GEOLOGY
OF
THE MEMPHREMAGOG QUADRANGLE
AND THE SOUTHEASTERN PORTION
OF THE IRASBURG QUADRANGLE
VERMONT

By

CHARLES G. DOLL

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CHARLES G. DOLL, *State Geologist*

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ABSTRACT

Among six formations mapped, three are new and, therefore, described in greater detail. They are the Ayers Cliff, Barton River, and Westmore formations, and embrace the belt occupied by the Waits River formation. More than 21,000 feet of strata ranging in age from Ordovician to possibly Devonian, constitute a variety of rock types exhibiting middle- and high-grade metamorphism.

A quartz conglomerate basal to the Cram Hill formation has been located, and the stratigraphic position of the Irasburg conglomerate is identified with the base of the Barton River formation. The Northfield slate continues as a distinct unit in Canada and lies below the Tomifobia slates and limestones of Clark.

Prominent structural features of late Devonian origin are the plutons composed of dark granite invaded by a light, coarser granite. On the west shore of Lake Memphremagog the Devonian granites are intruded by metabasic rocks consanguineous with those in the Owl Head Mountain region in Canada. The frequency of sills is related to the structures in the sedimentary rocks. Milky quartz veins of large dimensions are especially distinctive in the Barton River formation.

Northeast-trending folds and faults are related to the Acadian disturbance. Northwest of the axis of the Brownington syncline and including the tract containing the North Neighborhood anticline and

Indian Point syncline, the beds are predominantly overturned. The symmetry of the folds has been disturbed by the piercement of intrusives.

The Ware Brook thrust, lying between the Northfield slate and the Ayers Cliff formation, continues as the Bunker thrust in Canada and is traceable for more than 100 miles. A feature of importance associated with this fault, is a window in the Cram Hill formation exposing Ayers Cliff limestone. Genetically related to the Ware Brook thrust are the *en echelon* reentrants representing flaws in fanlike arrangement with respect to a major tectonic arc swinging toward the Gaspé from northern Vermont. The Black River fault merges with a fault mapped in the Tomifobia River valley in Canada which, projected northeastward, coincides with the flaw in the Massawippi reentrant. Noteworthy are the igneous masses on the southwest sides of the reentrants.

Superb examples of minor structures are impressive of their significance in the structural development of the region. Their effects are cumulative and are indicated by folds, faults, plastic flow, boudinage, and lineation.

INTRODUCTION

General Statement

Although small parts of the area described in this report received early attention, much of the geology remained unknown. The area, therefore, afforded an opportunity for detailed work which would serve as a beginning in the correlation of Vermont rocks with those on the other side of the International Border in Canada, thus adding to the knowledge of the Appalachian folded belt in this region. A prime consideration is the location of the area within a great tectonic arc whose geology is much better known to the southwest and northeast than it is here at the greatest bend. The present work might serve as a nucleus and reference for subsequent geological research in the region.

Location

The mapped area covers the Memphremagog quadrangle and a part of the adjacent Irasburg quadrangle, in northeastern Vermont (Fig. 1). On the Memphremagog quadrangle it touches upon the International Boundary between longitudes 72° 00′ and 72° 15′ west, extending south to latitude 44° 45′ north. The adjoining portion on the west in the Irasburg quadrangle, lies southeast of a northeast diagonal in the east rectangle and includes the southeast rectangle.

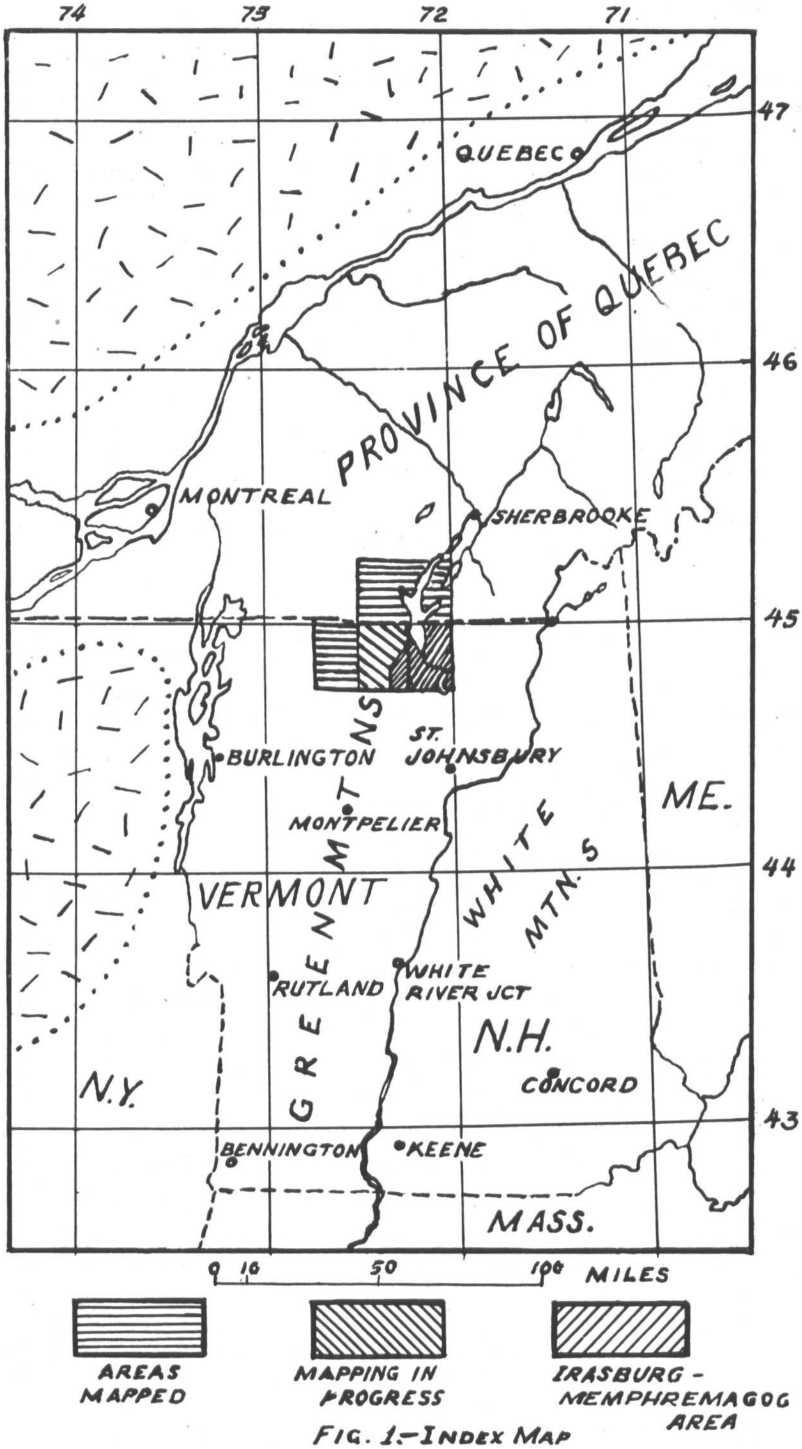

FIG. 1.—INDEX MAP

Acknowledgments

The writer is extremely grateful to the members of the Department of Geology in Columbia University, who have shown an interest in this problem. He is especially thankful to Professor Walter H. Bucher for his stimulating and valuable guidance and to Professor Marshall Kay for helpful suggestions. Dr. Robert K. Doten gave appreciable aid in the early part of the study of the thin sections and, in this connection, the writer wishes to express his thanks for the kindly advice given by Professor S. James Shand. Professor Elbridge C. Jacobs has made available, through the Vermont Geological Survey, financial aid for a part of one field season; his interest in the work is appreciated. Clarence G. Bailey and Harrison W. Cooke, Jr., assisted ably in the field for two seasons. Appreciation is expressed to Mr. L. S. Walker for rock analyses.

Previous Work

The earliest work pertaining to the geology east of the Green Mountains, appears in the first and second reports of the Vermont Geological Survey, by Adams (1845 and 1846, respectively). Thompson (1848) describes anew the "calcareo-mica slate" of Adams. A more comprehensive study of the rocks was made by Hitchcock (1861), who termed them "calciferous mica schist." Incidentally, Hitchcock (1861, p. 573) conceived the granite to be metamorphosed stratified rock. Early studies of the geology of the northern part of this region were made by Hall (1871, and in Hitchcock, 1861, p. 719). Parts of the area mapped were investigated by Richardson (1905–1906, 1907–1908, and 1911–1912) and by Jacobs (1921–1922).

In the terrane bordering on the International Boundary in Canada, pioneer work was done by Logan (1852 and 1863). Further studies were completed by Ells (1894), Kerr (1923), and Clark (1934). More recently, a geologic map accompanied by a brief description, by Ambrose (1943), shows the rocks equivalent to the three new formations herein proposed, undifferentiated.

Method of Investigation

The greater part of six summers was spent in detailed mapping of the Memphremagog and one-sixth of the adjacent Irasburg quadrangles, comprising in all an area of 262 square miles. Reconnaissance studies were made in surrounding regions, including Canada where some work has been done.

The standard topographical map, enlarged to a scale of 3 inches to the mile, served as a base in the field work. Most points and outcrops were located by the method of resection, excepting in wooded tracts where locations were established by means of pace-and-compass. Contacts between formations were traced in detail. A total of 172 thin sections were studied.

PHYSIOGRAPHY

Topography

The area is hilly to mountainous, rising from an elevation of 682 feet at the shore of Lake Memphremagog in the northwestern part to the greatest elevation of 2935 feet on the summit of Goodwin Mountain in the southeastern corner. The southeasterly increase in elevation is controlled largely by the progressively more resistant lithologies of the sedimentary formations in that direction. The lowest sedimentary surface is underlain by the relatively weak limestones of the Ayers Cliff formation and the highest sedimentary surface by the dominantly siliceous Westmore formation. By and large, stream dissection has attained a mature stage in a topography modified by continental glaciation.

The topography reflects the rock types and structures, as, for example, the domical granite hills extending northwestward from Elm Hill to Salem Hill in the central igneous body, and the hogback ridges exemplified by Shattuck and Dowling hills. The former reveal the massive structure of the more resistant igneous rocks and the latter the bedded structure of the more resistant among the steeply-dipping metamorphosed sediments. Barton Mountain is a conspicuous monadnock of granite near the southern margin of the area.

Elevated terraces in the vicinity of Lake Memphremagog record post-glacial uplift of the region. Two well-marked planes occur at elevations of about 700 feet and 1000 feet to the north and east of the city of Newport. The lower terrace is well-developed on the east side of the lake at Indian Point and Lake Park. The higher level is present on Pine Hill, directly east of Newport City, from whence it can be followed around the north end of the central igneous body and southeasterly to West Charleston. It probably represents, at least in part, an elevated delta of the Clyde River which was built into glacial Lake Memphremagog and has subsequently been dissected by stream erosion. Its surface contains the lakes peripheral to the central igneous body and is deeply incised by the Clyde River downstream from Salem Pond. Levels are manifest on the east side of the Barton River Valley south of Coventry Station at elevations of 760 feet and 960 to 1000 feet.

Weathering

Chemical weathering has penetrated along the bedding in the dipping limestones to depths of more than 2 feet, leaving a rusty-brown siliceous residue. On some limestone exposures it has etched a pock-marked surface, the elongate solution cavities delineating the bedding (Pl. 2, fig. 1). Spheroidal weathering is well-developed in the mafic granites on the slope east of West Charleston, and has produced subangular to rounded "cobbles" wrapped in extremely fissile shells, in a dense, hard phyllite in the Black River escarpment (Pl. 2, fig. 2).

A rather unusual display of differential weathering occurs in a pasture approximately half a mile northwest of Brownington Village (Pl. 2, fig. 3). A resistant, steeply-dipping stratum of dense, calcareous phyllite is here separated into segments resembling a row of tombstones, and thus, might aptly be termed tombstone weathering. Continued weathering has opened the pronounced vertical joints, causing individual slabs to become more or less isolated. The weaker adjacent layers lie concealed under a mat of grass.

Drainage

The region is situated in the Memphremagog basin, from which the drainage moves northerly to the St. Lawrence embayment. With the exception of occasional "cross-overs," the main streams are subsequent throughout the greater parts of their courses. The Clyde River meanders over thick deposits of outwash material in its upper course, but shows adjustment to rock structure in the vicinity of West Charleston where it flows through a picturesque gorge. Downstream from West Charleston it follows the contact between the central igneous body and the Barton River formation, curving around the northern border of the former and then descending rapidly from Clyde Pond across the regional structure in a series of cascades to the level of Lake Memphremagog. Two other major streams controlled by geological structure are the Barton and Black rivers, the latter following a faultline valley in the lower part of its course. The Willoughby River and Brownington Branch flow on glacial deposits throughout their lengths, the former crossing the regional trends of the bedrock. Waterfalls and rapids are not uncommon on the small tributaries that descend rapidly from the uplands. Of greater magnitude in the narrows of the major streams, they present spots of inviting scenery, as at Coventry on the Black River and at West Charleston on the Clyde River. The gorge of the Clyde River at West Charleston is an example of stream derangement due to glacial action (Doll, 1941–1942).

Excepting Johns River, which drains a restricted area north of Derby Pond, and Brownington Branch, the important streams have their sources in lakes, the Clyde River, in addition, flowing through a chain of lakes along its course. Seymour Lake and Echo Pond, which extend beyond the eastern margin of the quadrangle, discharge their waters into the Clyde River at East Charleston (Island Pond quadrangle). Noteworthy are the locations of Clyde, Derby, Salem and Brownington ponds in depressions with elongation on the contact between the central igneous body and the Barton River formation, indicating structural control. Clyde Pond is supported by a dam built for hydroelectric power.

Glaciation

The effects of continental glaciation are quite evident in the region generally. They appear as both erosional and depositional features, the former best displayed on outcrops at the higher elevations.

Glacial striae, often accompanied by grooves, indicate a general south-southeasterly movement of the ice, with local variations. On a single exposure the measured striae range from S 40° E to S 23° W, in consequence of the local topography. In places where the striae correspond in direction to the structural trend of the sediments, the direction of ice movement was probably guided by the flange- or rail-like edges of the steeply-dipping, resistant strata, to which the plastic bottom of the ice could have been molded. Such controlled glacial erosion could have modified valleys now containing subsequent streams. Chatter marks are sometimes associated with the striae on smoothed rock surfaces. A striking profile of large-scale *roches moutonnées* is exhibited by the southeasterly succession of eminences beginning with Barton Mountain on the northwest and including May Pond, Valley, and Wheeler* mountains, and Mt. Hor,* as viewed in a southwesterly direction from the pasture on the 2160-foot divide northeast of Westmore Village. In their

*Lyndonville quadrangle.

profiles of maximum asymmetry, the relatively gentle stoss and cliffed lee slopes illustrate clearly the southeasterly direction of glacial motion.

The northeast-trending Westmore mountain range of resistant rocks is breached transversely by glacial troughs, of which some contain lakes as, for instance, Crystal and Willoughby lakes. These erosional features resemble closely the geomorphic pattern on Mount Desert Island, Maine, described by Johnson (1925, pp. 95–99).

The southeastward trend of the Willoughby trough containing Lake Willoughby, particularly the southern half in the Lyndonville quadrangle, is a fine example of effective glacial scouring. The most impressive part of the trough lies between Mt. Pisgah* on the east and Mt. Hor* on the west, where the lake reaches a depth of 210 feet,† thus making the difference in elevation of the lake bottom and the summit of Mt. Pisgah some 1800 feet at this point. However, the location of the greatest sounded depth of 308 feet,† $1\frac{1}{5}$ miles to the north, and the huge talus piles sloping well into the lake from the commanding cliffs on both sides, suggest that overdeepening had added, at least, several hundred feet to the relief here before glacial and colluvial deposition. State Highway No. 5A is located for approximately 2 miles on forested talus along the east side of the lake. Incidentally, the highway affords a choice view of the imposing cliffs of the U-shaped trough, where it approaches the beach from the south at the south end of the lake. Jacobs (1919–1920, p. 297) has interpreted the Willoughby trough as a graben modified by glacial erosion.

Deposits of glacial till and stratified drift are strewn over the area. Thick boulder clay, underlain in places by intercalated sands and laminated clays, is exposed in high banks along the Willoughby River. In the river bank at Evansville an iron-stained sand layer between laminated clays exhibits thin laminations, especially at the border of a partially cemented seam separating 3 inches of coarser sand below from 13 inches of finer, somewhat clayey sand above. The sand is cross-bedded, and above the cemented seam it is folded in addition. The depositional sequence indicates the existence of a glacial lake dammed up by the ice front into which boulder clay was dumped during a temporary advance of the ice. The deformation of the upper lake beds below the till is apparently due to the unequal loading of the glacial till.

