GEOLOGY OF THE CONCORD-WATERFORD AREA, VERMONT

By

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ABSTRACT

The Concord-Waterford area is located in northeastern Vermont, between St. Johnsbury and the Connecticut River. Its geologic setting is the belt of predominantly low-grade metamorphic rocks of Paleozoic age between the Willoughby arch to the west and the White Mountains to the east.

Two sedimentary sequences crop out within the area: the Waits River-Gile Mountain sequence of Devonian, and perhaps in part Silurian age, and the Albee-Ammonoosuc sequence, of pre-Silurian age. The two sequences are separated by the Monroe contact which traverses the area in a northeasterly direction and is believed to be a fault, the Monroe fault. Devonian and possibly Silurian rocks in this area consist mainly of alternating fine-grained quartzite and schist, and subordinately of marble and volcanics. The pre-Silurian sequence east of the Monroe fault consists of alternating schist, quartzite, and volcanics. A few bodies of intrusive rock of various ages have been emplaced in the sedimentary rocks.

However, the nature of the break at the Monroe contact is controversial: tops of the Devonian and possibly Silurian rocks may face east at the Monroe contact, which would then represent a thrust of pre-Silurian over Devonian and possibly Silurian rocks; or tops of both sequences may face west, whereby the Monroe contact would be the Taconic unconformity. There is still insufficient evidence to decide between the two interpretations.

Dips are almost uniformly steep to vertical, and major structures are deduced solely from stratigraphic evidence. Minor structures, in this area, are useful guides to structural evolution, but they are not safe guides to major structural features. Two stages of deformation are deduced from the minor folds and from the secondary foliation in the rocks.

Metamorphic history is closely linked to structural history. An early, low-grade metamorphism was succeeded by a more localized medium- to high-grade metamorphism.

The two phases of structural and metamorphic evolution recorded in

the rocks may represent two distinct cycles, or stages in an essentially continuous history at the time of the Acadian (Late Devonian) orogeny.

INTRODUCTION

Location

The Concord-Waterford area, Vermont, comprises the Vermont portion of the Littleton quadrangle and is situated between the Connecticut River and latitude 44°30′, and longitudes 71°45′ and 72°00′.

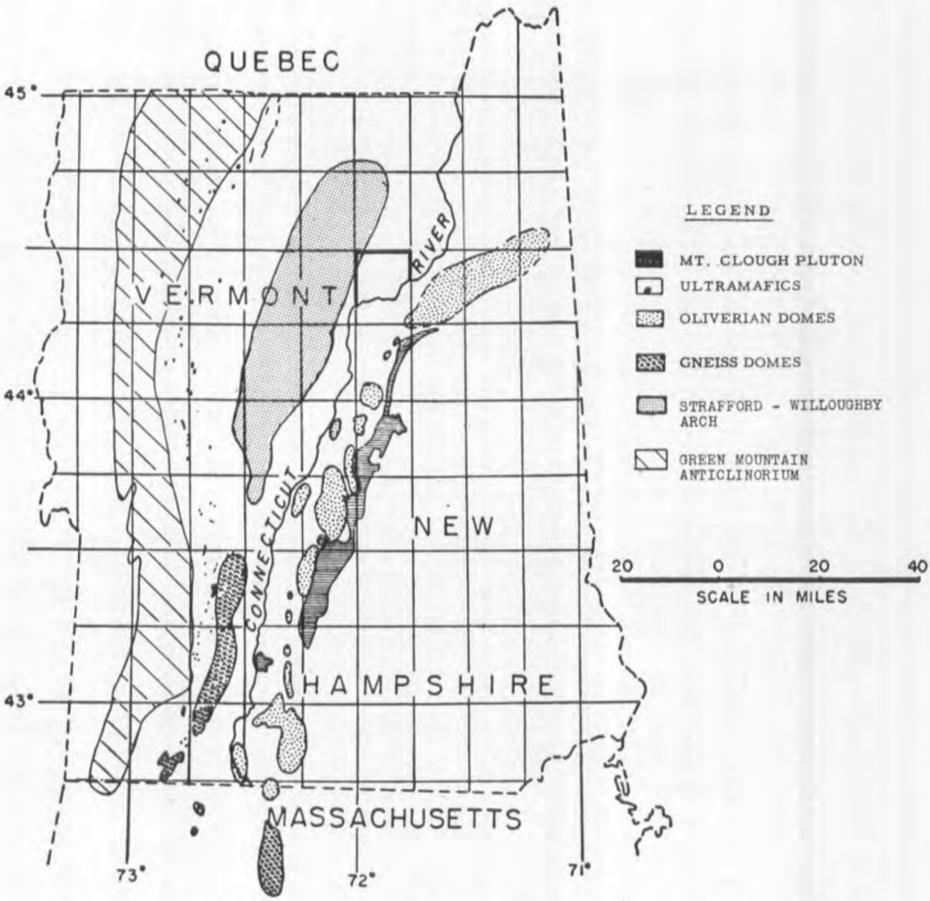


Figure 1. Index map, showing domal structures in northwestern New England. Concord-Waterford area in heavy outline. Modified from Lyons (1955, Figure 4).

Geologic Setting

The area lies in a belt of Paleozoic metamorphic rocks occupying the Connecticut River depression (Fig. 2). The northwestern part of the area is underlain by Devonian and possibly Silurian formations, known as the "Vermont sequence"; these rocks are separated from rocks of the Albee formation and Ammonoosuc volcanics, called the "New Hampshire sequence," by the Monroe contact or fault (Fig. 2). The structural significance of the break at the Monroe contact is controversial. A few small intrusive bodies occur in the area.

Previous Work

Early geologic reconnaissance work in the region was done by Jackson (1844), Adams (1845), E. Hitchcock (1861), C. H. Hitchcock (1874, 1877, 1878, 1884), Richardson (1902 and 1906), and Dale (1909, 1910, 1914, 1915, and 1923). Crosby (1934) studied the geology of the Fifteenmile Falls area in connection with the Littleton Hydroelectric Development. Detailed geologic work in adjacent quadrangles has been completed as follows (Fig. 1b): Littleton quadrangle (New Hampshire portion), Billings (1937); St. Johnsbury quadrangle (New Hampshire portion), Billings and Eric, *in* Eric (1942); Lyndonville quadrangle, Dennis (1956). At the time of writing (1957), most of the Burke quadrangle (B. Woodland) and of the Vermont portion of the St. Johnsbury quadrangle (L. Hall) had been mapped.

Background of Study

Past geologic work in western New Hampshire by Billings (1937) and others has established the fact that some of the rocks of that area are of Devonian and Silurian age. Unconformably beneath these rocks are several pre-Silurian formations, the exact age of which is unknown. In 1940 it was hoped that by carrying the detailed mapping westward and northeastward into Vermont, a definite sequence could be established between the fossiliferous Silurian and Devonian rocks of western New Hampshire and the alleged "Ordovician" terranes of central Vermont (Richardson, 1919). Similar, but independent, investigations were carried on at about the same time in central Vermont by W. S. White and R. H. Jahns (White and Jahns, 1950; White and Billings, 1951).

The "graptolites" upon which Richardson (1919) had based his determination of age in central Vermont were identified by R. Ruedemann, and were reported by him to be Early to Middle Ordovician. These "graptolites" had a preferred orientation in the planes of cleavage; and

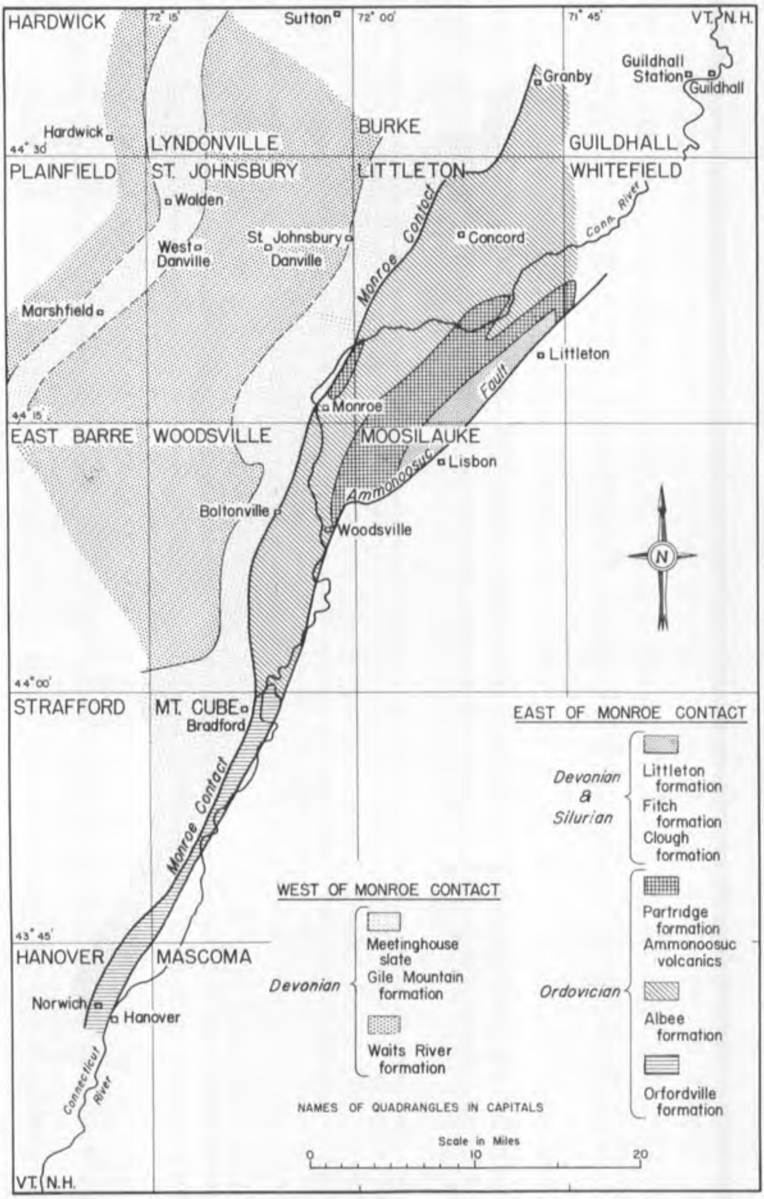


Figure 2. Generalized geologic setting of the Concord-Waterford area which is that portion of the Littleton quadrangle north of the Connecticut River.

these planes were not parallel to the bedding. Foyles (1931, p. 252) as well as Currier and Jahns (1941) showed that these markings are not organic, but are lineations caused by the parallel arrangement of micaceous minerals. This considerably complicated the problem, for the age of the rocks of central Vermont now became uncertain. It was hoped, nevertheless, to establish a stratigraphic sequence across eastern Vermont that would serve as a basis for the determination of relative ages. As detailed mapping proceeded westward and northwestward it soon became apparent that the complexity of the structures was further intensified by what seemed to be a great fault, called the Monroe fault (Eric, White, and Hadley, 1941), the presence of which precluded the establishment of a satisfactory sequence. Subsequent work in Vermont and western New Hampshire, in addition to the present work, has provided sufficient data for at least two hypotheses regarding the age and structure of the rocks in eastern Vermont.

Present Study

The senior author did the field work in the summers of 1940 and 1941, wrote the original report as a thesis at Harvard University (Eric, 1942) and revised it in 1949 after interruption by the war. The junior author spent some time in the area in 1954 and 1955 while working for the Vermont Geological Survey in the neighboring Lyndonville quadrangle; he returned to the Concord-Waterford area in the summer of 1956 for a brief field review, brought the maps and manuscript up to date, and prepared the material for publication by the Vermont Geological Survey.

Acknowledgments

The senior author was ably assisted in the summer of 1940 by George E. Kruger, and in the summer of 1941 by J. Francois de Chadenedes.

Prof. Marland P. Billings, of Harvard University, under whose direction the work was done, spent two weeks in the area during the summer of 1941 and made many valuable suggestions. Professor Billings also spent a considerable time reading and criticizing the manuscript of this paper, for which the writers are duly grateful. Prof. Esper S. Larsen, Jr., then of Harvard, gave much aid in the study of the thin sections, especially those of the intrusive rocks.

Special thanks are due to Prof. Charles G. Doll, State Geologist, for his support; and to Dr. Wallace M. Cady, of the U. S. Geological Survey, who gave much valuable advice.

STRATIGRAPHY

Introduction

The greater part of the Concord-Waterford area is underlain by metamorphic rocks of sedimentary origin. Igneous rocks, although abundant, are subordinate. No fossils were found in the area, but fossils have been found (Billings and Cleaves, 1934; Billings, 1937) in the New Hampshire portion of the Littleton quadrangle. The relation of these fossils to the regional geology indicates that most of the rocks of sedimentary origin in the Concord-Waterford area southeast of the Monroe fault are pre-Middle Silurian (Ordovician?). The present authors believe that the rocks northwest of this contact are Devonian, and possibly Silurian in part, in general agreement with Doll (1943 and 1944), White (1946), Billings (1948), Hadley (1950), Dennis (1956), and Cady (1956). Recent full discussions of the age question in eastern Vermont are by Billings (1956) and Dennis (1956).

Billings (1948) favored Devonian and Silurian age for the rocks west of the Monroe contact but, pending a clearer resolution of the problem,

SYSTEM	SERIES	WEST of MONROE CONTACT	EAST of MONROE CONTACT
DEVONIAN	LOWER	Meetinghouse Gile Mountain Standing Pond ? Waits River Northfield	Littleton
SILURIAN	MIDDLE	Shaw Mountain	Fitch
	MIDDLE or LOWER		Clough
ORDOVICIAN	MIDDLE and LOWER	Cram Hill - Moretown Stowe	Partridge Ammonoosuc Albee Orfordville

Figure 3. Correlation table. Modified from Billings (1956, Figure 7).

labeled these formations "Ordovician?" on his map of New Hampshire (1955), taking into account the divided opinion of the time at which the map went to press (e.g. Lyons, 1955 and Cady, 1951). However, in the accompanying text (Billings, 1956) he favorably discusses the possibility of Silurian and Devonian age. Similarly, Cady in a later publication (1956) believes that the weight of the evidence now favors Silurian and Devonian age.

Pre-Silurian Rocks

Parts of the Albee formation and of the Ammonoosuc volcanics (Billings, 1937) are exposed in the area. These were evidently laid down in the northern Appalachian early Paleozoic eugeosyncline of the Magog Belt (Kay, 1937, p. 290).

Albee Formation

General statement. Approximately half of the area is underlain by the Albee formation (Plate 1). Easily accessible exposures can be seen along State Highway 18, between Stiles Pond and the Connecticut River.

Lithology. The Albee formation consists of massive blue-gray quartzite, argillaceous quartzite, pure white quartzite, green slate, quartz-chlorite-sericite schist, quartz-mica schist, garnet-staurolite schist, garnet-sillimanite schist, and quartz-feldspar schists of volcanic origin. The chlorite, biotite, garnet, staurolite, and sillimanite zones of metamorphism (Harker, 1932) are represented: rocks of the chlorite zone are present in the southeastern part of the area; those of the biotite zone in the central and southwestern portions; the garnet, staurolite and sillimanite zones are limited to the vicinities of Miles Mountain and Kirby Mountain (Pl. 1).

The most conspicuous, although not the most abundant, lithologic type is massive blue-gray quartzite. It is very common in the Albee formation. The rock is extremely hard, breaks with a conchoidal fracture, and has a characteristic "sugary" texture. Most of the beds do not exceed 2 feet in thickness.

In thin section the quartzite is seen to consist almost entirely of quartz, with minor amounts of chlorite, sericite, biotite, and sodic plagioclase. One section showed an estimated mode of 98 percent quartz. The texture is granoblastic.

