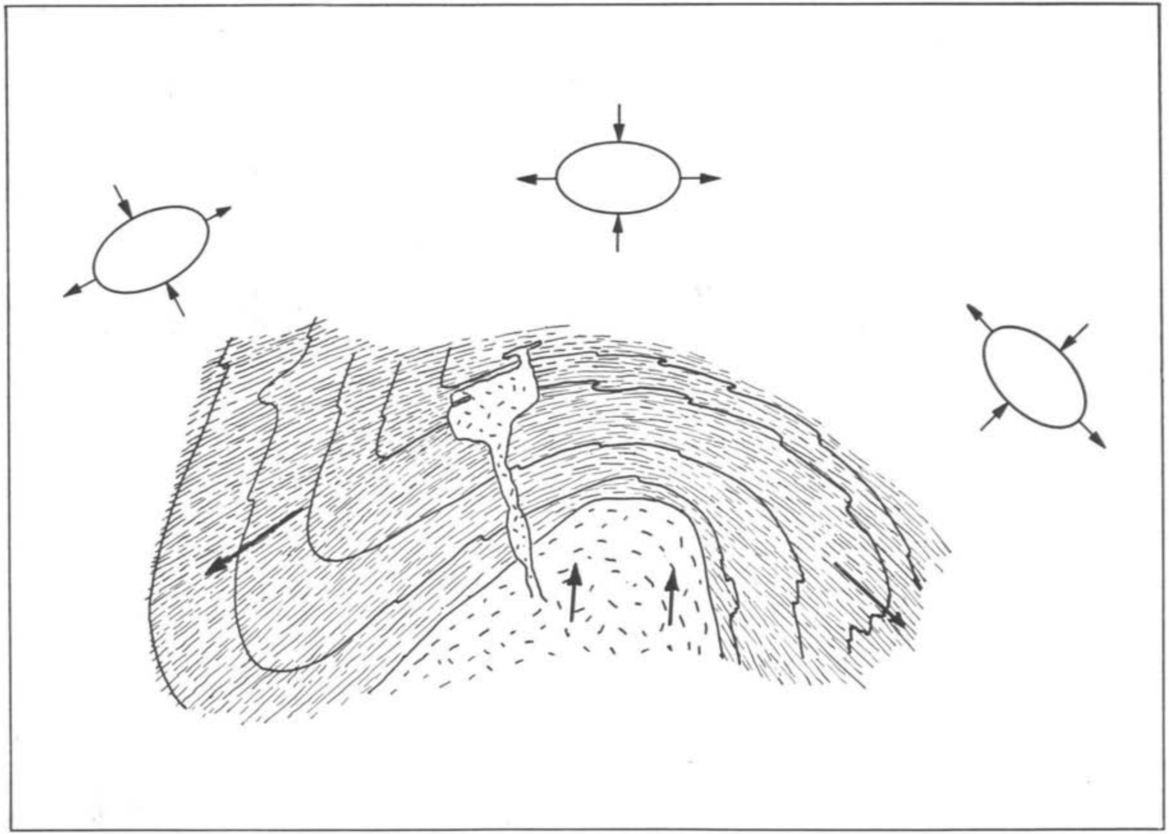


Figure 4. Generalized diagram showing relations between major and minor structures in the Lyndonville area. Heavy lines: bedding; fine, discontinuous lines; late secondary foliation.

gefüge, e.g. fissility in shales), to mineral parallelism (schistosity and slaty cleavage), or to closely-spaced slip planes: "slip cleavage" (Dale), "false cleavage" (Behre), "fracture cleavage" (Leith). There are gradations between all of these, and the Lyndonville area affords a number of good examples of foliation marked by a combination of any two of the above criteria.

The term foliation as here used comprises therefore both primary structures due to bedding, and structures of secondary origin. Schistosity is the result of recrystallization, while cleavage is a foliation due to mechanically produced planes of parting

(1) *Bedding*. The only safe way to identify bedding in the Lyndonville area is by lithologic contrast. Schistosity often simulates bedding to a remarkable extent, as shown, for example, on Pl. 7.

(2) *Schistosity* is that foliation which is due to the presence of more or less platy minerals. In the Lyndonville area the controlling mineral in pelites is sericite; in quartzites it is biotite. The foliation of the quartzites is frequently accentuated by quartz stringers a few millimeters thick. These features tend to give foliation in quartzite the appearance of bedding (Pl. 7). Marbles are sometimes foliated by parallelism of micas, but more often the foliation is produced by intimate banding of calcite and quartz layers on a microscopic scale. This is accentuated by a noticeable flattening of quartz and calcite grains parallel to the foliation planes. Narrow stringers of calcite and/or quartz parallel to the foliation are characteristic of the siliceous marbles; this gives the foliation of the marbles its macroscopic expression. The most immediately striking feature of the marble foliation is that it often delineates considerable internal deformation within the rock (see Fig. 5), while pelites in contact with the marbles show little or no disturbance. Frequently, even with the most intense deformation within the marbles, their contacts with other lithologies are undisturbed. More often than not the foliation within the marbles is discordant towards their contacts with other lithologies. This discordance is usually at a very small angle (2°–5° is a common range), but it is persistent. In thick bands of marble, undeformed foliation has the appearance of bedding.

East of the Lyndonville quadrangle, in the lower metamorphic grades of the Littleton quadrangle, the sericite schistosity of the argillites passes into "slaty cleavage." "Slaty cleavage" here merely refers to schistosity when it gives the rock the characteristic appearance and

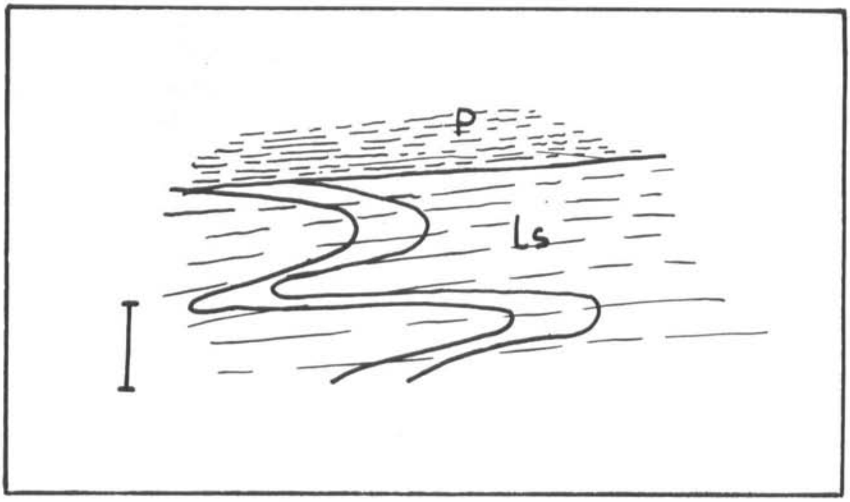


Figure 5. Minor folding in marble (ls), in tectonically unconformable contact with pelitic schist (p). In Fall Brook, south of Squabble Hollow School. East is at right. The short vertical line indicates one foot.

cleavability of slate; structurally it does not differ from the sericite schistosity of the Lyndonville quadrangle.

On the arch, and east of it, the sericite schistosity is almost invariably parallel or sub-parallel to bedding. This is a most characteristic relationship in the Lyndonville quadrangle, and in neighboring areas (Eric, 1942; White, 1949). Exceptions occur at comparatively rare isoclinal folds, whose axial planes are usually parallel to bedding.

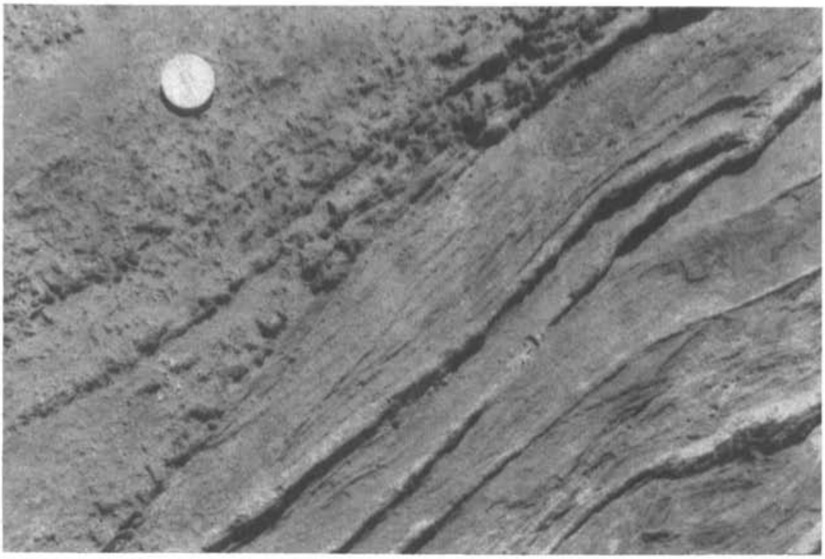
What is cleavage in the lower grade metamorphism is represented by schistosity in higher grades (cf. White, 1949); Pl. 8 shows both sericite schistosity and slip cleavage.

(3) *Slip cleavage.* In many cases the sericite foliation in the pelites is deformed by what Dale (1898) has called slip cleavage. The phenomenon is well illustrated in Pl. 7 and 8. This type of foliation is common in the pelites, where it often transects the sericite foliation; but it does not always visibly persist in marbles. It is usually expressed by biotite foliation and quartz veining in quartzites (cf. Pl. 7). In pelites, mica porphyroblasts in higher metamorphic grades may become oriented parallel to slip cleavage, and thus induce a second schistosity. In the Lyndonville area this second schistosity was never seen to entirely obliterate the

a. Foreground: staurolite schist, with staurolite porphyroblasts growing preferentially along cleavage, and along pelite-siltstone contact.
Background: siltstone (quartzite), with foliation emphasized by quartz veining (which partially continues into pelite). Note the manner in which the foliation cuts across the stratigraphic contact.

b. Cleavage (marked by ptygmatic veins) displaced by differential slip along bedding planes.

PLATE 7. Exposure (E 0508) illustrating the later structural history of Gile Mountain rocks.



c. Close-up of relations as in (a).

PLATE 7. Exposure (E 0508) illustrating the later structural history of Gile Mountain rocks.

original sericite schistosity; but in east-central Vermont White (1949) found schistosity due to development of aligned mica porphyroblasts completely masking the sericite cleavage. Visible expression of the slip cleavage appears to depend on the presence of earlier schistosity. Pl. 13 shows how the presence of sericite schistosity in a rock immediately brings out the cleavage. Phlogopite marbles with dimensionally oriented calcite appear to be an exception. The slip cleavage invariably terminates against predominantly granular rocks with subordinate or no mica. Evidently the deformation expressed by the cleavage is taken up in more granular rocks by lattice deformation in the grains, by intergranular movement and readjustment, or by movement along polycystalline planes with no crystallographic control ("fracture cleavage"). If rupture was one of the mechanisms involved in the deformation of these rocks, it has since been completely healed by re-crystallization.

(b) *The Pattern of the Secondary Foliations, and its Relation to the Minor Folds*—The sericite schistosity is a bedding schistosity, except where it acts as "axial plane cleavage" to a few isoclinal dragfolds whose

a. At G2728; looking northeast: Slate in Gile Mountain formation, showing bedding (emphasized by black line), forming one of the early minor folds, with "normal" drag relations. Sericite schistosity (early) is parallel to the steep limb of the fold. Slip cleavage (late) is in deep shadows, and dips toward the right of the picture.

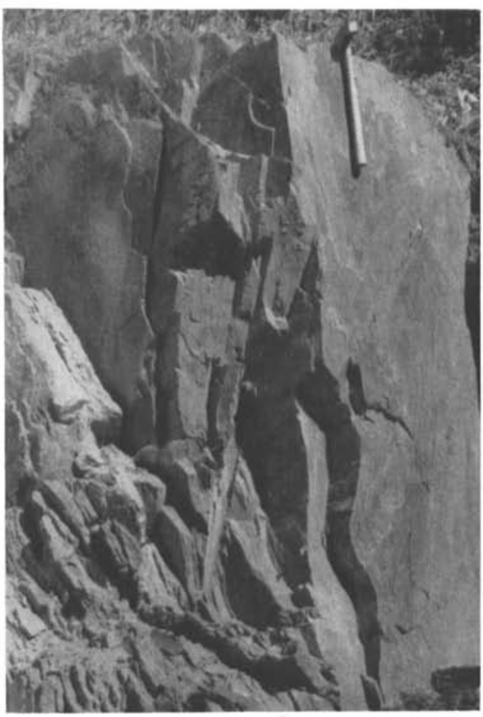


PLATE 8. Some minor structures in the Littleton quadrangle.

b. Close-up of same locality.

c. At G2206 (opposite Graves School): bedding displaced by slip cleavage forming "chevron folds" of the late generation of minor folds.

PLATE 8. Some minor structures in the Littleton quadrangle.

limbs are invariably parallel to the bedding. These minor folds face toward the top of the arch and out of the syncline in the normally accepted manner. On the eastern flank of the arch they are best exposed in the St. Johnsbury quadrangle, the Littleton quadrangle (Eric, 1942) and in east-central Vermont (White and Jahns, 1950).

The slip cleavage and later schistosity (here grouped as "later secondary foliation") are parallel to bedding near the crest of the arch, but dip away from it on both its flanks, regardless of the attitude of the beds (Fig. 4). The slip cleavage invariably deforms the sericite schistosity. The later secondary foliation acts as "axial plane cleavage" to a second generation of minor folds which deform the earlier sericite schistosity. These later minor folds face *away* from the crest of the arch, down the dip of the later foliation. Thus they correspond to what Bain (1931) has termed "flowage folds."

This general pattern is somewhat modified by the Brownington syncline, which appears to have acted as an "inhibitor" to flowage down the western flank of the arch. The attitude of the cleavage here and of a

PLATE 9. Bedding-cleavage relations and *b*-lineation, pelite-marble contact (pelite at left). Note concentration of tremolite-actinolite porphyroblasts in marble, near contact. At A 0458, looking north. Waits River formation.

few minor folds of the later (flowage) generation show that the deformation pattern is essentially analogous to that on the east flank. But cleavage is less pronounced than on the eastern flank, and often assumes divergent attitudes; late minor folds are also almost absent. Another anomaly is associated with the western flank of the syncline: the early sericite schistosity is almost absent from its steep limb. Evidently the rocks on that limb reacted differently to the deformation that caused the early schistosity; this suggests that the syncline had already formed, or was forming, when the early schistosity developed.

Fig. 6 is a reproduction of White and Billings' (1951) Fig. 8. This is a remarkable illustration of the relationship in map pattern of different types of cleavage. It was not possible to find this so clearly illustrated within the Lyndonville quadrangle. However, as a result of a brief reconnaissance in the eastern part of the town of Lyndon, the writer believes that very similar relationships will be found there.