Glacial erratics are scattered over the whole region. They are of many

*Lyndonville quadrangle.
†Vermont Lake Studies project, Vermont Geological Survey, Summer, 1948.

varieties and sizes, some of enormous bulk and weight overlooking the valleys from the higher elevations (Pl. 2, fig. 4). In addition to those of local origin are erratics that have been transported for known and unknown distances, but all showing the direction of glacial motion. Especially significant are erratics of the Bolton lavas from the Owl Head Mountain region in Canada on the west side of Lake Memphremagog.

SEDIMENTARY ROCKS

General Statement

Within the area covered, the sedimentary rocks can be divided into six formations comprising a variety of rock types, with a total thickness of approximately 21,400 feet.

They have been named, from top to bottom:

SUMMARY TABLE

Silurian or Lower Devonian	Westmore formation	Mainly phyllites; schists, limestones and quartzites.
Silurian	Barton River formation	Slates and limestones locally abundant; schists, quartzites, basal conglomerate.
	UNCONFORMITY	
Ordovician	Ayers Cliff formation	Gray to blue-gray limestones, arenaceous limestones, minor slates.
	Northfield Slates	Gray to black slates, thin sandy and calcareous beds.
	UNCONFORMITY	
	Shaw Mountain formation	Basal quartz conglomerate, sandy and tuffaceous rocks, thin calcareous beds.
	UNCONFORMITY	
	Cram Hill formation	Dominantly slates; basal conglomerate, quartzites, schists, tuffaceous rocks, greenstones.

Three formations are discussed at greater length, because they are herewith proposed for the first time. They range from Ordovician to possibly middle Silurian or even early Devonian. Metamorphism has not only altered their compositions to varying degrees by recrystallization, but also has coarsened their textures in many cases. The following descriptions of the sedimentary rocks include petrographic details.

Cram Hill Formation

General Description

The term Cram Hill formation was given by Currier and Jahns (1941, p. 1493) to a series of slates, phyllites, schists and quartzites in central Vermont. The complete breadth of outcrop of the Cram Hill formation does not appear on the map, because its western limit has not been mapped in this area. However, the writer has made a traverse up the bed of Ware Brook, finding a quartz conglomerate at the South Newport road crossing, composed of saccharoidal quartz pebbles drawn out in a lustrous, blue-gray, schistose matrix. This conglomerate probably marks the base of the Cram Hill formation, as it underlies rocks exhibiting typical Cram Hill lithology. It is very probably equivalent to the quartz conglomerate reported from the western edge of the town of Irasburg by Richardson (1911–1912, p. 150). In central Vermont greenstone intrusives are said to occur near the bottom of the formation (Currier and Jahns, 1941). From Walker Pond northward the formation is in contact with granite. The Cram Hill formation is regarded here as the upper part of the Missisquoi schist of Richardson (1919–1920, p. 62).

Type Locality

The Cram Hill formation has its type locality in central Vermont (Currier and Jahns, 1941, p. 1494).

Distribution

The Cram Hill formation has been followed from the southwest corner of the east rectangle (Irasburg quadrangle) northward to State Highway No. 105 leading to Newport City. North of this point the formation is concealed under surficial deposits and the waters of Lake Memphremagog. At Walker Pond it borders on the granite which underlies Coburn Hill to the northwest. On the southeast side of Walker Pond and again on the lower slopes of Coburn Hill on the Newport City farm, the slates of the Cram Hill formation are infolded with the granites. The formation is best exposed in the highlands from West Hill southward.

Relative Thickness

From the base of the Shaw Mountain conglomerate in Ware Brook to the horizon of greenstone intrusives the formation has a thickness of approximately 1000 feet, but from the base of its own conglomerate to the base of the Shaw Mountain conglomerate it is 2650 feet thick. At State Highway No. 105 directly northeast of Walker Pond the thickness is between 500 and 600 feet. The intense crumpling and shearing of these rocks preclude an accurate estimate of thickness, hence the thicknesses indicated above are approximations.

Age

Currier and Jahns (1941, p. 1496) placed the Cram Hill formation in the Ordovician. The writer concurs in this view, which identifies this formation with the fossiliferous Magog slates of Trenton age (Ruedemann, 1919, pp. 122–124, 130) at the Castle Brook locality west of Magog, Quebec, and considers its stratigraphic position directly below the Shaw Mountain formation yielding fossils with Ordovician affinities.

Lithological Detail

The Cram Hill formation is composed mainly of gray to black slates which are more or less persistent throughout its thickness. The section in Ware Brook begins with the basal conglomerate directly below black crinkly slates including dark greenstone sills. Although much of the lower half of the section is under cover, the visible exposures show a black fissile slate intercalated with quartzites, greenstone sills and tuffaceous rocks. The basal conglomerate is underlain by chlorite- and quartz-mica schists, quartzites and greenstones of undetermined age.

In the middle of the section gray, argillaceous slates grade downward into black carbonaceous slates, which in turn give place to thin, gray calcareous beds forming the 10-foot face of a waterfall. Argillaceous slates again appear above the waterfall. The horizon of calcareous beds is regarded by the writer as a window of the Ayers Cliff formation on the basis of lithologic similarity and stratigraphic position and is discussed in a later section. Stratigraphically below the argillaceous slates at the crest of the waterfall are sericite and chlorite schists, quartzites and greenstone intrusives. The formation is also intruded by granite sills and dikes; granite commonly cuts the formation along its outcrop, as on the south end of West Hill.

The upper part of the section consists of slates which are interbedded with quartzites, grits, tuffaceous rocks and greenish-gray breccias, the innumerable fractures of which are coated with brown iron stain.

Because of their occurrence within the Cram Hill formation and its observed northward extension, the greenstones are identified with those on the west side of Lake Memphremagog. The granite intrusions are considered by the writer to be of the same age as the granites generally in the region mapped; they are certainly well within the area occupied by the major granite plutons.

Shaw Mountain Formation

General Description

The name Shaw Mountain was given by Currier and Jahns (1941, p. 1496) to a formation of interbedded conglomerate, tuffs, and crystalline limestones in the Barre, Vermont, region. Because of the occurrence of a fossiliferous horizon in central Vermont, the formation becomes an important stratigraphic unit. However, no fossils have been discovered in this formation in northern Vermont.

Type Locality

The type section is in central Vermont at the base of Shaw Mountain (Currier and Jahns, 1941, p. 1496).

Distribution

The Shaw Mountain formation is traceable from a brookbed on the slope southwest of Kidder School, northeastward to the hill directly south of Walker Pond, a distance of 7¼ miles. It has good continuity of outcrop from about half a mile southwest of Ware Brook to the brook on the northeast side of Cleveland Hill, otherwise its exposures are limited and interrupted. The conglomerate member of the formation is well exposed where it is crossed by Ware Brook and gives the formation topographic expression because of its resistance to erosion. The fine-grained members higher in the formation are locally and poorly exposed.

Relative Thickness

The Shaw Mountain formation is not uniformly thick along its strike, but varies from 0 to about 500 feet in thickness. Since the tuffs and limy layers are extremely poorly exposed, it is impossible to get a measure of the true thickness of the formation. However, on the upper slopes about a quarter mile north of Ware Brook a few scattered exposures of the tuffs permitted a complete traverse across the outcrop of the formation, which gave a thickness of 300 feet. One mile north of Ware Brook a thickness of 500 feet was measured. In the bed of a brook half a mile south of Ware Brook the conglomerate is absent but the tuffs are 85 feet thick. In another brook bed southwest of Kidder School there are about 45 feet of tuffs and no conglomerate. The formation is thickest where the conglomerate is present and thins greatly where the conglomerate is missing. Because of its prominence where present, the thickness of the conglomerate can be fairly accurately measured as, for instance, where it is crossed by Ware Brook and is about 120 feet thick, and in a brook on the south side of Cleveland Hill, where it is approximately 200 feet thick.

Locally, along the strike, the formation thins visibly and even disappears completely as, for example, on the northeast side of Cleveland Hill. At a road junction half a mile north of this point it is 16 feet thick, intricately sheared and full of cavities, some filled with a limonitic powder and quartz grains. The variations in thickness of the Shaw Mountain formation are due in part to original deposition, in which the quartz conglomerate is the basal member in the thickest portions. However, brecciation, shearing, flow structure and faulting show that tectonic forces have played an important part in altering the apparent thickness. At the exposure southeast of Walker Pond, the formation is intensely sheared and brecciated, and abundantly porous due to the weathering of sulphides.

Age

Currier and Jahns (1941, p. 1501) placed the Shaw Mountain formation in the Ordovician, on the basis of crinoid stems found in a crystalline limestone member.

Lithological Detail

The most conspicuous member of this formation, a quartz conglomerate at its base, stands as a cliff on the north side of Ware Brook (Irasburg quadrangle), contrasting with the lower relief to the east underlain by the less resistant Northfield slates. Because of its prominent exposures, the quartz conglomerate makes an excellent mapping unit, despite its absence at intervals along the strike.

The pebbles consist of quartz through most of the thickness, but a darker phase approximately one foot thick at the top contains pebbles of a gray, banded slaty rock. Most pebbles are small, occasional ones having a length of 5 to 6 inches. They are elongated as though subjected to tectonic stress, possessing a schistosity which plunges steeply to the northwest, as, for example, a foliation plunge of 70° N 50° W a mile north of Ware Brook.

In places the conglomerate is cut by veins of milky quartz, generally along the strike but occasionally crossing it. Thin black lenses along the bedding are the result of the alteration of sulphides. Generally the schistose matrix exhibits a limonite stain which penetrates between the grains of quartz and sericite.

The quartz conglomerate is overlain by fine-grained, dark gray to sugary white, laminated sandy and tuffaceous layers which weather pinkish-tan to brown. These are usually poorly exposed owing to their susceptibility to erosion. Thin calcareous zones appear to have irregular and sparse occurrence near the top of the formation.

Northfield Slate

General Description

A group of slates as redefined by Currier and Jahns (1941, p. 1501) and lying between the younger Ayers Cliff and older Shaw Mountain formations, constitute the Northfield slate. Along its whole outcrop, it is overturned toward the southeast and therefore appears to rest unconformably on the Ayers Cliff formation. Convincing evidence, however, proves the contact to be a thrust fault. As pointed out by Currier and Jahns, the stratigraphic position of the Northfield slate has in the past lacked satisfactory explanation, chiefly because this slate was incorrectly correlated, by C. H. Richardson (1907–1908, p. 276), with slates definitely known to be within the Ayers Cliff and, very probably, the Barton River formations.

Type Locality

The type locality appears to be a quarry near the village of Northfield (Currier and Jahns, 1941, p. 1503).

Distribution

Good exposures of the Northfield slate can be traced northward along the prominent escarpment west of the Black River to the vicinity of West Hill School. Northward they are concealed by recent deposits but appear again southeast of Walker Pond, from whence they continue as widely scattered low outcrops to Lake Memphremagog. Along this traverse the formation is generally well exposed where it is crossed by brooks flowing down the escarpment, especially where the escarpment is steep. Its superior resistance to erosion when contrasted with the neighboring Ayers Cliff formation, is readily seen in the topographic expression of each.

Relative Thickness

Throughout most of its outcrop the Northfield slate varies between 850 and 900 feet in thickness. However, in the proximity of West Hill School its thickness increases to about 1200 feet, but north of this point the formation thins to about 860 feet as measured across the low hill southeast of Walker Pond. Although the formation is poorly exposed farther north, projections of outcrop data indicate that it apparently thins noticeably to the shore of Lake Memphremagog, not as the result of primary deposition, but of the gradual cutting out by a fault.

Age

Since the Shaw Mountain formation is regarded as of Ordovician age, the Northfield slate which lies stratigraphically above it cannot be older; because of fossil evidence in the underlying formation it is referred to the Ordovician.

Lithological Detail

The Northfield slate is gray to black, weathering to yellow- or reddish-brown on the cleavage surfaces. In places it contains bands of a lighter gray sandy material which in some instances composes thin layers. Calcareous beds occur at various horizons, becoming more abundant toward the top of the formation. An occasional massive, black layer of limestone is rich in pyrite grains, crushed in the course of deformation and drawn out into glossy streaks in the direction of greatest elongation. The slates are both smooth and highly crinkled on the bedding and cleavage surfaces and not infrequently do they show excellent bedding and cleavage relationships indicating the beds to be overturned to the southeast.

A thin zone exhibiting small, white quartz pebbles was found in the lower part of the formation where it is crossed by a brook on the north-east side of Cleveland Hill. The pebbles which occur in a predominantly limestone horizon are not at all abundant. This is probably equivalent to the nodular zone described by Currier and Jahns (1941, p. 1502). The basal conglomerate described by Currier and Jahns (1941, p. 1502) has not been located in this area. It has been seen, however, at the base of

Bunker Hill in Canada, not far above the fault and about 1½ miles southwest of Lake Massawippi. Its absence southward is probably due to cutting out by the fault.

Ayers Cliff Formation

General Description

C. H. Richardson (1905–1906, p. 86, 115) applied the term Waits River limestone to the calcareous members of a group of poorly differentiated strata traversing the length of the state east of the Northfield slate. Because a variety of other rocks are present the group became known later as the Waits River formation (Currier and Jahns, 1941, p. 1491). The writer has succeeded in delimiting three stratigraphic units in this belt of rocks and has named them Ayers Cliff formation, Barton River formation, and Westmore formation (Plate 1). Thus, because of the new interpretation of these rocks, the altogether too comprehensive terms Waits River formation (Currier and Jahns, 1941), and Tomifobia formation (Clark, 1934), become inadequate. The Ayers Cliff formation constitutes the westernmost of the newly proposed formations and comprises a belt of impure limestones, calcareous sandstones, and minor slates.

Type Locality

The type section from which the formation receives its name, is in a series of road cuts beginning half a mile west of the village of Ayers Cliff at the south end of Lake Massawippi in Canada. Good exposures within the area mapped may be seen at Coventry Falls in the Black River, one-fourth mile northwest of Coventry Village, on Stony Hill 2 miles northwest of the village of Irasburg, and on the shore of Lake Memphremagog at Lake Park. Other exposures are few and scattered.

Distribution

The general paucity of outcrops is due to the inferior resistance of the formation to erosion, consequently it lies generally in valleys. With the exception of a short distance across the more resistant slates at the nose of the Indian Point syncline at Coventry Village, the Black River is confined to this formation. Despite the common lack of exposures, the Ayers Cliff formation is traceable from the vicinity of Round Hill (Irasburg quadrangle) northward into Canada.

Relative Thickness

Owing to the dearth of outcrops, the thickness of the Ayers Cliff formation is difficult to measure, but from data available a thickness of 4500 feet is indicated.

Age

The Ayers Cliff formation overlies the Northfield slate which has been identified with the lower Tomifobia slates in Canada (Currier and Jahns, 1941, p. 1510). The Tomifobia slates which lie east of the Bunker thrust, are reported to have yielded graptolites of Trenton age (Clark, 1934, p. 12). In a later description they are called non-fossiliferous (Ambrose, 1943). Thus far no fossils have been found in the Ayers Cliff formation in the area mapped. According to Clark (personal communication) the graptolites were found in slates at the Tomifobia railroad station which is far to the east of the expected stratigraphic position of the Northfield slate in Canada. The writer's identification, based upon his own field work and reconnaissance in Canada, shows that the Tomifobia locality lies well within the area covered by his Ayers Cliff formation.

In the light of these facts, therefore, it appears that the correlation of the Northfield slate with the Tomifobia formation is incorrect. The Northfield slate in Canada composes the sole of the Bunker thrust at Bunker Hill.

The Ayers Cliff formation is at least middle Ordovician (Trentonian) or younger, since the Northfield slate over which it lies rests unconformably upon the fossiliferous Shaw Mountain formation of middle Ordovician age (Currier and Jahns, 1941, pp. 1500–1501).

Lithological Detail

The Ayers Cliff formation is commonly gray to blue-gray with interbedded light and dark gray banded layers. The beds are generally thin, slabby, fine- to medium-grained, and in places extremely fissile. At Stony Hill they are thick, massive and contain considerable sand. The lighter colored and white, marbleized varieties are normally coarser grained than the darker, and are, in many places, friable due to the large amount of sand present. The varying proportions of silica cause a range from arenaceous limestones to calcareous sandstones. Some of the friable limestones might more correctly be termed calcareous sandstones, judging from a study of thin sections and the chemical analyses shown in Table 1, even though the external appearance and vigorous effervescence suggest limestone.