Locally the quartzite is sufficiently abundant to be a mappable unit (Pl. 1). This quartzite unit is resistant to erosion, and holds up a range of high hills, the most important of which are Hurlburt Hill, Fuller Hill, Jackman Mountain, Shaw Mountain, and Goodreault Hill. The quartzite unit occupies approximately the crest of the Gardner Mountain anticline, which will be discussed in the section on structure.

In places the quartzite beds within the quartzite unit are unusually thick; beds 20 to 50 feet thick are common. The maximum thickness of any one bed of quartzite is approximately 100 feet, measured on Hurlburt Hill. In contrast, beds of schist within the quartzite unit are only 1 to 3 feet thick.

Pelites (mostly rocks that were originally shales) make up the greater part of the Albee formation. These rocks are now green slate, quartz-chlorite-sericite schist, quartz-mica schist, garnet-staurolite schist, and garnet-sillimanite schist. The structure is characteristically schistose.

Green slate is confined to the extreme upper part of the Albee in the chlorite zone. Accessible outcrops of green slate are located along the banks of the Connecticut River just below the base of the Ammonoosuc volcanics.

The rock is a greasy-looking, fine-grained green to greenish-gray slate. Under the microscope it is seen to consist predominantly of sericite (generally more than 50 percent), with 25-30 percent quartz, 15 percent chlorite, and minor amounts of pyrite and magnetite.

Quartz-chlorite-sericite schist is abundant in the Albee of the chlorite zone. This rock consists predominantly of quartz (more than 50 percent), and approximately equal amounts of sericite and chlorite, generally 15-20 percent each. The quartz-chlorite-sericite schist is commonly brownish-yellow due to the weathering of pyrite and magnetite, particularly in Waterford and southern Concord.

In the biotite zone quartz-mica schist takes the place of the quartz-chlorite-sericite schist of the chlorite zone. Biotite, sericite, and muscovite may each compose 20 percent of the rock; some chlorite survives; the remainder is mostly quartz. In the garnet and staurolite zones exposures of Albee are not common. Garnet-staurolite schist crops out on the low ridge 1 mile northeast of the village of North Concord. The garnets are less than 3 millimeters in diameter and the staurolites are not more than 4 millimeters long. Together they make up less than 5 percent of the rock.

Exposures of the Albee formation within the sillimanite zone are abundant on Miles Mountain, where the sillimanite is associated with garnet and biotite but apparently not staurolite. Estimated modes show approximately 50 percent quartz, 20 percent biotite, 10 percent muscovite, 10 percent garnet, 7 percent sillimanite, and 3 percent magnetite. One thin section, however, from the southwest side of the 1900-foot hill 1 mile west of the summit of Miles Mountain, showed an estimated mode of approximately 40 percent quartz, 25 percent garnet, 20 percent sillimanite, 10 percent plagioclase (Ab_{30}), and 5 percent biotite. But this sillimanite-rich rock is not typical. Most of the sillimanite porphyroblasts are approximately 5 millimeters long, but they range in length from microscopic to 7 centimeters. Commonly the orientation of these needles and columns of sillimanite bears no relation to the structure of the rocks.

A metavolcanic unit of the Albee formation is distinguished on Plate 1. It is located half a mile east of Brooks School in the southern part of the town of Concord. The rock consists essentially of soda-rhyolite tuff and is lithologically identical with much of the overlying Ammonoosuc volcanics, but is interpreted here as a lens of tuff within the Albee formation.

Thickness. The base of the Albee formation is not exposed in the Littleton quadrangle. It is, however, exposed in the Mt. Cube quadrangle to the southwest (Fig. 1b), where Hadley (1942) found that the thickness is approximately 5,000 feet. In the Littleton-Moosilauke area of New Hampshire, Billings (1937) gives a minimum thickness of 4,000 feet for the Albee.

Age. No fossils have been found in the Albee formation, but it is known from relations in the New Hampshire portion of the Littleton quadrangle (Billings, 1937) that it is of pre-Silurian age. The lower age limit is not so certain. White (1946) tentatively correlated the Albee with the lower part of the Cram Hill formation of central Vermont, which Currier and Jahns (1941) in turn had correlated with the Middle Ordovician Beauceville formation of Quebec. Failing other evidence, the Albee formation and the overlying Ammonoosuc volcanics are here tentatively assigned Middle Ordovician (?) age.

Ammonoosuc Volcanics

General Statement. The Ammonoosuc volcanics overlie the Albee formation conformably. The contact is drawn within a zone of interbedding between the two formations. The volcanics are exposed in two belts, one in southern Concord and southeastern Waterford, the other mainly in the town of Monroe, N. H. (see Fig. 2.).

The two belts of Ammonoosuc volcanics are not continuous along the regional trend: the Ammonoosuc volcanics probably occupy the cores of doubly plunging synclines.

Lithology. In this area the Ammonoosuc volcanics consist of white to gray soda rhyolite. Two types of felsic volcanics were distinguished in the field: those with and those without glassy bluish anhedral "phenocrysts" of quartz. Crystals of feldspar can be distinguished in some outcrops. Under the microscope the rock is seen to consist of albite-oligoclase, quartz, orthoclase, sericite, and chlorite, except that in the biotite zone small porphyroblasts of biotite are present. Most of the rocks were probably derived from tuffs, for the original tuffaceous character of the rocks is preserved in many outcrops. Some of the more aphanitic phases may have been flows. Bedding is obscure except along the Connecticut River in Concord, where the contact of the Ammonoosuc volcanics and the underlying Albee formation is exposed.

Agglomerate crops out on the New Hamshire side of the Littleton Dam, but exposures of this rock were not observed in Vermont. Possibly this agglomerate underlies parts of southeastern Waterford and is concealed by gravel and till.

Thickness. According to Billings (1937) the Ammonoosuc volcanics have a thickness of approximately 2,000 feet. The top of the formation may not be exposed on the Vermont side of the Connecticut River, but it is exposed to the south, in the New Hampshire portion of the Littleton quadrangle. No additional data concerning the thickness were obtained.

Age. No fossils have been found in the Ammonoosuc. It overlies the Albee formation conformably, and it underlies fossiliferous Silurian rocks in New Hampshire. Like the Albee, it has been considered pre-Silurian, probably Ordovician, in age.

Devonian and Silurian(?) Rocks

St. Francis Group

The Northfield slate, Waits River formation, Gile Mountain formation, and Meetinghouse slate correlate (Dennis, 1956, p. 32) with rocks which, in Quebec, have been grouped under the name of "St. Francis series" (Cooke, 1950). As it is convenient to refer to these four formations as a unit, the name St. Francis *group* has been applied to them by Dennis, in accordance with the terminology accepted in the United States.

Waits River Formation

General Statement. The Waits River formation (C. H. Hitchcock, 1878; Richardson, 1906; Dale, 1914; Currier and Jahns, 1941) is very extensive in other parts of Vermont, but in the Concord-Waterford area it is exposed only in the extreme northwestern corner. An excellent exposure is readily accessible south of the Lyndonville Dam in the extreme northwest corner of the map, Pl. 1.

Table 1.
SECTION OF WAITS RIVER FORMATION AT EMERSON'S FALLS

Lithology	*Thickness*	
	Feet	*Inches*
East (*top*)		
marble	2	8
phyllite	5	3
marble	4	1
quartzite	5	3
marble	2	3
phyllite	2	4
marble	1	4
phyllite	2	3
marble	2	6
phyllite	0	8
marble	7	5
phyllite	2	8
marble	7	5
phyllite	9	0
marble	7	8
quartzite	3	4
phyllite	3	10
marble	12	6
phyllite	7	5
marble	5	10
phyllite	7	3
marble	12	0
phyllite	4	8
marble	9	6
phyllite	1	9
marble	5	1
phyllite	7	0
marble	3	2
phyllite	1	3
quartzite	1	6
phyllite	2	2
quartzite	1	0
phyllite	1	3
quartzite	2	5
phyllite	1	8
marble	3	4
phyllite	3	0
marble	12	6
phyllite	7	5
marble	1	3
quartzite	0	4
marble	11	6
quartzite	0	8
West (*bottom*)		
	—	—
	Total: 197 feet	

Lithology. In the Concord-Waterford area the Waits River formation consists of impure bluish arenaceous marble, argillaceous marble, argillaceous quartzite, phyllite, and dark gray slate. Some idea of the lithology can be obtained from a section (Table 1) in the St. Johnsbury quadrangle measured by tape at the falls in Sleepers River (Emerson's Falls), near the U. S. Fish Hatchery 1 mile northwest of the village of St. Johnsbury and 2 miles west of the western border of the Littleton quadrangle. The section was measured normal to the bedding planes of each bed. The amount of repetition due to folding is not known, but is probably slight. All observed folds were avoided.

This section probably is fairly representative of the Waits River formation as it occurs in the Concord-Waterford area and in the eastern part of the St. Johnsbury quadrangle. The preponderance of calcareous rock is noteworthy. The sum of the thicknesses and the relative amounts of each variety of rock are as follows:

Lithology	*Thickness (feet)*	*Percentage of total*
quartzite	14	7
phyllite	71	36
marble	112	57
	197	100

Although approximately 50 percent of the Waits River formation in this section is composed of marble, it is largely an impure, siliceous dolomitic marble. In fresh specimens it is a relatively hard rock, light blue-gray, but on weathered surfaces it is soft and light to dark rusty brown, due to limonite. Quartz rather than carbonate is the chief constituent in many places: in one thin section an estimated mode showed 70 percent quartz, 20 percent carbonate, and 10 percent phlogopite. In general, however, the quantity of quartz is about 50 percent of the total. Carbonate is generally present as small disseminated grains or as shreds occupying the interstices between quartz grains.

In this area the Waits River formation lies entirely in the biotite zone. Brown phlogopite is abundant, forming up to 20 percent of the rock; pinkish grossularite and rosettes of pale green actinolite are also present. Some of the garnets are over 15 millimeters in diameter. In some places rims of magnetite have grown around the phlogopite.

The argillaceous quartzite, phyllite, and slate are similar to rocks in the Gile Mountain formation, and will be described below.

Thickness. There can be little doubt that Waits River formation is very thick. Only the extreme upper part is exposed in the Concord-Waterford area. The total thickness of the formation may be measured in tens of thousands of feet (Jahns and White, 1941; White and Jahns, 1950).

Age. The Waits River formation has been considered to be Ordovician in age. However, the "graptolites" upon which Richardson (1919) based his correlation have been shown to be inorganic (Currier and Jahns, 1941). The age of the Waits River formation, therefore, is not known. In the Concord-Waterford area no fossils were found in it. Currier and Jahns (1941) believed the Waits River to be Ordovician. Doll (1943a) found large crinoid calyces in the overlying Gile Mountain formation at Westmore, which led him to date it tentatively as Silurian or Lower Devonian. Later (1943b) he found the impression of a large spirifer in the Gile Mountain of the Strafford quadrangle, suggesting Lower Devonian age. Some geologists (e.g. White and Jahns, 1950; and White and Billings, 1951) did not accept this fossil evidence as valid. However, since 1956 most geologists working in western New England have accepted Silurian or Devonian age for the Waits River formation.

Gile Mountain Formation

General Statement. The Gile Mountain formation (Doll, 1944, p. 18) overlies the Waits River formation throughout much of eastern Vermont. It is exposed over a considerable area in the northern and western parts of the Concord-Waterford area, where it occupies most of the region between the Waits River formation on the west and the Monroe contact on the east. Readily accessible outcrops can be seen just north of U. S. Highway 2, particularly in the town of Kirby. Its contact with the Waits River formation is gradational by interbedding of lithologies. Though fairly readily located within 500-1000 feet in most places, it is particularly difficult to define in the northwest corner of the Concord-Waterford area because of what seems to be small-scale repetition by isoclinal folding.

Lithology. The Gile Mountain formation consists of interbedded dark-gray slate, light-gray argillaceous quartzite, phyllite, rhyolitic and amphibolitic metavolcanic rocks, a few thin beds of impure marble less than 1 foot thick and, near some of the intrusive bodies, quartz-mica schist and biotite-garnet-staurolite phyllite. In the northwestern part of the area two units that contain abundant argillaceous quartzite are separately designated on Plate 1.

The remainder of the Gile Mountain formation — that part which is not differentiated on Plate 1 — consists of rather thin alternating beds of dark gray slate or phyllite and light gray to brownish argillaceous quartzite. Beds more than 3 or 4 feet thick are not common; at many localities laminae 2 or 3 millimeters in thickness are very common, and some can be seen only under the microscope. Most of the beds are about 1 to 3 inches thick. The quartzite beds are usually thicker than the darker slaty beds.

Graded bedding was observed at several localities. In this area graded bedding appears to have little value as a top-and-bottom criterion, for no systematic orientation could be found; in some places the top seems to be to the east, in others to the west. This lack of systematic orientation could be caused by many isoclinal minor folds; but it might also be due to the fact that many beds are actually coarsest at the top, finest at the bottom, contrary to the usual sequence (Cooke, 1937, p. 18). It seems possible that in the Concord-Waterford area the Gile Mountain formation is, in general, not highly folded, but in any case, graded bedding in this particular area is unreliable for deciphering major structures.

Each of the two units in which argillaceous quartzite is abundant (Plate 1) holds up ranges of northeast-trending hills. The hills are not as high as those in the Albee formation because the quartzite is much less pure than that in the Albee, and consequently it is more easily eroded. Beds of light gray quartzite 5 to 10 feet thick are not uncommon, and beds 2 to 3 feet thick are abundant. Interbedded with the quartzite are layers of dark gray phyllite several inches thick. Besides quartz, the quartzite has sericite, biotite, and garnet.

Under the microscope the quartzite is seen to consist of anhedral grains of quartz, with an average diameter of 0.1 to 0.2 millimeter, scattered shreds of chlorite, irregular corrugated flakes of biotite and muscovite, euhedral porphyroblasts of garnet, and disseminated opaque matter. Quartz constitutes approximately 70-80 percent of the quartzite; the remaining 20-30 percent is chiefly biotite and muscovite, and 1 to 2 percent is garnet.

TABLE 2

SECTION OF VOLCANICS BETWEEN BIBLE HILL AND MT. PISGAH

	Lithology	*Thickness (feet)*
east		
	Dark gray slate	——
	Dense white felsite	15
	Medium-grained biotite gneiss	80
	Dense white felsite	55
	Fine-grained biotite gneiss	70
	Fine-grained amphibolite	40
	Dense white felsite with large blue-green hornblende crystals	35
	Dense white felsite	50
	(gap)	100
	Impure marble	——
west		

Volcanics. Several volcanic beds are intercalated in the Gile Mountain formation. A number crop out in the vicinity of Bible Hill, in the town of St. Johnsbury, but only two were of sufficient extent to be shown on Plate 1. Where present in beds of mappable width these volcanics are useful as markers. A typical section was measured about midway between Bible Hill and Mt. Pisgah (Table 2).

The felsite was originally rhyolite tuff. In some places numerous conspicuous euhedral feldspar crystals 2 or 3 millimeters in diameter stand out sharply against the aphanitic matrix. The biotite gneiss was probably originally coarser grained rhyolite tuff. It is not certain whether the amphibolite represents pyroclastic rock or flows, but the composition is basaltic. The fine-grained amphibolite is black, and breaks up into blocks of various sizes where it crops out, an unusual characteristic in this region and one which would make even a very thin layer an easily followed horizon marker. The amphibolite dies out along the strike, however, within a few hundred feet in both directions.

Thickness. Because the amount of folding is not known, the thickness of the Gile Mountain formation in the Concord-Waterford area cannot be precisely determined, but is estimated at about 5,000 feet.

Age. Structural evidence elsewhere (Doll, 1944; Dennis, 1956; Billings, Rodgers, and Thompson, 1952) indicates that the Gile Mountain formation lies stratigraphically above the Waits River formation; Dennis (1956, p. 32) reviews recent opinions on the age of the Waits River and Gile Mountain formations. The present authors agree with the majority opinion that these rocks are Devonian, and possibly in part Silurian. The Gile Mountain may be entirely Devonian.