(c) *Joints*—The most common and persistent joints in the area are steep to vertical joints striking near N50°W to N55°W. These are clearly shown on air photographs, and are represented as air photograph linears

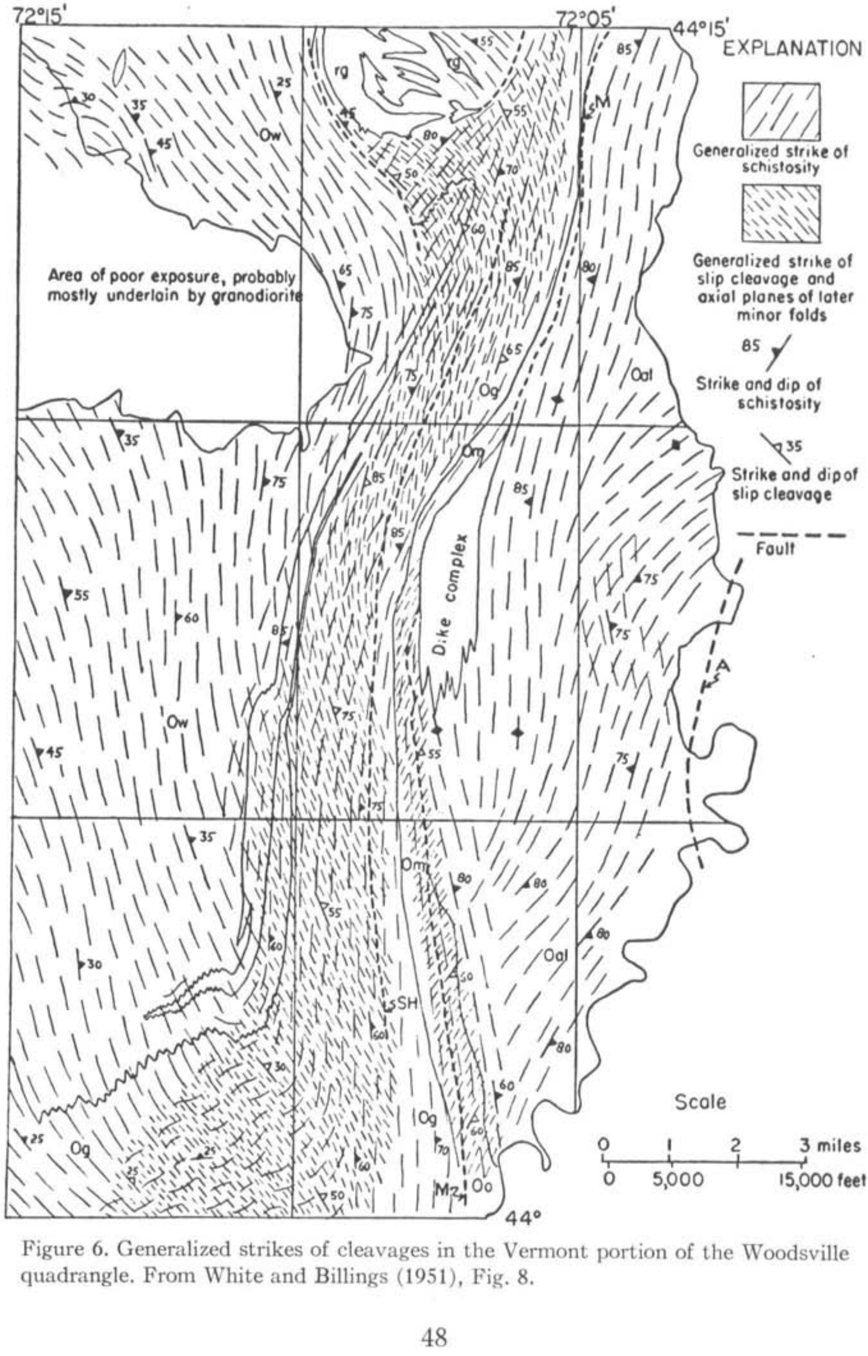


Figure 6. Generalized strikes of cleavages in the Vermont portion of the Woodsville quadrangle. From White and Billings (1951), Fig. 8.

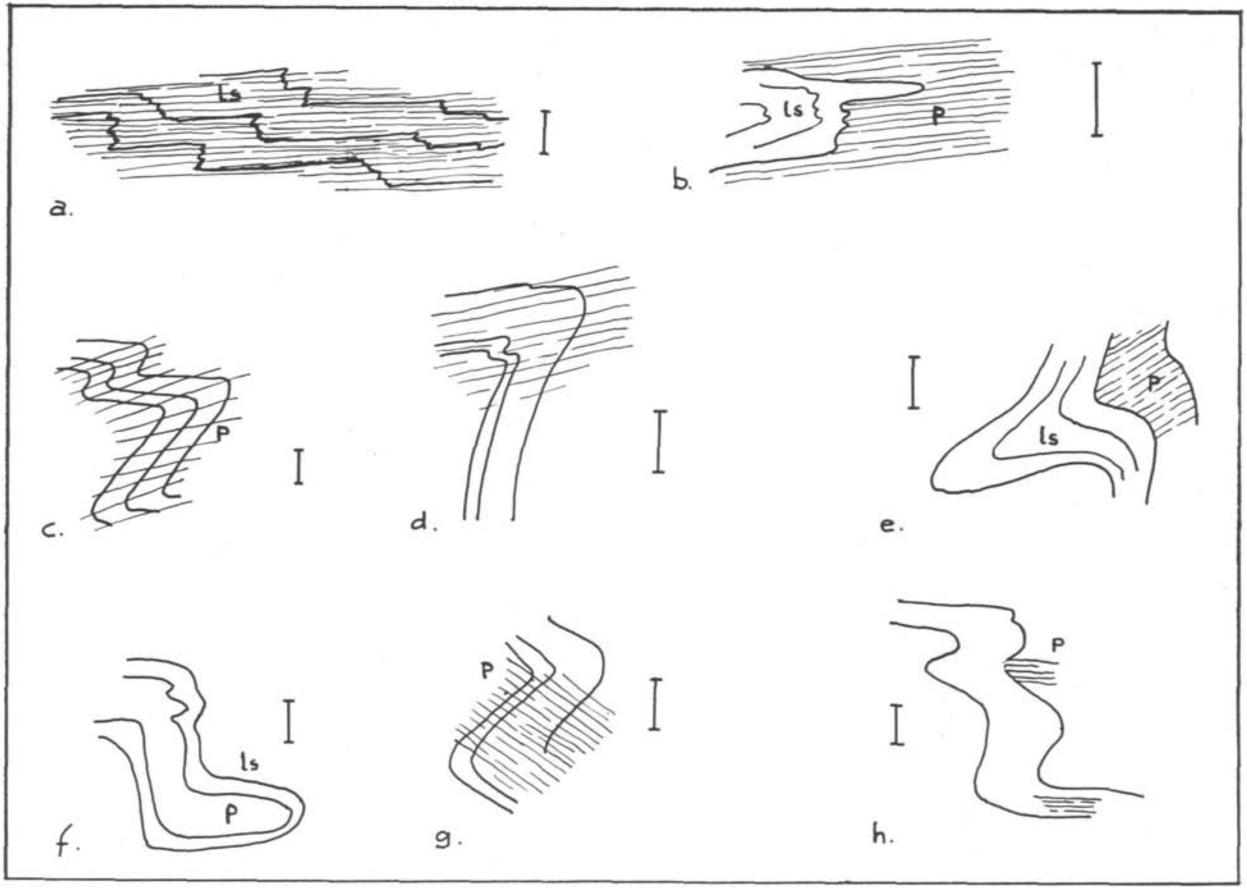


Figure 7. Typical minor folds of the second (doming) phase. West is at the left of the sketches throughout. The short vertical lines indicate the length of one foot. Firm lines represent bedding, fine lines cleavage. p = pelite, 1s = marble

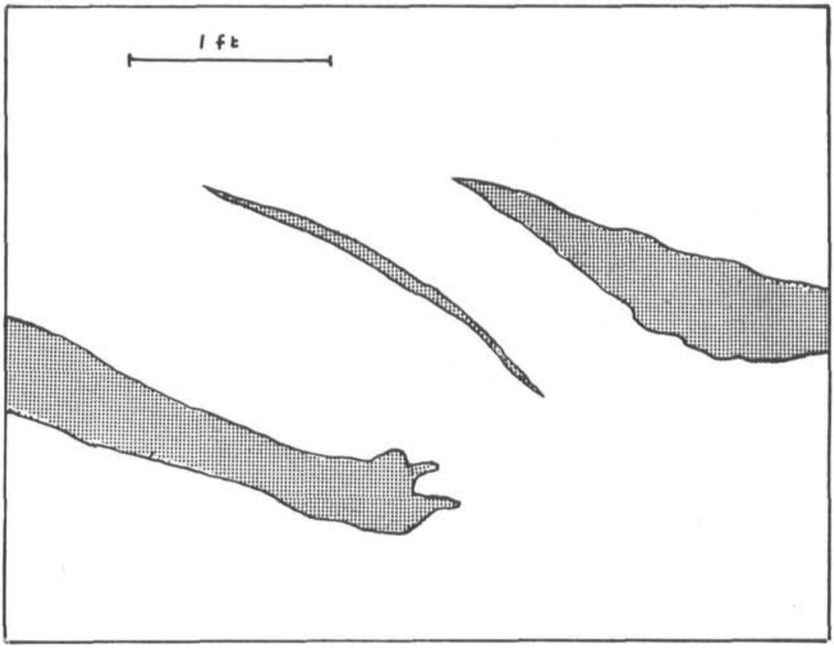


Figure 8. Post-kinematic basalt in en echelon tension joints. Calendar Brook, near edge of quadrangle.

on Pl. 2. A number of these joints are prominent on Pl. 5. They are rarely filled, though at F 4018, in Calendar Brook, basaltic dykes a few inches thick follow the joints. Fig. 8 shows how these fractures may become offset in an en echelon pattern. The granite of the Black Hills has a contact along one of these joints. The other common direction is N50°E to N60°E. Post-kinematic basaltic dikes appear to favor this set. Quartz veins are also found along these joints.

A third set is almost entirely confined to the granites. Like the other two sets, it is generally steep to vertical. In strike it ranges near N20°W to N30°W. The two prominent glacial lakes, Crystal Lake and Lake Willoughby, are aligned parallel to joints of this set.

Linear elements: (a) *Fold axes*—The trend of most minor folds is parallel to that of the Brownington syncline. The steep Waits River/Gile Mountain contact on its west flank is the only reliable guide to the axial trend of the Brownington syncline. It is confirmed by the study of air photographs. The trough of the syncline cannot be located accurately, and would in any case not be a good guide to the axial trend.

The trend of the Brownington syncline roughly parallels the flat top of the Willoughby arch and is taken as the regional *b* direction. The plunge, as recorded on minor folds parallel to *b*, is about 20°N in the northern part of the quadrangle, but changes to horizontal, and even a few degrees south in the town of Glover. At the same time the axial trend changes from NNE to NNW. In Sheffield the trend changes back to NNE and the plunge is northerly once again, though much shallower than near Barton (tectonic map, Pl. 2). Near the southern border of the map the trend becomes almost due N.

There are a number of deviations of trend. Just south of Point B 0000, axes of minor folds trend almost at right angles to the regional *b*, and plunge down dip. In the village of Lyndon, below the bridge at C 3311, axes of minor folds are oriented completely at random within one exposure. At Lyndonville dam, just off the extreme SE corner of the quadrangle, fold axes plunge steeply down-dip (see Pl. 16). These irregularities are not surprising, if arching is considered as the prime cause of deformation. In structures with dominant vertical stress, such as diapirs (Wegmann, 1930), $b' \perp b$ is common. The Willoughby arch seems to be an attenuated structure of this kind; thus $b' \perp b$ is here confined to prominent changes in trend. The widening of the arch south of Stannard, and the conspicuous bends near Barton and Lyndonville are such changes in trend.

(b) *Intersections of s-planes*—Intersections of foliations, and of foliations and bedding, are nearly always parallel to the minor fold axes, wherever these features occur together. Often these intersections show as small puckers in the dominant foliation. However, not infrequently there is divergence between intersection lineations and fold axes, as in Pl. 10. Sometimes there are two intersection lineations (revealed by puckerings or crenulations), one parallel to the *b* direction of the fold axes, and the other appreciably steeper.

(c) *Boudinage*—Boudinage is rare, and was observed only in Roaring Brook, at G 2752 (Pl. 11). Outside the Lyndonville quadrangle, boudinage was observed in the Passumpsic river just north of St. Johnsbury Center, and also in the Barton River at the locality described by Doll (1951, p. 66).

In each case, it is the more competent pelite which is stretched between beds of less competent siliceous marble. The long axes of the

PLATE 10. Road-cut in Rt. 122, west of Wheelock village (E 2804): Divergence of axis of minor fold and lineation. Lineation produced by cleavage planes intersecting a curved bedding plane. The lineation curves with the bedding plane, suggesting that it had evidently developed before the fold had its present geometry.

PLATE 11. Boudinage. At G 2752. Competent pelitic schist stretched by flowage in incompetent marble. Quartz has crystallized in the construction. Quartz-calcite veins delineate the foliation in the marble. Hammer handle is parallel to glacial striations.

boudins plunge steeply down-dip. This agrees with the attitude of boundinage in the Chester dome (J. B. Thompson, personal communication 1956).

A stretching phenomenon related to boudinage was observed in a granitic sill in the Barton River at G 3024 (Pl. 12). The granite here was stretched and fractured, with limestone flowing into the notch.

It is not surprising that boudinage is encountered near a convex bend in the trend of the arch, while the down-dip fold axes are found where the mantling formations are crowded in a concave part of the arch trend (Pl. 3).

Grain Fabric: Strictly speaking, petrofabric analysis describes and measures the entire fabric of the rock, macroscopic as well as microscopic. Macroscopically, the attitudes of all linear and planar elements visible in outcrop or in oriented hand specimen are plotted on an equal-area projection (Lambert projection). In addition, the grain orientation of certain critical minerals may be measured and plotted in the same way on a statistical basis. Such grain orientation may be dimensional or by crystal lattice or both. In either case measurements of attitude are carried out on the universal stage. The lattice orientation is usually found by measuring the attitude of the optic axis; the normals to certain preferred crystal planes, such as cleavages and twinning planes, may also be plotted.

While dimensional orientation can be immediately interpreted as schistosity or lineation, the interpretation of lattice orientation where not directly related to dimensional orientation presents certain problems. Crystals may orient their lattices to growth along available planes, for instance, biotite (001) along sericite (001), or structures ("inherited" structure); they may present their least densely packed faces toward the direction of greatest stress; or they may attempt to orient their planes of least shear strength (glide planes) parallel to the greatest resolved shear stress.

The pattern obtained by plotting of petrofabric data is then related to the tectonic axes *a*, *b* and *c*, and an attempt is made to interpret this in terms of one or several movement patterns.

A preliminary microscopic fabric analysis on four specimens was carried out in order to see if the grain fabric would give any additional clues to the movement pattern.