By and large, the formation possesses a fair degree of crystallinity due to metamorphism. Certain horizons are especially dark and carbonaceous, lacking somewhat in luster. Throughout the formation are narrow beds of dark gray slates and occasional phyllites which become more abundant in the lower and upper parts, thus suggesting a gradational contact and possible conformable relationship to the subjacent and superjacent formations, both of which are composed largely of slates.

In thin section, calcite ranges from less than 15 per cent to as much as 65 per cent which, with proportionate differences in the amounts of quartz, indicates that the formation consists of highly impure carbonate rocks. Minerals present besides those already mentioned as composing the major part of the rocks, are porphyroblasts of tremolite, chlorite, muscovite, diopside and occasional titanite. The chlorites inclose zircons, many of which are surrounded by pleochroic haloes; the chorites appear to have replaced biotite. Diopside is developed in appreciable quantities near igneous contacts, particularly in porcelainized zones. Sulphides in parallel arrangement are confined largely to carbonaceous bands. Some of the bands are rich in sericite and alternate with others abounding in carbon. As is evidenced by the composition of these rocks, the original sediments were probably calcareous sands and muds and impure limestones.

TABLE 1—Chemical analyses of limestones

	A	B	C	D	E	F	G
Water	0.04	0.03	0.10	..	..	..	..
Organic matter	0.52	0.26	0.64	0.55	0.64	0.40	0.46
Insoluble matter	76.78	39.34	65.56	33.63	62.98	38.19	32.23
Calcium carbonate	20.99	51.25	29.26	54.63	33.47	56.84	61.10
Magnesium carbonate	1.67	3.88	3.24	3.64	0.92	1.02	5.69
Iron & aluminum oxides	0.00	2.80	1.04	..	..	..	..
Potassium silicate	..	..	..	1.15	trace	0.42	trace
Sodium silicate	..	..	..	5.97	trace	2.79	trace
Undetermined	0.00	2.44	0.16	0.70	1.99	0.34	5.52

A. Barton River formation. Southwest slope of Cargill Hill.
B. Barton River formation. Beside abandoned road, 1⅛ miles west of Orleans.
C. Barton River formation. Road cut 3¼ miles northwest of Barton Village.
D. Ayers Cliff formation. Shore of Lake Memphremagog, Lake Park.
E. Ayers Cliff formation. Stony Hill, 2 miles northwest of Irasburg Village.
F. Ayers Cliff formation. East shore of Lake Memphremagog at International Boundary.
G. Ayers Cliff formation. Black River Valley, 3¾ miles southwest of Newport City.

Analyses by L. S. Walker, Chemist in charge, Regulatory Service, Agricultural Experiment Station, University of Vermont.

Barton River Formation

General Description

The Barton River formation consists of intercalated impure calcareous rocks ranging from limy quartzites to limestones, amphibolitic layers, slates, phyllites, quartzites and schists. They are gray to black, the calcareous members containing thin, porcelainized bands parallel with the bedding. Brown iron stain is a characteristic weathered coating, particularly of the slates and phyllites.

Calcareous beds of the Ayers Cliff formation are involved with the folds of the overlying Barton River formation and display the same structural behavior. Thus a structurally conformable relationship appears to exist between both formations. However, in the village of Irasburg and southwestward the formation is underlain by a basal conglomerate. This conglomerate was discovered by C. H. Richardson in 1904 and named by him Irasburg conglomerate from the town of that name. Because of its historic as well as geologic significance, this conglomerate is discussed in a separate section. The contact between the Barton River and Ayers Cliff formations is largely under cover of recent deposits, but at a few places along the strike, as on the west slopes of Shattuck and Dowling hills and in a pasture at the south end of the Newport airport 2¼ miles east-northeast of Coventry Village, they occur in juxtaposition.

Type Locality

The type locality of the Barton River formation is located in Newport City in a low cut along the main highway to Derby and on the south flank of Shattuck Hill. Here is a section exhibiting rock types typical of the formation. South of the city of Newport in the five-mile long escarpment bordering the Black River on the west, is a series of rocks composed largely of slates which constitute the lower and basal portions of the Barton River formation. Stratigraphically higher in the railroad cut, immediately north of the business center in the village of Orleans, the formation is dominantly limestone. In the bed of Stearns Brook at Tice near the top of the formation, limestone is again the prevailing rock.

Distribution

The formation covers the area between the Ayers Cliff and Westmore formations, and in addition includes the youngest beds of the Indian Point syncline. Like the Ayers Cliff formation it is continuous from the southern border of the map-area northeastward into Canada, but with outcrops more abundant. Exposures are scarce, however, along portions of igneous contact zones where they lie concealed under a blanket of Pleistocene and later sediments. The more resistant sections of the formation form prominent ridges and steep slopes, a good example of which is the cliffed escarpment on the west side of the Black River extending from Newport City to Coventry Village. Because the formation is generally well exposed on the higher slopes of the Barton River Valley, particularly on its west side, the name Barton River has been adopted for it.

Relative Thickness

The beds of the Barton River formation are prevailingly overturned. Moreover, the examinable exposures in the entire width of outcrop between the west slope of the Barton River Valley at Coventry Station to the eastern boundary at Brownington Center, show inversion of the strata. However, even in this succession of overturned beds the width of outcrop is not the true measure of thickness of the formation, because minor folds, flow structure and longitudinal shear give evidence of thickening. The presence of these structures indicating intense deformation and, in addition, the discontinuity of exposures, make the determination of thickness an approximation at best. Notable additions to the apparent thickness of the formation must be attributed to the many sill intrusions.

Since the structural behavior of the strata here is very similar to that in central Vermont, the writer has based his calculations of thickness on the tried factor of 0.4 which was used by Currier and Jahns (1941, p. 1495). Employing this average factor, the relative thickness of the formation becomes 8800 feet.

Age

Recognizable fossils have not yet been found in the Barton River formation, but elongate lustrous markings on the bedding surfaces of a thin horizon of fine-grained, medium-gray limestone in the bed of Trout Brook in Brownington at the Coventry townline, might, from their shapes, be suggestive of graptolites. In general, it may be assumed that these markings represent nothing more than mineral streaks, especially when they are found also on cleavage surfaces. The markings in the limestone in Trout Brook, however, are of several distinctive shapes and are restricted to the bedding surfaces. Furthermore, they are confined to one narrow horizon, although the lithology of this horizon, where repeated elsewhere in the formation, is barren. They bear a striking resemblance to many of the individuals from the classic Castle Brook locality at Magog in Canada, which are likewise mere smears but have the advantage of being associated with easily identifiable graptolithic impressions. It may be that strong movements along the bedding planes and fine recrystallization have eradicated the structural characteristics of the original impressions and transformed them into featureless smears with a predominantly parallel orientation. Doctor Ruedemann holds similar markings in the rocks from the eastern part of the State to be graptolites (1947, p. 63 and personal communication; Doll, 1943, p. 59). E. C. Jacobs has reported "crushed graptolites" from the bed of Trout Brook (1921–1922, p. 101).

Regardless of the interpretation of these markings, they are in such a poor state of preservation that the age of the formation can be determined at present only by stratigraphic methods. The formation overlies the Ayers Cliff formation unconformably, but, unlike it, is referred to the Silurian, because the unconformity underlies a conglomerate exhibiting a variety of constituents along its strike that are believed to have come from the Taconic highlands at the close of the Ordovician, thus indicating a major break. This age assignment is also based upon the position of the formation stratigraphically below strata yielding fossils of Silurian age or younger. In this connection, it is significant that "slip cleavage" is recorded in the strata of this formation but was not observed in the beds of the underlying formation (Pl. 5, figs. 1 and 2).

Lithological Detail

The lower part of the formation consists of about 1600 feet of dominantly slates and phyllites intercalated with lesser amounts of limestones and calcareous schists. In the middle and upper parts the calcareous rocks increase perceptibly, locally constituting the major rock type. The top of the formation is characterized by a noticeable change to siliceous and argillaceous rocks composed preponderantly of phyllites. Five main rock types have been identified and include slates, phyllites, quartzites, limestones and schists. Although these rocks are distributed fairly uniformly throughout the formation, each is notably more abundant locally. Certain beds of limestone are weathered to a depth of 2 feet and more, leaving a dark brown, crumbly, siliceous residue.

Slates—The slates are fine-grained, banded, and extremely fissile in places. They are gray to black on fresh surface, but on the weathered surface the color is characteristically a rusty brown due to iron stain. Thin section study shows the bands to vary from those made up nearly of pure quartz grains and carbonaceous material to those very rich in biotite. There are a few bands of quartz and sericite. Biotite is present both as fine scales and large masses. Fine-grained sulphides occurring as stringers parallel to the banding, are possibly of syngenetic origin. Sulphides also occur in large masses throughout the rock and in quartz veins. One band carries from 30 to 40 per cent of a reddish-brown, translucent mineral associated with the sulphides, which may possibly be hematite. Blotches of hematite have been determined on fracture surfaces in the slates. Large porphyroblasts of a colorless garnet and scattered grains of detrital, greenish-brown tourmaline, are present.

Phyllites—In a number of instances it was very difficult to distinguish between phyllites and slates, and in such cases the specimens possessing lustrous surfaces showing a better crystallization, were included with the phyllites. In other respects both rock types have essentially the same features, since they differ only in the degree of metamorphism suffered.

Besides the minerals described from the thin sections of the slates, the phyllites carry additional varieties in accordance with a higher degree of metamorphism. As in the slates, the groundmass is composed mainly of quartz, sericite and carbonaceous material and, in addition, often biotite. Porphyroblasts of a colorless garnet show inclusions in an unusual pattern, in which the cut of the thin section is such as to disclose the porphyroblast with a clearly delineated hexagonal outline bounding a six-rayed set of remarkably straight fractures corresponding to crystallographic planes (Pl. 3, fig. 1) (Cohen, 1900, Pl. 17, fig. 3; Renard, 1882, p. 16, Pl. I, fig. 1). These planes expose the garnet to alteration, so that in the more advanced stage the fractures have developed into a six-rayed set of inclusions, giving a cartwheel pattern to the porphyroblast (Pl. 3, fig. 2). Occasional garnets are edged by a thin band of sulphide.

Biotite flakes often show parallel arrangement and at an igneous contact they compose small veins. They are sometimes lath-shaped or irregular in outline and pale-colored when associated with chlorite. Numerous porphyroblasts of andalusite are rimmed by biotite and are now much altered to quartz and muscovite (Pl. 4, fig. 1). Rectangular areas of biotite, muscovite, calcite, and quartz are suggestive of completely altered andalusite crystals. Large porphyroblasts of a pale yellow-green amphibole are bent and broken where they lie across the schistosity (Pl. 3, fig. 1).

A coarsening texture and noteworthy mineral development have been observed at igneous contacts and at the borders between the sedimentary rocks as well. Where the phyllite is banded with impure limestone there is a generous growth of biotite and brown tourmaline. At a quartzite contact the phyllite carries large biotites frequently in parallel orientation, muscovite flakes, quartz and masses of radiating amphiboles. Veins parallel with the contact contain quartz, biotite, sulphide and a little orthoclase. Along a granite contact a zone of muscovite mixed with calcite occurs in the phyllite. Zones of carbon-rich material continue across the contact from the phyllite into the granite.

Quartzites—The quartzites are not abundant rocks in the formation. They are normally thin-bedded, are seldom massive thick-bedded and are fine- to medium-grained, sometimes displaying cross laminations. They range from medium-gray to black and are frequently streaked with minerals zoned parallel to the bedding. None of the quartzites are pure, consequently they bear a variety of complex metamorphic minerals. Some of the quartzites are calcareous and might represent highly silicified limestones. Patchy, nonuniform effervescence of the hand specimen might be suggestive of such an origin, as might also be the presence of lime-silicate minerals. However, since the quartzites often are interbedded with the limestones, the carbonate substance necessary to form the lime-containing minerals could conceivably have been introduced by selective diffusion from the adjacent limestone during metamorphism. A few of the calcareous quartzites contain up to 15 per cent of calcite.

Microscopic sections show small to large biotites in varying amounts and frequently with parallel arrangement. Orthoclase and plagioclase are present in small amounts. Both brown and green tourmaline in irregularly-shaped masses and specks, is disseminated widely in some slides. Sillimanite is abundant locally in igneous contact zones where it occurs in both sheaf-like bundles and isolated crystals (Pl. 4, fig. 2). Yellow-green, pleochroic amphibole contains scattered epidote grains which in part are replacing the host mineral. In other instances the amphibole shows alteration to zoisite or clinozoisite and contains many quartz grains. Nearly colorless amphiboles occur in contact zones. The minerals tremolite, zoisite or clinozoisite, scapolite, and plagioclase, are found predominantly in calcite-bearing quartzites. Sulphides are present in masses and as stringers and are frequently in parallel orientation. Minerals in minor amounts are titanite, rutile and chlorite.

Limestones—Like the quartzites, the limestones are impure and greatly altered. Lithologically they resemble closely the limestones of the Ayers Cliff formation, excepting that, due to their association with other kinds of stratified rocks and granite plutons, they contain a greater variety and larger quantity of complex minerals produced by metamorphism. They are light- to dark-gray, blue-gray, mottled, and white in narrow bands affected by hydrothermal metamorphism. The texture becomes coarser with increase in metamorphism. Calcite composes as much as 70 per cent of the rock, more rarely constituting the only mineral along with more or less carbonaceous matter. Quartz has about the same ratio that calcite has in the quartzites, generally as high as 20 per cent; occasionally it reaches 60 per cent, in which case calcite makes up about a third of the rock. Calcite veins have been noted in some of the limestones. They follow both bedding and fractures and are ordinarily thin. However, one coarse-textured vein in the bed of upper Johns River measured 7 inches in thickness and was traceable for 8 feet; its strike is N 60° W and dip vertical.

Microscopic study shows the limestones to be largely lime-silicate rocks. Among the lime-bearing silicates are wollastonite, tremolite, diopside, scapolite, zoisite, clinozoisite and, possibly, vesuvianite. Tremolite sometimes appears in brush-like groups often attached to rectangular-shaped muscovite porphyroblasts and scattered throughout the slide. Muscovite is occasionally abundant and in a fine-grained rock it appears as phantom crystals visible only in crossed nicols. Biotite flakes are often elongated parallel to the banding of the rock, and phlogopite is present as lath-shaped crystals. Fan-shaped masses of needle-like sillimanite occur along the contact with a fine-grained rock composed of quartz grains and much carbonaceous material. Diopside appears to be restricted to igneous contact zones. Much of the zoisite and clinozoisite is polysynthetically twinned and the clinozoisite is present sometimes as interstitial material. Nearly colorless garnets appear in some layers as large metacrysts. Porphyroblasts of a yellow-green, pleochroic amphibole are much corroded and replaced by quartz and calcite. Occurrences of sulphides are framed by calcite crystals. Specks and grains of tourmaline are green-brown and brown. Other minerals in smaller amounts are epidote, titanite, chlorite and zircon.

Schists—These rocks comprise quartz-mica schist, andalusite schist, hornblende-garnet schist, biotite-garnet schist and phyllite schist. The predominant minerals are quartz and mica, and the shades of gray to dominantly black are due to various concentrations of carbonaceous material in the rock. The schists are prevailingly fine-grained and lustrous due to disseminated micas. They are not confined to any particular part of the formation, but are found widely distributed. On the whole they are strongly banded and foliated.

The quartz-mica schist contains either biotite or sericite as the maximum mica. Both micas are frequently elongated parallel to the banding in the rock and also occur as large irregular and well-defined porphyroblasts. In some schists the biotite shows partial resorption and in others it is strongly twinned. In one schist possessing strikingly developed herringbone structure (due to incipient fracture cleavage) in the fine laminae, large porphyroblasts of biotite are bent and broken to conform with this structure (Pl. 5, fig. 1). Veins of secondary biotite also occur. Porphyroblasts of tremolite appear as fragments and phantom crystals. Large porphyroblasts of yellow-green, pleochroic amphibole contain much quartz. A few calcite grains show elongation with the schistosity. Olive-colored and brown tourmaline in single grains and irregularly-shaped masses, is fairly common. Sulphide occurs in the form of blebs and stringers parallel to the banding and as specks in augen of quartz. Rods of sulphide are surrounded by haloes of secondary biotite (Pl. 5, fig. 2). Rectangular porphyroblasts of chlorite are in random orientation, often cutting across the structure and also associated with partially altered biotite.