Meetinghouse Slate

General Statement. The Meetinghouse slate, named by Doll (1944), consists very largely of dark gray slate which at one time (Richardson, 1906) was quarried three-quarters of a mile southwest of Stiles Pond in the town of Waterford (marked on Pl. 1). Richardson (1906) had named it the Waterford slate.

Lithology. The Meetinghouse slate consists predominantly of quartz, sericite, shreds of chlorite, conspicuous porphyroblasts of biotite and, in a few places, garnet and staurolite. In the town of Victory a single outcrop (which may not be exactly in place) was found containing needles of sillimanite. Carbon is present as small disseminated grains scattered through the rock, and in some places is so abundant that it gives the thin sections a black, opaque appearance. In the northern part of the area, where metamorphism has been more intense, the Meetinghouse slate is a phyllite or schist.

On fresh outcrops the sericite of both the Meetinghouse slate and thin slate units in the Gile Mountain formation reflects light brilliantly from cleavage surfaces, so that in sunlight it has a shiny, almost metallic luster. The shiny luster of the slates in the Meetinghouse slate and Gile Mountain formation is one of the best criteria for rapidly distinguishing these formations from the much duller slates of the Albee formation, and therefore for locating the Monroe contact, which will be discussed in another section. Besides the characteristic luster, the slate has many closely spaced plications or crinkles that are useful as keys for rapid distinction between Meetinghouse and Albee lithology.

Volcanics. A few isolated lenses of rhyolitic tuff occur in the Meetinghouse slate. It is light yellowish buff in outcrop, and extremely tough. It is fine-grained but irregularly textured. At the abandoned Waterford slate quarry, fragments of gneiss, limestone, and shale are embedded in a matrix of this tuff (Pl. 7). In thin section, it is possible to recognize quartz, biotite, apatite, and zircon. A feldspar is also present but, owing to the extremely fine grain, it has not been possible to determine its composition. The main outcrop of this tuff (too small to be shown in Plate 1) is in the center of the Meetinghouse slate, at the abandoned slate quarry.

Inasmuch as this tuff contains a fair amount of zircon, it may be possible to obtain lead-alpha ages from it; this would contribute to the solution of the age problem in the area.

Age. The Meetinghouse slate may either overlie the Gile Mountain formation and thus be at the top of the St. Francis group (thrust hypothesis of the Monroe contact) and Devonian in age, or correlate with the Northfield slate which is at the base of the St. Francis group (unconformity hypothesis of the Monroe contact — see under Tectonic Synthesis) and thus be Devonian or Silurian in age.

PLUTONIC ROCKS

Highlandcroft Granodiorite

General Statement. The Highlandcroft granodiorite (Billings, 1937) is found in one large body in southwestern Concord, and in two much smaller bodies to the northeast. A small area of diorite, considered to belong to the Highlandcroft plutonic series, but not separately mapped, is exposed just above the falls in Halls Brook near Joslin Turn, and another area of similar rock, also not separately mapped, is located in the southernmost of the two smaller bodies of granodiorite.

Petrography. The Highlandcroft is a granular medium- to fine-grained rock consisting essentially of light gray to greenish plagioclase, white to bluish glassy quartz, white microcline, and dark green hornblende. Most of the feldspar crystals range in diameter from 2 to 4 millimeters. The granodiorite is generally massive, but locally has a weak secondary foliation.

The microscope reveals these minerals: colorless albite-oligoclase, colorless anhedral quartz, colorless euhedral microcline, light-green hornblende, chlorite, epidote, sericite, calcite, magnetite, pyrite, and apatite. The plagioclase has been saussuritized, and consists of intergrown albite-oligoclase, epidote, sericite, and some calcite. Although much of the quartz shows strain shadows, and some granulation was observed, the texture is in general not cataclastic.

Age. The Highlandcroft granodiorite was emplaced in Ordovician(?) rocks, but is overlain elsewhere by the Silurian Fitch formation (Billings, 1956, p. 48 and 106; Lahee, 1913) ; lead-alpha age determinations by Lyons et al. (1957, p. 543) , assign an age of 385 ± 32 million years to the Highlandcroft. This would indicate emplacement at about the time of the Taconic orogeny.

Moulton Diorite

General Statement. Within the Concord-Waterford area metamorphosed diorite is exposed in a large body in the town of Barnet. A second locality, too small to be shown on Plate 1, is in the bed of Kirby Brook at an altitude of 1,020 feet, just northwest of the Monroe contact; a third locality, also too small to be shown on the map, is on the north slope of the hill 3,000 feet east of B. M. 1228, 3 miles northeast of the village of St. Johnsbury. Correlation of these plutons with the Moulton diorite of New Hampshire (Billings, 1937) is based upon lithologic and structural similarity.

Petrography. The Moulton diorite of this area can be divided into two types: a massive coarse-grained green and white metadiorite, and a finer grained green metadiabase. Aplitic stringers cut the bodies at many localities.

The coarser metadiorite has an ophitic texture and consists essentially of large white rectangular crystals of plagioclase and large dark green crystals of hornblende. Most of the feldspars are tabular and range in length from 2 to 10 millimeters, in width from 1 to 2 millimeters. The hornblendes are dark green in hand specimen, and range in length from 1 to 30 millimeters. The average length is 10 millimeters.

The finer-grained green metadiabase contains crystals about half the size of those in the coarser type. The hornblende is the same dark green color, but the plagioclase is olive green rather than white.

Under the microscope the diorite is seen to consist predominantly of albite-oligoclase averaging An_{10} and of hornblende ranging in color from very pale yellowish green to deep blue green. Another common mineral is epidote, which constitutes up to 10 percent of the rock. Ilmenite is abundant in grains 0.5 to 3 millimeters in diameter, commonly with rims of leucoxene. Accessory minerals are calcite, sericite, biotite, chlorite, and apatite.

Age. The metadiorite at Barnet and related rocks are correlated with the Moulton diorite of New Hampshire which intrudes the Littleton formation (Billings, 1937, p. 502). The rock is therefore Early Devonian or younger. It was affected by Acadian deformation, unlike the main New Hampshire plutonic series (see below). It may, therefore, be regarded as an early member of the New Hampshire plutonic series (Billings, 1956, p. 54).

Diabase

General Statement. Sills and dikes of metadiabase are predominant in the Albee formation just southeast of the Monroe contact (Pl. 1), where they constitute approximately 80 percent of the exposures in an area approximately 2 miles wide. Scattered bodies are also present at other localities within the Albee formation. They range in average width from 5 to 20 feet. Most of the bodies are conformable, and some may be flows.

Petrography. These greenstones are dark green to gray fine-grained completely recrystallized igneous rock of several distinctive textural types. The main types are: ophitic, aphanitic, and porphyritic with phenocrysts of albite-oligoclase up to 10 millimeters in diameter. Intermediate textural types also exist. These rocks contain albite-oligoclase, yellowish green to olive-green hornblende, blue-green hornblende, pale green actinolite, carbonate (probably largely ankerite), chlorite, epidote, and sericite. Ilmenite is abundant, and in some places has been completely altered to leucoxene. Other minerals are magnetite, apatite, rutile, pyrite, and a little quartz.

These rocks may be emplaced in what really is a continuation of the Ammonoosuc volcanics mapped by Billings in Monroe, N. H. This possibility is considered unlikely for reasons already given but, if true, tops of the pre-Silurian rocks face west at the Monroe contact.

Age. The metadiabase dikes and sills in the Albee and Ammonoosuc were emplaced in Ordovician rocks and were deformed by the Acadian orogeny. They are petrographically very similar to the Moulton diorite, and were probably emplaced at about the same time as that intrusive.

Granitoid Dikes and Sills

General Statement. Granitoid dikes and sills crop out on and near Miles Mountain, where they constitute approximately 50 percent of the outcrops. The rocks range in composition from light gray granite to dark gray tonalite.

Petrography. The granite consists of white potash feldspar as well as plagioclase, clear quartz, and biotite. The average grain size is 1 to 2 millimeters. The texture is hypidiomorphic granular; in hand specimen the rock closely resembles the Kirby quartz monzonite (described below), but in thin sections the potash feldspar is seen to be mostly orthoclase rather than microcline. Accessory minerals are magnetite, associated with the biotite; apatite, associated mainly with the plagioclase; and a little muscovite.

The tonalite is much darker than the granite. It is medium-grained, and consists of white to slightly pinkish subhedral andesine, clear quartz, green hornblende, and biotite. Accessory minerals are magnetite, sphene, apatite, and a little rutile.

Rocks intermediate between granite and tonalite are also abundant on and near Miles Mountain.

Age. The age problem of the granitoid dikes and sills is the same as that of the Kirby quartz monzonite, discussed below.

Kirby Quartz Monzonite

General Statement. The Kirby quartz monzonite is exposed in 3 small bodies, two of them at the north border of the quadrangle in the town of Kirby, the third near Stiles Pond in the town of Waterford. The rock has been described by Richardson (1906) and, in more detail, by Dale (1909, 1910, 1923).

Petrography. The rock is a light gray to white hypidiomorphic quartz monzonite consisting essentially of gray glassy quartz, biotite, microcline, albite- oligoclase, and, in some specimens, muscovite. The average grain size is 0.5 to 2 millimeters. At most localities the rock is a quartz monzonite, but in a few places where microcline is abundant (more than 67 percent of the feldspar) it is a true granite.

Age. The Kirby quartz monzonite was emplaced after the main period of Acadian deformation. A potassium-argon age determination by L. Long (personal communication, 1956) on a mica sample from the related Willoughby granite (Dennis, 1956) gave an A^{40}/K^{40} ratio of 0.0219 which corresponds to an age of 339 ± 15 million years. Assuming 95 percent retentivity of argon, the age would be boosted to 355 million years. These figures are appreciably higher than those given by Lyons et al. (1957, p. 543-544) for the New Hampshire and Oliverian plutonic series. However, the margins of error overlap. P. Damon obtained an age of 330 ± 14 million years for related pegmatite micas from Beryl Mountain, N. H.

The Kirby quartz monzonite and related rocks therefore may reasonably be correlated with the Oliverian and New Hampshire plutonic series.

Post-Metamorphic Mafic Dikes

Four post-metamorphic mafic dikes were observed in the Concord-Waterford area. They are believed to belong to the Mississippian (?) White Mountain plutonic series (Billings, 1956), although they may be Triassic. One, 3 feet wide, is exposed near B. M. 1472, 1 mile south of Shadow Lake; another, 2 feet wide, may be seen in the brook just northwest of B. M. 998 on the Caledonia-Essex county line; a third, 2 feet wide, crosses Ranney Brook at an altitude of approximately 1,100 feet; the fourth is just below Lyndonville Dam.

STRUCTURE

Major Folds

General Statement. From southeast to northwest across the area, 4 major tectonic units have been recognized:

Partridge Lake syncline
Gardner Mountain anticline
St. Johnsbury homocline
Willoughby arch (east flank)

Partridge Lake Syncline. The Partridge Lake syncline is named from Partridge Lake in the town of Littleton, N. H., because the lake is situated in the trough of the fold. Actually, it is a minor fold between the Walker Mountain syncline and the Gardner Mountain anticline (Billings, 1937). From Partridge Lake the syncline trends about N. 35° E., entering Vermont just south of the now abandoned village of Waterford, whence it trends northeast toward East Concord. It is occupied in the north by the Albee formation, in the south by the overlying Ammonoosuc volcanics. Still farther south, in New Hampshire, the stratigraphically higher Partridge, Clough, and Fitch formations are exposed in the core of this syncline (Billings, 1937); therefore, the fold plunges southwest. Associated minor folds (Pl. 2) plunge southwest at approximately 40°. The axial planes of these minor folds and the flow cleavage parallel to them dip chiefly toward the northwest, at angles of 65°-90°, although some dips of 85° SE. were observed. The Partridge Lake syncline is therefore somewhat overturned toward the southeast.

Gardner Mountain Anticline. The Gardner Mountain anticline was first described by Billings (1937) in the New Hampshire portion of the Littleton quadrangle. It trends N. 35° E., parallel to the axis of Gardner Mountain. The crest of the fold enters Vermont just south of the village of Lower Waterford and continues to the northeast, presumably into the Guildhall quadrangle. The structure is difficult to decipher in the vicinity of Miles Mountain because of the disoriented attitude of the strata and because of the paucity of outcrops. In the Littleton quadrangle, the core of the Gardner Mountain anticline is occupied by the Albee formation; Ammonoosuc volcanics are exposed on the northwest flank and a somewhat larger area of the same formation is exposed on the southeast flank. Many minor folds are superimposed upon the major structure; their axial planes dip mostly 55-90° SE. Local steep northwest-dipping axial planes were observed. Most of the minor folds plunge 25-40° SW., although one fold near B. M. 1154, 1 mile northeast of Lower Waterford, plunges 50° NE. It may therefore be assumed that the Gardner Mountain anticline is slightly overturned toward the northwest (in contrast to the Partridge Lake syncline, which is overturned toward the southeast), and that it has a moderate southwesterly plunge. Reconnaissance mapping in the southern part of the Guildhall quadrangle showed outcrops of what may be Orfordville formation, underlying the Albee, on the Granby-Guildhall road about 2 miles west of Guildhall Station; if this interpretation is correct, it indicates that the structure continues to rise toward the northeast.

St. Johnsbury Homocline. Northwest of the Monroe fault the structure is somewhat different from that to the southeast. Lack of repetition of lithologic units indicates the probable absence of major folds, and the structure is interpreted as an eastward dipping homocline, here referred to for convenience as the St. Johnsbury homocline. The strike of beds and flow cleavage is northeastward, ranging from north through northeast to east. Except for the extreme northern part of the town of Concord the dips are uniformly eastward at angles chiefly of 65-85°, although at places they may be as high as vertical or as low as 40°. The western part at least of the homocline appears to be the southeast limb of an anticline or arch, the apex of which is to the west.

There are two sets of minor folds: one early and nearly isoclinal, trending northeast with gentle northeast or southwest plunges of generally less than 2°; and one late and more open, northwest- or north-trending with north to northeast plunges of generally 20-60°. With one exception the late set of folds was seen only in the rocks northwest of the Monroe contact. The minor folds will be described in more detail later.

The trend of the St. Johnsbury homocline in the Concord-Waterford area is shown by the trace of the Waits River-Gile Mountain contact, which in the adjacent St. Johnsbury quadrangle is practically a straight line with a northerly bearing.

However, the St. Johnsbury homocline could be a syncline, if the Monroe contact is an unconformity. This possibility is discussed in the section on Tectonic Synthesis.

Willoughby Arch. The St. Johnsbury homocline which occupies the northwestern part of the Concord-Waterford area is here interpreted as the southeast limb of a great northeast-trending arch locally called the Willoughby arch (Dennis, 1956), the crest of which, at this latitude, is in the town of Danville, Vermont, some 7 miles west of St. Johnsbury village. Mapping in the Lyndonville and St. Johnsbury quadrangles has shown that the eastward dips in the Waits River formation become progressively gentler toward the west. In the quarry at West Danville just east of the outlet of Joe's Pond, the strata are nearly horizontal and the arch plunges gently northeast. The beds dip to the west farther west in Marshfield and Walden. The flat crest of the arch can be traced northeastward through the town of Sutton, Vermont. The major arch in the Strafford quadrangle to the southwest (Doll, 1944) is part of the same regional structure, named the Strafford-Willoughby arch (Fig. 1).