Specimen O 24, collected in Miller Run at E 3010, is a contact between pelite and siliceous marble (see Pl. 13a). Sericite schistosity in the pelite

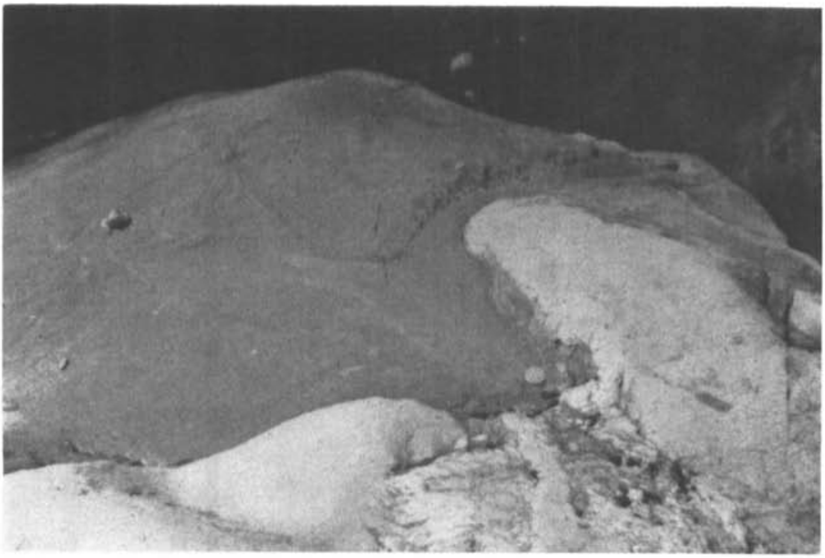
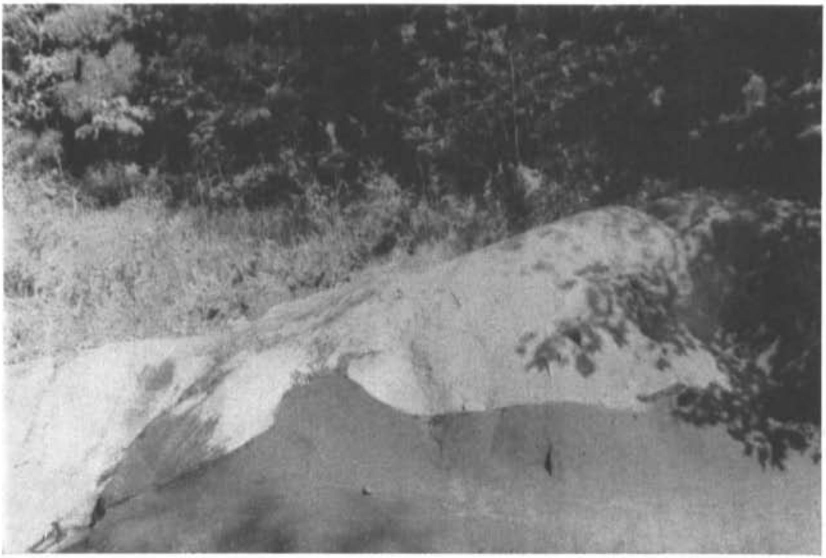
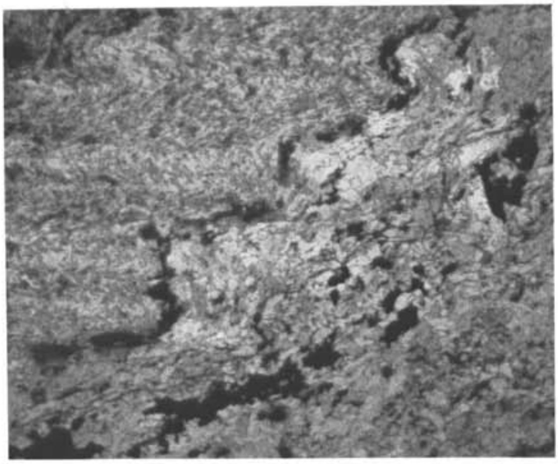
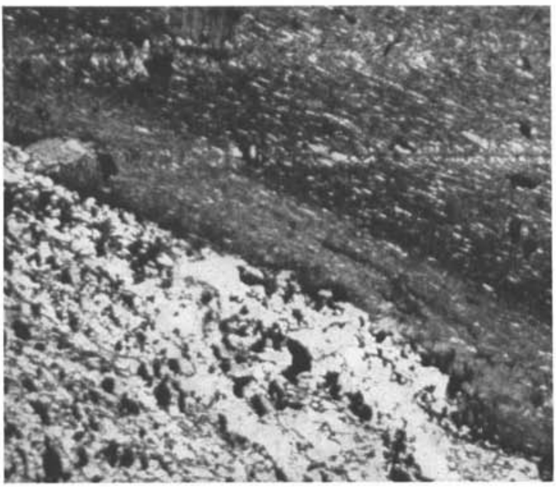


PLATE 12. Boudinage-like constriction in granitic sill, Barton River south of Glover village. Marble has flowed into a constriction in the sill, and quartz has crystallized in the granite at the constriction. Note tremolite/actinolite porphyroblasts parallel to contact in the marble.





a. Specimen O 24. Slip cleavage in pelite terminates against bedding schistosity in siliceous marble. Waits River formation.



b. Specimen L 27. Contact between slate and comformable quartzite lens, at G2728 in the Littleton quadrangle. Sericite schistosity in slate parallel to contract. Slip cleavage along 2 bands crosses the picture from left to right. Quartz segregates in the micro-folds, and produces intimate composition banding parallel to the cleavage. Gile Mountain formation.

PLATE 13. Foliation in thin section. Magnification 30X. Polarizer in position.

c. Specimen O 32. Contact between staurolite schist and re-crystallized siltstone (quartzite), from E 0508. Schistosity, outlined by biotite in the siltstone, and by sericite and biotite in the schist, cuts across the contact. What looks like a darker layer along the upper edge of the picture is a single large staurolite porphyroblast grown out into the siltstone. Gile Mountain formation. (Photographs by Rae L. Harris, Jr.)

PLATE 13. Foliation in thin section. Magnification 30X. Polarizer in position.

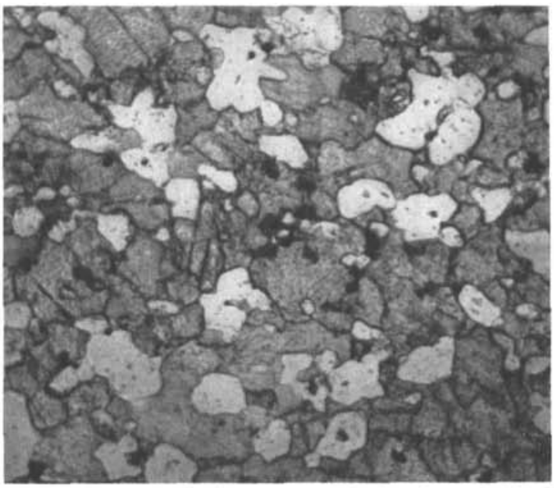


PLATE 14. Photomicrograph of siliceous marble. Specimen O 19. Magnified 30X, polarizer in position. (Photograph by Rae L. Harris, Jr.)

and calcite-phlogopite-quartz foliation in the marble are both parallel to the contact, and hence parallel to bedding. Slip cleavage in the pelite ends at the marble contact. Two thin sections at right angles were checked for dimensional orientation of quartz and calcite, using 50 grains in each case. There is distinct flattening parallel to the foliation of the limestone here taken as the *ab* plane, but no preferred linear dimension. The mean flattening (thickness over diameter) is 0.56 in quartz and 0.65 in calcite. In both cases the standard deviation is 0.15. Biotite porphyroblasts in the pelite do not show an obvious preferred orientation when examined in thin section. Typically, they carry as inclusions discordant relicts of the groundmass. These relicts show that the porphyroblasts have not been disturbed since their formation.

In specimen O 24 (Pl. 13a and Fig. 9a) both quartz and calcite *c*-axes were plotted in two sections mutually at right angles. (The intersection of bedding and slip-cleavage, as shown on the diagram, is taken as *b*, the slip-cleavage as the *ab* plane; hence *a* is at the center of the projection.) The quartz diagram, Fig. 9b, showed little preferred orientation. The calcite diagram (Fig. 9a) shows a distinct girdle of *c*-axes at about 60° to *ab*. Prominent maxima make angles of 30°, 50° and 66° with the pole of s_1 (=*ab*) while one maximum almost coincides with *a*. There are few cleavages and no twin lamellae in the thin sections.

Mica porphyroblast orientation was also measured. Wherever mica cleavages could be adequately measured on the universal stage, cleavage poles were recorded. In grains with cleavages at difficult angles, the optic axis was oriented.

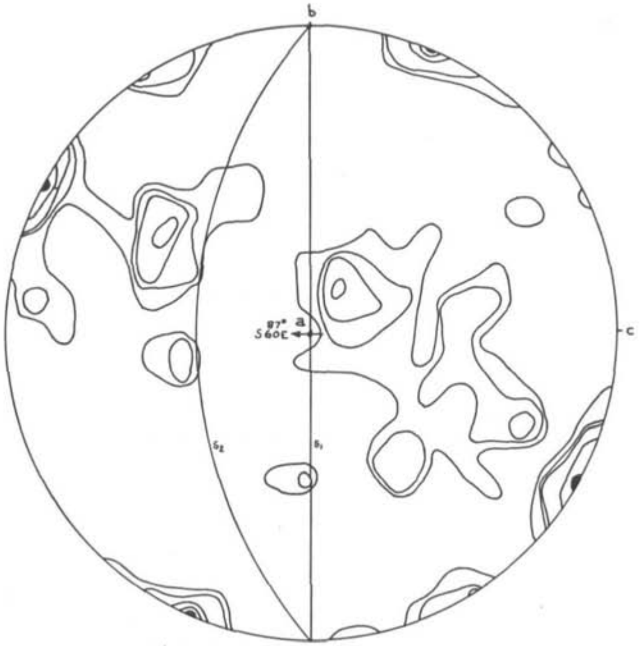
The mica diagram shows an *ac* girdle of cleavage poles and *c*-axes. Good maxima are developed normal to the long limbs of the sericite cleavage, s_2.

Axes of micro-folds in the sericite schistosity are parallel to the cleavage/bedding intersection.

Specimen O 19 (Pl. 14 and Fig. 9e) from Fall Brook at C 2041 is a siliceous marble similar to the calcareous part of O 24 but lacking mica. The foliation is taken as the *ab* plane, and the regional *b* as the fabric *b*.

The diagram shows imperfect orthorhombic symmetry. Two maxima make angles of about 50° with the pole of *s*, and there is another maximum at about 64° to the pole of *s*.

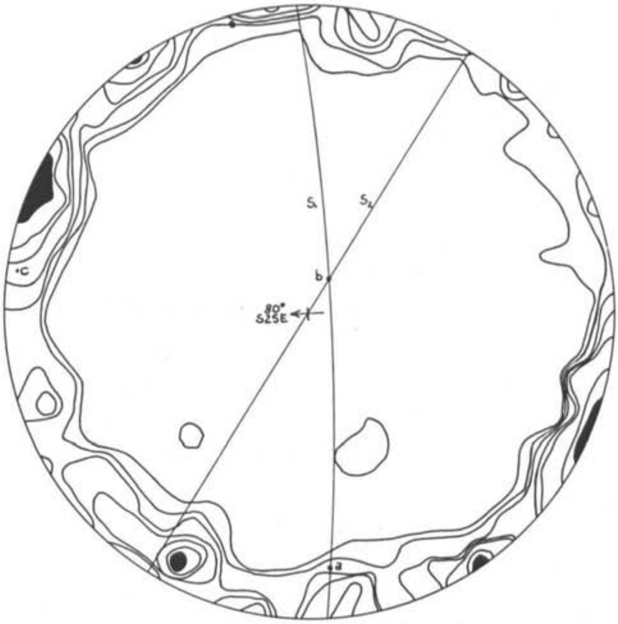
Specimen O 18 was collected in the St. Johnsbury quadrangle, at K 3649. The more prominent foliation plane (schistosity) is taken as *ab;* the intersection of the two foliations is the fabric *b*. This is down dip



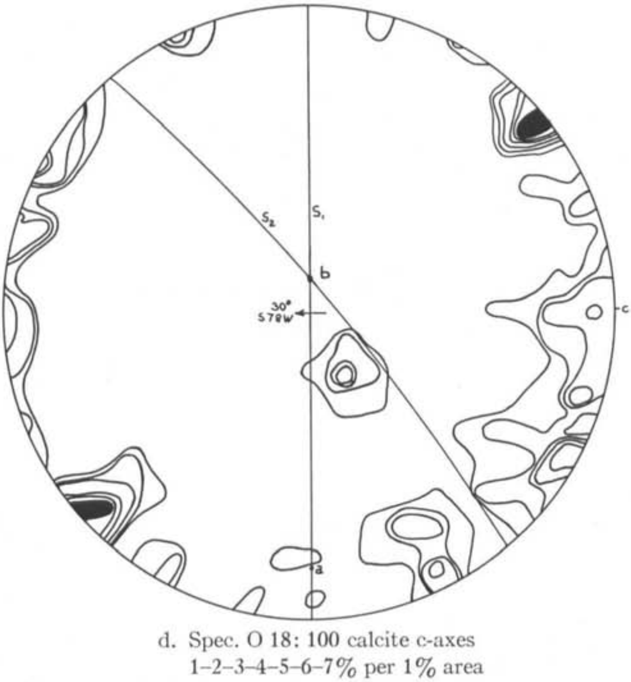
a. Spec. O 24: 100 calcite c-axes
1–2–3–4–5–6% per 1% area

b. Spec. O 24: 200 quartz c-axes
½–1–1½–2–2½% per 1% area

Figure 9. Petrofabric diagrams.

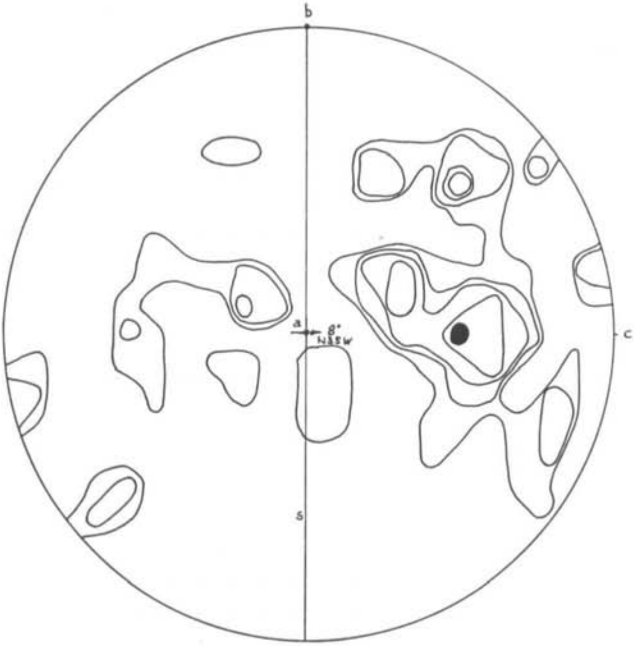


c. Spec. O 24: 200 mica (001) poles
1–2–3–4–5–6–7% per 1% area



d. Spec. O 18: 100 calcite c-axes
1–2–3–4–5–6–7% per 1% area

Figure 9. Petrofabric diagrams.



e. Spec. O 19: 100 calcite c-axes
1–2–3–4–5–6% per 1% area

Figure 9. Petrofabric diagrams.

in the field, i.e. at right angles to the regional *b*. In Fig. 9d a discontinuous *ac* girdle has prominent maxima at angles of about 30°, and 40° with the pole of *s*. This is the only thin section of a limestone in which cleavages were reasonably abundant. Most of these cleavages are parallel to the principal foliation. Twinning was not observed.

Interpretation of the grain fabric: Interpretation of preferred orientation is based on recent work by Turner *et al.* (1954). These authors showed that deformation of calcite crystals was most likely to take place by translation along $\{10\bar{1}1\}=r$; another possible translation direction given is $\{02\bar{2}1\}=f$. Gliding along $\{01\bar{1}2\}=e$ could only be explained as twin gliding, not as translation.

The angles between the glide planes and $\{0001\}=c$ of calcite are given by Turner et al. (1954, p. 886):

$$c \wedge r = 44\tfrac{1}{2}°$$
$$c \wedge e = 26\tfrac{1}{4}°$$
$$c \wedge f = 63°$$

The broad maximum, almost a small circle girdle, of specimen O 19 could well be related to translation parallel to *ab*, along *r* planes. All 3 specimens have such maxima with respect to the postulated *ab*, within 6° of the theoretical angle of 44½°. Specimens O 24 and O 18 have prominent maxima at 30° to s_1, which may be related to gliding on *e*. However, no twinning was observed.