Large interlacing porphyroblasts of andalusite resembling diabasic texture make up an appreciable part of a few andalusite schists. Occasionally the andalusite is pleochroic in a rose-colored variety (Pl. 6, fig. 1). Much of it is considerably altered to quartz and muscovite; some of it contains clinozoisite. The space between the andalusite porphyroblasts is occupied largely by biotite, brown tourmaline, quartz and carbonaceous material. One thin section contains only pseudomorphs of the pleochroic andalusite porphyroblasts composed of a mixture of muscovite, biotite and quartz grains. Another slide shows abundant large porphyroblasts of the variety chiastolite containing geometrically arranged carbonaceous inclusions (Pl. 6, fig. 2). They are also considerably altered to quartz and muscovite. Staurolite showing inclusions, is present as porphyroblasts.

The hornblende-garnet schist has large porphyroblasts of a pale-colored garnet and green-yellow amphibole uniformly distributed in a groundmass of quartz and carbonaceous material. In the biotite-garnet schist the frayed biotite porphyroblasts become invisible when their cleavage directions do not coincide with the vibration direction of the lower nicol, but the cleavage lines alone are clearly discernible. The garnets are noticeably fractured and somewhat shattered, and partially rimmed with sulphides. The foliated structure of the phyllite schist is strongly emphasized by bands much richer in mica than others. The rock is cut by quartz veins containing chlorite, biotite, and sulphides.

Irasburg Conglomerate

The type locality of the Irasburg conglomerate is in the bed of Lords Creek and on the adjacent slope to the south, where Richardson (1905–1906, p. 82) first described it. Richardson (idem, p. 83) placed it at the base of the Ordovician while Currier and Jahns (1941, p. 1509) consider it to be an intraformational conglomerate in the Ordovician series.

The writer regards the Irasburg conglomerate as basal to the Barton River formation, chiefly on the basis of its stratigraphic position. It has been traced along the boundary between the Barton River and Ayers Cliff formations from the bed of the Black River in Irasburg Village southwesterly to the steep slope half a mile directly east of the village of Albany. Richardson (1917–1918, p.106) has reported this conglomerate as far south as Northfield, but from his description of the locality it is evident that he was dealing with the conglomerate basal to the Northfield slate (Currier and Jahns, 1941, p. 1502). Furthermore, among the five phases into which Richardson (1927–1928, p.108) divided the Irasburg conglomerate on the basis of matrix composition, only the Irasburg and possibly the Albany phases belong to this conglomerate. The stratigraphic position of the "Northfield phase" is cited above and the "Coventry phase" is very probably the Shaw Mountain conglomerate.

The Irasburg conglomerate has not been studied in detail, but field observations have brought to light some interesting facts regarding it. The limestone matrix, which strongly resembles the limestones in the underlying Ayers Cliff formation, is constant in composition, but the constituents vary considerably in variety, size and amount along the strike. The greatest variety of particles is exposed at the type locality and range in size from dominantly pebbles to sparsely distributed boulders. Indeed, in places along the strike only single cobbles betray the presence of the conglomerate. The largest boulder recorded measures 9 feet long by 3 feet wide and, like the other elongated constituents, was oriented with the long dimension parallel to the axes of minor folds on the bedding planes of the matrix. Some of the boulders are bent with the folds. The constituents are well rounded to sub-angular and, according to Richardson (1927–1928, pp. 108–109), consist of granite, andesite, diorite, diabase, serpentine, chlorite- and sericite schists, sericitic marble and micaceous quartzite. To this list the writer would add dark, crystalline limestone, black mudstone and phyllite. The amygdaloidal structure of some of the aphanitic constituents might be suggestive of volcanics. A conglomerate with so large an assortment of constituents can hardly be intraformational, as classified by Currier and Jahns (1941, p. 1509). Field relations show definitely that the Irasburg conglomerate is a basal conglomerate and constitutes the earliest deposits of the Barton River formation. Northward from Irasburg Village, the conglomerate is probably concealed under the widespread loose deposits.

Westmore Formation

General Description

The Westmore formation derives its name from the village of Westmore on Lake Willoughby in the southeastern corner of the Memphremagog quadrangle. The rocks are largely phyllites and schists, with smaller amounts of limestones and quartzites, all interbedded. Phyllites of whetstone quality have been quarried at several places in the formation, but the largest opening is located 1¼ miles due north of Lake Willoughby in southeastern Brownington. The light-colored rocks are proportionately greater in amount than in the formations to the west and the sulphides appear to be greatly reduced in comparison. Field observation and petrographical study seem to indicate that the sulphides are much more common in the darker, more carbonaceous rocks, hence their wider dissemination in the underlying formations.

Type Locality

A good section is exposed in the whetstone quarry cited above. The quarry is about 150 feet long and has been opened approximately along the strike of the layers. The rock is massive and well bedded. Other exposures affording additional sections, are located directly north and south of the village of Morgan, in a pasture half a mile west-southwest of Echo Pond, and in cuts along the state highway leading from West Charleston to Island Pond.

Distribution

The formation has its greatest width of outcrop in the southeastern portion of the map and narrows north of the reentrant at Pensioner Pond south of West Charleston, continuing northward along the eastern margin of the map into Canada. Much of the formation is obscured by glacial and recent deposits, as for example, north of the Barton Mountain-May Pond Mountain pluton. Outcrops are more abundant at the higher elevations.

Relative Thickness

Although the breadth of outcrop of the Westmore formation exceeds that of the Barton River formation along the line of section, its thickness is less because of a wide terrane of gentle dips in the southeast portion of the map-area. Much of the formation is cut by igneous intrusions in the same manner as the Barton River formation. Measured from the axis of the Brownington syncline eastward, the thickness is estimated to be 4300 feet.

Age

In a paper published not long ago (1943), the writer described a few specimens believed by him to represent fossils. They were found in a mottled limestone stratum in the bed of Stony Brook just above the farmhouse and at a point under the "S" in the town word "WESTMORE" on the map. The fossils have been identified as cystoid and crinoid calyces whose structural features indicate genera belonging to the Middle Silurian or possibly the Lower Devonian.

However, since there is some disagreement among authorities concerning the true nature of the specimens, the age of the formation is at present based on its stratigraphic position above the Barton River formation which makes it the youngest formation in the area mapped. Some 50 miles to the south in the Strafford quadrangle, the Barton River formation appears to have its counterpart in the Memphremagog formation (Doll, 1943–1944). The latter is overlain to the east by the Gile Mountain formation of Devonian age, which resembles broadly the lithology of the Westmore formation. The corresponding lithologies and stratigraphic positions of these respective formations are indicative of similar age relationships. The Westmore is therefore considerably younger than Middle Ordovician and, in the writer's opinion, more likely Silurian or conceivably Lower Devonian.

Lithological Detail

The rocks have been metamorphosed to crystalline types. The most abundant of them are phyllites which are interbedded with smaller amounts of schists, limestones, and quartzites. These dominantly siliceous rocks are the southeasternmost in a series becoming progressively more siliceous in this direction. Single beds range in thickness from less than an inch to uncommonly 4 or 5 feet. Bedding is generally well defined, but where the phyllites are successively thick in section and dip at low angles, the bedding is determined with difficulty because interbeds of a different lithology are rare. In general the brown weathered surfaces of the Barton River formation are not so conspicuous in this formation.

Phyllites—These rocks are on the whole comparatively of lighter color and more arenaceous than the same rocks in the Barton River formation. They are light- to medium-gray and fine- to medium-grained. Cleavage is absent or very poorly developed in the sandy phyllites, except perhaps where it coincides with the bedding, as in the whetstone types in which the perfect planes of parting, considered to be bedding planes, fairly glisten with mica. Some of the beds are finely laminated.

Varying mixtures of fine-grained quartz and mica are the most essential minerals present. In the different specimens examined, sericite and biotite exceed one another in amounts, but in some, quartz is present far in excess of either mica, practically conforming in composition to a quartzite. The uniform, fine grain and shining luster distinguish them as phyllites. In some the biotite crystals are lath-shaped and linearly arranged with the long axis parallel to the banding in the rock, while in others the biotite appears as porphyroblasts in part altered to chlorite. Muscovite and sericite occur as small disseminations and the former as fairly large crystals in random orientation and often corroded.

Tourmalines, both dark-brown and olive-colored, are well represented as patches and as isolated crystals, sometimes elongated parallel to the banding. Minerals in smaller amounts are zircon, chlorite, garnet, calcite up to 10 per cent, magnetite, sulphides, possibly talc, hornblende, and titanite. One section is cut by quartz veins and in another a vein is composed of quartz, calcite, and sericite or talc. Carbonaceous matter and sulphides occur in small amounts.

A microscopic study of a specimen of whetstone quality from the type locality, shows abundant sericite, most of which is oriented parallel to the structure of the rock, associated with quartz. Accessory minerals present are feldspar, irregular brown tourmaline grains, lath-shaped biotite, apatite and a little chlorite.

Limestones—The limestones are fine- to medium-grained, rarely coarse-grained, and range from light-gray, sometimes mottled, to black. They are sometimes schistose or phyllitic, and in all cases react vigorously to acid. Calcite is present from about 20 per cent to considerable amounts. In contrast to the phyllites, the limestones contain more carbonaceous matter. Like the limestones in the formations already described, they are considerably altered by metamorphism, resulting in the production of such new minerals as diopside at igneous contacts, clinozoisite or vesuvianite, orthoclase and microcline, titanite, epidote, and amphibole. In one slide large porphyroblasts of amphibole are considerably corroded and broken, and in part altered to chlorite and epidote. Clinozoisite is occasionally interstitial to the other minerals. Numerous corroded crystals of an olive-green biotite and rather large, irregular patches of dark-brown tourmalines, are present in one section.

Quartzites—Some of the quartzites are banded, with such minerals as rectangular biotites and needles of tourmaline parallel to the banding. A few calcareous quartzites contain about 10 per cent of calcite grains. The micas are represented by small to moderate amounts of biotite, chlorite, and muscovite with intergrown quartz. In some rocks rutile needles are numerous, and there are minor quantities of apatite, zircon, a clay mineral, magnetite, kaolinized orthoclase, garnets and titanite partially altered to leucoxene. One specimen shows large porphyroblasts of a colorless garnet containing about 50 per cent of quartz in a poikilitic texture. These suites of minerals produced by metamorphism show that the quartzites were originally impure. An occasional bed shows cross-bedding.

Schists—The schists are fine- to medium-grained and generally exhibit a well-developed schistosity. They are light-gray to black, often displaying a high luster due to the micas present. The kinds of schists are quartz-mica, garnet-mica, amphibole, garnet-staurolite and staurolite, all containing quartz and mica as the chief minerals. The schistose structure is largely due to the parallel orientation of the micas, and often to the porphyroblasts as well.

The quartz-mica schists frequently contain muscovite plates and crystals, in addition to sericite flakes and augen of biotite. The augen are arranged with their long axes parallel to the schistosity and are composed of separate crystals of biotite and in a few cases small amounts of muscovite (Pl. 7, fig. 1). In some the rock is crowded with biotite crystals in random orientation.

Considerable carbonaceous matter darkens the garnet-mica schist. The garnets are large, colorless porphyroblasts containing numerous inclusions. Also present are some large, much-corroded, yellow-green amphibole porphyroblasts (Pl. 7, fig. 2). In addition are many random oriented biotite crystals, some of them wholly or partially altered to chlorite. The amphibole schists carry large, yellow-green amphiboles more or less corroded or replaced. In one specimen the amphibole is considerably replaced by a mixture of quartz, calcite, epidote and clinozoisite.

In the staurolite schists the staurolite is often shattered and replaced by quartz, and occasionally appears as remnants where it is strongly corroded. It occurs in euhedral prisms and as cross sections, oriented and unoriented with regard to the schistosity (Pl. 8, fig. 1). It frequently contains carbon impurities and in some porphyroblasts the carbon along with the sulphide inclusions show a linear parallelism corresponding in direction to that of the groundmass, regardless of the orientation of the porphyroblasts. Garnets and sulphides are found surrounded by bleached zones of quartz, muscovite and biotite (Pl. 8, fig. 2). Quartz and carbon appear in some garnets possessing a pinkish tinge. Tourmaline is common to all schists, as are also quantities of titanite, rutile and zircon.

IRRUPTIVE ROCKS

General Statement

The largest intrusions are plutons of granite which forced their way into the strata by pushing them aside. These granites have also invaded an earlier dark, hornblendic granite. With the exception of the northeastern portion of the area, granite sills and dikes fairly abound. They are especially numerous in the southwestern and eastern portions of the area. Pegmatites and aplites are common in the granites, occasionally extending into the sediments. Among the multitudinous quartz veins, the huge veins of milky quartz which follow the general regional strike, deserve mention. The resistant granite masses support the dominant elevations and the sediments occupy the lowlands generally.

In the northwest corner of the Memphremagog quadrangle, metabasic rocks have been studied in detail and indicate a relationship to rocks of similar composition in the Owl Head Mountain region to the north in Canada. The igneous rocks are younger than the sediments and are believed by the writer to have been intruded sometime during the Acadian orogeny.

The Bolton Igneous Group and Associated Granites

An area of intrusive metabasic rocks borders the west side of Lake Memphremagog north of Holbrook Bay. These rocks are associated with granites, the granites becoming much more predominant south of Holbrook Bay. The rock types indicate that the area is largely in a transition zone between basic rocks to the north in Canada, and granites to the south. They consist of fine- to medium-grained metadiorites and related rocks, and more or less contaminated granites, some of the latter in places exhibiting porphyritic textures in which the feldspars are as much as 12 mm. long.

Northward in Canada these rocks merge with the metagabbros and metabasalts of Clark and Fairbairn (1936), called by them the Bolton igneous group. The map by Clark and Fairbairn (idem, p. 15) indicates a southward extension of the Bolton igneous group across the International Boundary and that by Ambrose (1943) shows its continuation southeasterly along the lakeshore in the direction of the area of metadiorites. In addition, a petrographic sequence from the metagabbros in Canada through the metadiorites to the granites on the south, is suggestive of the consanguinity of these rocks. Because of these field relationships, the writer has adopted the term Bolton igneous group for this strip of basic rocks.

The diorites and related rocks are gray to dark green. Thin section study of the diorites shows pale green to greenish-yellow amphiboles of varying intensities of pleochroism, plagioclase, and biotite as predominant minerals. The amphiboles are somewhat shattered and chloritized. Some sections contain smaller amounts of orthoclase in addition to plagioclase. The plagioclase grains display the customary twinning, are anhedral to subhedral, and are often greatly altered and saussuritized. In some cases zoisite and tremolite are prominent among the alteration products. The biotites are commonly altered to chlorite, especially in the highly chloritic, schistose rocks and they range from pale green to greenish-brown. A great deal of secondary carbonate is present in some of the rocks and large masses of twinned calcite as well, the latter often occurring as amygdules. Sulphides are present in varying quantities, having considerable representation in both the acid and basic rocks and ranging from dust particles, distributed largely at mineral boundaries and along cleavages, to rather large individuals. Minerals in minor amounts generally are quartz, muscovite, leucoxene, titanite, apatite, zircon, and epidote. Epidote occurs also as thin veins.

The granites have been altered to varying degrees by the invading basic rocks, as is shown by their compositions. They are light- to dark-gray, sometimes having a greenish cast. In texture they are fine to coarse, and commonly porphyritic. Besides the essential minerals of granites which are of themselves present in varying amounts, are plagioclase, carbonate, greenish-brown biotite, muscovite, chlorite, apatite, titanite, epidote, and sulphides. Some of the rocks contain considerable secondary carbonate, probably as an alteration product of the plagioclase. The feldspars show much alteration to secondary products and are somewhat shattered as well. The potash feldspars are often kaolinized, and quartz contains numerous inclusions. Considered as a group these rocks represent hybrid types for the most part, as might be expected in an igneous contact area. Purer granites are found directly south of this area.

The structural relationships and textural differences show definitely that the basic rocks are intrusive into the granites. That the basic rocks are younger than the granites, is revealed conclusively by dikes of the former penetrating the granites and containing granite xenoliths (Pl. 9, fig. 1). These dikes attain widths of more than 16 feet, and are known to extend outward from the main mass of basic rock and to strike in the general direction of the Owl Head Mountain region to the north in Canada. Schistose drag at the borders of some of them is oriented so as to confirm the suggested direction of intrusion. The dikes display contact effects such as chill zones and amygdaloidal bands 5 inches wide in places. The amygdules, filled with calcite and light-colored feldspathic material, appear to be due to the escape of gases at the time of intrusion of the dike. Associated with the feldspathic amygdules in places are extremely thin to thread-like veinlets of the same material. Some of the veinlets ramify, but none of them extend to the granite contact. Since they are associated with the feldspar amygdules and have the same composition, it is concluded that they originated at the same time and in the same manner as the amygdules.