Minor Folds

General Statement. Two types of minor folds were observed in the area (Fig. 4). Because they were produced at different times they may be referred to as early (Fig. 4b) and late (Fig. 4a) folds respectively. Associated with these folds are two types of axial-plane cleavage, a schistosity or flow cleavage contemporaneous with the early folds, and a slip cleavage related to the late folds. In the northwestern part of the area the two types of fold are readily distinguished, but in the southwestern portion the two are more nearly parallel, so that it is not always possible to differentiate one from the other. Where bedding-plane cleavage has been folded, the folding is of the late generation. Almost all the minor folds southeast of the Monroe contact belong to the early set; most of the minor folds northwest of the contact are of the late type. With one exception, late folds are confined to the area northwest of the Monroe contact.

The two types of fold were distinguished independently in other parts of Vermont by both W. S. White and R. H. Jahns before the present study began, but M. P. Billings has expressed the opinion (personal communication) that the evidence is clearest in the Concord-Waterford area.

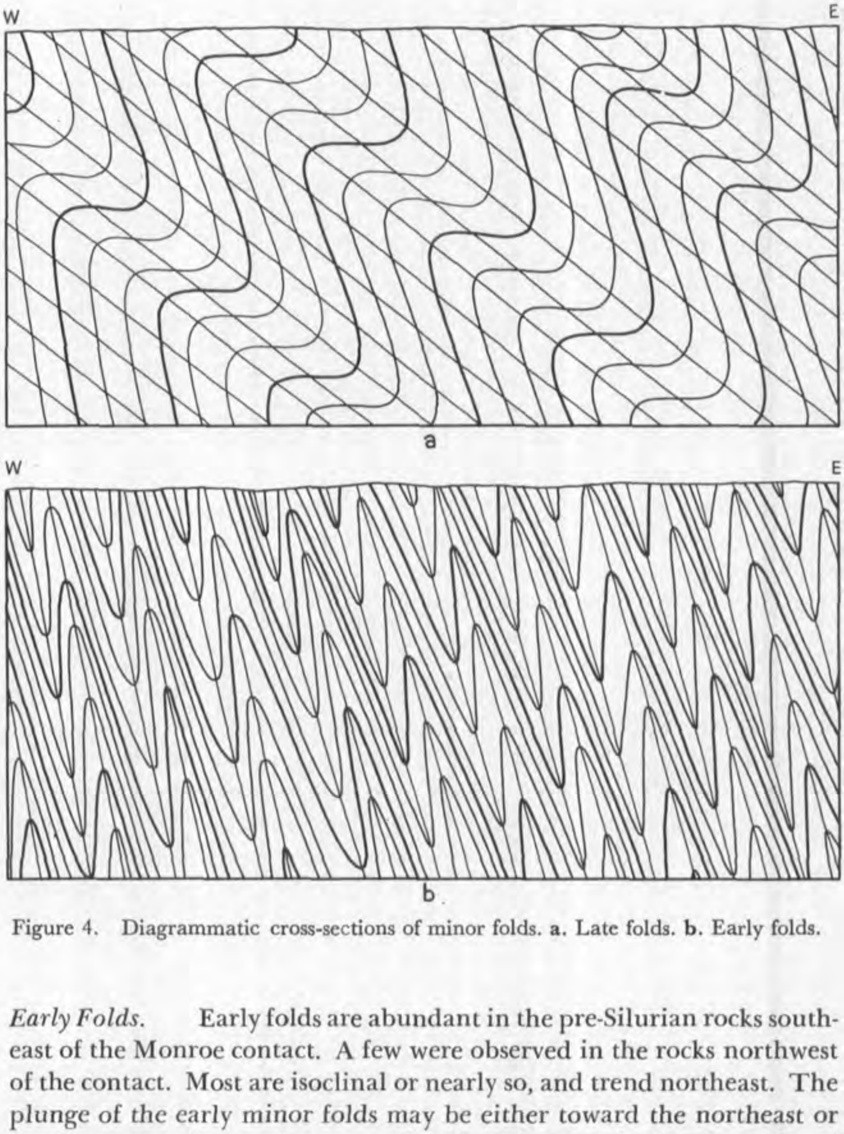


Figure 4. Diagrammatic cross-sections of minor folds. **a.** Late folds. **b.** Early folds.

Early Folds. Early folds are abundant in the pre-Silurian rocks southeast of the Monroe contact. A few were observed in the rocks northwest of the contact. Most are isoclinal or nearly so, and trend northeast. The plunge of the early minor folds may be either toward the northeast or southwest; southwest plunges predominate in accord with the regional structure southeast of the Monroe contact. Plunges of early folds southeast of the contact tend to be steeper than those northwest of it. Plunges greater than 20° are rare in the Gile Mountain and Waits River forma-

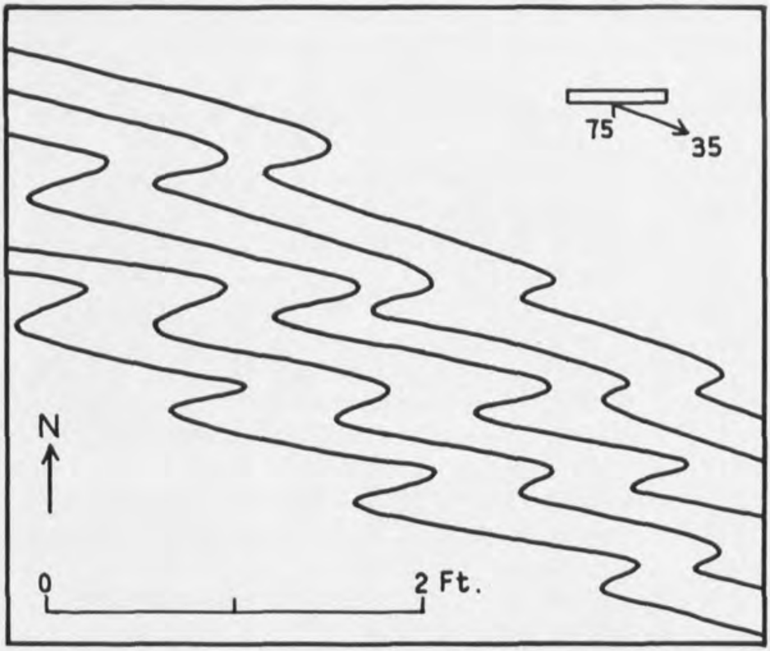


Figure 5. Diagrammatic map pattern of minor folds in sillimanite schist. West of summit of Miles Mountain. Symbol indicates strike of axial plane of fold, with direction and plunge of axis.

tions, but plunges of 40° to 50° are common in the early minor folds in the Albee formation.

Steeply plunging early folds are rare. Three thousand feet southwest of B. M. 938 on the Chesterfield Valley road there is an early fold in the Gile Mountain formation with a plunge of 85° SW. Near B. M. 1090, 2 miles south of Miles Pond, an early fold in the Albee formation plunges 85° NE. This locality is only 2.5 miles south of Miles Mountain, and the exceptional attitude of the plunge may be a result of emplacement of the granitic dikes and sills. Another divergence from the usual moderate southwest plunges in the Albee is approximately 3,200 feet south of the summit of Goodreault Hill, where a plunge of 70° SE. was observed.

The great majority of abnormal plunges in the Albee formation, however, occur on and adjacent to Miles Mountain, where many east- and southeast-plunging folds were seen, especially on the 1,900-foot hill 1 mile west of the summit of Miles Mountain (Fig. 5). Along the summit of this hill rocks are well exposed, and observation of 14 folds showed 13 of them plunging east or southeast at angles of 20°-60°, and one plunging west at 45°. It may be significant that along the summit of this particular hill, where the alignment of folds is systematic, granitic dikes and sills are rare.

Early minor folds were observed west of the Monroe contact at several localities. On the south slope of the 1834-foot hill, 4,000 feet northeast of Sugar Hill School in the town of Kirby, the axial-plane cleavage strikes slightly east of north and dips 80-85° E.; the axes plunge 20° NW. Early minor folds are well exposed on the south-facing hillside northeast of the cemetery, near the point where U. S. Highway 2 crosses the Kirby-St. Johnsbury town line. Here they trend slightly east of north. The axes plunge 10° NE., through horizontal to 10° SW. At both localities most of these folds seem to indicate a syncline to the southeast. The wave length ranges from several inches to 20 and even 50 feet. The folds are nearly isoclinal, and bedding is approximately parallel to the schistosity (axial-plane flow cleavage), except on the apices; because of their gentle plunges they rarely appear in plan. They are best displayed on south-facing *ac* joints where glacial plucking has exposed natural cross-sections (Fig. 6; Pl. 3) approximately normal to the axes of the folds.

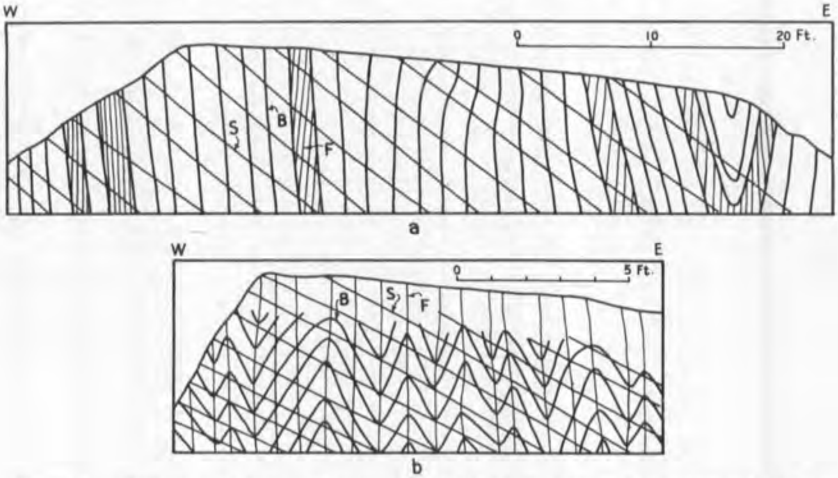


Figure 6. Diagrammatic cross-sections of early folds. Exposed on south-facing *ac* joints, 1834-foot hill northeast of Sugar Hill School. B, bedding; F, flow cleavage, parallel to axial planes of early folds; S, slip cleavage, parallel to axial planes of late folds, one of which is shown in upper part of Figure 6a.

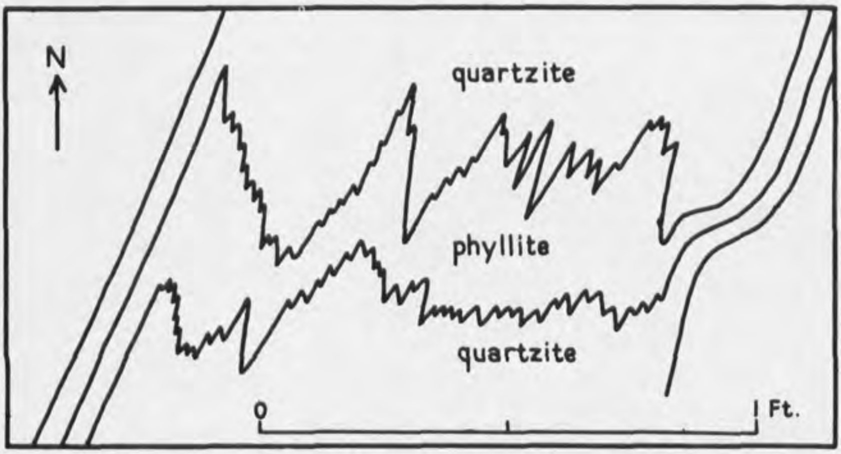


Figure 7. Map pattern of early fold. In side valley of Chesterfield Valley (see text). Axial plunge southeast. Late fold on east.

Other early folds were observed under the U. S. Highway 2 bridge near the western border of the area; 3,000 feet southeast of B. M. 938 on the Chesterfield Valley road; 2,400 feet northwest of B. M. 683 on U. S. Highway 2; on the summit of the more northerly of the 1,800-foot hills northeast of Sugar Hill School; just below Comerford Dam in the St. Johnsbury quadrangle; 1,000 feet south of South Kirby School; on the 1,080-foot hill south of the "E" in East St. Johnsbury (Pl. 1) ; 500 feet southeast of B. M. 1228 near the Spaulding Brook-Roberts Brook drainage divide; at an altitude of 1,150 feet (Fig. 7) in the small valley tributary to Chesterfield Valley, just southeast of the 1,347-foot hill; and on the western summit of Saddleback Mountain, northeast of the village of St. Johnsbury.

Of 15 localities where early minor folds were observed in the Gile Mountain formation, 13 indicated a syncline to the east. Jahns and White (1941) report similar observation for central Vermont. White and Jahns (1950) noted that the patterns of early minor folds in central Vermont generally are sinistral when viewed in cross section from the south, indicating older rocks to the west, younger rocks to the east.

The early minor folds and the flow cleavage are undoubtedly contemporaneous with most of the major folds of the region. The Willoughby arch, however, was probably formed somewhat later (Dennis, 1956).

Late Folds. Late minor folds (Fig. 3a) were observed only in the for-

Plate 3a. Early syncline exposed on south-facing *ac* joint. Away from trough both limbs are parallel to flow cleavage. Photograph by J. F. de Chadenedes.

Plate 3b. South-facing *ac* joint plane showing trace of bedding (hammer handle), flow cleavage (pencil left of hammer, and chain), slip cleavage (pencil right of hammer, and scratch mark). Photograph by J. F. de Chadenedes.

Plate 4. Late (chevron) folds, showing axial-plane slip cleavage, exposed on *ac* joint in road cut opposite Graves School. Axial plunge, northeast. Photograph by J. F. de Chadenedes.

mations of the St. Francis group. Generally they have a dextral pattern (White and Jahns, 1950) when viewed in cross section from the south. Most of them are open folds, some so open that the axial planes would be difficult to locate were it not for the axial-plane slip cleavage, which is prominent in most of the country west of the Monroe contact (see below). The late folds plunge north and northeast at 20°-60°, but near the intrusives of Kirby Mountain the axes have been deflected to the northwest in some exposures west and south of the mountain. At a point 1,200 feet due north of B. M. 814 on the Maine Central Railroad a late fold has a vertical plunge. The fold is exposed in a small cliff, and the plunge changes within 2 feet from steep (85°) south through vertical to steep north.

The late folds appear to be of three main types: (1) flexural folds, such as the chevron folds shown on Plate 4; (2) slip folds, formed by

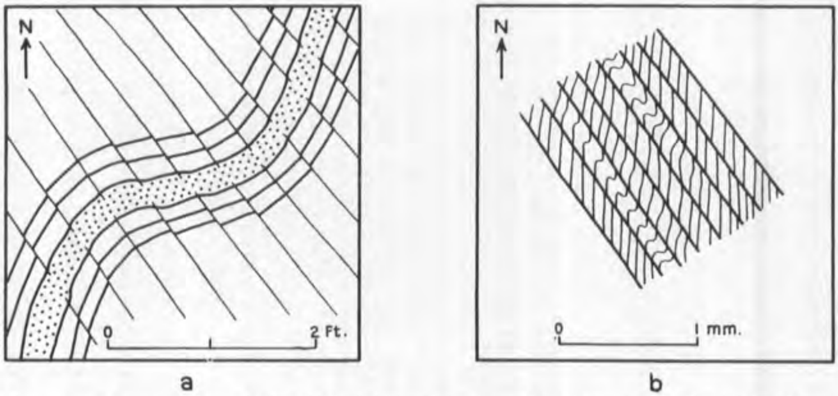


Figure 8a. Map pattern of late fold with northwestward striking axial-plane slip cleavage. Cleavage is well developed in pelites (blank), but only in margins of quartzite bed (dotted). Note slipping in pelite, slight flexures in quartzite.