The maximum at 66° to the pole of s_1 in specimen O 24, may be related to translation on *f*. Only in specimen O 24 is there a check of the attitude of slip cleavage planes against that of sericite schistosity planes. In this case the calcite orientation appears better related to the slip cleavage than to the limestone foliation, which is parallel to the sericite cleavage. Only the mica orientation seems related to the earlier fabric. This is probably due to control by the pre-existing sericite fabric—a mechanism which is generally recognized as valid (Fairbairn, 1949).

It appears, then, that the orientation of the calcite lattice is not directly related to the pronounced dimensional orientation inherited from the earlier deformation phase. Rather, the orientation pattern would seem to be related to later slip cleavage, which is so prominent in the pelite but which ends at the limestone contact (Pl. 13a). Much of the deformation of the calcite must have taken place along *r*-planes. Twinning along *e*-planes of suitably oriented grains probably occurred at the same time. The fact that no twin lamellae can now be observed must be taken to indicate recrystallization after the last deformation phase. In specimen O 18 cleavages may be due to slight post-kinematic slippage along foliation planes.

This fabric study was initiated as an exploratory project. Circumstances permitting, the writer intends to carry out further work to check and amplify the interpretations given, and, if possible, to investigate the fabric of the quartzites. It is interesting to note that in the quartzites, biotite orientation is dimensionally parallel to the (later) slip cleavage (cf. Pl. 13c).

The Movement Pattern

(a) *Criteria accepted as valid:* The *sense* of movement in any one segment of a rock structure can be deduced by several lines of evidence. Elongation of fabric elements was not observed in the Lyndonville area, and will not be discussed. The axes of most folds (*b*) are at right angles to a direction of tangential shortening (*a*); the axes of boundins are at right angles to a direction of stretching. Schistosity and cleavage, where not purely mimetic, are along planes of plastic flowage (*ab*) and hence denote thinning at right angles to those planes.[1] All occurrences known to the writer can be adequately accounted for by plastic flow on *ab*, both theoretically and in the field. In other words, cleavage and schistosity (non-mimetic) appear at right angles to the direction of maximum shortening. This is the original concept of Sorby and Sharpe. For a good recent summary, see Goguel (1945).

Conjugate fracture systems intersect in a line parallel to the *b*-direction at the time of their formation. Also, cleavage/bedding intersections are parallel to the *b*-direction at the time of initiation of the cleavage.

The glide planes of grain deformation are generally taken to be parallel to the *ab* plane at the time of deformation.

The *timing* of several succeeding phases of movement can sometimes be re-constructed from the deformation of *s*-planes, from deflections of *b*-axes (if according to a recognizable movement plan); and from the chronology of recrystallization (if correctly interpreted). The latter often assumes importance in petrofabric analysis.

(b) *Recapitulation of geometric premises*—In the Lyndonville area the axes of folds and lineations formed by intersection of *s*-planes are nearly all parallel to the regional *b* as previously defined. Important exceptions occur west of Barton village (down-dip boudinage) and in the south-eastern part of the quadrangle (fold axes and lineations plunging down-dip, i.e. close to the regional *a*).

There are clearly two successive generations of important *s*-planes: the sericite schistosity and the slip cleavage; the latter invariably deforms the former wherever both occur together.

In the Brownington syncline, the early minor folds face out of the syncline along each limb. East of the arch, mainly outside the area investigated, a number of minor folds related to the (earlier) sericite schistosity face toward the crest of the arch (Eric, 1942; White & Jahns, 1950). Schistosity is predominantly parallel to bedding on the crest of the arch, only cutting across bedding at the apices of minor folds. Slip cleavage dips away from the crest of the arch on both sides; related minor folds face away from the crest of the arch, principally on the east flank, but also on the west flank of the arch. And there is appreciable thinning of beds on the arch, while its east flank has an abnormal thickness of limestones.

(c) *Kinematic Interpretation*—The *Willoughby arch* is unlike a normal

[1] The writer does not subscribe to interpretations of cleavage as related to oblique shear directions because this view lacks field support.

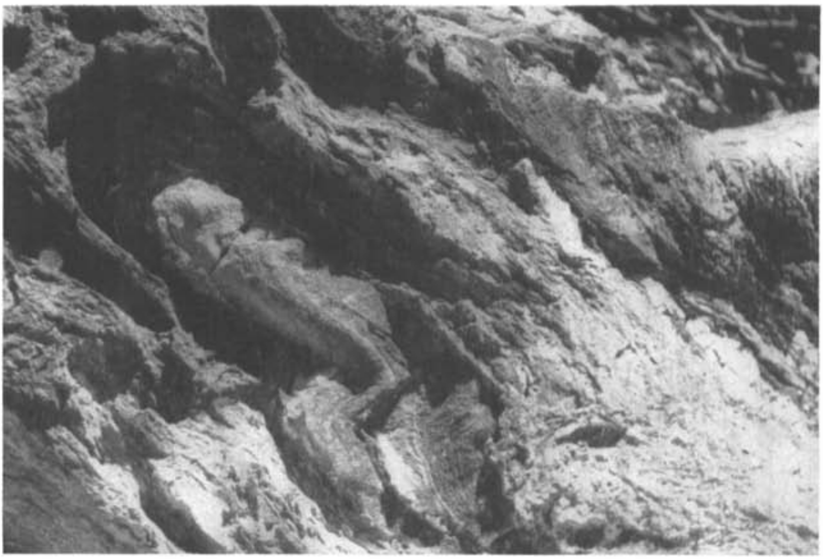
anticline. True, the earliest dragfolds show the normal pattern to be expected with an anticline, i.e. a relative movement toward the crest of the higher beds. But such folds are rare, and are superseded by later minor folds which face away from the crest of the arch and deform the sericite schistosity associated with the earlier phase of deformation. While the earlier dragfolds are strictly parallel to the regional *b* wherever seen, the later folds may locally diverge to varying extents (although they are normally parallel to *b*). Thinning of the Waits River formation on the crest of the arch, coupled with thickening of limestones on the east flank, down-dip facing of the later minor folds, and prevailing "wrap-around" cleavage together indicate that the rocks flowed away from the crest and down the flanks of the arch in a later phase of deformation.

The role of the secondary foliation in this style of deformation is of some interest. The analysis given above would seem to indicate that the slip cleavage is due to the arching. It expresses the planes along which the movement away from the crest of the arch took place. This movement could not be homogeneous laminar flow throughout, and so the east-facing minor folds developed. *But the ab planes were the dominant feature, marking a thinning and flowing-out, and the folding was quite incidental to the process.*

It would seem that the (earlier) sericite schistosity marked essentially a distributed thinning of the beds without directed flowage.[1] Where the bedding was oriented at an unfavorable angle to the main direction of thinning, no thinning, and therefore no cleavage developed. This is well illustrated by the steep limb of the Brownington syncline: absence of sericite schistosity here can best be explained by the above reasoning. The (later) slip cleavage was superimposed as soon as actual flowage away from the crest began to develop. Apparently the Brownington syncline took a passive part in the later stages of the arching; this is undoubtedly the reason why it is overturned to the east. The pattern of the minor folds (facing out of the trough) and their style suggest that

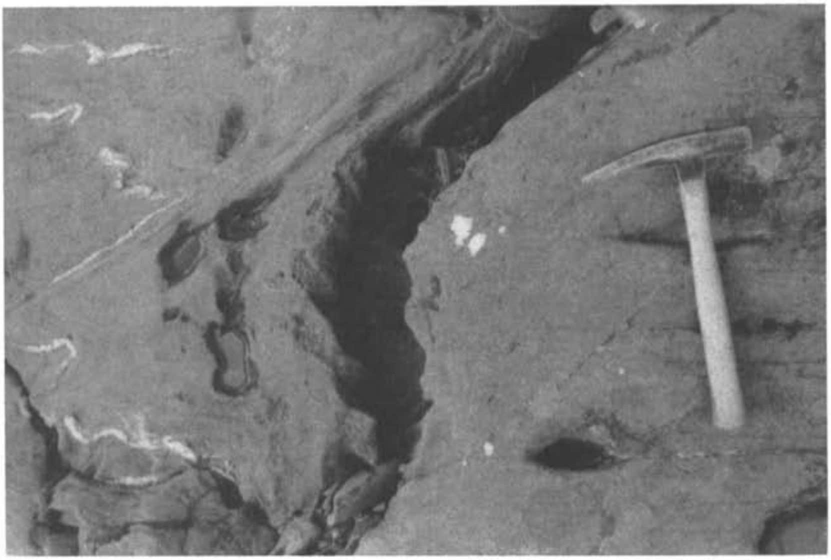
[1] This view is not far from van Hise's (1904, p. 716) interpretation of deformation in the mantling formations of the New England domes. Brock (1934) interpreted deformation in the Shuswap terrane of British Columbia in very much the same way. However, he was unable to give direct evidence that in the Shuswap terrane bedding planes acted as *ab* planes. Gilluly (1934) found such evidence by petrofabric analysis, but interpreted it differently. In eastern Vermont bedding schistosity becomes axial plane schistosity in isoclinal minor folds, leaving little doubt that here bedding schistosity represents *ab* planes of the earlier deformation phase. This adds weight to Brock's (*loc. cit.*) interpretation of deformation in the Shuswap terrane. In fact Brock's and Daly's (1915) descriptions suggest that the Shuswap terrane has good examples of mantled gneiss domes.

a. Below Lyndonville dam. Pelite in center and at extreme top and bottom. Limestone in between.



b. In Calendar Brook, at F 4018. Northwest is at left.

PLATE 15. Minor folds in Waits River formation.



a. Bedding outlined by quartz-calcite veins at left. Foliation visible below hammer.

b. Bedding brought out by a zone of slightly different composition, at center of picture. Foliation emphasized by quartz-calcite veins, left bottom of picture.

PLATE 16. Steeply plunging minor folds in Waits River Marble. Below Lyndonville dam, just beyond the SW corner of the Lyndonville quadrangle. Hammer handle points east.

they are normal dragfolds related to the formation of the syncline. It has not been possible to relate them to the later deformation.

Referring to Fig. 6 once again, and remembering the dominant northward plunge, this map pattern suggests rather strikingly the relation between the later slip cleavage and the east-facing flow-folding in profile, as revealed in the Woodsville quadrangle. Evidence for westward flowage off the arch in the Lyndonville area is limited (fold in Fig. 7e) but is confirmed farther south by structures described by White and Jahns (1950).

Outside the Lyndonville quadrangle, west of Island Pond, the conspicuous steep to vertical beds of Gile Mountain striking across the trend of the arch can only be accounted for as marking the abrupt northern termination of the Willoughby arch at the Brighton line. Only some form of diapirism can result in such a structure.

(d) *Sequence of Movements*—The time sequence sericite schistosity-slip cleavage has been an inevitable part of the foregoing discussion. Evidence of displacement of the former by the latter is good, both in the field (cf. Pl. 8) and miscroscopically (cf. Pl. 13). No direct evidence as to the amount of time-lag between these phases has been found.

Continued deformation caused distortion of the slip cleavage, essentially parallel to the bedding. At this stage, evidently, the metamorphism had taken effect; it caused a partial "healing" of the cleavage, while accentuating mechanical contrasts between lithologies. Thus, the cleavage planes became mechanically ineffective, both by metamorphism and through having become unfavorably oriented in the course of deformation. Movement was once more controlled largely by bedding. This is illustrated by the wildly deformed schistosity in many limestone exposures, and by striking ptygmatic folding in the cleavage shown in Pl. 7b. Here the original slip cleavage has become mechanically ineffective.

No appreciable movement occurred after the initiation of growth of the biotite, garnet, chloritoid and staurolite porphyroblasts; this is clear under the microscope from the undisturbed relicts of groundmass in these grains (cf. Pl. 17). The final crystallization of the marbles must have taken place after the last movement phase, for there are no twins and very few partings occur in them. The boudinage must have developed after recrystallization of the marbles and pelites, for boudins of more competent pelite are stretched between beds of more plastic marble; in unmetamorphosed sediments, the limestone would be the more competent.

Tectonic Synthesis

On the basis of the foregoing analysis, it becomes possible to attempt a dynamic interpretation of the phenomena observed in the Lyndonville quadrangle and adjoining areas.

Thinning on the crest of the arch, flowage folding facing away from the crest of the arch, and "wrap-around" schistosity, taken together, are good evidence for doming as the cause of deformation. Also, fold axes and boudinage in *a* are more characteristic of domal structures than of folds due to tangential deformation.

The role of doming in the deformation pattern of eastern Vermont has been discussed by workers in east-central Vermont, where tectonic style resembles that of the area here discussed.

Bean (1953) made a gravity survey of central Vermont and New Hampshire. He found residual negative anomalies under all known gneiss domes covered by his survey, and under two domes in the Waits River arch: the Strafford dome (Doll, 1944; White and Jahns, 1950) and the Pomfret dome (Lyons, 1955). In addition, he found a negative anomaly in the Woodstock area which may also underlie a domal structure. Bean (p. 536) concluded that these negative anomalies could result from a mass of relatively low density rock beneath the higher density St. Francis rocks at the surface. His calculations show that these low density core rocks persist to depths of the same order of magnitude as those of the New Hampshire gneiss domes (Lebanon and Mascoma domes).

Lyons (1955, p. 125) pointed out that Bean's work renders improbable White and Jahns' (1950) hypothesis of upward flowage of calcareous rocks to form the domal structures. Lyons suggested that upward flowage of granitic rock would be consistent with all the known facts. Minor structures within the Hanover quadrangle continue the pattern described farther north by Doll and White and Jahns with certain local exceptions due mainly to the Lebanon dome.

J. B. Thompson (personal communication 1956) observed marked thinning of the formations mantling the Chester gneiss dome. Boudinage here occurs with its long axes down dip; minor folds face away from the crest of the dome. Thompson interprets all the deformation in this area as linked with the doming.