The contacts between the basic and acidic rocks are well defined, due to strong textural and erosional contrasts. The granite wall rock is coarse to porphyritic, with feldspars half an inch long and smaller quartz insets. At the contact with the basic rock it stands out in sharp relief and has a decided greenish color because of exomorphic effects.

The granites are locally brecciated and contain fragments which are both angular and subangular. The fragments range in size from a few inches to huge inclusions. Good examples of the brecciated granite may be seen in a pasture bordering the Lake road on the east and half a mile due south of Maxfield Light. Here two enormous granite inclusions measure 60 feet by 10 feet and 45 feet by 20 feet, respectively. One inclusion contains an apophysis of the basic rock matrix. Granite xenoliths are also well displayed in the exposure on the lakeshore at Maxfield Light.

In places the basic rocks are traversed by ridges of resistant material which follow fracture systems. The ridges are from a fraction of an inch to 4 inches high, and are composed of basic material with injected siliceous seams running lengthwise of the ridge. In following the fracture systems they enclose triangular, rectangular and rhombic polygons of various dimensions, thus giving the surface of the outcrop a reticular pattern (Pl. 9, fig. 2). In general the major directions of the ridges are north-south and east-west, with minor variations. Those trending north-south show east-west displacements of an inch to as much as 4 inches. No displacements were seen in the north-south direction. The intricate fracturing of the basic rocks and the siliceous injections along these fractures, appear to be late effects in the intrusion of the basic magma.

From the foregoing paragraphs it can be seen that the Bolton igneous group is younger than the granites. Nowhere have these rocks been observed in contact with sediments younger than the Cram Hill. However, the granites contain infolded slates in places and farther to the south the Lowell Mountain pluton has intruded the Cram Hill, Shaw Mountain, Northfield, and Ayers Cliff formations, sending dikes out into the lower members of the Ayers Cliff formation. Because the Shaw Mountain formation is at least Ordovician in age (Currier and Jahns, 1941, p. 1500), the granites must be younger. According to Clark (1934, p. 13) the Stanstead granite, which is the age equivalent of the granites in this area, is Devonian. The writer concurs in this view and, since the Bolton igneous group is intrusive into these granites, he is forced to the conclusion that the Bolton igneous group is at least Devonian in age. Cooke (1948, p. 17) has reached the same conclusion with regard to the age of the Bolton lavas, which Clark (1936) believes to be the extrusive equivalents of the basic intrusives.

Structural features already described in both the Bolton igneous group and the granites, suggest that the former was intruded during a period of orogeny. Petrographical studies show further evidence of the activity of distortional forces in bent, shattered, and intensely folded mineral grains. Supported by the age relations already determined, the induced structures are probably indicative of the Acadian orogeny.

Granite Plutons

The granites constitute several varieties and are distributed widely in the area. They occur as large bodies covering considerable portions of the central and southern rectangles. Smaller exposures occur near and at the eastern and northern margins of the map. Some of these are continuous with larger bodies on the adjacent quadrangles. The domical elevations, such as Salem, Hopkinson, Sugar, and Elm hills in the central igneous body, and Barton Mountain in the southern part of the quadrangle, are surface expressions of the granites.

By and large the granites are medium- to coarse-grained, granitoid rocks and light to dark gray depending upon the amount of biotite and hornblende present. Dark hornblende granite occurs in two places at the border of the arcuate body extending northward from West Charleston. Granites containing abundant biotite are locally prominent, especially in the masses in the eastern part of the map-area. In places the granites are porphyritic, containing feldspar phenocrysts as large as 1 centimeter.

A medium-grained, light gray, equi-granular, binary granite is quarried at the Willey Granite Quarry northwest of Derby Center. A microscropic study of this granite reveals as major minerals orthoclase, microcline, quartz, muscovite, and biotite. A few grains of titanite and apatite, and some zircon needles occur as accessory minerals. The quartz contains numerous inclusions. Some of the feldspars are kaolinized and there is a small amount of chlorite, probably as an alteration of the biotite.

The mineralogic composition of a specimen from Sugar Hill consists of orthoclase, microcline, oligoclase, quartz, and biotite. The biotites are large and contain apatite and zircon inclusions, the latter surrounded by pleochroic haloes and frequently elongated with the cleavage in the biotite. Much of the biotite "frames" many of the feldspars producing a pattern resembling mortar structure. An occasional plagioclase is zoned. Minerals in lesser amounts are muscovite, garnet, tourmaline, titanite, and some carbonate. The muscovite is associated with quartz. A few grains of quartz possess myrmekitic intergrowths. Some of the biotites show alteration to chlorite and many of the feldspars are partially kaolinized. In a hand specimen the lighter feldspars stand out prominently against the darker background of quartz and biotite, thus giving the rock a slightly porphyritic effect.

A specimen of granite from the south shore of Echo Pond is very similar in general appearance to the Willey Quarry granite, except that the quartz grains are much more conspicuous. In thin section it shows orthoclase, microcline, oligoclase, quartz, and biotite, with smaller amounts of muscovite, apatite, zircon, tourmaline, and sulphide. The biotite encloses both apatite and zircon, and is occasionally associated with sulphide, which appears to be replacing it. Some of the plagioclase exhibits zonal structure. There is a little quartz with graphic intergrowths. Kaolin is present as a coating on some of the feldspars.

About a mile southeast of the village of West Charleston, at a road junction, is an exposure of a dark, coarse-grained, mafic granite composed of quartz, microcline, a little sodic plagioclase (oligoclase), biotite, pale yellow and light green, pleochroic hornblende, a few zircon crystals, and a little magnetite. The biotite is partially altered to hornblende and contains streaks of magnetite often following the cleavage of the biotite. Magnetite streaks also follow the cleavage directions of the replacing hornblende. The hornblende attains a length of 1 centimeter.

Orbicular granite occurs on the Hitchins farm on the northeast slope of Prospect Hill in Brownington Village. This locality was mentioned by E. C. Jacobs (1921–1922, pp. 107–108) and very probably is the locality referred to by S. R. Hall (1871, pp. 74–75). The nodules are spheroidal with average dimensions of about 2.5 cm. x 1.5 cm. They are composed largely of black, lustrous, crinkly biotite which envelopes grains of feldspar and quartz. The granite is light gray, medium-grained, and contains as prominent minerals quartz, orthoclase, sodic feldspars, muscovite, and biotite. The micas tend to surround the feldspar and quartz grains in a manner similar to that described above in the specimen from Sugar Hill. This mineral arrangement is suggestive of the early stages in the development of the nodular forms in the continued thickening or growth of the biotite around one, or several, of the feldspar and quartz particles. Areas of corrugated biotite studded with feldspar and quartz grains occupy the more basic portions of the granite. Orbicular granites have been reported from other localities (E. Hitchcock, 1861, pp. 563–565; Adams and Barlow, 1910, pp. 127–139). The latter discusses the origin of the orbicules. The literature on orbicular granite can not be gone into here (see citations given).

Sedimentary Inclusions

Remnants of the sedimentary rocks are numerous as inclusions in the igneous bodies. Their structures in most cases do not conform to those of the region, but in the central granite mass a large majority of recorded measurements show that the structural orientation of the inclusions is conformable with the long axis of the mass as well as with the structure of the sediments along the southwest and northeast contacts. Although the tectonics of the granites have not yet been studied in detail, it is believed that this structural conformity might possibly be correlated with the flow direction in the granite.

With few exceptions, the bedding of the inclusions has gentle to moderate dips. In places, small areas of sediments lie in shallow synclinal contacts with the granite; in these the bedding is, generally, also synclinal (Pl. 9, fig. 3). Some of these sedimentary remnants, still retaining a structural conformity with the regional rocks, might conceivably be preserved portions of the original roof of the igneous body.

The inclusions possess distinct outlines and are lens-shaped and tabular generally, the shapes being determined largely by the bedded structure (Pl. 9, fig. 4). On the west slope of Sugar Hill near the Brownington-Derby Center road, interbedded limestone and phyllite at several levels, dip beneath the granite cliff at relatively low angles (Pl. 10, fig. 1).

The inclusions show well the effects, both structurally and mineralogically, of the intrusions on the sedimentary rocks. The limestones frequently possess a conspicuous flow structure, while the more resistant beds yielded by breaking, bending, and faulting. Shallow folds, some of them overturned, are not uncommon. At the actual contact contaminated, highly micaceous granite or "granitized" sediments exhibit pronounced flow structure, filling in the spaces between ruptured sedimentary layers (Pl. 10, fig. 2). Occasional smaller inclusions are hollow shells in which the less resistant interior has been weathered out, the dense, hard shells attesting to the efficacy of the intrusive (Pl. 10, fig. 3). Saucer-like depressions bordered by hard, prominent rims, ordinarily half an inch thick, compose the surface of many of them.

Granite sills in lit-par-lit fashion occur in patches of the sedimentary rocks having steep dips. Like the sedimentary beds beside them, they thicken and thin and are drawn apart along the strike, producing a boudinage-like structure. Disjointed beds, often displaced across the strike and tilted at varying steep angles, are prominent features in the foliated groundmass of the contaminated granite. Besides disclosing the former extent of the sedimentary rocks the inclusions, by their structures and to a certain extent orientations, may ultimately shed light on the manner in which the granite came into being.

The forceful intrusive nature of the magma is also shown by the structures in the wall rock bordering the central igneous mass, which they parallel. Furthermore, the inward dips of the country rock around the perimeter of this body, suggest a funnel-shaped intrusive. In a limited number of places at the edge of the intrusion, foliation was discernible, but over its areal extent in general the rock appears massive. Additional evidence of the pervasiveness of the intruding magma is again shown by the central igneous mass whose expansive force has deflected the regional northeast strike of the country rock, resulting in bulges in the sediments toward the northwest and southeast (Plate 1).

The sediments flanking the intrusives in the southern and southeastern sections of the map-area, dip radially outward. While the main part of the plutonic body southwest of Orleans is concordant, its eastward extension appears discordant, as revealed by its structural relationship to the bordering sediments. With this possible exception, these bodies are conformable and laccolithic in behavior. The narrow, concordant body longitudinal between West Charleston and Cargill Hill, has the structural attributes of a sill, or possibly a phacolith.

Age of the Granites

An earlier dark, biotite and, in places, hornblende granite, has been invaded by a light gray, coarser granite. Apophyses of the younger granite traverse the dark granite in various directions and inclusions of the earlier occur in the later granite (Pl. 10, fig. 4). On the west slope of the 1600-foot hill southeast of West Charleston, an intrusive breccia contains rounded and angular fragments of the dark phase (Pl. 11, fig. 1).

Both granites intrude the sedimentary rocks, and are therefore younger than the latter. The youngest sediments constitute the Westmore formation, which is considered by the writer to be Middle Silurian or possibly Lower Devonian in age (Doll, 1943). On the basis of the above relations a late Devonian age is favored for the granites. Such an age assignment is in accordance with the determinations made by earlier investigators.

Sills and Dikes

Sills are much more widespread than dikes. They are parallel to the bedding and foliation of the sedimentary rocks, but in an occasional vertical or horizontal section they have been observed to cut the structures with a slight obliquity. They are common in all of the formations, but especially abundant in the Barton River formation. Great concentrations of them, such as occur in the southwestern part of the quadrangle west of the Barton River Valley, might be indicative of their close proximity to the underlying granite mass. Locally, sills follow in rapid succession across the strike and, in a few instances, a sill continued as thinning ramifications interfingering with the sediments.

The great preponderance of sills over dikes throughout the area might be explained structurally. The strata are relatively thin and steeply dipping, both conditions being highly favorable to the formation of sills. Individual transverse fractures, which might accommodate dikes, are rarely continuous, due largely to the inhomogeneity and high inclination of the beds. In contrast, the steep bedding planes and accompanying cleavage have considerable continuity in depth and longitudinally as well, thus affording readily accessible openings. Barrell (1902, p. 294) calls attention to inclined strata as avenues of escape in a discussion of infiltration of strata. This same structural weakness is here extended to include invasion of strata. Excepting for the property of homogeneity, the situation is analogous to the up-ended layers in ice-packs.

The sills range in thickness from 8 inches to 33 feet and are frequently traceable for considerable distances along the strike. One sill 16 feet thick was traced for more than 1800 feet along the strike, and another 14 feet thick, for 1000 feet.

Prominent joints across the strikes of both sills and dikes are common. Many of them contain quartz veins, while others served as surfaces of faulting along which successive segments were given *en echelon* arrangements. In a few cases the segments were drawn apart and rotated (Fig. 2). An occasional dike is bordered by a fault contact along which the sediments exhibit drag and displacements of a few inches to several feet. Sediments at sill contacts are often strongly contorted and the sills infolded with the sediments (Pl. 11, fig. 2).

Neither the sills nor the dikes were studied petrographically, but those observed in the field were essentially granitic in composition, ranging from fine to coarse in texture, sometimes porphyritic. Some of them become aplitic along their extensions. The contact zones in some

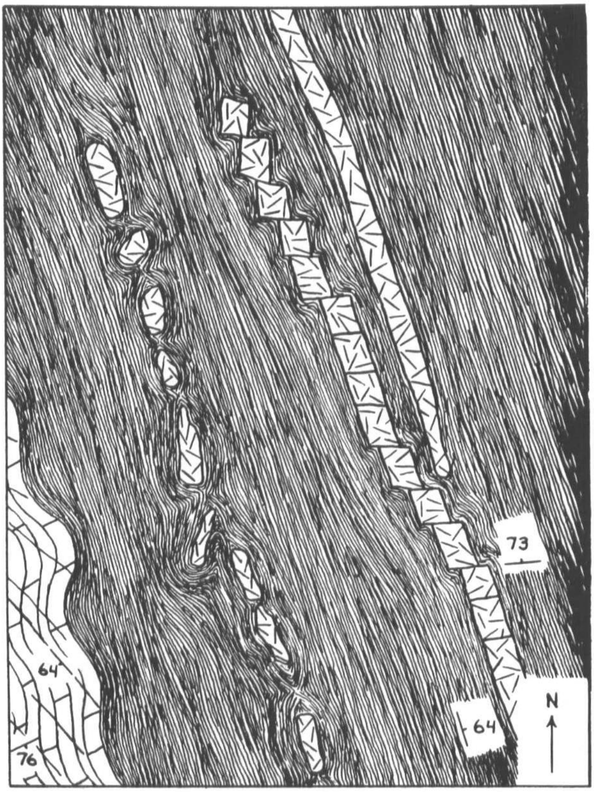

Figure 2. At Cobb Brook on west side of Pine Hill road. Three stages in disruption of granite streamers in slates of the Barton River formation. Disruption of one streamer has advanced to the boudinage stage. The sketch covers a width of 8 feet.

are chilled, in others pegmatitic. Conspicuous minerals are feldspars, quartz, muscovite, biotite, and black tourmaline. The tourmaline occurs in small dendriform patterns. In a few sills the prominent minerals are parallel to the adjacent sedimentary structures. Likewise in places in the sediments a copious development of coarse muscovites shows a linear parallelism with the contact. Slate inclusions are infrequent in the sills and dikes.

Only one basic dike was encountered in the field cutting the sediments. It crosses the road about half a mile north of the "T" in the township word "WESTMORE." It is 3½ feet thick and strikes N 52° W.

The sills and dikes are presumably comagmatic with the granite plutons, and since they also cut the sediments they are considered younger than the latter. Structural evidence indicates that some of the sills were affected by the orogeny which folded the regional rocks, thus tending to show that they are older than the orogeny. The writer, however, believes that the sills were emplaced at intervals during the late Devonian.

Pegmatites

Pegmatites are associated with the granite as dikes and irregular masses. They occur also in the sediments as dikes and sills, but are much more numerous within and nearest the igneous bodies. The pegmatitic areas are gradational into the granite and occasionally send off apophyses. They represent segregation centers of residual magmatic liquids. The pegmatites contain as prominent minerals quartz, potash feldspars (cleavelandite in some), muscovite, black tourmaline and garnet. In some dikes the tourmaline and garnet are concentrated at the granite contact, the tourmaline attaining a length of one inch. The feldspars reach lengths of 6 inches and the muscovites one and one-half inches. Generally there are no visible contact effects at the borders of the sedi-

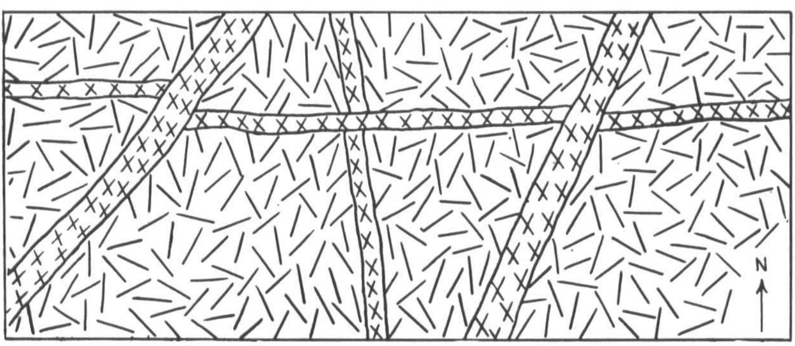

Figure 3. Southeast slope of Barton Mountain on north side of Barton-Lake Willoughby highway, 1¼ miles west of elevation 1404. Three generations of pegmatite veins in granite. The sketch covers a width of 5½ feet.

mentary rocks, but some of the limestones exhibit blanched contacts an inch or more wide.