Figure 8b. Younger microfolds in thin section. Orientation approximate. Heavy lines, slip cleavage; light lines, flow cleavage (sericite schistosity). Note sericite oriented parallel to both cleavages.

movement along a closely spaced system of planes (slip cleavage) parallel or approximately parallel to the axial planes of the folds; and (3) flexural-slip folds, formed by flexure of the more competent, arenaceous layers, and by slipping of the less competent, more argillaceous layers along slip-cleavage planes (Knopf and Ingerson, 1938, pp. 137-162). Displacements parallel to the slip cleavage appear in almost all late folds in the slaty units of the Gile Mountain formation and in the Meetinghouse slate, and although in some places they are too minute to be distinguished by the naked eye they can usually be observed in thin section (Fig. 8; Pl. 6.). Displacements may be as great as 5 millimeters, but are generally less than 1 millimeter. The porphyroblasts formed after the movement along the slip-cleavage planes, for garnet and biotite have grown directly across them without deformation.

Cleavage

General Statement. Two types of cleavage occur in the area, flow cleavage (schistosity) and slip cleavage. Wherever they are present in the same outcrop, the slip cleavage displaces the flow cleavage.

Flow Cleavage. Flow cleavage or schistosity is of the usual type described by Leith (1923). It consists of a dimensional orientation of muscovite and sericite, and more rarely other minerals, in approximately parallel surfaces, and is in many places the most conspicuous structure observed. It appears throughout the area, but is most common southeast of the Monroe contact. It is approximately parallel to the axial planes of the early folds. Because most of the early folds are closed folds, isoclinal or nearly so, the flow cleavage is practically parallel to the bedding (Fig. 4b) on the limbs of these folds, and cuts across the bedding near the apices. Flow cleavage and bedding are parallel, or approximately so, in most outcrops.

The strike of the flow cleavage is northeast or north, parallel to the regional trend, except near the granitic intrusives of Miles Mountain and Kirby Mountain. These igneous rocks, which were probably emplaced at about the same time as the New Hampshire plutonic series (Billings, 1956), are younger than both the flow cleavage and the slip cleavage. The regional trend, normally north to northeast, is deflected to a northwesterly direction near both plutons. This may either be due to the emplacement of the plutons, or else their emplacement may have been controlled by a pre-existing anomalous trend.

Possible crossbedding was observed on the small knob 2,000 feet west of B. M. 1260, between Kirby Brook and Ranney Brook in the town of Kirby (Plate 5). Cross-laminations are nearly tangent to the normal bedding on the west. This would seem to indicate younger rocks to the east. The fact that sericite has recrystallized parallel to and along these slightly curved surfaces might indicate the possibility that they represent flow cleavage and not crossbedding. On the other hand, recrystallization may well have been controlled by primary (bedding) foliation.

Slip Cleavage. Slip cleavage (Dale, 1914; Hadley, 1942) is confined almost exclusively to the area northwest of the Monroe contact. It resembles fracture cleavage as described by Leith (1923), except that there has been some crystallization of sericite parallel to the cleavage surfaces, and slipping has occurred along them. Individual movements along each plane of slip are slight, amounting to less than 5 millimeters, generally to less than 1 millimeter. Where these displacements cannot be seen with the naked eye, they generally can be distinguished under the microscope, particularly in the more argillaceous layers in which they are characteristic (Pl. 6). In such places it is seen that the sericite, originally parallel to the earlier flow cleavage, has been partially reoriented and recrystallized along the surfaces of slip cleavage. Planes of slip cleavage are generally 1 to 25 millimeters apart.

Plate 5. Possible cross-bedding in quartzite of Gile Mountain formation (see text). East is at right. Left pencil: slip cleavage. Center pencil: normal bedding. Right pencil: cross-laminations. Note how cross-laminations become tangent to normal bedding at left (west). Owing to extensive isoclinal folding in the area, such primary structures are of little value in determining the attitude of major structures.

Plate 6. Thin section of cleavage from Gile Mountain formation, road cut 1 mile north of U. S. Route 2, Chesterfield Valley, showing contact between slate and quartzite lens. Flow cleavage (sericite schistosity) in slate parallel to contact. Slip cleavage (2 bands) crosses picture from left to right. Quartz segregates in the micro-folds and produces intimate composition banding parallel to the cleavage.

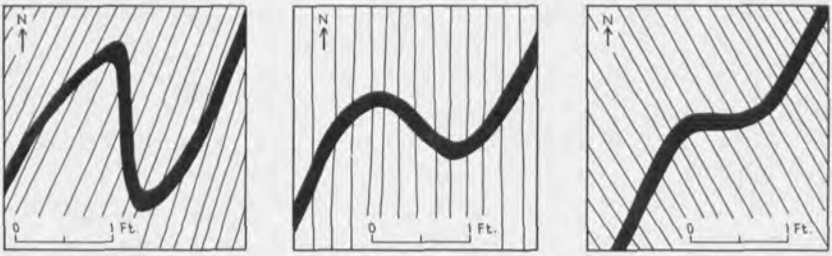


Figure 9. Map patterns of late folds. Showing dependence of shape of these folds on angle between bedding and slip cleavage in any particular area.
a. Southwestern part of area. Resembles early-type fold.
b. West-central part of area.
c. Northwestern part of area. Characteristic open flexure.

The slip cleavage is approximately parallel to the axial planes of the late folds. Nowhere in this area is it parallel to the bedding. The strike of the slip cleavage is most commonly northwest. In almost all places where both flow cleavage and slip cleavage are in the same outcrop, the flow cleavage strikes more easterly than does the slip cleavage. The attitude of the slip cleavage appears to be related to the shape of the late folds with which it is associated at any particular locality (Fig. 9). Thus, in the southwestern part of the area, many of the late folds are closed, so that in places it is impossible to distinguish them from early folds. The slip cleavage trends northeast, parallel to the axial planes of the late folds (Fig. 9a). In the west-central portion of the region the late folds are more open, and here the slip cleavage trends northward (Fig. 9b). In the northwestern part of the area the late folds are extremely open, and the slip cleavage trends northwest (Fig. 9c). There are many exceptions, but generally the late folds tend to decrease in tightness toward the north.

At many localities, particularly in the Gile Mountain formation, arenaceous and argillaceous beds alternate with one another. In these places slip cleavage is almost invariably confined to the argillaceous beds, and dies out in the more quartzose ones (Pls. 3b and 6; Fig. 8a), even where the phyllite and quartzite beds are but a few millimeters thick. Flow cleavage in the Gile Mountain formation, on the other hand, is more conspicuously megascopically in the arenaceous beds. In the argillaceous beds, the flow cleavage is obscured by the later slip cleavage, but is plainly visible under the microscope as sericite schistosity.

Schwartz (1942) mentions both the lack of cleavage in some of the more quartzose beds and the change in attitude of the cleavage as it passes from argillaceous to arenaceous to argillaceous beds. This change in attitude, or refraction of cleavage, has been observed in the Gile Mountain formation; but it is by no means a common phenomenon, and the refraction amounts only to about 10°.

Refraction of cleavage in rocks of different competence should not be confused with the appearance of different generations of cleavage in adjacent rocks. Thus, along the hillside north of U. S. Highway 2 in the town of Kirby there are several outcrops showing what at first sight appears to be refraction of cleavage (Fig. 10a). Upon closer inspection, however, it proves to be merely a series of alternating quartzite and slate beds, with flow cleavage in the quartzites and slip cleavage in the slates. A few beds of intermediate composition have both kinds of cleavage, so that these rocks split with a pencil structure.

Planar Structures of Igneous Rocks

Planar structures of igneous rocks here include primary and secondary foliation, schlieren, and oriented inclusions.

Primary foliation, or flow structure, is prominent in the darker phases of the granitoid rocks of Miles Mountain, and along the borders of the many dikes and sills. It is commonly expressed as a preferred dimensional orientation of micas and, in some places, feldspars, and is associated with schlieren and inclusions.

On the northern summit of Miles Mountain, just above B. M. 2430, all these types of planar structure are present. Here the inclusions range in size from 1 inch in diameter up to 1 foot wide and 4 feet long, but the third dimension could not be measured. The strike is N. 20° E., the dip 75° NW., parallel to the flow cleavage in the adjacent sillimanite schist of the Albee formation. Along U. S. Highway 2, just north of the west end of Miles Pond, and along the same road just above the bridge across Carr Brook, foliated granitic dikes are well exposed.

The Moulton diorite is weakly foliated within about 50 feet of its contacts. The foliation is parallel to the contacts, whether they are cross-cutting or not.

In the Highlandcroft granodiorite in Waterford, secondary foliation strikes northeast, parallel to the regional trend, and any parallelism to the contacts is coincidental.

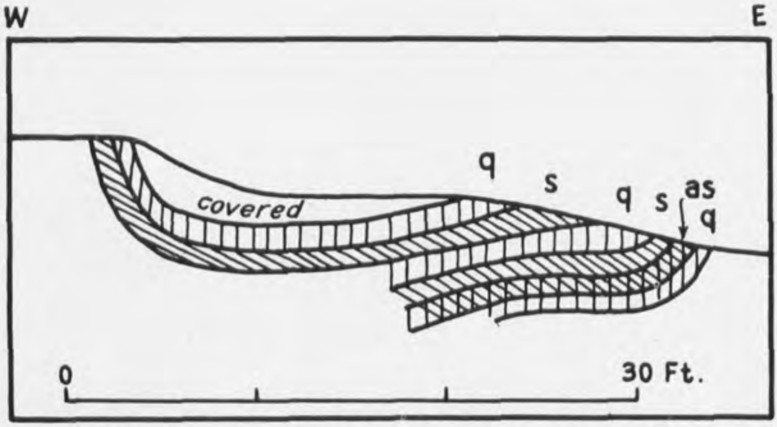


Figure 10a. Cross-section on south-facing *ac* joint showing pseudo-refraction of cleavage. Flow cleavage developed in quartzite beds (q), slip cleavage dominant in slate beds (s). Both types of cleavage developed in beds of intermediate composition, as in arenaceous slate (as).

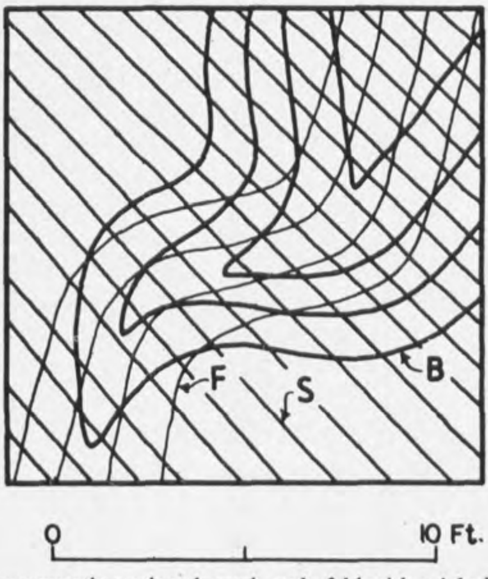


Figure 10b. Diagrammatic section through early fold with axial-plane flow cleavage. Deformed by late folding developed on axial-plane slip cleavage. F, flow cleavage; S, slip cleavage; B, bedding. Lineation due to intersection of bedding and slip cleavage is nearly parallel to that formed by intersection of flow cleavage and slip cleavage.

Lineation

General Statement. The three main types of lineation in the area are caused by the intersection of planes or surfaces. The three types are:

(1) Intersection of bedding and flow cleavage;

(2) Intersection of bedding and slip cleavage;

(3) Intersection of flow cleavage and slip cleavage.

All are *b* lineations (Cloos, 1946).

Intersection of Bedding and Flow Cleavage. Flow cleavage is approximately parallel to the axial planes of the early folds. In the rocks of the St. Francis group it strikes northeast and dips steeply, generally 70° or more, to the southeast. In the Albee formation and Ammonoosuc volcanics it strikes northeast and dips in general vertically or very steeply northeast. The plunge of the early folds is generally 40-50° in the Albee, less than 20° in the Gile Mountain, and is either to the northeast or southwest.

The lineation due to the intersection of bedding and flow cleavage is parallel to the plunge of the early folds. Because the axes of the early folds were deformed somewhat by the late folds, the early lineation has been similarly deformed.

Intersection of bedding and slip cleavage. Lineation caused by the intersection of bedding and slip cleavages is confined almost exclusively to rocks of the St. Francis group. The slip cleavage is parallel to the axial planes of the late folds. It strikes north or northwest and generally dips 40-65° NE. The plunge of the late folds in general is steeper than that of the early folds, usually 20-60° NE. Where no early folds exist, the lineation caused by the intersection of bedding and slip cleavage is parallel to the plunge of the late folds. However, in outcrops where bedding, flow cleavage, and slip cleavage are observed in early folds (Fig. 7), the slip cleavage is always later than the other features, and cuts across them. In the St. Francis group early folds are not common; consequently the attitude of this type of lineation is much more regular than it would be if early folds were more abundant.

Intersection of Flow Cleavage and Slip Cleavage. The flow cleavage is earlier than the late folds; consequently it has been deformed by the movements that caused the late folds. Inasmuch as most of the early folds are nearly isoclinal, bedding and flow cleavage are nearly parallel except along the apices of folds. Therefore, either folded bedding or folded flow cleavage, or both, may be present in the late folds, and the lineations caused by the intersections of either bedding or flow cleavage with the late slip cleavage are very nearly parallel to one another.

The small cremulations and crinkles commonly observed in many of the rocks, particularly in the slaty beds of the Gile Mountain formation and in the Meetinghouse slate are caused entirely by one or more of these intersections. This is because the sericite is parallel to both flow cleavage and slip cleavage, and is curved or slightly folded where the two cleavages intersect. This curvature of the sericite, which is readily seen under the microscope, is expressed megascopically as small crinkles.

Other lineations result where bedding, flow cleavage, or slip cleavage intersect joint planes.

In some of the igneous bodies crude primary lineation is caused by the parallel orientation of inclusions. Lineation caused by the dimensional orientation of minerals was not observed.

Joints

Of the many sets of joints in the area, *ac* joints are perhaps the most conspicuous. These joints are normal to the lineation due to the intersection of bedding and flow cleavage, which in turn is parallel to the axes of the early folds (where this lineation is not deformed by late folds). Hence they strike west-northwest, and dip steeply. They are later than the period of deformation and metamorphism, for they cut across flow cleavage, slip cleavage, and biotite and garnet porphyroblasts.

Structure of Igneous Rocks

General Statement. The intrusive rocks of the area are: (1) three bodies of Highlandcroft granodiorite; (2) two bodies of Moulton diorite; (3) three small bodies of Kirby quartz monzonite; (4) mafic dikes and sills in the Albee formation; and (5) granitoid dikes and sills in the Albee formation.

Highlandcroft granodiorite. The Highlandcroft granodiorite plutonic series is pre-Silurian (Billings, 1956, p. 106). Consequently, bodies of this rock were affected by the Acadian orogeny at least to the extent that their present form is probably not primary. The main body in this area is approximately half a mile wide and 3.5 miles long. It trends N. 40° E., approximately parallel to the regional trend. The most southwesterly (New Hampshire) portion of this body was mapped by Billings (1937).

The contact between the Highlandcroft pluton and the adjacent Ammonoosuc volcanics is not seen at most places, because of the scarcity of critical exposures and because the two rock types are lithologically remarkably similar in many respects. The trace of the contact between the pluton and the Albee formation is, however, easily recognized. The attitude of the contacts is not known and cannot be inferred.

At some localities a secondary foliation was observed in the Highlandcroft rocks. This foliation is parallel to the schistosity (flow cleavage) in adjacent formations. Even at the extreme northeast end of the pluton, where the contact trends northwest for a short distance, the foliation strikes northeast directly across the contact. This foliation is therefore secondary.

Two much smaller bodies belonging to the Highlandcroft plutonic series were found northeast of the main body. Exposures were seen only in the beds of two small brooks, so that the shape and size of these bodies is unknown.

Moulton Diorite. The main body of this metadiorite is 5 miles long by less than three-quarters of a mile wide, and trends N. 30° E. On the northwest it is bounded by the Meetinghouse slate. The southeast boundary is the Monroe contact.