From the foregoing, it would appear reasonable to relate the Willoughby arch to the gneiss domes of southern Vermont. Eskola (1949) indicates what he believes to be the mode of origin of mantled gneiss domes (Fig. 10). He finds that mantled gneiss domes occur only in

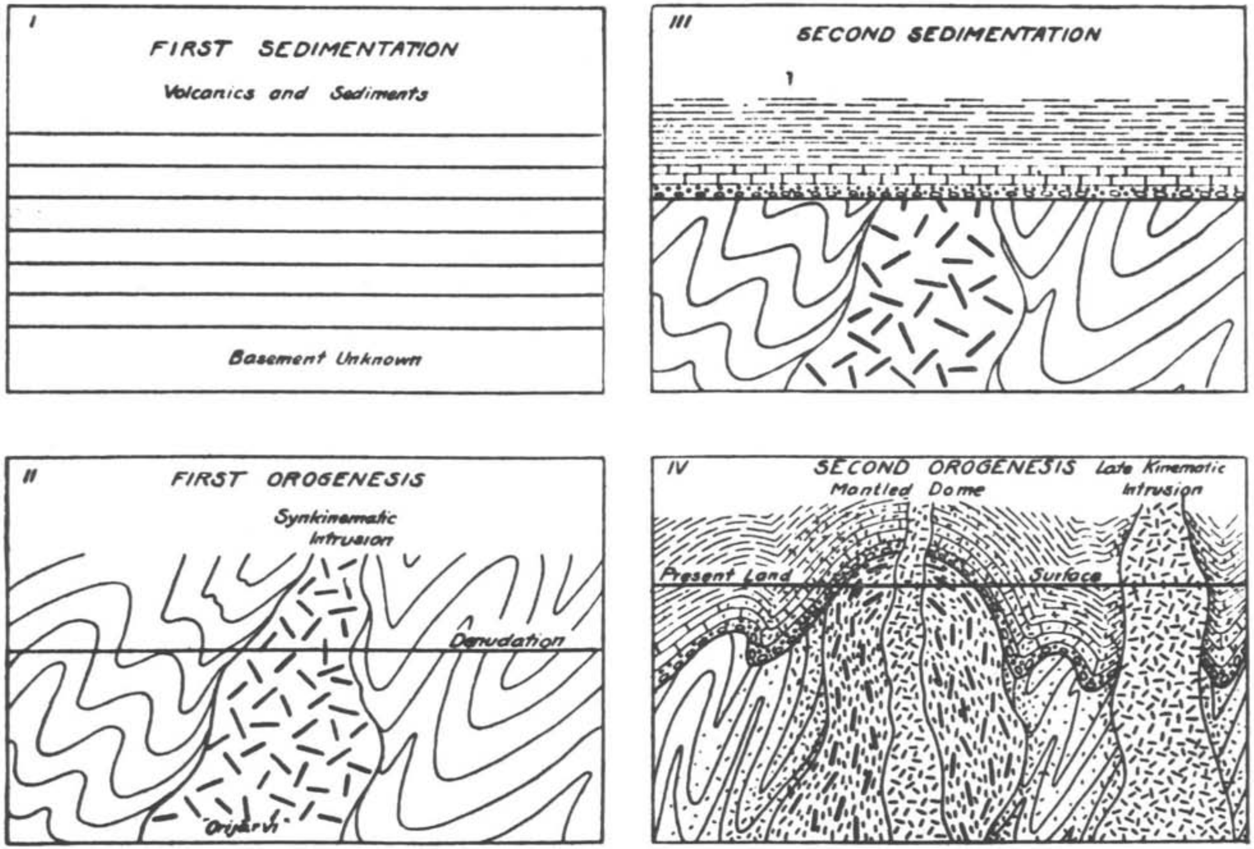


Figure 10. History of a mantled gneiss dome. From Eskola (1949), Fig. 7.

areas where the crystalline basement of an older orogenic belt has been involved in a subsequent orogeny. The gneiss domes represent granitic basement material of the older orogeny that has become remobilized by the younger orogeny. Eskola interprets the mantled gneiss domes of the Appalachians—in Maryland and New Hampshire—as being of such an origin. He had no knowledge of the Vermont domes at the time, but their mode of occurrence (Thompson, 1952, and personal communication 1956) undoubtedly fits these domes into the same class.

An adequate mechanism for doming in Vermont is therefore available. The mobilized basement would be the pre-Cambrian basement which crops out in the Green Mountains. The mantling formations of the Chester dome are too old for the Taconic basement—if it exists—to provide the raw material. Remobilization is evidently related to the Shickshockian orogeny, and is synkinematic to late kinematic.

It now remains to establish the time relationships of the deformation phases. The earlier sericite schistosity is parallel to the bedding over a large area. In view of the interpretation given earlier, it must have been caused by predominance of vertical over tangential stresses. It is more widespread than the later doming. It would seem therefore that early upwarping was spread over a wider area, but was more attentuated than the final expression of the arch. This is borne out by the wide regional distribution of the granites which appear to be genetically related to the mobilized basement.

With increasing tangential stresses due to the Shickshockian orogeny, doming probably became confined to the most mobile zone, where it also became more accentuated and individualized, producing the second generation of cleavage and minor folds. Finally the second generation of cleavage itself became deformed as movement continued. Thus distinct and successive styles of deformation may have been produced by essentially the same mechanism. Thompson (1952, p. 20) appears to have come to very much the same conclusion for doming in southern Vermont. On the other hand, the same phenomena could have been caused by two separate cycles. If, as the writer believes, the St. Francis group is mainly Lower Devonian, both these cycles must have been initiated within a relatively short period.

METAMORPHISM

Introduction

A brief account of metamorphism in the Lyndonville area is given mainly in relation to structure. As work proceeded, it became clear that investigation of metamorphism in the zone of doming of eastern Vermont would have to be treated as a separate project, outside the scope of this report.

An attempt was made at first to define zones of metamorphism by means of index minerals. However, no consistent distribution of common index minerals could be outlined. There are two possible reasons for this: one is that the distribution of granite is quite erratic. The plutons shown on the map are the only two that could be mapped. Elsewhere granite may occur as dikes and sills too small to map with confidence in view of prevailing exposure conditions. There is also the real possibility that many subsurface granite bodies and apophyses affect the mineral assemblage in sediments seen in outcrop. The third reason is the highly variable composition of the pelites, which it is often hard to appreciate in the field, but which affects the formation of many index minerals.

A description of significant observations will be followed by a discussion of principles and tentative conclusions.

Description of Rocks

Metamorphism in three groups of rocks will here be considered: pelites (fine-grained graywackes), siltstones (fine-grained biotite quartzites), and pillow lavas, now amphibolites.

The pelites have retained the best record of the metamorphic history in the area; they are, therefore, examined here in more detail than the other groups.

1. Pelites. The term pelite is here used for rocks having a grain size of less than 0.005 mm, regardless of composition. In hand specimen and in outcrop the pelites are gray, tough when fresh, with a moderate to poor foliation. Sericite generally imparts a silky sheen to fresh surfaces. The only porphyroblastic mineral commonly visible is biotite, which characteristically appears in clusters. Garnet and staurolite are conspicuous in some localities, but more often they can be identified only in thin section or, at best, after careful study with a hand lens.

The extreme toughness of the rocks is characteristic, and not unlike that of typical hornfelses. There are some very fine-grained pelites, which contain little or no biotite and in thin section show a groundmass in which individual minerals cannot be identified.

The typical pelite in the area has a groundmass of quartz and sericite, with minor feldspar in varying proportions. Biotite may also appear in the groundmass. Where sericite is present, it is commonly oriented so as to impart a well-defined schistosity to the rock.

Common accessory minerals include graphite, pyrrhotite, and pyrite, each of which may locally become conspicuous. Porphyroblasts of minerals characteristic of regional as well as contact metamorphism have grown in the groundmass, and often contain relicts of it. These minerals are biotite, garnet, staurolite, chlorite, chloritoid, andalusite, kyanite and sillimanite. Relict slip cleavage preserved in the biotites is evidence that the crystallization of sericite was separated from the formation of porphyroblasts by a structural episode. As has been seen in the section on structural geology, this episode consisted in the main phase of doming. There is admittedly no evidence as to the length of time involved. Nevertheless, the structural evolution was such as to make a change in environment likely; the writer therefore feels justified in taking the textural evidence at its face value, and in postulating at least two phases of metamorphism: pre-cleavage and post-cleavage. The former is characterized by the groundmass assemblage, the latter by the various assemblages of prophyroblasts. In the section on evaluation these assemblages will be considered in the context of the facies classification.

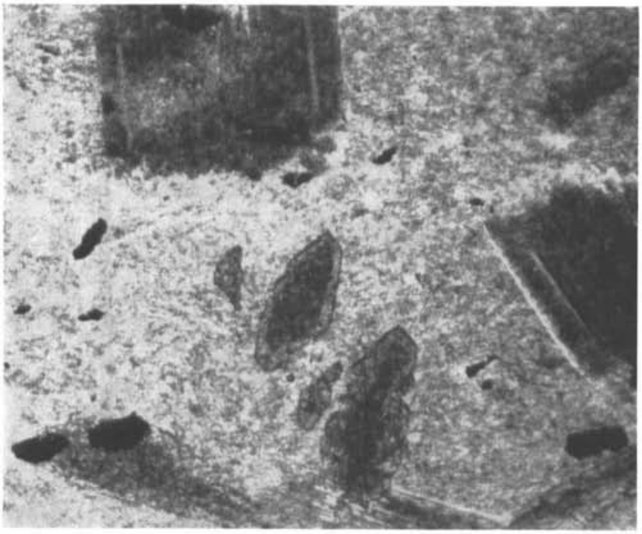
Biotite is the most common porphyroblastic mineral and is found throughout the quadrangle. It usually contains trends of carbon which outline the structure of the groundmass. These trends show that no biotite has been displaced with respect to the groundmass.

The biotite appears to grow preferentially around rods of graphitic material and clusters of ilmenite or other opaques. It has a moderate dimensional orientation, generally guided by any sericite foliation present.

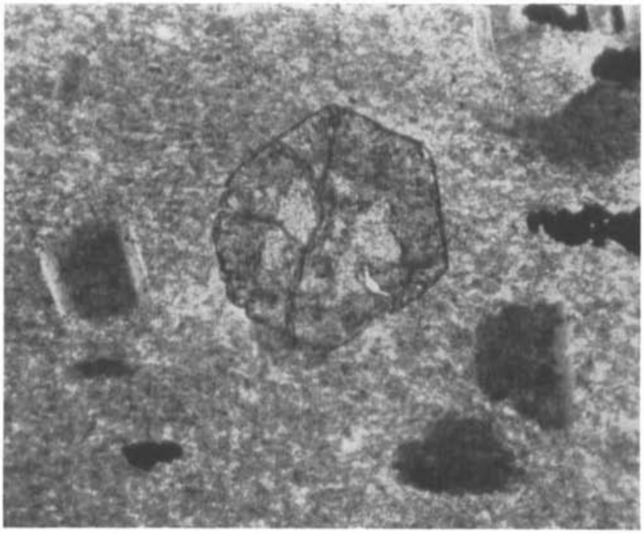
Garnets vary greatly in size, from 0.5 mm and less to several millimeters. They are pink in thin section, and have a refractive index of 1.805 ± 0.004, suggesting a somewhat magnesian almandine. In many thin sections garnets appear to have grown in two stages: they show "zoning," i.e. a core of slightly different texture from the outer rim, with generally a sharp contact between the two. It may perhaps be significant that these "cored" garnets are never chloritized, nor found in the same rock as chlorite.

The cores suggest that there were two distinct phases of garnet growth, the second of which may have been contemporaneous with the formation of chlorite in lower grade rocks.

No evidence of displacement of garnets relative to the groundmass

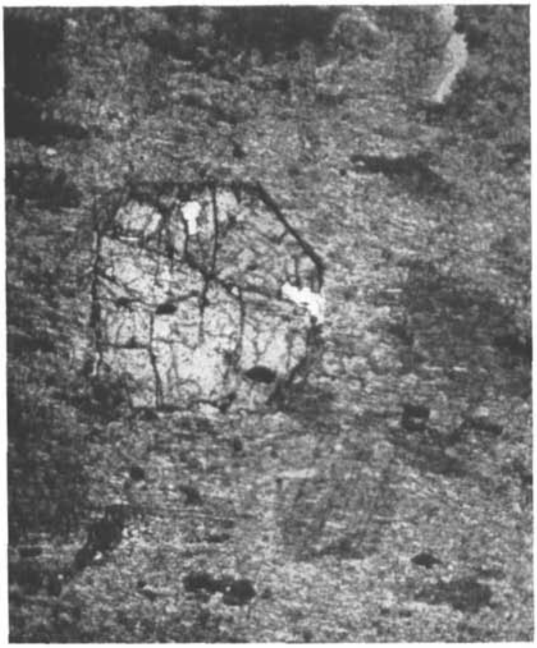


a. Specimen M 21. Large, stubby porphyroblasts of biotite in exceedingly fine-grained groundmass; groundmass relicts (probably carbon) carry its texture into the prophyroblasts. White rims on biotite are quartz. Euhedral staurolite in the lower center of the picture appears to replace biotite. Waits River formation.

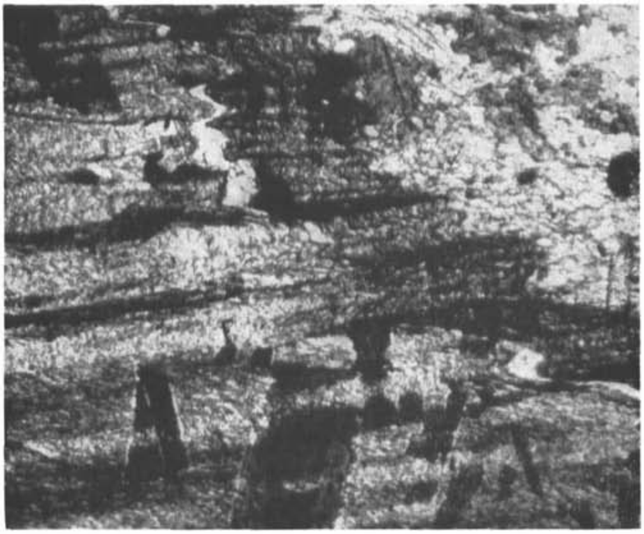


b. Specimen M 20. Porphyroblasts of almandine and biotite, showing two different zones in individual crystals; these zones are brought out by contrasting degrees of groundmass relict preservation. The clear rims on biotite are quartz. Gile Mountain formation.

PLATE 17. Photomicrographs. Magnification 30X; polarizer in position.



c. Specimen M 53. Euhedral garnet (center), biotite containing groundmass trends (lower right) and chloritoid (at top). Waits River formation.



d. Specimen L 96. Pelite-marble contact. Relict cleavage clearly outlined in the biotite porphyroblasts. Gile Mountain formation.

PLATE 17. Photomicrographs. Magnification 30X; polarized in position.

was found. A few garnets contain relict trends of groundmass, and these are clearly undisturbed.

The distribution of *staurolite* is highly erratic; this is not surprising, in view of its sensitivity toward the iron content of the rock (Williamson, 1953). On the whole staurolite is confined to the same zone as garnet. It may or may not occur together with the latter, depending on the bulk chemistry of the host rock.

Staurolite is often seen as poikiloblastic crystals, or in granular aggregates. Both aggregates and poikiloblasts tend to assume euhedral outlines. In some instances staurolite seems to replace biotite.

Chloritoid often takes the place of staurolite in the lower grades, especially in the northwestern part of the quadrangle. This mineral sometimes clearly replaces biotite, indirectly suggesting that staurolite in the higher grades also grew at the expense of biotite.

The chloritoid is almost invariably euhedral, in slender flakes of a dirty green color, and normally has random orientation. Carbon trends clearly outline the groundmass within the mineral, once more showing no post-crystallization disturbance. In a few cases chloritoid has grown in ill-defined anhedral patches.

Chlorite occurs in two ways: replacing biotite or garnet, or as independent euhedral porphyroblasts, rather like the chloritoid in habit, though generally less slender in outline. It is randomly distributed over the whole quadrangle, but comparatively rare.

Kyanite was observed only in 6 thin sections, all from localities within 1500 feet of a granite pluton contact. It forms subhedral laths, often slightly poikilitic.

Andalusite is rare. It was observed in only 5 thin sections, as colorless subhedral laths. Except for an isolated occurrence near Wheelock village, it is confined to the area between the two granite plutons.

Sillimanite was observed in two thin sections as very small individual needles—not in aggregates: the localities (north of Parker Pond, and near the source of the Passumpsic river) are both in the area between the two granite plutons.

Sphene generally occurs as an alteration product of ilmenite.

Plagioclase porphyroblasts are rare, but were observed in at least two instances: on the west spur of Big Rock Hill and in Annis Brook: both localities are also in the area between the two plutons.