In the granite the pegmatites were injected along tension joints, certain sets of which show displacements of small magnitudes, usually measurable in inches (Pl. 11, fig. 3). These displacements probably indicate expansion of the igneous mass while the magma was still actively rising. Three generations of pegmatites have been noted in the granites, an exposure of which may be seen beside the road between Barton Village and Lake Willoughby on the southeast side of Barton Mountain (Fig. 3). The later pegmatites appear to be coarser grained than the earlier, and preceding the pegmatite injections are narrow intrusions of granite (Pl. 11, fig. 3). Some pegmatite dikes continue as aplites along the strike.

The pegmatites are considered to be genetically related to the granites and are therefore presumably of practically the same age as the granites.

Quartz Veins

Countless numbers of quartz veins traverse all of the rocks in the area. There are several generations of them. They follow the bedding, fractures, cleavage, and schistosity, and in many instances exhibit "book structure" in the fissile rocks. Many of the veins are drag-folded along with the sedimentary rocks, while others cut across the folds. Cross-cutting veins may be restricted to a single stratum or they may traverse a series of strata. Not infrequently are veins faulted on a small scale, usually in a succession of step faults. Quartz also occurs as veins between the walls of small faults, often showing slickensided surfaces.

Outstanding are the thick milky quartz veins of considerable extent along the bedding of the sedimentary rocks. In some cases they attain thicknesses of more than 10 feet, and can be traced along the strike for distances exceeding 150 feet. One vein located on a hillside just off the main highway to Lake Willoughby, and half a mile northeast of the East Brownington School, has a visible width of 40 feet. Another vein about 10 feet thick is exposed for 145 feet along the strike and 14 feet down the dip, in a field on the west side of the Barton River Valley a mile and a half south of Coventry Station (Pl. 11, fig. 4). These veins occur in the siliceous and calcareous sediments alike, and are most abundant in the Barton River formation.

Veins occurring in the buff gray limestone are in places very irregular in outline and are frequently broken into lens-like segments. The hackly borders and discontinuity of the veins along the strike are probably due to the dynamic effect during plastic flow of the limestone; plastic flow is a conspicuous feature of the limestones. The quartz veins in the siliceous sediments generally possess well-defined sharp and smooth borders.

Many of the quartz veins are not pure, but carry such minerals as calcite, mica, or less commonly feldspar. These minerals usually show good crystallization and fair size. Occasionally iron sulphides are present. Hackly-fringed quartz veins in the limestones sometimes contain muscovite scales and frequently irregular clusters of ironstained calcite. More rarely do quartz veins contain feldspar, in which case they approach pegmatite in composition and may even grade into the latter.

Not only are the quartz veins folded conformably with the sediments, but they also show ptygmatic folding. Good examples of ptygmatic folds occur in the limestones possessing flow structure (Fig. 4). As thin slate bands in the same exposure also exhibit ptygmatic folds (Pl. 17, fig. 1), it is the writer's opinion that the quartz veins were injected prior to the movements producing this type of fold. Ptygmatic folding occurs in rocks that have undergone plastic flowage during regional deformation and igneous intrusion. Thick lenticular masses of quartz occasionally occur as bulges around which the strata bend and thin (Pl. 12, fig. 1). Several of the quartz veins carry inclusions of slate or limestone.

The observations cited above show that the great majority of the quartz veins and masses are definitely related to the processes which folded the sedimentary rocks, and some may have antedated the orogeny. Quartz veins also cut the pegmatites, thus revealing late injections. Quartz in all of its forms may be considered then to represent several ages.

STRUCTURE

General Statement

The rocks of the Memphremagog area are strongly folded and faulted, and the whole further complicated by the injection of igneous masses. A striking feature is the excessive amount of overturn to the southeast throughout the whole stratigraphic section northwest of the axial region of the Brownington syncline, the degree of overturn increasing toward the southeast. The dominant structural elements from northwest to southeast are: the Ware Brook thrust and associated Stony Hill and Lake Memphremagog reentrants, the Indian Point syncline, the Black River fault, the North Neighborhood anticline, and the Brownington syncline.

The rocks show an abundance of minor structures which reveal the great intensity of the orogeny. With the exception of the northwest-trending reentrants, the prevailing direction of the folds and faults is northeast-southwest. The structure-building movements were generated during late Devonian time and, thus, were practically synchronous with the Acadian orogeny.

Indian Point Syncline

The Indian Point syncline is located in the extreme western part of the area. It extends northeastward from the vicinity of Lowell Mountain through the village of Coventry and the city of Newport into Canada, a distance of 16 miles. Between Lindsay Beach and a point 1¼ miles southwest of Beebe Plain, the fold is concealed by a heavy cover of drift. The syncline is delineated best where the lower Barton River formation occupies its core, while south of Coventry Village the scarcity of exposures in the underlying Ayers Cliff formation has greatly obscured the structure. The axial plane of the fold strikes northeastward and dips moderately to the northwest.

The synclinal axis is convex toward the northwest, which appears to be due to the expansive effects of the central igneous mass not far to the southeast. That the fold is greatly overturned to the southeast, the Indian Point segment excepted, is indicated by the extreme easterly position of its axis which is located in the escarpment of lower Barton River slates along the Black River immediately south of Newport City. The position of the axis indicates an asymmetrical fold whose east limb is either considerably thinned or cut out in part by a fault. An appreciable width of the west limb is concealed by the Ware Brook thrust.

Parallelism of the bedding and axial planes of the minor folds, shows the syncline to be a tight fold. The minor folds plunge from 13° to 45° NE, indicating a northeasterly plunge of the major syncline. Bedding and cleavage relationships are difficult to determine generally in the slates of the Black River escarpment, as the former is more often absent. Occasional beds of different lithologies make possible determinations of these relationships, and where they have been observed they show a succession of overturned beds with tops to the east. Vertical transverse fractures, indicating elongation, are common as prominent smooth faces.

Indian Point Segment

An abrupt reversal of dips on Indian Point suggests a sharp twist or a transverse break in the Indian Point syncline where it crosses Lake

Memphremagog. The intensity of the minor structures and cleavage-bedding relations indicate a strong westward overturn approaching recumbency in which the east limb has been largely stripped by erosion. No lateral displacement is apparent, and so it is believed that the contrary dips originated when the structures were rotated by the deep-seated effects of the central igneous intrusion on the southeast. The line separating these two areas of opposite dips runs south-southeasterly through Newport City and South Bay, and appears to be the southeasterly extension of the major shear zone in the Lake Memphremagog reentrant. Between Indian Point and the 1080-foot hill southwest of Beebe Plain the fold is concealed by drift cover, but from this hill it can be followed into Canada.

Brownington Syncline

The Brownington syncline is a broad, flat, practically recumbent fold with a strong overturn to the southeast. It covers approximately the eastern three-fifths of the map-area, bordering the North Neighborhood anticline on the west and extending eastward into the Island Pond quadrangle. Its axial plane strikes northeast and is strongly overturned to the southeast, thus dipping gently to the northwest. The fold is characterized by a steep, inverted northwest limb in contrast to a southeast limb with gentle inclination. Exposures typical of the west limb structural relations are amply displayed at Coventry Station (Pl. 16, fig. 1). On a traverse southeasterly from Coventry Station the beds of the Barton River and Westmore formations show pronounced overturn to the southeast and on the long northwesterly slope of Goodwin Mountain they are gently inclined and in normal sequence. Beds on the east limb are exposed on the open slopes directly east of the East Brownington School and in the abandoned scythestone quarry three-fourths of a mile to the southeast and indicated on the map.

The dips of the beds become progressively less toward the southeast, which appears to indicate that the brunt of the regional movements building the structures was borne by the rocks in the western portion of the area where they are broken by faults. The plunges of the minor folds in the Barton River and Westmore formations indicate a general plunge of 30° NE for the major fold. Locally the plunges are variable both in direction and amount due to tectonic effects, an occasional one on the steep northwest limb plunging as high as 80° NE. The axis of the fold lies in the proximity of East Brownington School. From here it trends northeasterly around the bulge of the plutons on the northwest and continues beyond the eastern border of the Memphremagog quadrangle where its position is somewhat obscured in an area of igneous rocks.

The fold is further complicated by overturn of beds to the northwest on the section of the west limb between Pine Hill and Johns River. The dips are noticeably steep, most of them being not far from vertical. The anomalous structures in this sector are probably the result of rotation from original northwest dips during emplacement of the central igneous pluton whose northwest side they flank. Movement along the Lake Memphremagog reentrant could also have had a part in reversing the attitudes of these structures. The fold is punctured by several large-sized igneous bodies which have distorted its symmetry and complicated the structures.

North Neighborhood Anticline

The presence of this fold is inferred largely from its narrowing outcrop belt northeastward between the Indian Point syncline on the northwest and the Brownington syncline on the southeast. The North Neighborhood anticline is indicated by bedding-cleavage relations in the adjacent Barton River formation at Coventry Station (Pl. 16, fig. 1). It plunges to the northeast, with its axial plane steeply inclined to the northwest. With one exception, the extremely small number of exposures are located well out on the flanks and, although possessing dips not far from the vertical, they show an overturn to the southeast. The exception is a single outcrop of Ayers Cliff limestone at the North Neighborhood School, which, on the basis of its location and structural attitude, is suggestive of its position in the axial region of the fold. Its much gentler dip in comparison to those on the flanks, tends to support this opinion and, in addition, might be indicative of the fold's plunge. The northwest limb of the fold is cut out by the Black River fault.

Ware Brook Thrust

This fault is the southward continuation of the Bunker thrust in Canada and comprises the boundary between the Northfield slate on the west and the Ayers Cliff formation on the east. Its strike is northeast, with relative movement from the northwest. It follows the base of an east-facing escarpment which gradually loses its topographic expression northward and disappears under the waters of Lake Memphremagog (Plate 1).

In the Ware Brook section the fault plane has a low southeasterly dip and is exposed by erosion in a window at an elevation of about 975 feet. Here the minimum amount of horizontal displacement is 4000 feet, which increases greatly northward. The thrust cuts across the limbs of folds and, therefore, is believed to have developed after the folded structures.

Throughout its length the fault is not easily recognized because of the apparent gradation between the formations bordering it and, in addition, its general concealment beneath a heavy spread of drift. Indeed, were it not for evidence elsewhere than along its trace, this fault might have remained unnoticed. In central Vermont the Northfield-Ayers Cliff (Waits River) contact is said to be conformable (Currier and Jahns, 1941, p. 1492).

The existence of the Ware Brook thrust is shown by the following criteria:

(1) The outcrop breadth of the Ayers Cliff formation in Canada is as much as twice that in the area mapped. This difference probably is not a primary phenomenon but tectonic. That the Ayers Cliff formation possesses the same degree of structural disturbance everywhere it is exposed negates any possible interpretation that the formation in the area mapped was excessively thinned in comparison to its outcrop in Canada. Furthermore, it has been observed that the Ayers Cliff formation diminishes noticeably in breadth of outcrop southward where it passes successively the northwest-extending reentrants (discussed separately). Additional evidence suggesting a structural explanation for the apparent variations in outcrop width, is a window of Ayers Cliff limestone well within the Cram Hill formation.

(2) About half way in the section at a waterfall in the bed of Ware Brook, a window of thin-bedded Ayers Cliff limestones is "framed" by black, carbonaceous slates. This same horizon has been located in an exposure of slates just off the east side of State Highway No. 105, north-northeast of Walker Pond. In Canada it is found to the east of the thrust in a road cut in the Ayers Cliff formation near the upper end of Fitch Bay and, again, in the prominent limestone cliffs at the southeast end of Lake Massawippi.

This horizon is readily detectable, once it is recognized, because of uniformity in thickness, striking lithologic similarity in the known exposures and its stratigraphic position relative to the fault. In the Ware Brook and Fitch Bay exposures the resemblances are even more marked by small lenses of calcite common to both and identical. A detailed study of a lens from the Ware Brook locality has revealed strongly twinned, dark calcite with minor amounts of quartz grains and, in addition, greatly strained and recrystallized organic structures. The fossil affinities were determined by comparisons with a series of thin sections showing several stages in the defacement of brachiopod fragments by metamorphism. This sequence, of which the Ware Brook specimens show the most advanced stage, is a part of a slide collection in the Department of Geology at Columbia University.

(3) In the northwest corner of the Memphremagog quadrangle the greenstones adjoin the Ayers Cliff formation while along its entire length south of Lake Memphremagog and into the Irasburg quadrangle, the Northfield slate is close to the fault. Reconnaissance studies in the vicinity of Craftsbury Village (Hardwick quadrangle) appear to indicate that the Ayers Cliff formation is cut out by the southward extension of the Ware Brook thrust. It has also been observed that southwest of the village of Irasburg the Barton River formation rapidly converges upon the Ware Brook thrust. The Irasburg conglomerate, which is considered by the writer to be basal to the Barton River formation, is exposed on the steep slope on the east side of the Black River Valley directly opposite the village of Albany (Hardwick quadrangle) and less than three-fourths of a mile from the thrust. This discontinuity of formations along its strike points to the existence of the fault.

(4) On the summit and northeast side of Round Hill, 1½ miles southwest of the village of Irasburg, is an area of slates closely resembling the Northfield slate. These slates are underlain by limestones of the Ayers Cliff formation. This patch of presumably Northfield slate, which is surrounded by the younger limestones of the Ayers Cliff formation, is interpreted by the writer as a klippe.

(5) In the light of the preceding criteria, the escarpment which rises above the trace of the fault and gains prominence southward, is considered as additional evidence for the fault.

The Ware Brook thrust can not be older than Devonian, since it traverses rocks of lower Devonian age in the Dudswell region in Canada. It is younger than the folding, which is considered to be late Devonian, as it cuts across the folds. It involves the basic intrusives of the Bolton igneous group, of late Devonian age, at Holbrook Bay and northward on Lake Memphremagog. The writer, therefore, believes the Ware Brook thrust to be of late Devonian age, originating during the late Devonian orogeny but after the folding. It is interesting to note that the so-called

Devonian granites which are east of the fault in Canada, transect it south of the International Boundary. Likewise, the ultra basics, which occur to the west in the village of Lowell, gradually approach the fault northeastward and come quite close to it in Canada.

"En Echelon" Reentrants

A cursory inspection of the Hardwick, Irasburg, Memphremagog (U. S.), Memphremagog (Canada), Sherbrooke, Dudswell and Disraeli quadrangles discloses *en echelon* reentrants along the line of the Ware Brook-Bunker thrust. They trend northwesterly, becoming more northerly to the northeast. From southwest to northeast in the following summary they are:

Reentrant	*Linear extent of reentrants approx. miles*	*Distance between reentrants approx. miles*
Craftsbury (Hardwick quad.)	2	
		13
Stony Hill (Irasburg quad.)	2	
		13
Lake Memphremagog (Memphremagog quads., U. S. and Canada)	3½	
		16
Lake Massawippi (Memphremagog quad., Canada)	3½	
		30
Dudswell (Dudswell quad., Canada)	3	
		20
Lake Aylmer (Disraeli quad., Canada)	7	

The tabulation above shows that the lengths of the reentrants increase northeastward and the distances between them as well. These facts are significant since they suggest crustal displacements of progressively greater magnitudes in a northeastward direction. These structures are in the section of greatest curvature of a major tectonic arc which swings from a northerly course in Vermont to a northeasterly direction in Canada. Apparently it is the same arc recognized by Keith (1923a, pp. 313–314) and described by him as one of "four great westward salients" in the Appalachian system. This arc, here called the Vermont-Quebec arc, is probably structurally related to the Massachusetts arc in southeastern New England, described by Billings (1929, pp. 131–132). In contrast, the Massachusetts arc has a decidedly sharper bend, which appears to be due to its position well within the main structure and nearer the source of the compression.