A crude primary foliation in the marginal regions of the main pluton dies out within 50 feet of the contact. The strike of the foliation is parallel to the contact with the Meetinghouse slate. For example, just southwest of B. M. 816 on the Barnet-Waterford town line the foliation strikes nearly N. 90° E., and is vertical (implying that the contact is vertical), although the bedding and flow cleavage in the adjacent Meetinghouse slate strike N. 45° E. and dip 55° SE. This attitude is in marked contrast to that of the secondary foliation in the Highlandcroft pluton.

A much smaller body of the same rock is exposed on a steep hill in the southeastern part of the St. Johnsbury quadrangle, just west of the main body. The contacts of the northeast and southwest ends cut across bedding and flow cleavage in the Meetinghouse slate; elsewhere the contacts are parallel to bedding and flow cleavage.

Kirby Quartz Monzonite. Three small bodies of Kirby quartz monzonite are exposed in the Concord-Waterford area. Two of them are located on the northern boundary of the area, on the south slope of Kirby Mountain.

Each of the three bodies exposed in the area is less than a mile long and approximately 1,000 feet wide. A few northwest strikes were ob-

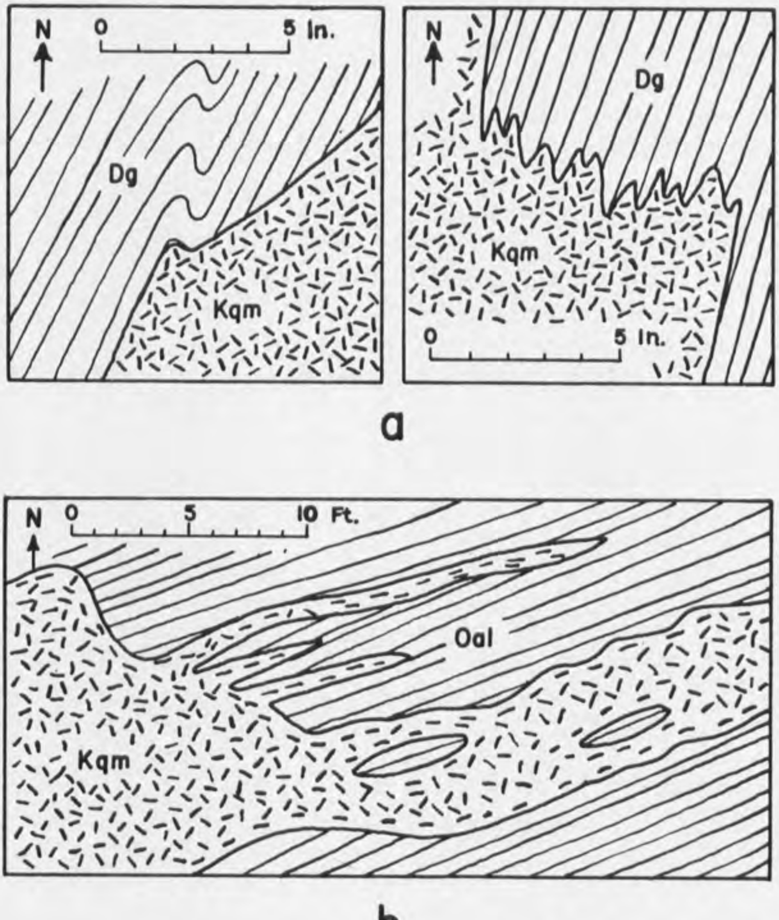


Figure 11. Map diagrams showing local cross-cutting relations of granitoid rocks—(a) near Stiles Pond, and (b) on Miles Mountain. Note that intrusive is foliated near contacts only. Dg, Gile Mountain formation; kqm, Kirby quartz monzonite; Oal, Albee formation.

served in the Gile Mountain formation southwest of the northeastern body. The other two bodies trend northward and northeastward, parallel to the regional trend.

In general the intrusive bodies are concordant, but local cross-cutting (Fig. 11a) was observed at several localities. At the southern end of the southernmost body, near Stiles Pond, the contact swings around in a nearly perfect arc, but the strike of bedding and flow cleavage in the adjacent Meetinghouse slate is northeast wherever observed.

The Kirby quartz monzonite is a massive rock, having little if any flow structure. It probably belongs to a late phase of the New Hampshire plutonic series (Billings, 1956), and is tentatively correlated with the Willoughby granite (Dennis, 1956).

Metadiabase Dikes and Sills. The rocks just southeast of the Monroe contact have been injected by dikes and sills of intermediate to mafic composition. Many varieties of different texture but approximately the same mineralogy were observed; no attempt was made in the field to distinguish between them. Most of the dikes and sills range in thickness from 5 to 20 feet.

Granitoid Dikes and Sills. In the vicinity of Miles Mountain the Albee formation has been injected by many granitoid dikes and sills ranging in composition from tonalite to granite. In that area perhaps 50 percent of all observed exposures are of dike rock, the other 50 percent being Albee. These intrusives range in thickness from about an inch to slightly more than 100 feet. A few of them are dikes; most of them, however, are essentially sills, although they may locally be cross-cutting (Fig. 11b). The more mafic phases contain primary planar structures and are believed to be earlier than the massive, felsic phases, which are correlated with the Kirby quartz monzonite. All these intrusive rocks are probably members of the New Hampshire plutonic series (Billings, 1956).

METAMORPHISM

Zoning

Chlorite Zone. The chlorite zone occupies a belt roughly parallel to the Connecticut River. The river is approximately in the center of this zone, and from it the metamorphism increases both toward the southeast (Billings, 1937; Hadley, 1942) and toward the northwest (Dennis, 1956). Throughout most of the region where the Connecticut separates New Hampshire from Vermont, the river is approximately parallel to the regional structure (Plate 2); in the Concord-Waterford area, however, it flows across the structure. It is suggested that the mineral assemblage of the chlorite zone is more readily eroded than the assemblage of the higher grade zones, and that the Connecticut River, in this area, is a subsequent stream; subsequent, however, not upon the structure but upon the metamorphism. The situation is perhaps analogous to that in the Mt. Washington area, New Hampshire (Billings, 1941, p. 924), where the altitude of the mountains is directly controlled by the metamorphism.

Minerals that are common in pelites of the chlorite zone are chlorite, quartz, albite, epidote, calcite, and sericite. Brown biotite, garnet, staurolite, and sillimanite are characteristically absent.

Biotite Zone. The biotite isograd marks the upper limits of the chlorite zone. Beyond it brown biotite can be observed in most of the rocks, cutting across the planes of cleavage. Near the biotite isograd the biotite is in small flakes, barely visible to the naked eye, but toward the north and west flakes 1 or 2 millimeters in diameter are common. In the pre-Silurian rocks, chlorite is present in the biotite zone near the biotite isograd, but near the garnet isograd it is very rare. The plagioclase is albite-oligoclase, epidote disappears, but quartz, calcite, and sericite remain. Hornblende appears in metamorphosed mafic rocks. The texture of the rocks is slightly coarser than in the chlorite zone.

Garnet Zone. Beyond the garnet isograd the texture of the rocks is essentially the same, except that porphyroblasts of almandite are present, and the plagioclase is usually oligoclase. Quartz, calcite, sericite, biotite, and hornblende are present. The garnets cut across the planes of cleavage.

Staurolite Zone. Staurolite appears in most of the outcrops between Kirby Mountain and Miles Mountain. In some places the crystals are twinned. The porphyroblasts are up to 5 millimeters long and show no suggestion of alteration. East of Miles Mountain, along Carr Brook, poikilitic porphyroblasts of staurolite up to 4 centimeters long and 5 millimeters wide are present. These fairly large crystals enclose muscovite and quartz. In general they are randomly oriented, but a few have their long axes parallel to the planes of flow cleavage. Flakes of muscovite can be distinguished megascopically within a few hundred feet of the Kirby quartz monzonite. Besides staurolite, the common minerals in pelites are quartz, oligoclase, garnet, biotite, sericite, and in places, calcite. Hornblende is characteristic of mafic rocks in this zone. The phyllites in the St. Francis group are very much more crinkled within the staurolite zone than they are in the lower grade zones.

Sillimanite Zone. The sillimanite isograd surrounds Miles Mountain on at least three sides, and within it the rocks are by definition high grade (Billings, 1937). Most of the sillimanite porphyroblasts are about 5 millimeters long, but they range from microscopic size to 7 centimeters. These white to gray, acicular or columnar crystals commonly have random orientation. In a few places, however, their long axes lie within the planes of flow cleavage, but not parallel to one another; nowhere are their long axes preferentially oriented parallel to any direction. The porphyroblasts grew after deformation had ceased.

All the sillimanite crystals are fresh; no evidence of alteration was observed. Other minerals of the sillimanite zone are quartz, oligoclase-andesine, garnet, biotite, muscovite (in pelites), and hornblende (in mafic rocks). Calcite and staurolite were not observed. The texture is coarser than in the staurolite zone. Bedding is obscure, and in many places is absent, in contrast to the lower grade zones where bedding is prominent at numerous localities.

The general appearance of the Albee formation in the sillimanite zone is strikingly different from that in the other zones. On the other hand, the sillimanite-zone greenstones in the Albee look very similar to those in the chlorite zone. The principal difference is in the composition of the plagioclase, which is oligoclase-andesine in the sillimanite zone and albite or albite-oligoclase in the other zones.

Other Metamorphic Minerals

Ottrelite. Small porphyroblasts of ottrelite up to 2 millimeters in length were observed in some of the rocks of the garnet zone northwest of the Monroe contact, particularly in the phyllite of the St. Francis group. One accessible locality is along U. S. Highway 2 in the town of Kirby, but the mineral was observed in many outcrops in both Kirby and St. Johnsbury. It occurs associated with porphyroblasts of both biotite and garnet, but not staurolite. The occurrence appears to be identical with that reported by Hadley (1942) from the Mt. Cube quadrangle, and Dennis (1956) from the Lyndonville quadrangle.

Actinolite. Pale green needles and rosettes of actinolite are found in some of the impure marble in the extreme northwestern part of the area. This actinolite may be the result of the reaction of dolomite with quartz. A fibrous actinolite was observed in some of the thin sections of the Moulton diorite in the biotite zone not far from the biotite isograd.

Hornblende. Blue-green hornblende is present in most of the mafic intrusive and volcanic rocks of the area. Billings (1937) states that this hornblende has not been observed in the chlorite zone in the Littleton-

Moosilauke area, New Hampshire. He believes that the mineral is not stable in that zone. Dodge (1942) has observed a similar blue-green hornblende in the Black Hills of South Dakota, where apparently it is only in the garnet and staurolite zones. In the Concord-Waterford area, however, blue-green hornblende first appears in the uppermost part of the chlorite zone, and is abundant in the biotite, garnet, staurolite, and in part of the sillimanite zones.

Study of thin sections indicates the following genetic sequence for hornblende: yellowish green ⟶ olive green ⟶ bright green ⟶ blue green. This sequence seems to be in agreement with Dodge's data. But the presence of blue-green hornblende in the chlorite zone is apparently not in such agreement. Dodge states that small crystals are more strongly colored than large ones, which are blue green only in the staurolite zone. In the Concord-Waterford area, however, large hornblende crystals, up to 3 centimeters long, are blue green in the biotite zone, and small ones occur in the upper part of the chlorite zone. Porphyroblasts of this mineral are present in the greenstones of the sillimanite zone on the lower east slopes of Miles Mountain.

The evidence seems to indicate that, whereas the blue-green hornblende is more stable in higher grade zones than are the older, paler varieties, it nevertheless has a somewhat wider range of stability than had been believed. Because of the absence of rocks of suitable lithology at certain critical localities, the blue-green hornblende isograd could not be accurately drawn; probably it coincides very nearly, but not precisely, with the biotite isograd.

Time of Metamorphism

It is clear that the metamorphism is closely related in time to both the orogeny and to some of the igneous intrusions. The two types of folds were produced at different times, and each kind had its own associated type of cleavage. The flow cleavage is contemporaneous with the early folds, and the slip cleavage is contemporaneous with the late folds. The formation of the flow cleavage (schistosity) involved recrystallization, hence the corresponding metamorphism is contemporaneous with the early deformation.

The Moulton diorite and the diabase dikes and sills in the Albee formation were emplaced between the two stages of folding. An insignificant amount of contact metamorphism, with growth of small green hornblende crystals, took place at that time. These porphyroblasts of hornblende are up to 2 millimeters long, and are present only within 10 feet of the diorite.

The main period of porphyroblast growth came later, with the intrusion of granitoid dikes and sills in the Miles Mountain area. The porphyroblasts cut across all generations of cleavage and are clearly later. These porphyroblasts are, however, indisputably related geographically to the granitoid dikes and sills, particularly in the higher grade zones of metamorphism. Hence the later metamorphism is contemporaneous with the emplacement of the New Hampshire plutonic series. The main period of metamorphism was, therefore, contemporaneous with the main period of intrusion, and later than the main period of deformation (see Table 3).

Discussion

Metamorphism in this area consists essentially of a higher grade mineral assemblage (porphyroblasts) superimposed on a lower grade mineral assemblage (pre-slip-cleavage assemblage). In the Concord-Waterford area the spatial relation of the porphyroblast assemblages to rocks of the New Hampshire plutonic series is brought out by the pattern of the isograds on Plate 1. The main characteristics of the two phases of metamorphism are as follows:

1. Early phase (pre-slip-cleavage). In this phase sericite and chlorite crystallized parallel to the direction of flow cleavage, imparting a true schistosity to the rocks. The Devonian and possibly Silurian rocks have little or no surviving chlorite associated with this phase. In the pre-Silurian rocks immediately east of the Monroe contact chlorite is the characteristic mineral. This chlorite is mostly replaced by biotite wherever the pre-Silurian rocks enter the biotite zone, though it may survive in equilibrium with biotite and give some rocks of the biotite zone a characteristic greenish tinge.
2. Late phase (post-slip-cleavage). Here poorly oriented porphyroblasts are superimposed on the groundmass inherited from the first metamorphism.

Most of the mineral assemblages seen in rocks of the area are equilibrium assemblages. Minerals of the early phase of metamorphism survive only if they are also in equilibrium in the later phase. In particular, sericite did not recrystallize readily in the higher grades, but normally retained its original texture; the failure of sericite to recrystallize to muscovite may in part be due to lack of available water (Dennis, 1956, p.

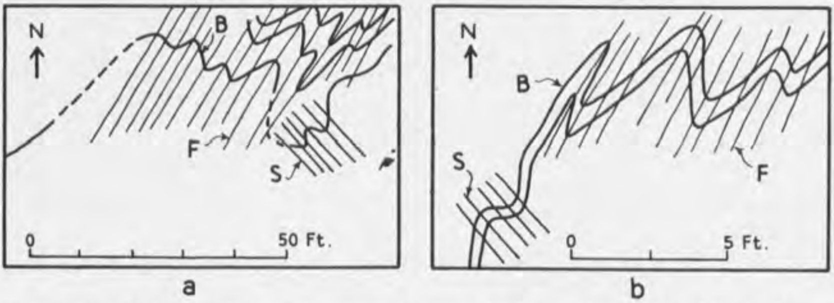


Figure 12. Diagrammatic map pattern of minor folds. Southwest of B. M. 938, Chesterfield Valley. B, bedding; F, flow cleavage (parallel to axial planes of early folds); S, slip cleavage (parallel to axial planes of late folds). Early folds plunge southwest. Late folds plunge northeast.

82) ; but this author no longer believes that lack of water is necessarily the principal cause of "texture inertia." The porphyroblasts form readily enough. It is certainly no coincidence that the surviving minerals are those that are chemically stable in the new assemblage; these minerals are, in particular, sericite (in all observed metamorphic zones of the area, excepting the sillimanite zone), and chlorite (in the biotite zone). The survival of chlorite is rather well exemplified in some of the volcanics near Bible Hill, which had been metamorphosed to chlorite schist in the early phase, and in which poorly oriented actinolite porphyroblasts of the late phase are superimposed on the original schistosity.