2. *Siltstones*. These are commonly dense, gray rocks with moderate fissility parallel to a secondary foliation. The bedding, where not parallel to the foliation, can sometimes be recognized by dark, graphitic banding

not unlike varving. The foliation is frequently accentuated by parallel quartz stringers. Biotite flakes and clusters can sometimes be seen aligned parallel to the foliation.

In thin section, the siltstones consist of almost equidimensional grains of quartz about 0.01 mm in diameter, with biotite flakes outlining the foliation. Small, randomly oriented sericite flakes may occur between the quartz grains. Apatite is an accessory which becomes enriched locally, especially near granite contacts. Near pelite contacts porphyroblasts characteristic of the pelites occasionally appear within the siltstones close to the contact. The great uniformity in grain size of the siltstones throughout the area is remarkable. At H 1303, microcline porphyroblasts have formed in the siltstone close to aplitic veins. The veins themselves contain phenocrysts of oligoclase.

3. Siliceous marbles. The grain fabric analysis shows that the marbles must have recrystallized after most of the deformation had ceased. No twinning was observed. Both calcite and quartz crystals are clear and undeformed, almost without exception. There is no difference, either megascopically or microscopically, between limestone metamorphism in different parts of the quadrangle, or in relation to granite contacts, except right close to the contacts where the grain size is coarser, approaching 2 mm in diameter.

Impurities within the marbles generally recrystallize to sericite, tremolite, or phlogopite; but such impurities are not too common. The occasional tremolite may be derived from primary dolomite and calcite.

4. Mafic volcanics.[1] The Standing Pond lava has become metamorphosed to amphibolite. Hornblende often occurs in long needles, dark green in hand specimen, bluish green in thin section, making up over 60% of most specimens. The plagioclase is now a finely granulated oligoclase-andesine; in view of the fact that the Standing Pond is a pillow lava, the original plagioclase may have been more calcic.

Accessory minerals include quartz, calcite, apatite, chlorite, and occasional biotite. The classification of this rock will be considered in the next section.

Evaluation of Metamorphism

1. Siltstones. These rocks adjusted to the second (post-cleavage) phase of metamorphism by assuming a foliation parallel to the slip cleavage. The only newly-formed mineral appears to be biotite, and even

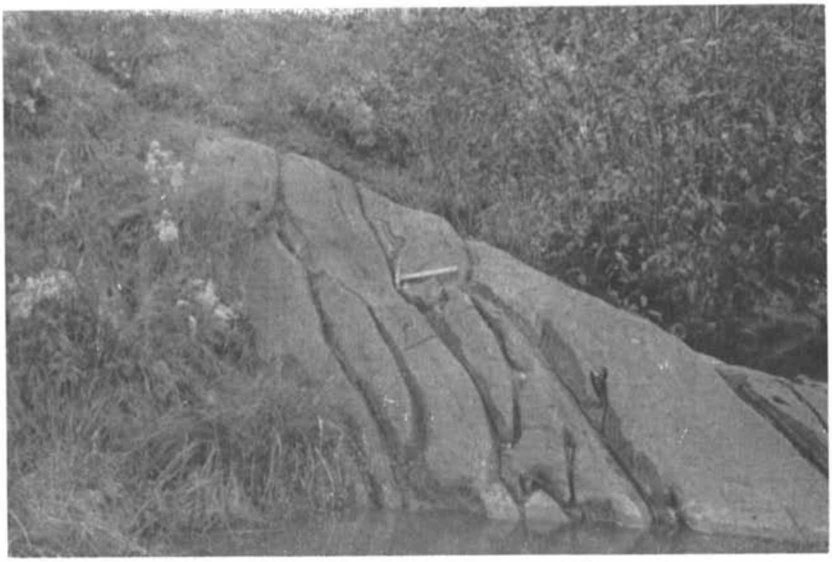
[1] A band of Standing Pond volcanics was discovered immediately south of the quadrangle, parallel to the Passumpsic River (Pl. 3).

PLATE 18. Standing Pond lava. Below St. Johnsbury dam. Green Laths of hornblendes, in groundmass consisting mainly of fine-grained calcic oligoclase.

a. Pillow lava in Passumpsic River near St. Johnsbury dam. Hammer handle points east.

PLATE 19

b. Single pillow, at same locality.



c. Pillow lava, Robert Brook, east of St. Johnsbury Center.

PLATE 19 (Continued)

PLATE 20. Pelite overlying marble, in a minor asymmetric anticlinal flexure. Calcite-quartz veins mark the trend of the foliation in the marble, which carries through into the pelite. In Barton River, at Glover Village.

this may have recrystallized from an earlier generation. There has been no important crystalloblastesis of the quartz.

2. Marbles. Many marbles adjusted in the same way as the siltstones, by assuming a schistosity parallel to the later cleavage. However, many others, especially micaceous marbles, have retained the older schistosity, suggesting that the micas are products of the earlier metamorphism. Grain orientation analysis shows that the calcite grains have assumed a lattice orientation conforming to the later cleavage. Occasional tremolite is undeformed, and is evidently later than the slip cleavage.

3. Mafic Volcanics. Oligoclase amphibolites, such as are found in the Lyndonville area, have been shown to be possible products of the metamorphism of basaltic rocks (Sutton and Watson, 1951). These authors recognized that regional metamorphism of basaltic rocks may assume different trends: the end-product of the metamorphic history of a given basaltic rock depends on that history, as much as on the final temperature-pressure environment.

It is known from the groundmass assemblage of the pelites in the

Lyndonville area that the rocks have passed through a stage of low-grade (sericite-quartz) metamorphism. Basaltic rocks are very sensitive to metamorphic change, and so it may be assumed that the lava has passed through the greenschist facies at the time of this early metamorphism. No evidence of this is preserved in the rock. However, there is an indirect line of evidence: it would seem that the second (post-cleavage) phase of the metamorphism was characterized by water deficiency, as deduced from the metamorphism of the pelites; yet the assemblage hornblende-oligoclase-quartz is characteristic of "wet" metamorphism (Nockolds, 1937, p. 114–115). A basaltic rock reduced to the greenschist facies, or even to the epidote-amphibolite facies contains more water bound in hydrous minerals than an oligoclase amphibolite. Hence, transformation of a greenschist facies rock into an oligoclase-quartz amphibolite would result in the release of excess water (Yoder, 1955, p. 509). If this reasoning is correct, one would expect the lava to release excess water in the course of the second phase of metamorphism. This water should locally modify the "dry" metamorphism of the adjoining pelites; for, as will be shown, the surrounding pelitic rocks provide evidence that in their case the later metamorphism has been water-deficient. The writer believes that Doll's (1944) "coarse garnet-mica schist" which adjoins the Standing Pond volcanics in the Strafford quadrangle is evidence for such a local modification: the texture is granoblastic, and garnet porphyroblasts attain sizes of the order of 6 cm. However, the mineral assemblage does not differ from that of the normal metamorphosed pelite of garnet grade in this region. The locally increased water concentration seemingly affects only the grain size, not the mineral facies. The assemblage bluish hornblende-oligoclase-quartz-(calcite-epidote-biotite-chlorite) fits into the chloritoid-almandine subfacies of the epidote-amphibolite facies (Turner, 1948c, p. 89). The Standing Pond lava south of Lyndonville would fall within field 4 of Fig. 12, close to the hornblende corner. This would correspond to the almandine grade in pelitic rocks (fields 2 and 3 in Fig. 13). However, south of Lyndonville no almandine was found in pelites in the neighborhood of the lava, though Eric (1942) places this general area within his garnet zone.

4. Pelites. The normal pelites of the area contain appreciable amounts of free quartz, and no calcium minerals. Their composition, therefore, comes under the classification "excess SiO_2 and Al_2O_3" of Turner (1948c). For such compositions, Turner (*loc. cit.*, p. 82) has constructed

AKF diagrams (Al_2O_3, K_2O, FeO/MgO) analogous to Eskola's (1920; 1946) ACF diagrams. Three of these diagrams are reproduced in Figs. 11, 13 and 14.

All post-cleavage assemblages in pelites of the area coexist with sericite (muscovite) of the groundmass and with negligible exceptions all include biotite. This limits their position on the AKF diagrams. In much of the eastern part of the quadrangel,[1] no other porphyroblasts occur; these rocks, therefore, fall in field 1 of the greenschist facies (Fig. 11). West of the eastern limit of the band of Gile Mountain schist, garnet becomes abundant. These rocks fall within field 3 of the AKF diagram representing the epidote-amphibolite facies for the rocks here considered (Fig. 13). Minerals of this facies are sometimes partly altered to chlorite. In the same general area, staurolite occurs either together with garnet and biotite, or with biotite only. Such rocks fall within fields 2 and 3 of Fig. 14 representing the AKF diagram for pelites in the amphibolite facies. According to Fig. 14, biotite and staurolite cannot coexist as major constituents: they are separated by the almandine-muscovite line. It is possible that the staurolite is later than the biotite-garnet assemblage; it may be more directly associated with emplacement of the granites. In fact, a narrow zone of staurolite-bearing rocks can be traced along the northern contact of the Black Hills granite, suggesting a close relationship. However, J. B. Thompson (private communication 1956) has suggested that FeO and MgO be considered as separate components, instead of combining both in the "F" corner of the triangular diagrams. In that case four-component tetrahedra can be constructed which allow for the coexistence of biotite, staurolite and garnet.

Another post-cleavage assemblage is that composed of biotite, garnet and chloritoid. This assemblage falls in fields 2 and 3 of Fig. 13. This presents the same problem as the previous case. However, chloritoid replaces biotite in several thin sections. Also, chloritoid has more random orientation than biotite. There is thus evidence that chloritoid was formed after the biotite. Geographically, chloritoid is almost completely confined to the area between the two granite plutons. The probability that chloritoid is the low-grade equivalent of staurolite lends weight to the assumption that staurolite has formed independently of the biotite-garnet assemblage, and (in rocks with suitable composition) is more directly associated with emplacement of the granites. Alumino-

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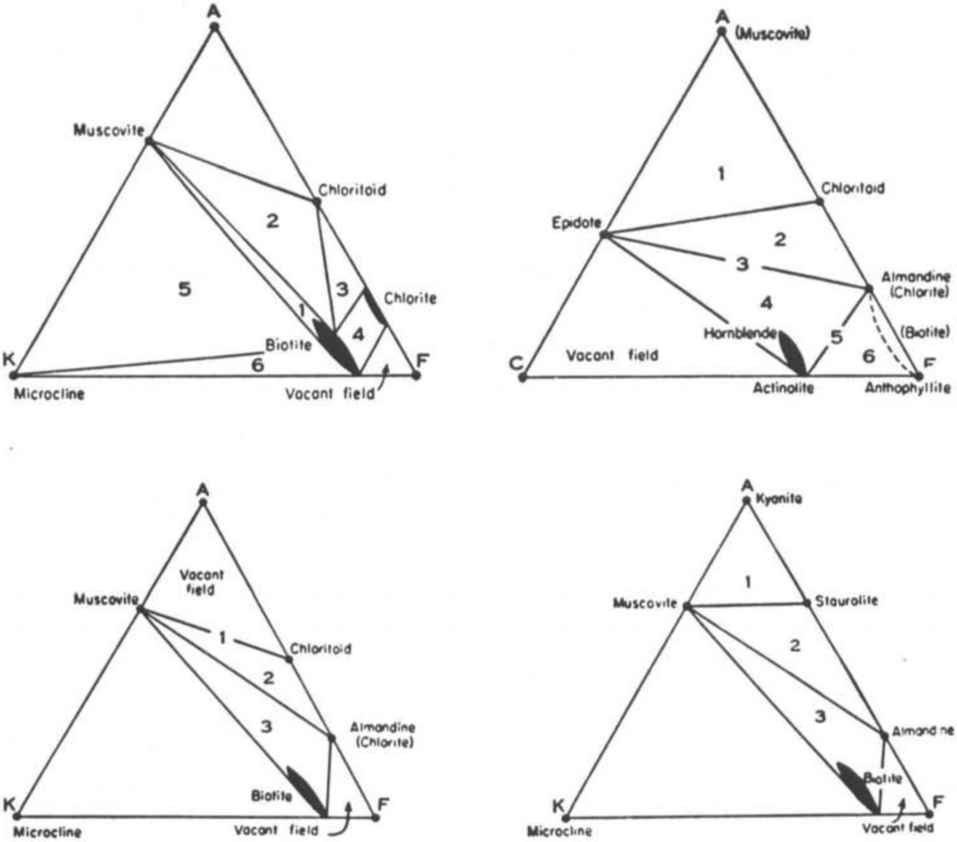
[1] As mentioned before, distribution of index minerals is too erratic to justify the drawing of isograd lines.



Figures 11–14. Facies Diagrams. From Turner (1948c).

Figure 11. Greenschist facies. Biotite-chlorite subfacies AKF diagram for rocks with excess Al_2O_3 and SiO_2.

Figure 12. Albite-epidote amphibolite facies. Chloritoid-almandine subfacies. ACF diagram for rocks with excess SiO_2 and deficient K_2O.

Figure 13. Albite-epidote amphibolite facies. Chloritoid-almandine subfacies. ACF diagram for rocks with excess SiO_2 and Al_2O_3.

Figure 14. Amphibolite facies. Staurolite-kyanite subfacies. AKF diagram for rocks with excess SiO_2 and Al_2O_3.

silicates are not common in the area, but are generally found in the vicinity of granite contacts or in the area between the plutons. It is normal for sillimanite and andalusite to be associated with granite contacts. The similar pattern of occurrence of kyanite is unusual, but well established in this area. Fig. 15 is reproduced from Yoder and Eugster (1955).

The equilibrium curve is for the ideal case when total pressure equals water pressure. In that case the curve passes outside the stability field of kyanite. However, Yoder (1955, p. 514) has shown that with reduced water pressure (maintained in an open system), an equilibrium curve of this kind will assume a negative slope, and will tend to be the farther left, the lower the water pressure. It follows that kyanite may be a stable product of the reaction illustrated in Fig. 15, when water pressure is low. Thompson (1955, p. 97–98) has shown how this can explain the formation of kyanite under the "open" conditions of regional metamorphism. In the Lyndonville area, the pelites became water-deficient as a result of an earlier phase of metamorphism. When the granites were emplaced, both water pressure and water concentration must have been low, and kyanite was able to form as a contact aureole mineral.

The relative abundance of the alumino-silicates in the area between the two larger plutons suggests that contact aureole conditions must have been fairly continuous between the two bodies.

Discussion

This area presents an example of polymetamorphism in which a high-grade mineral assemblage (porphyroblasts) supersedes a low-grade assemblage (the pre-cleavage assemblage). Such a sequence is not common in regional metamorphism. Read (1949, p. 105) suggests that high-grade rocks do not pass through the low-grade stage. In other words, the environment changes faster than the rocks can adjust. Evidence from the Lyndonville area indicates that here the rocks had time to recrystallize at a low grade of metamorphism, before porphyroblast growth. In this low-grade assemblage water available for any subsequent metamorphism was drastically reduced relative to the water originally contained by the sediments (Yoder, 1955, p. 508), and it seems that little if any water was introduced by metasomatism.[1]

Evidently little or no water was available to circulate along grain boundaries and to aid even such ready readjustments as grain growth. Internal movement (*Durchbewegung*) has often been cited as a necessary condition of regional metamorphism. It seems that this is only a prereq-

[1] Contrast this with conditions obtaining nearby, in the southern part of the Island Pond quadrangle. Here feldspathization has resulted in the formation of augen gneiss in the Gile Mountain formation.