It is generally agreed that the deforming forces producing the Appalachian structures were exerted from the southeast, as indicated mainly by folds overturned to the northwest and overthrusts with easterly dipping fault planes. However, in the area mapped the folds are overturned to the southeast, which presents a structural anomaly insofar as the long-accepted conception of Appalachian structural origins is concerned. Inasmuch as the sediments here are associated with large-scale intrusions their structural irregularities are probably explainable by local aberrations in the regional deformation. The Vermont-Quebec arc lies with a westward convexity between the pre-Cambrian Adirondack dome on the southwest and the Canadian shield on the north. From an examination of Keith's map (1923a, facing p. 309) it appears that the pre-Cambrian areas acted as buttresses at both ends of the arc while the Paleozoic rocks were being subjected to strong pressures from the southeast, thus resulting in the farthest westward advance between the stationary pre-Cambrian protuberances.

During the period of compression, relief within the arc seems to have assumed a fanlike pattern in *en echelon* reentrants. The change in trend of the reentrants from northwest in Vermont to a more northerly direction in Canada, appears to be due to major structural controls. These structural lines are betrayed topographically by the marked northwest and northeast orientations of many lake basins and river courses, Lake Memphremagog with its embayments and the St. Francis River system being especially good examples (Fig. 1). From their patterns, it can be seen that these structural lines of weakness comprise two systems with acute angle intersection facing southeast in Vermont and south in Canada. This is suggestive of stress application from these directions in conformance with the rule of conjugate shearing planes (Bucher, 1920) and, in addition, harmonizes with the direction of Appalachian pressures. Bucher (1933, pp. 245–248) calls attention to a conjugate system of shear planes in the Jura tectonic arc described by Heim. Just as in the Jura Mountains, the northeast side of the reentrant has moved northwestward relative to the southwest side (idem., p. 247). This is well illustrated by the Lake Memphremagog reentrant along which the rocks on the northeast side do not match those directly opposite on the southwest side. In Canada, Ambrose (1943) has mapped a fault which follows the valley of the Tomifobia River and heads into the Lake Massawippi reentrant. Cooke (1944) has indicated a fault in the Dudswell reentrant.

Divergences in the regional trend occur also in places other than those cited. Recent field work in the Jay Peak quadrangle in the northern Green Mountains, has disclosed a northwestward swing of 4 miles in the regional strike in the vicinity of Hazens Notch. This westward swerve has involved the Green Mountain anticlinal axis which is in turn shifted to the west. Cooke (1937, p. 23) has found similar structures of corresponding magnitudes in the Eastern Townships of Quebec. These instances show that northwest deviations in the structural trends are widespread in the region generally. In fact, related structures can be traced westward beyond the northern Green Mountains to the eastern part of the Champlain Valley, where a series of thrust slices involving mostly Cambrian rocks, has been pushed westward (Keith, 1923b). These structures lie within the scope of the Vermont-Quebec arc and, in this relation, it is noteworthy that structures of great complexity appear to be extensively developed in areas encompassed by these structural arcs or "salients," as Keith called them. The southernmost structural arc in the Appalachian system, which bends with a pronounced convexity westward into Tennessee, also embraces a terrain of highly complicated structures (Keith, 1907).

The location of the Monteregian Hills on the foreland of the Appalachian folds in the greatest bend of the Vermont-Quebec arc and with a linear arrangement to the west, is particularly significant (National Research Council, 1944; Dresser and Denis, 1944, p. 456, maps 703A and 704A; Adams, 1913, map facing p. 32). The pattern resembles similar structures in the Ouachita foreland, in which simple tension fractures on a regional scale were developed along structural lines of weakness, due to warping and longitudinal stretching. In the case of the Monteregian Hills, the tension fractures are conceived to have had a similar origin and, in addition, to have served as avenues of escape for the igneous materials.

Of significance is the presence of intrusives on the southwest sides of the reentrants (Plate 1). The intrusives in the several reentrants are unlike in composition, increasing in basicity northeastward. Considered individually their rock types are:

Reentrant	*Rock Type*
Stony Hill	Granite
Lake Memphremagog	Granite and metadiorite
Lake Massawippi	Serpentine[1]
Dudswell	Rhyolite and basic lavas[2]

[1]Ambrose, 1943 [2]Cooke, 1944

Laverdière (1935) also has mapped an intrusive on the southwest side of the Dudswell reentrant, but gives a composition differing somewhat from that of Cooke.

The part played by the intrusives in the faulting is not clear, but it is believed by the writer that the segments which they occupy might have acted as relatively resistant, though not stationary, masses of the crust while the segment bordering on the northeast moved in a northwest direction. If the intrusions and faulting were synchronous, a differential movement could be expected, in which case the intrusive side of the fault must have taken an active part.

The *en echelon* displacements came into existence during the late stages of the folding and were probably simultaneous with the development of the Ware Brook thrust. The amount of displacement at the Stony Hill reentrant is about 2 miles and considerably more in the Lake Memphremagog reentrant, possibly between 3 and 4 miles.

Black River Fault

This fault is subparallel to the Ware Brook thrust and represents a subsidiary thrust slice. It follows the base of the 5-mile escarpment south of Newport City and continues northeasterly close to the Barton River core of the Indian Point syncline to the International Boundary at Beebe Plain.

The existence of this fault is inferred mainly from the deficiency in width of the North Neighborhood anticline bordering it on the east. The thickness of the Ayers Cliff formation on the west limb of the Indian Point syncline is too great to be accommodated in the narrowed area occupied by the North Neighborhood anticline. This indicates that a part of the western limb of the anticline has been cut out and, similarly, much of the eastern limb of the Indian Point syncline, thus bringing the synclinal axis close to the fault.

Along the fault younger beds have been thrust over older and piled into a folded series presenting a precipitous escarpment to the east, south of the city of Newport. Because it is nowhere exposed, nothing is known about the fault itself, nor is its extent south of the village of Coventry traceable. The writer is of the opinion that the Black River fault is of the same age as the Ware Brook thrust.

Summary of Arguments for Structural Interpretations

1. a) Field evidence had long convinced the writer that the limestone at Lake Park is similar to that of the Brownington belt, but the

latter contains more slate and other metamorphic rock types in addition.

b) Similarly, the limestone west of Irasburg Village is essentially the same as that of Lake Park.

2. East of the Brownington belt low dips prevail. West of it the dips are higher.

3. The width of the low dip belt is so great that it suggests a dominant right side up condition.

4. a) Points 1–3 demand that the Brownington belt be a syncline, and that the Black River escarpment also be a syncline.

b) Without any structural hypothesis in mind the writer mapped the Brownington section as growing wider northward, which checks with the plunge of all observed minor folds.

c) The width of the Black River escarpment belt is also greater in the north than farther south, suggesting a synclinal structure.

5. The highly sandy Westmore formation has no counterpart in the western part of the area.

6. In the Cold Brook section in the southwestern rectangle of the Island Pond quadrangle, the Westmore formation is seen to pass downward into beds of whitish sandstone weathering into very thin beds. These resemble the sandstone of the Shaw Mountain formation in the Ware Brook section.

7. Points 5 and 6 suggest that:

a) The Westmore group is stratigraphically near the top of the Ware Brook section southeastward.

b) If "a" is correct, then the slate beds and equivalent metamorphic rocks should appear intercalated in the sandstone of the Cold Brook section as one descends.

8. Points 6 and 7 fit perfectly into the structural picture developed in points 1 to 5, and are essentially independent of the latter.

9. a) If 1 to 8 are correct, then the sequence of the larger pattern of the stratigraphic succession of this area is cut out by an overthrust in the northwestern part of the area.

b) This overthrust was not invented, for on the basis of the writer's own correlations it was at first assumed that this contact was a stratigraphic overlap, but he was driven by points 1 to 8 to the interpretation that it is an overthrust and finds that in Canada an overthrust had already been discovered. Other lines of evidence for this overthrust have already been discussed.

MINOR STRUCTURES

Folds

Excellent folds may be seen in many places. Several well-defined folds are exposed in the Ayers Cliff formation on Cove and Bell islands in Lake Memphremagog (Pl. 12, fig. 2). A large fold in interbedded slates and limestones of the Barton River formation, displaying bedding in color differences on a glacially smoothed and striated surface which, to some extent, is due to weathered pyrites in certain layers, is exposed in a pasture 1 mile due south of Coventry School (Pl. 12, fig. 3). Bedding is also disclosed here by rows of minute pits left by the decomposition of sulphides.

Excepting in the tract directly to the northwest of the central igneous body where overturning is to the northwest, the folds are overturned to the east and southeast. Axial planes dip anywhere from positions of recumbency to verticality. The regional cleavage in the southern half of the area changes from dips of more than 80 degrees in the town of Coventry to dips as low as 13 degrees in the town of Westmore, indicating the progressively more pronounced overturning of the structures to the southeast. Away from igneous contacts the folds follow a general northeasterly strike with an average plunge of 28° predominantly northeast. An occasional plunge is as high as 80° or more, particularly in the vicinity of the plutons. The changes in the orientations and attitudes of the folds locally reflect the influence of the intrusions. Convergence of the fold axes and plunges toward Beebe Plain in the northern portion of the area is corroborative evidence that the intrusions are related to the deformation of the sediments.

Slip of the beds along bedding planes in folding has produced drag folds in the less competent strata. Where slates or phyllites lie between competent strata they are often folded in this manner. In some cases thin beds of hard phyllite, lying between massive beds of limestone, are squeezed into sharp-crested drag folds, the limestone on one side showing perceptible flow structure leading into the troughs of the folds (Pl. 13, fig. 1). Apparently, the drag folds were originated before the adjacent limestone attained the status of complete flow, as the flow structure in the limestone shows no deformation to conform to that of the drag-folded phyllite.

Folds are sometimes cut off longitudinally by shear planes, the beds, especially in the region of the nose, ending abruptly against layers following the regional strike. Where they are closely spaced, as in some folds in the Ayers cliff formation, the shear planes traverse the many tightly-flexed layers. Commonly small offsets are shown clearly by layers of brittle quartz, simulating those in shear folds. Occasional shear planes contain thin veins of quartz of a later date than that which is conformable to the bedding (Pl. 13, fig. 2).

Quartz veins frequently exhibit complex folds of the ptygmatic variety. Good examples occur in limestones of the Ayers Cliff and slates of the Barton River formations. They are prevalent in beds possessing flow structure, occurring in a series of narrow loops resembling a con-

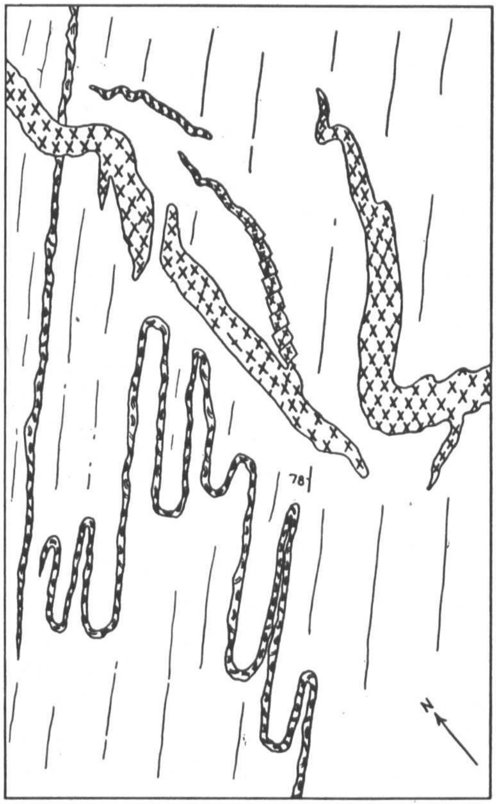

Figure 4. Shore of Lake Memphremagog at Lake Park. Ptygmatically folded quartz veins in Ayers Cliff limestone. pegmatite. The sketch shows 8 feet of width.

voluted length of rope (Fig. 4). Their association here with rocks that have been in a plastic condition is significant, as it suggests an origin similar to that advanced by Kuenen (1938). Here the previously emplaced quartz vein appears to have suffered compression when the enclosing strata behaved plastically. That the movements producing the ptygmatic folds are restricted in scope, is indicated where veins remain undisturbed in the adjoining beds (Fig. 4). Hills (1943, p. 84) classifies this type with the flow folds.

Faults

Numerous transverse faults exist along strongly-developed northwest-trending fractures. They involve beds, veins, sills and dikes, with lateral displacements ranging from a few inches to several feet. Fault surfaces are often coated with slickensided quartz and enclose quartz of varying thicknesses, or even dikes. The faults are well defined in the more competent beds but soon die out after extending a short distance into the adjacent limestone, indicating that the limestone must have been in a plastic condition when faulting occurred. Where drag in the abutting beds could be observed, it appeared to show that movement was predominantly northwestward. Cumulatively, these faults represent extensive movement and they might be related in time to the major offsets associated with the Ware Brook thrust, this in addition to a suggestive structural connection.

At the margins of the large igneous bodies and in the neighboring sediments are faults of small magnitudes caused by expansive action of the intrusions (Pl. 13, fig. 3). These faults are usually distinct and, in places, contain quartz veins. They involve aplites, pegmatites and strata, the lattermost especially in the manifold sedimentary patches lying within the areas of the plutons.

Longitudinal faults of the thrust type and contiguous beds indicate an interesting series of events. Fig. 4, Pl. 13 is illustrative of this type of fault and shows a 6-foot longitudinal displacement of a thick layer of phyllite. Thickening of the horizon by reverse faulting has caused the thin bordering strata to bend. The marked attenuation of the confining strata is, in part, due to lateral pressure exerted by the wedgelike movement of faulting and, in part, to regional stretching during the period of folding. Regardless of competency, all the beds are thinned, some pinching out altogether.

Noteworthy are the diagonally truncated ends of the faulted stratum, indicated by the hammer and steel tape. The truncation at the chisel end of the hammerhead fits that marked by the steel tape, suggesting that both ends were joined here prior to faulting. The fact that the truncations no longer face each other, precludes an explanation of deformation by thrusting alone. Rather, the displaced bed was brought into its present position by separation along the strike sufficiently to permit clearance of the ends before movement in the reverse direction was instituted. Normally, the segments would be expected to lie with their fault surfaces facing each other. Deformation was accompanied both by stretching and shortening, during which the ruptured competent bed had complete freedom of movement in the intervening plastic limestone. If the fault was produced by dominantly linear movements, tension stresses must have preceded those of compression, otherwise the diagonal fault surfaces would not face away from one another as they do at the end of each segment. Faulting may have been even more complex but evidence for it is lacking. The longitudinal faults disclose both elongation and shortening. One has only to imagine the behavior of thinly-bedded, more or less, heterogeneous strata with innumerable inherent planes of weakness than are present in heavy-bedded, homogeneous strata, to appreciate the effects of longitudinal stresses on such rocks.

Fig. 1, Pl. 14, which is located in the same outcrop and not many feet away from that shown in Fig. 4, Pl. 13, shows a feature that, in the writer's opinion, depicts an early stage in the development of the longitudinal fault discussed above. The diagonal fracture has begun to form, its open end containing a wedge of the adjacent limestone. The V-shaped fracture probably has an origin analogous to the fractures resulting when a stick of shivery wood is bent to the breaking point, thus allowing the plastic limestone to enter. An alternative explanation views the plastic limestone, which is under confining pressures, as aggressively wedging the fracture apart as it is forced to intrude.

Bedding

Bedding is shown in different ways throughout the area mapped. It is commonly recognizable, particularly in a succession of strata exhibiting different lithologies (Pl. 14, fig. 2). However, in thick sections of slates and phyllites bedding has been either totally obscured or rendered extremely difficult of detection by metamorphism. It is vaguely visible upon close examination in shade differences of color and slight changes in texture as, for instance, a crossbedded, sandy layer in the slates of the Black River escarpment south of Newport City.

In some slates bedding appears as fractures (Pl. 14, fig. 3). The writer believes that, in some cases, curving fractures in vertical sections of cliffs of slate horizons quite possibly represent bedding. In Fig. 4, Pl. 14, bedding is disclosed by a corrugated fracture and paralleling line of solution pits formerly occupied by pyrites. The bedding surface exposed in another part of the outcrop, displays a washboard structure. Contrasting examples in which bedding is manifest, are shown in Pl. 15, figs. 1 and 2; in the former the beds of slate are delineated by a series of parallel grooves while the latter shows a succession of thin beaded ridges in an exposure of hard, massive amphibolite, as though the beds were welded together. The exceedingly smooth bedding planes in the whetstone phyllite of eastern Brownington, are parallel to the cleavage (Pl. 15, fig. 3).