TECTONIC SYNTHESIS AND GEOLOGIC HISTORY

Dynamic Interpretation of Structures

In the St. Johnsbury homocline the apices of late minor folds face east and down, away from the crest of the Willoughby arch (Fig. 1). The dextral pattern (White and Jahns, 1950), viewed in cross-section from the south, indicates that the rocks on the west have moved up relative to the rocks to the east, and the late folds are best interpreted as "reversed" drag folds associated with the formation of the arch.

The arch-like pattern of the late cleavage (White and Jahns, 1950) is not interpreted as deformation of a pre-existing horizontal structure, but as an expression of flowage associated with the doming (Dennis, 1956, Fig. 4).

The isoclinal early folds and associated cleavage, therefore, would have formed before the rise of the arches and domes of eastern Vermont. The divergence between the two generations of folding in this area is brought out in detail in Figure 12, as well as in the map pattern of the cleavages (Pl. 2). See also Figure 8 in White and Billings, 1951.

As early minor folds are comparatively rarely observed in outcrop, they are not reliable guides to the major folding, which has to be reconstructed from stratigraphic evidence. A key problem in the evaluation of the regional structure is the significance of the Monroe contact.

The Monroe Problem

Eric, White and Hadley (1941, p. 1900) discovered a break between the "New Hampshire sequence" and the "Vermont sequence" of Paleozoic rocks in western New England. They cited evidence, later amplified by Eric (1942), that this break was a fault and named it the Monroe fault. Later work by Kruger (1946), Lyons (1955, p. 133), and J. B. Thompson (personal communication, 1956) showed that, along certain sections of this break south of the type area the evidence seemed to favor the absence of a fault. It was therefore originally decided to adopt a non-committal terminology in this report, referring to the break as the Monroe contact. However, since no decisive evidence has come to light against the fault hypothesis in the area here discussed, the break is referred to as the Monroe fault, as originally named by Eric et al. (op. cit.). In a more regional context, the break is non-committally referred to as the Monroe contact.

Evidence Favoring Faulting

Stratigraphic Evidence. Throughout the entire distance of 65 miles between Granby and Norwich (Fig. 2), the same formation (the Meetinghouse slate) is present on the northwest side of the contact, except for about 5 miles of intrusive rock. Different formations — the Orfordville and Albee formations and the Ammonoosuc volcanics — crop out on the southeast side at different places. The attitude of the bedding and flow cleavage in adjacent formations is at most places essentially parallel to that of the Monroe contact, and it is only by regional mapping that the truncation of formations on the southeast side of the contact becomes apparent. Such truncation of formations may be due either to an unconformity or to a fault; changes in lithology along the strike may also cause apparent truncation.

In the Mt. Cube quadrangle, near the village of Bradford, Vermont, Hadley's map (Hadley, 1942) shows that the Albee formation is cut out by the Monroe contact and is replaced farther south by the Orfordville formation. Moreover, several different members of the Orfordville are successively truncated along the southeast side of the contact.

Truncation of Plutons. In the towns of Waterford and Barnet, Vermont, there is an elongate body of Moulton diorite. This body is approximately 5 miles long and less than 1 mile wide. It is bounded on the northwest by the Meetinghouse slate, which it intrudes; and on the southeast by the Monroe contact. Either the intrusive was emplaced prior to faulting and was in turn faulted into its present position; or it was emplaced along a pre-existing plane — either a fault or an unconformity. Now the metadiorite has cut across the structure of the Meetinghouse slate and has caused contact metamorphism in these rocks, resulting in a metamorphic aureole 5 to 10 feet wide containing small porphyroblasts of hornblende. No such contact metamorphism was observed in the Albee formation to the southeast; this circumstance would seem to favor the first hypothesis.

Furthermore, the northwestern marginal regions of the intrusive, comprising approximately the outer 50 feet, are crudely foliated wherever exposed, and the foliation is parallel to the contacts with the Meetinghouse slate, even where these contacts make a large angle with the bedding and flow cleavage. On the other side of the pluton, several exposures of metadiorite are believed to be within 50 feet of the Monroe contact, and at none of these localities was foliation observed.

Truncation of Structure in Detail. Truncation of structure in detail can be seen approximately 1½ miles northeast of the village of North Concord, Vermont. Northwest of the contact (Pl. 1), on the two small hills between Dudley Brook and Moose River, near North Concord School, bedding and flow cleavage in the Meetinghouse slate strike northeast, approximately parallel to the Monroe contact; but southeast of the contact, on the 1900-foot hill 1 mile west of the summit of Miles Mountain, bedding and flow cleavage in the Albee formation strike northwest, at right angles to the contact. These two localities on opposite sides of the contact are almost 2,000 feet apart. Truncation of structure in minute detail was observed three-quarters of a mile north of the village of Monroe, New Hampshire, where bedding and flow cleavage in both the Meetinghouse and the Albee strike at small angles toward a breccia occupying the contact plane.

Displacement of the Biotite Isograd. In the vicinity of Boltonville, Vermont (Fig. 2), in the northern part of the Woodsville quadrangle (White and Billings, 1951), the biotite isograd is offset approximately 2 miles horizontally along the Meetinghouse slate-Albee contact. In the Littleton and St. Johnsbury quadrangles the biotite isograd does not cross the Monroe contact, but no displacement was observed of the three isograds which do cross it.

Breccia. Three-quarters of a mile north of the village of Monroe, New Hampshire (St. Johnsbury quadrangle), there is a thin breccia with a gray cherty matrix. The breccia was observed at two localities within a few hundred feet of one another and in one place had a width of 6 to 8 inches, in the other 3 inches. Northwest of the breccia is the Meetinghouse slate; southeast of it the Albee formation. Field observations indicate there can be no doubt that this breccia occupies the Monroe contact. The flow cleavage in both formations strikes toward the breccia and is cut off by it; there is no flow cleavage in the breccia, except within the fragments of slate in the breccia which are 1 or 2 inches in diameter. The formation of the breccia was therefore a late phenomenon, later than the formation of flow cleavage.

Possible Arguments Against Faulting

The following points may be raised against the faulting hypothesis: the stratigraphic evidence is equally applicable to an unconformity of Devonian or Silurian over Ordovician. The "truncation" of the Moulton diorite pluton is inferred from observations at one small locality (see above); moreover, contact aureoles are notoriously erratic, especially where different lithologies are involved. The truncation of structure in detail could be due to an unconformity and emphasized by the severe deformation near Miles Mountain. Isograds are not very precise boundaries at best. They are subject to individual interpretation, and the appearance of biotite is appreciably influenced by rock composition. The rocks southeast of the Monroe contact do not have the same composition as those to the northwest. The displacement of the biotite isograd across the contact is not, therefore, a valid criterion for movement along it. The breccia is a weak criterion, by itself, for a fault of large displacement.

Evidence Favoring an Unconformity

Evidence for an unconformity has been accumulating since 1946. At certain localities Lyons (1955) and Kruger (1946) have shown that there is an unbroken succession across the Monroe contact. This circumstance agrees with the hypothesis that the Monroe contact is the Taconic unconformity between the Ordovician "New Hampshire sequence" and the Devonian (and possibly partly Silurian) St. Francis group.

Workers in the area (e.g. White, 1946, and W. M. Cady, personal communication, 1955) have for some time noted the similarity between pre-Silurian rocks of the Connecticut River valley and those west of the Shaw Mountain formation of central Vermont. In particular, there is a striking resemblance between the Cram Hill formation (and its correlative, the Moretown formation) on the one hand, and the Albee-Ammonoosuc succession on the other. And White (1946, p. 132) notes the resemblance between the Northfield slate and the lower part of the Littleton formation, particularly by the presence of soda rhyolite tuff in the lower part. Now the Northfield slate is very similar lithologically to the Meetinghouse slate, and the latter, as shown in another section of this Bulletin, also contains lenses of rhyolitic tuff. This circumstance would seem to favor the relations shown in Fig. 14b.

Discussion

There are difficulties inherent in both the fault and the unconformity hypotheses for the Monroe contact. A fault along the Monroe contact would have to be a thrust. Throughout its 85 miles strike length (Billings, 1956, p. 117) its attitude is parallel to that of the adjacent Meetinghouse slate, a circumstance that would be extremely unusual for a normal fault. With rare exceptions (Billings, 1933) older rocks overlie younger rocks along thrust planes. This would likely be true of a thrust fault with a strike length of the order of magnitude of that of the Monroe contact. Thus, structurally, the pre-Silurian rocks would have to overlie the Meetinghouse. But, in the case of an overthrust, the hanging-wall formation is more likely to remain continuous, and the footwall is more likely to be truncated. This is the opposite of conditions at the Monroe contact, if it represents a thrust.

If, then, the Monroe contact is an unconformity, the tops of beds should face west across it. This would bring the west limb of the Gardner Mountain anticline (Billings, 1937) to the Monroe contact, eliminating the Monroe syncline (Eric, 1942). Field relations of the Ammonoosuc

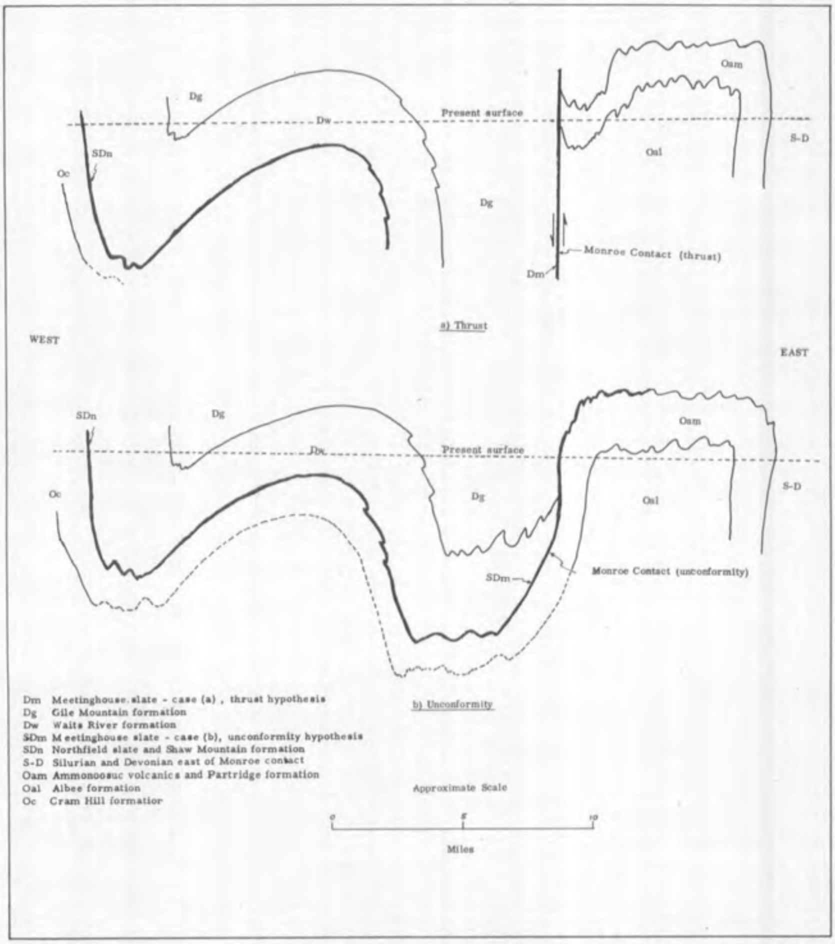
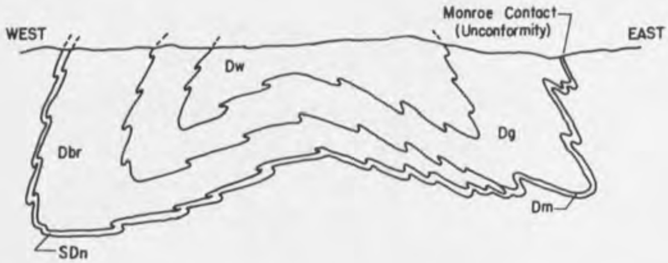


Figure 13. Diagrammatic cross-sections, illustrating the two main hypotheses for the structural significance of the Monroe contact.

volcanics here make this very unlikely (M. P. Billings, personal communication, 1958).

The difficulties are greater west of the Monroe contact. The Meetinghouse slate would have to be the correlative of the Northfield slate as stated above; that this might be so has already been suggested on several occasions (e.g., W. M. Cady, personal communication, 1956; Billings, 1956, p. 97-98). There can be little doubt that the Waits River to the



Dw Waits River formation
Dg Gile Mountain formation
Dbr Barton River formation
SDm Meetinghouse slate
SDn Northfield slate

Figure 14. Diagrammatic cross-section, illustrating another interpretation of the regional structure. Modified from Murthy (1957, pl. 3).

west of the area is at the core of an arch (Dennis, 1956). If, as seems likely, this arch is a true anticline, there would have to be a syncline (see Fig. 13) between the arch and the Monroe contact, occupying the position of the St. Johnsbury homocline. The problem is: what has happened to the Waits River, which should appear in the east limb of the postulated syncline, west of the Meetinghouse? Various explanations might be suggested. Facies change; differential squeezing out of the more mobile marble of the Waits River formation. None of these possibilities are attractive, but none of them introduce greater difficulties than the fault hypothesis.

An interesting variant of the unconformity hypothesis has been presented by Murthy (1957). This author believes that the western band of Gile Mountain, which crops out on the west flank of the Strafford-Willoughby arch (Fig. 2), is in a homocline. He bases his opinion on observations in the East Barre area. Since it is now established that rocks become progressively older going west from there (Currier and Jahns, 1941; Cady, 1956), the Waits River east of the western band of Gile Mountain would have to be *younger* than the Gile Mountain. Hence Murthy separates the Barton River member of the old Waits River formation into an older western formation, the Barton River, and a younger eastern formation, and Waits River; the latter would here be deformed into an arch (the Willoughby arch) flanked by two synclines; stratigraphically, this younger Waits River would then be the core of a syncline (see Fig. 14).

TIME	SEDIMENTATION	IGNEOUS ACTIVITY	METAMORPHISM	DEFORMATION	
POST-MIDDLE DEVONIAN		Scattered basaltic dikes of White Mountain Plutonic Series		Fracturing	
MIDDLE DEVONIAN		Main phase of plutonic igneous activity	Medium- and high-grade metamorphism	Rise of Willoughby arch; late minor folds and cleavage.	Acadian Orogeny
			Low-grade metamorphism	Early folds and cleavage.	
SILURIAN AND EARLY DEVONIAN	St. Francis group and Littleton fm. deposited	Intermittent volcanic activity (volcanics of Bible Hill and Slate Quarry). Intrusion of intermediate dikes & sills mainly in pre-Silurian rocks; Moulton diorite		Continued subsidence	
LATE ORDOVICIAN TO EARLY SILURIAN	Unconformity	Emplacement of Highland-croft pluton		Taconic orogeny, only weakly reflected in Area.	
EARLY AND MIDDLE ORDOVICIAN	Albee fm deposited, grading upward, with increasing volanic activity into	Ammonoosuc volcanics		Eugeosynclinal subsidence in Magog belt.	

Table 3. Geologic History of the Concord-Waterford and adjacent areas.

The authors are convinced that this interpretation is untenable in the Concord-Waterford and Lyndonville areas. From Danville to Sutton the Brownington syncline (Doll, 1951; Dennis, 1956) opens out quite remarkably, with near vertical dips in the west and westerly dips of around 30° in the east. This is not easily interpreted as a homocline. Reconnaissance work in 1956 by Dennis and detailed mapping in the Burke quadrangle by B. Woodland (personal communication, 1957) showed that the Gile Mountain is indeed closing over the Waits River in the neighborhood of Brighton township (Island Pond quadrangle), where the arch plunges north. Furthermore, the Gile Mountain band of the East Barre quadrangle appears to trend into the syncline west of the Chester dome (J. B. Thompson, personal communication, 1957).