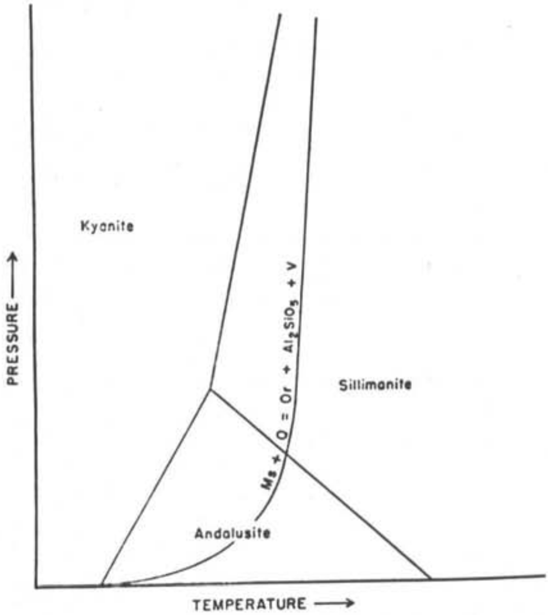


Figure 15. Probable relationship of the $A1_2Si0_5$ polymorphs to the muscovite (Ms) ÷ quartz (Q) = orthoclase (Or) ÷ $A1_2Si0_5$ ÷ water (V) stability curve. From Yoder and Eugster (1955), Fig. 19.

uisite to the extent that it facilitates the circulation of solutions (Bearth, 1952, p. 346). When there are no solutions to circulate, metamorphism is handicapped; this results in textures closer to contact metamorphism than to accepted textures of regional metamorphism (compare plates 17 and 18 with typical "hornfels" or "spotted schist" textures, shown in textbooks, e.g., Williams et al., 1954, p. 182–183). The grade indication given by the garnets cannot be far wrong, in view of the metamorphism of the Standing Pond lava.

The above considerations indicate that rocks which have gone through lower grades should retain a record of their early history: lack of water during their later history impedes recrystallization. In other words, the metamorphic rocks of the Lyndonville area are an exception which lends weight to the rule formulated by Read (*loc. cit.*), that high-grade rocks do not normally pass through the low-grade stage.

Summary of Metamorphic History

It seems, then, that the rocks of eastern Vermont have had a somewhat unusual metamorphic history. This history is best elucidated when considered in conjunction with the structural evolution of the region.

Textural and structural evidence show that the early phase of metamorphism was associated with the early broad upwarping, which caused the earlier schistosity. With continuing orogenesis in the east the mobilization of the basement became more pronounced along certain favored zones. In New Hampshire, this activity probably gave rise to the Oliverian and New Hampshire plutons. In Vermont a zone of domes became individualized parallel to the present Connecticut and Passumpsic rivers; from previous considerations, it seems that the post-cleavage biotite-garnet porphyroblastic assemblage was related to this phase. The basement finally became sufficiently activated to cause migmatization and intrusion in New Hampshire (Billings, 1937, Chapman, 1939) and cross-cutting granites in Vermont. From what is at present known it seems likely that the appearance of andalusite, kyanite, sillimanite, staurolite and chloritoid, was associated with this episode. As far as chlorite is concerned, all that can be said is that it formed either at the same time, or later than chloritoid: textural relations of both minerals are often similar. In Unst, Read (1934), has described chlorite and chloritoid as occurring in the same episode. If they are contemporaneous in the Lyndonville quadrangle, the emplacement of the plutons not only caused contact metamorphism, but also had more regional effects.

The above may be stages of essentially a single diastrophic event. Although the rocks reacted to changes in environment in distinct steps, both as regards structure and mineral assemblages, this does not necessarily mean that the dynamic causes acted as separate events. The rocks probably only had to overcome what may be called a certain "inertia of response" to proceed from one step to the next. Smoluchowski (1952) has studied the nature of this inertia of response. In recrystallization in steel, he observed a pronounced time-dependence of the rate of nucleation. He noted, in particular, a long "incubation period" during which no reaction seems to occur.

There can be no doubt that the chronology of metamorphism must be approached with extreme care. Where no age determinations are available, only unequivocal structural criteria and the superposition of assemblages of proven incompatibility can establish a safely founded time sequence.

THE GRANITES

Introduction

The rocks referred to as "granites" in this report are, strictly speaking, *granodiorites* and *quartz monzonites* (Johannsen). An estimated mode (specimen L 78) is: plagioclase: 50%, microcline: 20%, quartz: 20%, biotite: 6%, others: 4%. Accessory minerals include muscovite, apatite, sphene, zircon, zoisite, almandine and allanite.

The composition of most of the granodiorites is close to that of the estimated mode given above. However, great variations were observed at cross-cutting contacts and in vein rocks.

Mineralogy

Plagioclase is oligoclase. It occurs either untwinned, or twinned according to the albite, pericline, Baveno and Carlsbad laws. Both progressive and oscillatory zoning are common. The feldspars are almost always clear; turbidity is rare. Sericitization along cleavages is often present, but limited in extent. At contacts with microcline, plagioclase frequently develops myrmekite.

Microcline is typically cross-twinned and sometimes perthitic (irregular patch perthite).

Biotite is usually brown, with pleochroic haloes but in at least one granite (D 3030) the biotites are green. Plagioclase and biotite are euhedral or subhedral; microcline and quartz are generally anhedral.

Garnet is common near some contacts. It has the same refractive index as that in the sediments, 1.805 ± .004. In pegmatites close to cross-cutting contacts, garnets up to 1 cm in size have been seen. As a rule, garnet occurs in comparatively biotite-poor parts of the granite, suggesting migration of potassium. This phenomenon was observed only at Lake Willoughby where contacts are exceptionally well exposed.

The Contacts

Only two good contacts of the larger massifs are exposed within the quadrangle: On route 5A, along Lake Willoughby; and above the village of Barton. The Barton contact is practically conformable and sharp. The sediments immediately in contact have been recrystallized to the extent of having an almost granitic aspect in thin section, but there is no evidence of metasomatism. The contacts at Lake Willoughby show a great deal of assimilation and granite veining in the sediments.[1]

Impure calcareous rocks of the large mass of Gile Mountain formation exposed in the road cut have become calc-silicate rocks; these generally have a quartzofeldspathic groundmass and large poikilitic porphyroblasts of hornblende and grossularite, with, occasionally, diopside and

[1] A brief guide to the most interesting outcrops along the lake road is given in Appendix I.

a. Breccia of biotite-rich granitic rock in light, aplitic granite.

b. Breccia of transformed country rock (now an amphibolite) in aplitic matrix. 1110 feet north of reference point (see Appendix I, exposure No. 11).

PLATE 21. Granite contact phenomena. Route 5A, Lake Willoughby.

a. Specimen L 71, Black Hills granite. Twinned and zoned crystals of oligoclase; also quartz, biotite, and a little microcline.

b. Specimen M 59. From disused quarry at D 3030. Microcline, oligoclase, green biotite. *Myrmekite at the contact between the zoned oligoclase near the center and* the large microcline. Untwinned feldspars are oligoclase.

PLATE 22. Photomicrographs of granite. Magnification 30X, nicols crossed. (Photographs by Rae L. Harris, Jr.)

sphene. This contrasts with the reaction of the siliceous marbles of the Waits River formation. Where such marbles are in contact with the granite, as for instance in the Mount Pisgah cliff, the only change is an increase in grain size of calcite and quartz. With no additional impurities these two minerals are stable in each other's presence up to fairly high temperatures. This illustrates the general observation that pure limestones, siliceous marbles and sandstones are relatively resistant to metasomatism. A few calc-silicate rocks occur well away from any known granite contacts, for example, three-inch diopsides south of Perry Station, a circumstance which cannot be explained on present evidence.

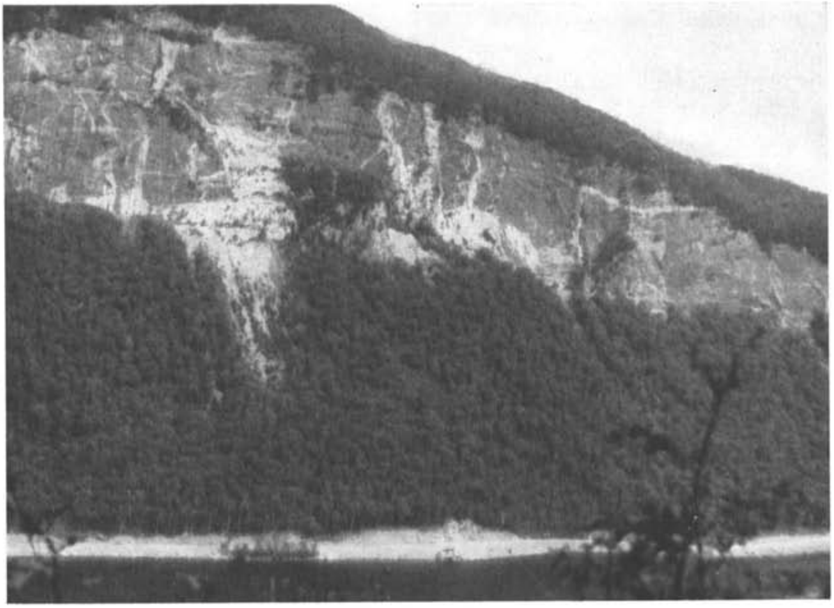
Structural Relations of the Granites

Many of the vein rocks and larger massifs are inferred to be conformable or sub-conformable from their topographic expression. Contacts observed in the field (mainly those of vein rocks) are most commonly conformable either to bedding or cleavage of country rocks. Where contacts are cross-cutting, the surrounding sediments are not often disturbed. This is well shown in the face of Mt. Pisgah (Pl. 19). However, in some cases the structure of the sediments appears to curve around a granite mass. This is particularly evident near A 0016. Xenoliths of sedimentary rock are frequent and are often conformable in alignment with the country rock. "Ghosts" of xenoliths, represented by biotite-rich portions in the granite, are also sometimes dimensionally aligned parallel to the attitude of the beds in near by sediments.

The granites seem to have unmistakably "intrusive" relations toward the country rock. They carry a fairly large volume of assimilated sediments, judging by the number of xenoliths and "ghosts."

According to Wegmann (1929), within a homogeneous stress field, intrusion should take place along planes which are normal to the least "closing component" (*Schliesskomponente*). This condition quite evidently does not obtain in eastern Vermont. These granites are late to post-kinematic, for they are undeformed (apart from some local readjustments; cf. Pl. 12). Hence the stress field can reasonably be assumed to be homogeneous. Weak west-directed stress must have continued throughout the granite emplacement (Balk, 1926). The *ac* planes of the regional folding would then be the logical intrusion controls (Bucher, 1920). Such controls were not observed in the Lyndonville area.

Any further discussion of this dilemma at the present stage of the work in eastern Vermont would be merely speculation, and a well-based



a. View of the cliff face of Mt. Pisgah. The bedding in the sediments is almost horizontal.

b. Detail of a cross-cutting contact.

PLATE 23. Contacts of the Mt. Pisgah granite and calcareous sediments of the Waits River formation.

opinion must await further detailed mapping and laboratory study of the granites.

SUGGESTIONS FOR FURTHER WORK

For the age question, short of new fossil evidence, structural and stratigraphic studies in two critical areas are recommended:

1. In the Lake Memphremagog area: detailed investigation of the relationship between the Shaw Mountain formation and the Glenbrooke and St. Francis rocks north of the Canadian border.

2. In the Connecticut valley area: detailed study of the relations of the Orfordville formation.

New evidence that would throw light on the structural synthesis here presented should be sought in adjoining quadrangles. The geometric pattern of cleavage and schistosity in the Burke and Hardwick quadrangles would be of particular interest. The petrofabric evidence presented in this report is not more than indicative. More petrofabric studies of pelite/marble contacts (and, possibly, quartzite/pelite contacts) are needed.

Further afield, future structural studies must include the abrupt end of the Willoughby arch in the Island Pond quadrangle, and the tectonic significance of the Brighton line.

The petrologic work carried out in connection with the present project was hardly more than exploratory. More detailed information is needed about the contact relations of the granites. The writer's conclusions concerning metamorphic history should be tested by a closer thin section coverage across and beyond the zone of doming.

The metamorphism of the pillow lava should be checked against that of the surrounding pelites.

APPENDIX I.

A few interesting and easily accessible exposures.

1. G 2752, in Roaring Brook (cf. Pl. 11). Minor structural features in Barton River lithologies: boudinage, cleavage, schistosity, minor folds. Also glacial striations and markings of possibly organic nature in the marble.

2. G 3859, above Barton village: Well exposed sharp granite contact, with partly recrystallized siliceous schist. Very small lens of Standing Pond lava, a little below the contact.

3. G 3231, in the bed of the Barton River, in Glover village: Fold in Standing Pond lava, showing bedding-cleavage and contact relations (cf. Pl. 20).

4. A 0458, on Route 12: Minor structural features in Barton River lithologies.

5. A 2035, west of trail: good exposure of staurolite schist.

6. Wheeler Mountain: best exposed conveniently accessible granite in the area.

7. Stream above B 1333: A good sequence of exposures in marble-rich Gile Mountain lithology.

8. Miller Run in Sheffield and Wheelock villages: Excellent exposures of typical Barton River lithologies.

9. Lyndon village, under covered bridge, at *west* end of village: Minor fold axes with many diverging attitudes.

10. Lyndon dam (just off the SW corner of the Lyndonville quadrangle): Excellent exposures below the dam, showing steep axes of folding, and bedding-cleavage relations; cf. Pl. 16.

11. Road cut of Route 5A on Lake Willoughby. Figures indicating individual localities are distances in feet from the southernmost of two streams shown entering the lake on its east shore. At 0 ft: The stream runs over a large biotite-rich portion of granite which is broken up by more leucocratic and pegmatitic veins. The wall rocks of some of the latter show evidence of relative mutual displacement. In some of the lighter parts of the granite small red "pinhead" garnets can be seen.

160–500 ft.: Large biotite-rich portions of granite in a matrix of more aplitic granite; this latter generally contains pinhead garnets.

525–550 ft.: Large irregular pegmatite segregation; minerals include microcline (specimens ranging up to 6 inches in length); muscovite in unusual radiating habit; large pyrite crystals; and quartz.

575–1000 ft.: Large exposure of Gile Mountain schist. Contains many examples of calc-silicate rocks, and intensely recrystallized siliceous pelite, as well as a number of granitic sills.

1110 ft.: Breccia of calc-silicate rock fragments in aplitic cement. Shows various stages of assimilation by the granite.

1170 ft.: Brecciated xenoliths in pegmatitic matrix.

1210 ft.: Xenolith of Gile Mountain schist.

1460 ft.: Pegmatite with large garnets (up to 1 cm in diameter).

1795 and 1635 ft.: biotite-rich schlieren, possibly xenoliths in the process of assimilation.

1990 ft.: (cf. Pl. 21a): Breccia of biotite-rich granite fragments in aplitic matrix. The aplitic matrix is studded with pinhead garnets, biotite-poor fragments carry a moderate amount and biotite-rich fragments do not contain garnet. These fragments are probably xenoliths that have been partially assimilated by the granite. No granite of this biotite-rich composition occurs in the adjoining region.

2005 ft.: Biotite-rich "ghosts."

2010 ft.: Northernmost of the two streams. Biotite-rich "ghosts."

3025–3270: Northernmost exposure along the lake road. Many pegmatites; pinhead garnets throughout, except in "ghosts" rich in biotite.