Murthy's main evidence for a homocline in the Gile Mountain formation in the East Barre area is the different proportion of marble on the two sides of the band. However, differential tectonic thinning of the marbles, demonstrated elsewhere (Dennis, 1956, p. 36), might account for this.

Another interpretation might suggest itself: the Brownington syncline could be the core of a recumbent anticline (tauchfalte), whose root zone is the St. Johnsbury homocline and whose axial plane is draped over the Willoughby arch. Recumbent folds are known to exist farther south along the arch (White and Jahns, 1950; Thompson, in Billings, *et al*, 1952). However, in the Glover-Hardwick-Danville area, rocks become uniformly older going west from the western band of Gile Mountain; there is no evidence here of a nappe front of Waits River embedded in "autochthonous" Waits River. Furthermore, the north-plunging structures nowhere bring out this situation in map pattern; yet in the more southerly areas referred to above, the recumbent folds are strikingly recognizable in map pattern.

Geologic History

The geologic history of the area, as interpreted by the authors, is set out in Table 3.

ECONOMIC GEOLOGY

Granite

The following information is based on Dale (1923, p. 110-123): Three quarries have been worked on Kirby Mountain at various times prior to 1916. The granite has been used mainly for rough and cut monuments. The granite at the quarries varies from biotite quartz monzonite to biotite granite with a fine, even-grained texture. Feldspars have a grain size of up to 0.25 in., micas up to 0.1 in.

Copper

Green stains on the ridge southeast of Concord Corner attracted prospectors over 100 years ago. In the 1860's the Essex Mining Company was incorporated in New York City. It is not known whether this company ever found ore of commercial grade, and several later attempts to find and mine ore at that location (shown on Pl. 1) appear to have failed. The site is marked by a number of small pits and waste dumps. The sulfides are mainly pyrite, bornite, and chalcopyrite, the latter producing green malachite stains upon weathering.

The following is an extract from a contemporary report by mining engineers of the Essex Mining Company, New York, quoted by Woodbury and others (no date), p. 970-971: "The Essex Mine is situated in the town of Concord and is commonly known as the Moulton and Darling farms. The rocks of this district are of a highly cupriferous character, consisting of the talcose schists. There are also exposed to view parallel bands of quartzite, though the schist seems to be the characteristic rock of the district, and belongs to the lower Silurian system. Their position is as near vertical as possible, and they contain the copper-bearing veins of the mine: these veins are composed chiefly of iron pyrites, quartz and feldspar, and are richly charged with the yellow sulphurate of copper. They are conformable with the stratification, and take a course of N. 55° E. by S. 55° W. The upturned edges of the strata which are abundantly exposed on the property, exhibit incrustations of gossan throughout." As far as is known, the mine never was of economic value.

Slate

"Waterford Slate" (Richardson, 1906) was quarried on the southeast slope of Fairbanks Mountain, at the locality marked on Plate 1. This operation ceased about the time of the Civil War.

Gravel

Several gravel deposits were left as a result of the Pleistocene glaciation. Deposits along Chandler Brook are being exploited by the Caledonia Gravel and Sand Company.

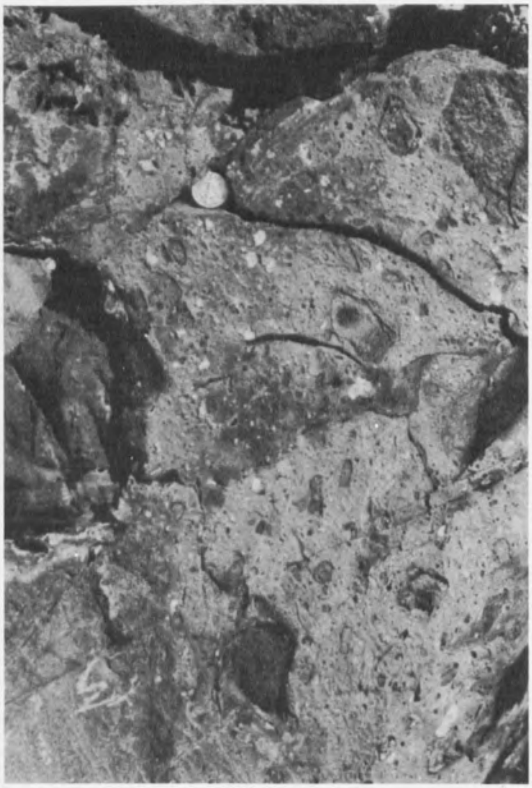


Plate 7. Volcanic agglomerate lens in Meetinghouse slate. Vertical face of Waterford slate quarry (abandoned), looking northwest. Bedding near vertical. Matrix: rhyolitic. Fragments: rhyolitic and sedimentary, the latter mainly quartz schist; also one gneiss fragment (not shown).

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UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

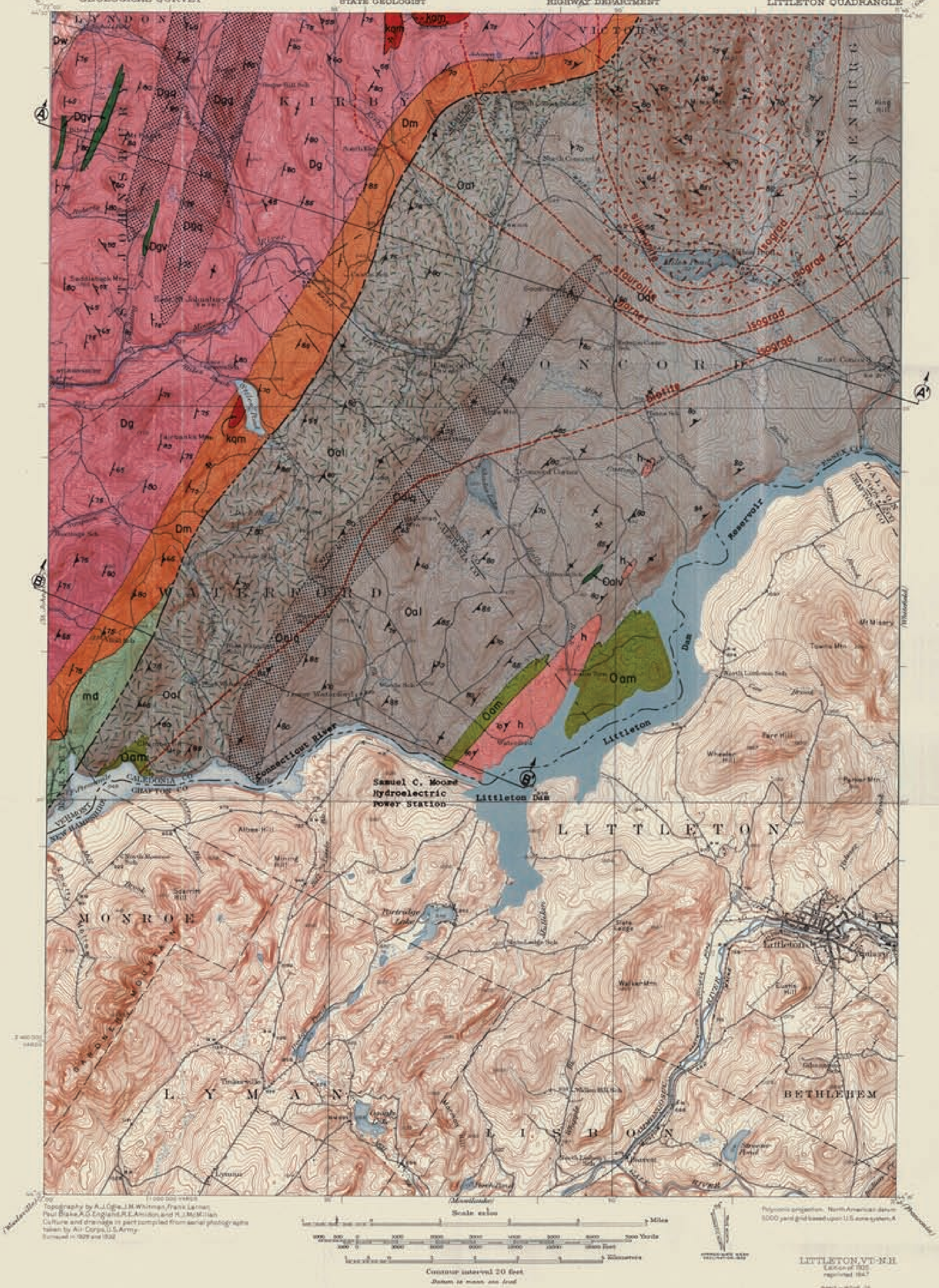
STATE OF VERMONT
REPRESENTED BY THE
STATE GEOLOGIST

GEOLOGIC MAP

STATE OF NEW HAMPSHIRE
HIGHWAY DEPARTMENT

PLATE I

VERMONT - NEW HAMPSHIRE
LITTLETON QUADRANGLE



LEGEND

DEVONIAN — New Hampshire plutonic series

- kqm — Kirby quartz monzonite
- Granitoid dikes and sills abundant
- Metadiabase dikes and sills abundant
- md — Moulton diorite; Gray-green metadiorite

SILURIAN (?) and DEVONIAN — St. Francis Group

- Dm — Meetinghouse slate
- Dg — Gile Mountain formation: Gray slate, phyllite, and quartzite; Dgq: quartzite abundant; Dgv: volcanics
- Dw — Waits River formation

UNCONFORMITY

U. ORDOVICIAN(?) — Highlandcroft plutonic series

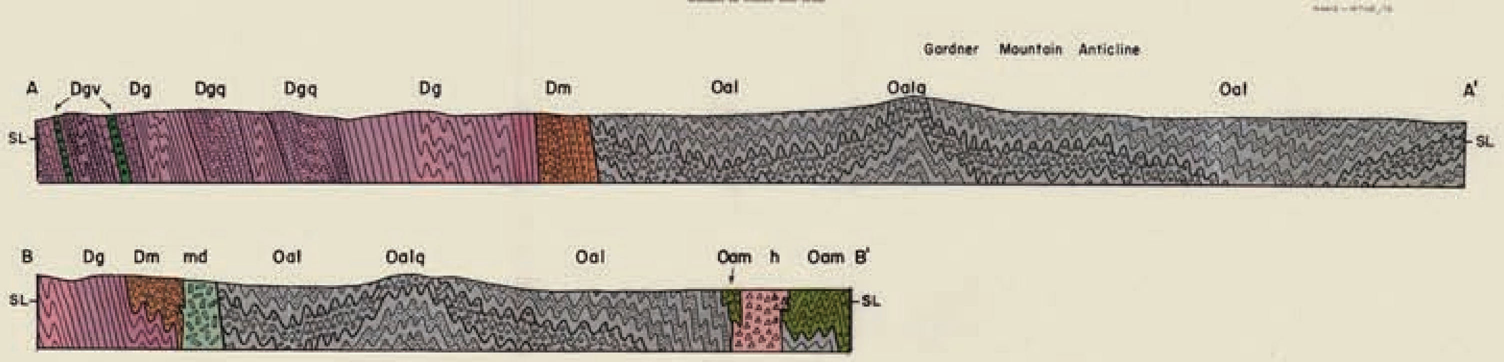
- h — Highlandcroft granodiorite

ORDOVICIAN(?) — Pre-Silurian formations

- Oam — Ammonoosuc volcanics: Soda-rhyolite tuff, breccia and volcanic conglomerate
- Oal — Albee formation: Phyllite, mica schist, quartzite and slate; Oalv: volcanics; Oalq: quartzite abundant

SYMBOLS

- Accurate / Approximate contacts
- Isograds
- Strike and dip of bedding
- Strike of vertical bedding
- Strike and dip of flow cleavage and foliation of igneous rocks
- Strike of vertical flow cleavage
- Strike and dip of slip cleavage

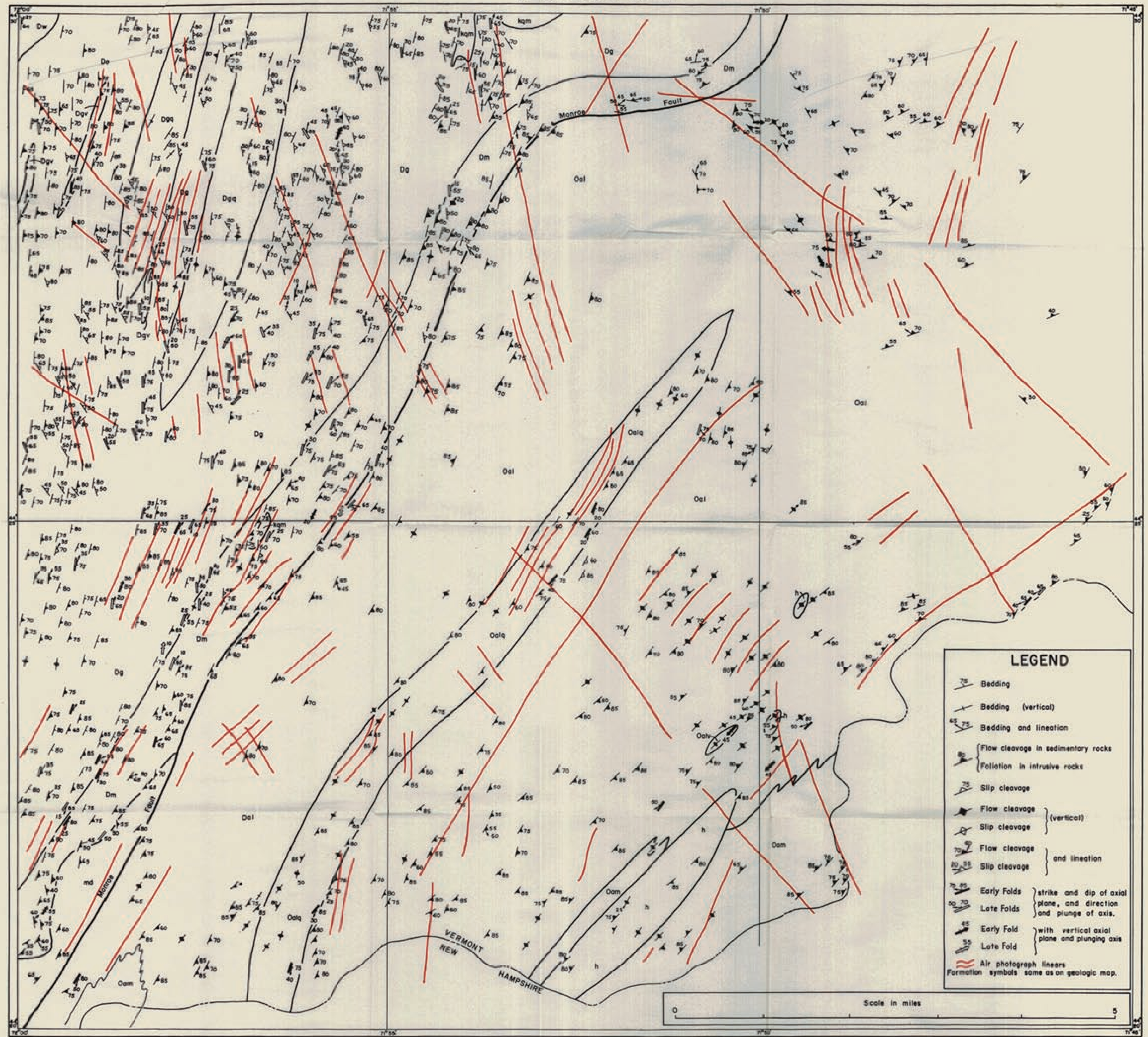


GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE CONCORD-WATERFORD AREA, VERMONT

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist

Geology by John H. Eric 1940-41
Revised by John G. Dennis 1956

TECTONIC MAP



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