12. E 0508: Rocky opening in north slope of hill, at about contour 1800. Not too easily accessible. Good example of relation between late schistosity (and slip cleavage) and bedding. Staurolite growth controlled by cleavage. Pseudo-bedding aspect of schistosity can be compared with true bedding shown by lithologic contrast. See Pl. 7 and 13.

APPENDIX II.

Explanations for plates 1–3.

Pl. 1. Red hachuring represents areas in which granitic dykes are common. The contact between the hachured area and the plain red for "granite" is arbitrary, for in the field this is a transition rather than a sharp contact.

The cross-sections give the impression of a comparatively quiet, large-scale structure. As shown in the text, there is much deformation in detail. Such deformation cannot be represented in cross-section without distortion of scale. Diagrammatically, the relations of the minor structural features are shown in Fig. 4.

Pl. 2. Air photo linears are shown in red. These are interpreted as lines of structural significance, and transferred from air photographs to a 4x enlargement of the one-inch quadrangle map. It is not always possible to discriminate between lines of fracture and lines due to lithology. However, in many cases trends are obviously parallel to bedding, while in others they follow joint directions measured in the field. It has, therefore, been possible to complement measurements of the strike of both bedding and jointing by referring to air photographs.

Pl. 3. Owing to the scale of the map, it was impracticable to represent certain thin formations. The most important omission is that of the Shaw Mountain formation. This follows (with many gaps) the contact between the Waits River formation shading and that used for the pre-Shaw Mountain formations. An inlier (anticline) of Stowe formation south of the 45th parallel is outlined, but shown in the same shading as the pre-Shaw Mountain formations. The Orfordville formation has a separate symbol, though possibly "pre-Clough," because its position is not quite certain.

All the cross-cutting granites have received the same symbol, because in Vermont and Quebec it is sometimes not possible, with available information, to distinguish between the New Hampshire and White Mountain plutonic series. Under the circumstances it was considered advisable not to differentiate.

The Highlandcroft plutons are included under the general "pre-Clough" symbol. Code numbers to sources are given below with the corresponding reference in the bibliography.

In many cases information taken from the sources cited was modified by the writer.

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2. Morin, 1954.
3. Billings, 1955.
4. A. Albee, unpubl. compilation.
5. Ambrose, 1943.
6. Cooke, 1948b.
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9. Dennis, reconnaissance.
10. ——, this report.
11. Woodland, oral information, 1955.
12. Hall, oral information, 1955.
13. Eric, 1942.
14. Billings, 1937.
15. White and Jahns, 1950.
16. White, 1951.
17. Doll, 1944.

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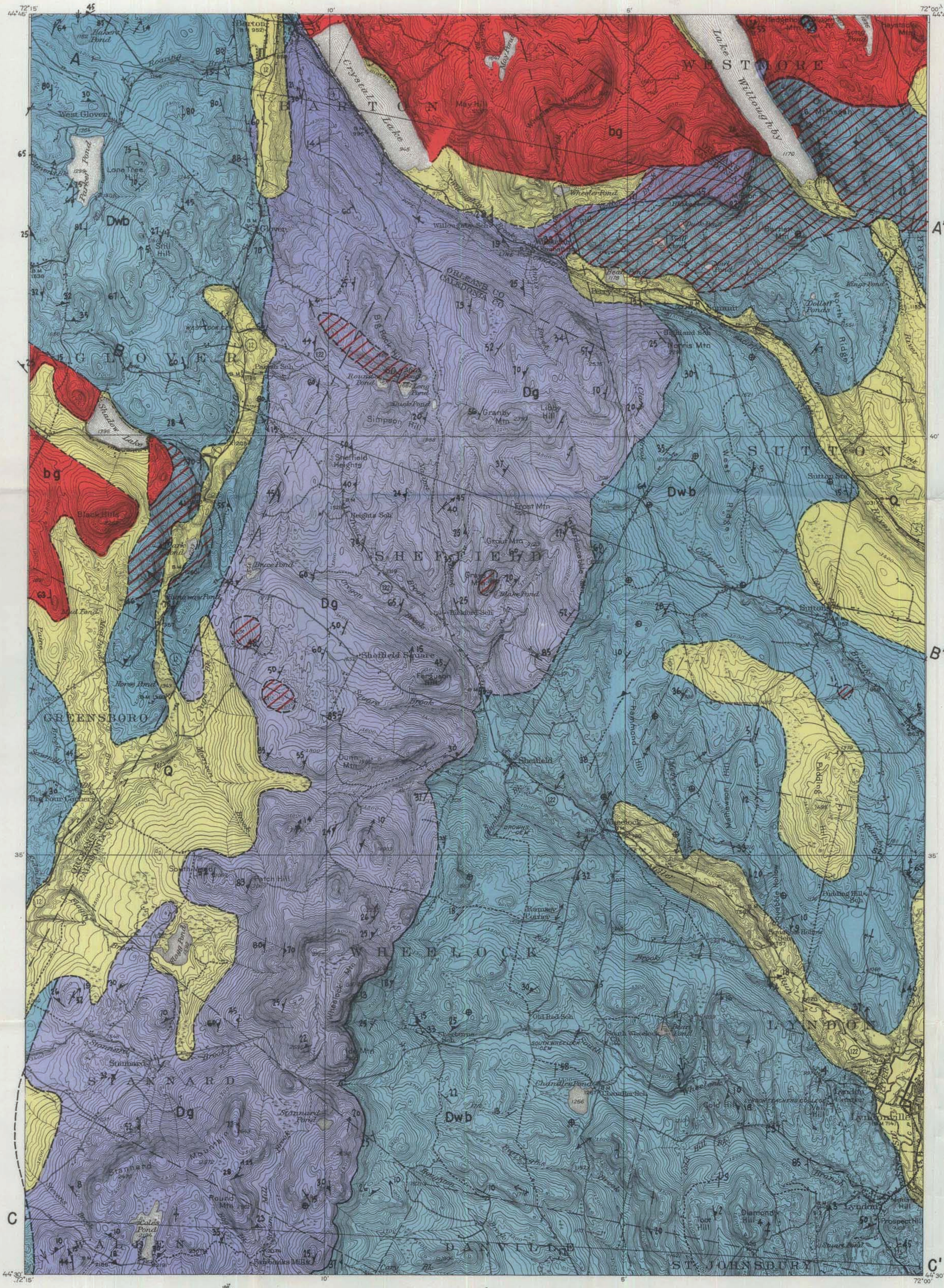


Topography by U.S.G.S. 1934-1935
Revised 1951

Scale 1/62500

Datum is mean sea level

LEGEND

Quaternary

Q — Pleistocene cover

?Devonian

bg — Biotite granite

? Lower to Middle Devonian ?

Dg — Gile Mountain formation

Dwb — Waits River formation
Dwv: Standing Pond Lava
Dwb: Barton River member

Abundance of Granitic dike rocks

Strike and dip of beds

Horizontal beds

Strike and dip of foliation or schistosity

Strike of vertical foliation or schistosity

Bearing and plunge of lineation in b

Bearing of horizontal lineation in b

Direction of glacial striations

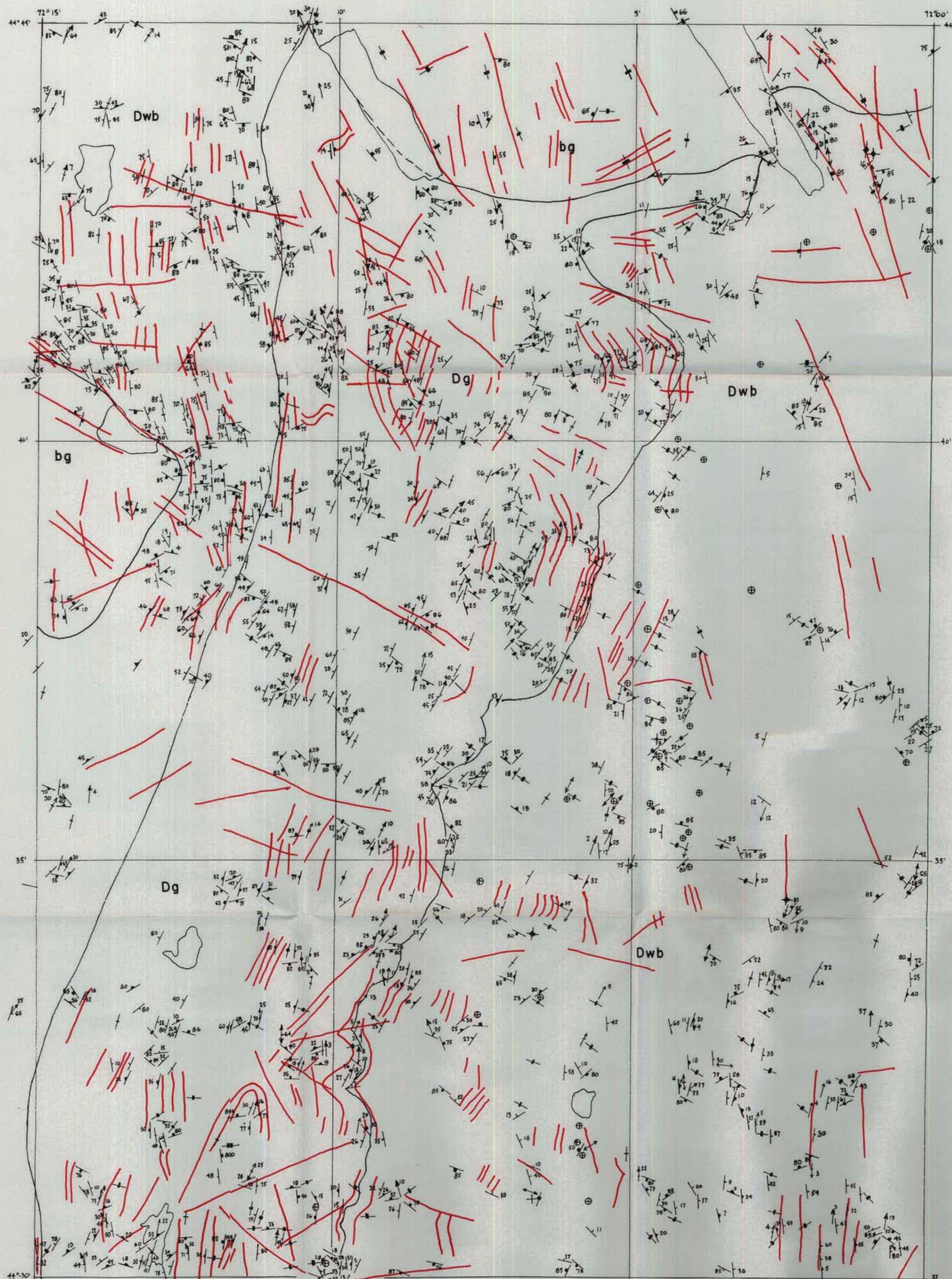
GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE LYNDONVILLE QUADRANGLE, VERMONT

PLATE I

Geology by John G. Dennis
1954 - 1955
Published 1956

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist

TECTONIC MAP OF THE LYNDONVILLE QUADRANGLE, VERMONT



LEGEND

STRUCTURAL SYMBOLS

- Strike and dip of beds
- Strike and dip of foliation or schistosity
- Strike of vertical foliation or schistosity
- Horizontal schistosity
- Horizontal beds
- Bearing and plunge of lineation in b
- Bearing of horizontal lineation in b
- Strike and dip of joints
- Air photograph linears

Formation symbols as on geologic map

1 1/2 0 1 2 3 4 MILES

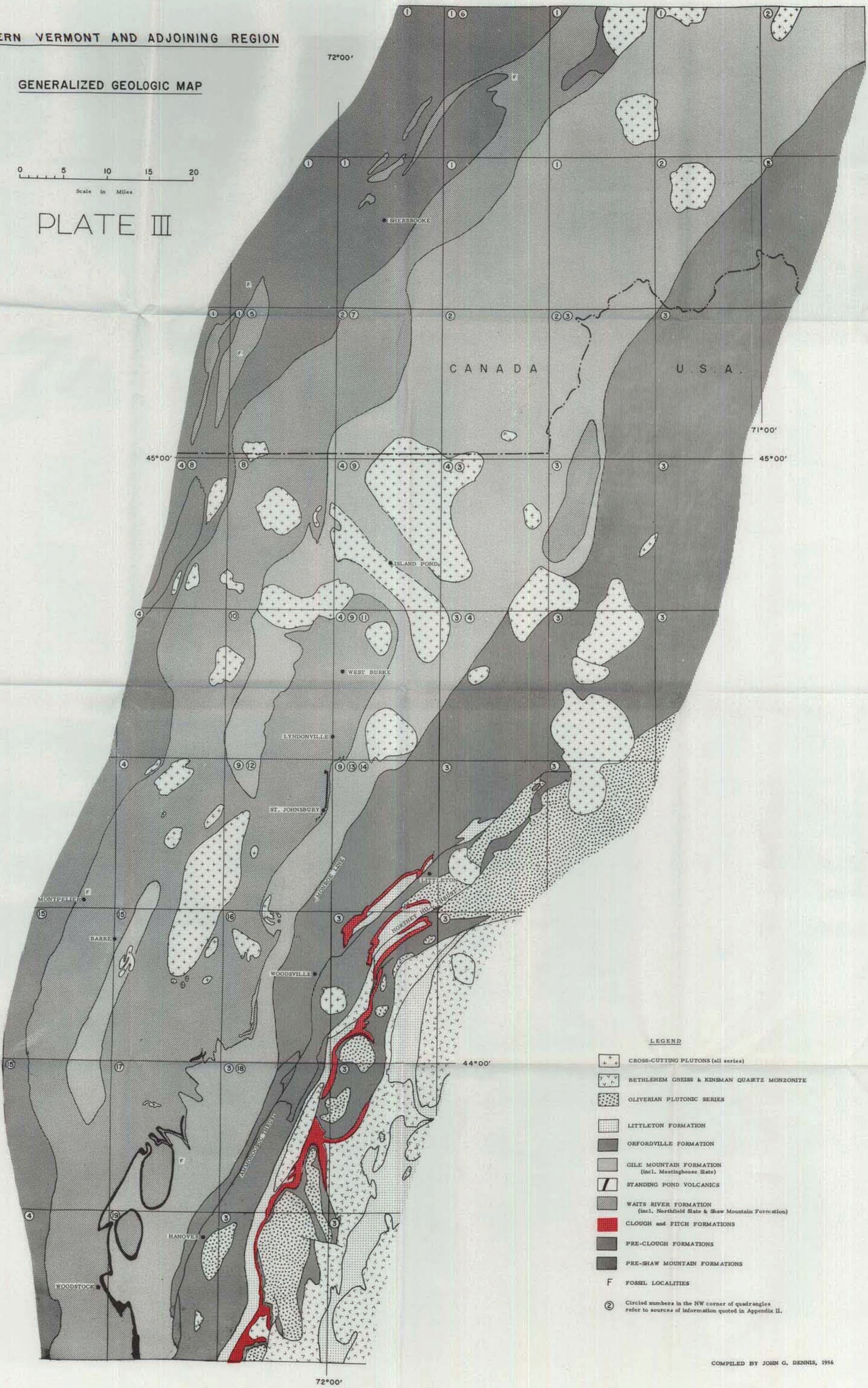
SCALE

PLATE II

EASTERN VERMONT AND ADJOINING REGION

GENERALIZED GEOLOGIC MAP

PLATE III



COMPILED BY JOHN G. DENNIS, 1956

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