Bedding—Cleavage Relations

Bedding and cleavage comparisons are highly important in structural studies in metamorphosed regions. They have aided enormously in the determination of the major structures in this area and showed the beds to be overturned with tops to the east in considerably more than half of the area. The degree of overturning and the position on the fold is clearly demonstrated where slates and phyllites are interbedded with massive layers such as limestones, or in thick sections of slates in which bedding is distinct (Pl. 16, figs. 1 and 2). In slates where bedding and cleavage relations are not well defined as, for example, on the limbs of tight folds where both are essentially parallel, it became necessary to base structural interpretations largely on observations made in adjacent tracts.

Sinuosity of Beds

The strata frequently display local rolls both along the strike and down the dip, which, if not recognized, might seriously affect the accuracy of strike and dip measurements (Pl. 16, fig. 3). Sinuosity may be expected in a region of tightly folded, comparatively thin, incompetent strata such as those dealt with here. Locally, sinuous structures are the result of expanded masses of quartz (Pl. 12, fig. 1). Generally, however, they result from unequal stresses of folding and dilation of igneous intrusions. The small-scale sinuosities reflect those of greater magnitude flanking the plutons and curving with the major folds.

Plastic Flow of the Calcareous Rocks

Plastic flow in the limestones is widespread. It is expressed in the bedding planes which appear as lines, often described by thin ridges of the more resistant material due to weathering, and color distinctions probably attributable to varying degrees of recrystallization. The flow lines normally follow the bedding and occasionally curve around the noses of folds. These folds are frequently cut off on one side by a shear plane and apparently represent drag (Pl. 16, fig. 4). Thin, steeply-dipping slates sometimes occur torn apart and the slender pieces rafted to new positions, still retaining their steep attitudes. In other instances, the thin slate layers are molded into extremely complex small-scale folds, the flow lines in the limestone core directed into the strongly crinkled, slate-bordered salients. As suggested by its pattern, this peculiarly intricate type of fold might appropriately be described as "birdsfoot" (Pl. 17, fig. 1).

In Fig. 2, Pl. 17, the somewhat bulbous cores of two synclines are composed of limestone displaying flow structure. The limestone, rendered plastic under the intense stresses of folding, was forced into the synclinal cores, producing the bulblike shape. Although most are ordinary drag folds, a few folds of the flowage type were developed in the limestone during its migration into the cores under confining pressures (Bain, 1931). Where overturned, some of the anticlinal folds are sharp crested and bear a resemblance to the profile of a machete blade. Their upper limbs sag and are thinned while the lower limbs display a more pronounced sag and are greatly thickened. These features are all illustrative of a flow condition in which there was a tendency for material to move downward. Folds similar to these are described from the Vermont marble belt by Dale (1902, p. 9). In places the flow structures resemble cross-bedding where one set is truncated by another, or they appear as drag at the contact with the enwrapping slates.

Not only were the limestones in an environment of deep burial but they must also have been strongly affected by the igneous intrusions, since most of the examples cited are located near their contacts. In some folds at igneous contacts certain thin calcareous layers are completely recrystallized to a white marble, while adjacent beds are partially marbleized.

An occasional fragment of the intercalated, more competent rock occurs embedded in limestone (Pl. 17, fig. 3). Similar occurrences have been reported from other localities (Balk, 1936, pp. 720–723; Adams and

Barlow, 1910, pp. 220–221). The presence of these erratic pieces in the limestone matrix appears to be due to their dismemberment during folding movements and subsequent inclusion. Behre (1933, p. 148) has described sedimentary fragments isolated in other strata and attributes the phenomenon chiefly to settling movements sometime during diagenesis. That the "floating" fragments herein described have a tectonic origin is evidenced by the cleavage and clean-cut transverse fractures which they possess, the latter terminating segments which, therefore, could readily be dislodged.

The free moldability of the limestone is further seen in the thickening and thinning of individual beds along the strike. Although it is prominent in the limestones, thickening and thinning is not confined to them but is found in all types of layers, including granite sills and quartz veins. Some beds are pinched out along the strike into lenses resembling horses in faults; commonly slates and phyllites show this behavior between limestone beds. Thicknesses of beds at the noses of folds are as much as 10 times greater than on the limbs.

Flowage of the limestone and associated structures are significant as evidence showing that the rocks of the region were subjected to great stress under conditions of considerable depth and to the influence of igneous action.

Boudinage

Beds, veins and thin sills or streamers are frequently found drawn apart into separate units and the ensuing gaps filled with material from the adjoining beds or by quartz. This structure was termed *boudinage* by M. Lohest (1909), many examples of which have been cited in the geological literature since. E. Cloos (1947) has reviewed the literature on boudinage with a brief discussion.

In the bed of the Clyde River between the highway and railroad bridge, 1.1 miles northwest of Barton Village, slate boudinage occurs in limestone (Fig. 5). The slate beds are relatively thin and lie between thick beds of limestone. Ostensibly, the weaker rock composes the boudins, which is contrary to the competency relationships in the many occurrences observed, but upon examination of the thicker limestone beds it is found that they are thinly banded. The banding is much thinner adjacent to the boudins than farther out in the limestone and parallels them (Fig. 5). Frequently the bands are bordered by thin, sinuous quartz veins which occur singly or in parallel sets.

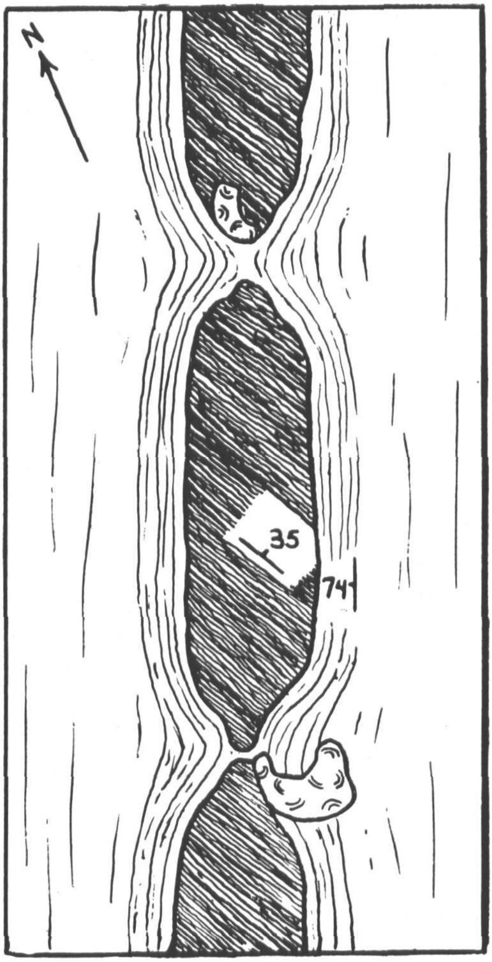

Figure 5. Bed of Barton River between highway and railroad bridge, 3/8 mile east of Country Club. Slate boudinage in limestone. Barton River formation. quartz. The sketch shows 4 feet of width.

The banding in the limestone represents a flow structure, which is an important factor in the explanation of the competency relationships between the limestone and the slate. The slate is actually the more competent while the limestone, which was effectively weakened by the close succession of shear planes, was prevented from behaving as a unit when under stress. That a plastic condition prevailed during formation of the boudinage structure, is exemplified, not only by the limestone, but also by the shapes of the boudins themselves. They are drawn out into tapering ends, much the same as taffy candy when pulled apart. The cleavage in the slate is apparently older than the boudinage structure.

Two major stages in the formation of boudinage structure are illustrated in Fig. 2. In this exposure, located where the Pine Hill road crosses Cobb Brook, three granite streamers, of which two exhibit varying degrees of disruption, are elongated with the cleavage of the enclosing slate. One of the streamers has suffered very little deformation, while its neighbor has undergone segmentation on fairly evenly spaced transverse fractures. Along its northwestward extension this streamer is progressively more disrupted, the farthest segments showing incipient rotation. The remaining streamer is completely disjointed into a series of boudins, several showing rotation. In this developmental sequence the competent layer is transversely fractured, along which the segments are displaced and, apparently, rotated as they are longitudinally separated. Due to friction caused by linear and rotational movements, the short, angular blocks of the early stage have become lengthened and rounded in form. The flow cleavage wraps around the isolated boudins and is lobed in the gaps between them. It has already begun to bend around the displaced angular segments in the middle streamer.

It is somewhat perplexing that competent layers lying only a matter of feet or even less apart, should behave so differently under stress. Selective differential movements, however, within the slates could account for the structural differences in the three granite streamers. Again, the time of intrusion relative to movements in the slate, might explain the marked contrast between the practically undeformed and completely dismembered streamers. The tension joints and offset rhombs in the resistant layer in Fig. 2 suggest stresses causing a bending motion as in folding. Boudinage is indicative of elongation and is a phenomenon associated with layers that have been subjected to folding stresses.

In Fig. 6, which is from the same location as Fig. 5, a stratum of slate shows both transverse offsets and boudinage, the faults and the boudinage gaps being occupied by quartz. Although the flow structure in the limestone curves into the gap between the boudins, it shows no tendency to do so at the faults. Rather, it ends abruptly against them and is even noticeably displaced; the displacements extend out into the limestone only as far as the thin quartz filler extends. In this case, the limestone appears to have possessed greater plasticity where the boudinage was produced than it did along the faults. Apparently, the faults are later in origin.

Quartz often occurs as boudins (Pl. 15, fig. 4). The thin-bedded limestones and slates have been bent inward to close the gaps between the quartz lenses. The limestone layers in the gaps have been thickened by flowage while those contiguous to the boudins have suffered attenuation. However, these effects do not extend far into the enclosing strata

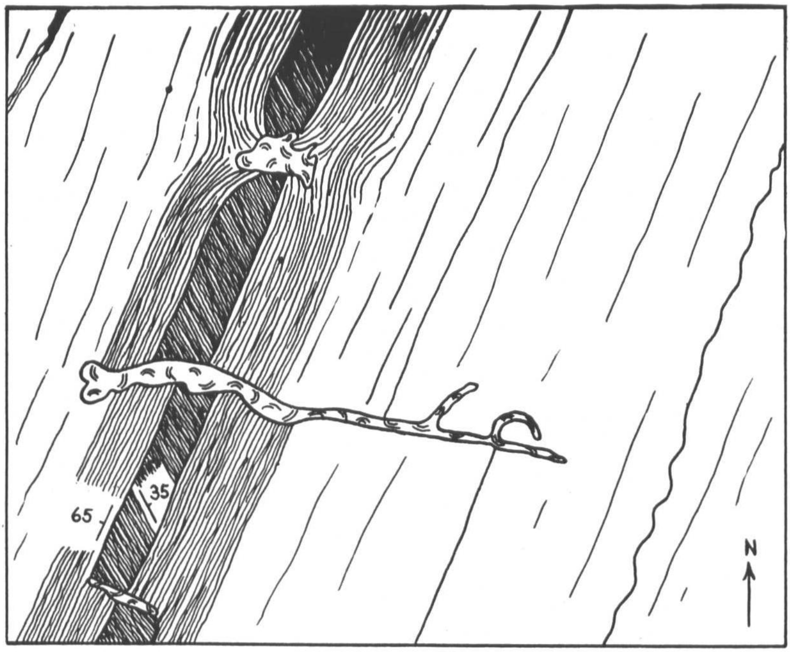

Figure 6. Bed of Barton River between highway and railroad bridge, ⅜ mile east of Country Club. Faulted bed of slate and incipient boudinage formation. Barton River formation. quartz. The sketch shows 7½ feet of width.

(Pl. 15, fig. 4). Boudinage is strikingly shown in the numerous patches of thinly-bedded sedimentary rocks in the areas of the granite plutons. Here the coarse, micaceous granite paste has flowed between the separated beds.

Linear Element

Parallel streaks of minerals or their weathered products are common on the planes of cleavage, schistosity and bedding. This linear parallelism is shown by elongated mica flakes, amphiboles, sulphides and rust streaks resulting from weathered sulphides. The sulphides are often drawn out into a well-defined lineation and in some cases the biotites and an undetermined, steel-colored mica exhibit definite oval shapes with parallel elongation. The linear element is generally well developed, but in many of the slates it is discerned with difficulty.

Orientation of the linear features over the whole area is shown in Plate 1. The direction of the linear element is not the same everywhere in the area, but for the most part it plunges northeast in the western half and northwest in the eastern. Northwest plunges in the northern part of the towns of Irasburg and southeastern Coventry, appear to be the northwestward continuation of a similar plunge orientation in the towns of Brownington and Westmore. As a rule, the linear element plunges more to the west in the southern than in the northern portion of the eastern half of the area. Its many deviations have probably been caused by nonuniform stresses generated during folding of the sediments and emplacement of the plutons.

A study of the map reveals two major directions of the lineations, namely, northeast and northwest. It has occurred to the writer that, since there is a great deal of overturning in the area, the lineations also might be overturned in one of the principal directions noted. Where the beds are overturned to the southeast, which is the dominant attitude of the beds northwest of the axial region of the Brownington syncline, the trend of the lineation is markedly northwest. Near igneous contacts it may deviate from the directions observed. Cloos (1946, pp. 21–22), in discussing the age relations of the linear element, traces the probable sequence of structural changes during the process of folding. He suggests a stage when the process producing flowage and lineation is arrested and followed chiefly by fracturing. These two succeeding stages are well illustrated in Plate 5, figs. 1 and 2, in which a later "slip cleavage" has been superimposed over an earlier flow cleavage. In strongly folded rocks, such as those characteristic of this region, the lineation could have become fixed in the fabric of the rock before overturning, so that a lineation possessing an original northeast strike, was given a northwest strike through reversal of the dip of the strata. For example, a bed of slate whose linear element plunges to the northeast on a surface dipping to the southeast is, with continued or renewed deformation, overturned to the southeast, thereby bringing the previous northeast plunge of the linear element into a northwest orientation. Turner (1948, p. 282) mentions changes in attitude of lineation by folding. In connection with the above discussion, the question of the possible use of linear element in determining reversal of beds where the usual criteria are totally lacking, presents itself.

Lineation is both down the dip and parallel to the fold axes. It sometimes occurs on the same surface in more than one set, as in the axial region of the Brownington syncline one mile southwest of Toad Pond. Crenulations in at least two sets line the bedding surfaces and sometimes are parallel to the plunge of the minor folds.

METAMORPHISM

General Statement

The rocks of the area were affected by regional metamorphism in which the intrusives played a prominent part. They are assigned to metamorphic zones according to grades on the basis of certain mineral indicators (Billings, 1937, pp. 540–543) and also according to the zonal subdivisions of Harker (1932, p. 209). All the formations east of the Ware Brook thrust are in the middle-grade zone of metamorphism. The minerals diagnostic of the intensity of metamorphism indicate an increase toward the southeast. The high-grade zone is present in the aureoles bordering the plutons. Northwest of the Ware Brook thrust the rocks have not been studied petrographically, hence their place in the zonal delimitation is not known, although general observation is suggestive of middle-grade at most.

Minerals

Chlorite—Chlorite is widely distributed in the area and, thus, has no zonal significance. Much of the chlorite is the result of retrograde metamorphism, replacing amphibole, augite and biotite. Where it contains inclusions of zircon its origin from biotite is confirmed. Pale-colored or "bleached" biotite flakes are associated with chlorite and a partial alteration of biotite to chlorite is not uncommon. Chlorite occurs as an alteration product in the fractures of garnets, as in the six-rayed garnets cited earlier in this paper.

Rectangular, well-defined porphyroblasts of chlorite, often of large size, transgress the parallel structures, thus indicating a late development of this mineral. They were formed either as the normal product of progressive metamorphism, or in a lower temperature during retrogressive metamorphism.

Biotite—Biotite occurs in the Ayers Cliff formation but is found in greater amounts in the Barton River and Westmore formations. It is present both as irregularly-shaped porphyroblasts and as well defined crystals parallel to the banding in the rock or in random orientation. Occasional large phantom crystals broken up parallel to the banding and large porphyroblasts bent and broken to conform to a herringbone structure, indicate the orientation effects of deformation. Flaser structures, many with frayed ends, composed of both oriented and unoriented biotite, evince the effectiveness of dynamic metamorphism on these rocks.

The concentration of biotite in rich bands and its development in contact zones between sedimentary layers in the presence of muscovite, calcite and quartz, is suggestive of hydrothermal action and possible reactions among these minerals.

Garnet—Garnet occurs in the Barton River and Westmore formations, including their contacts with intrusives where the high-grade metamorphic zone prevails. It is prominent as porphyroblasts which are especially numerous in certain horizons, occurring in groups or as single crystals. It is found with perfect crystal outline and also in irregularly shaped masses which are often considerably corroded. In some porphyroblasts corrosion has produced a poeciloblastic or "sieve" structure in which quartz inclusions constitute as much as 50 per cent. Undistorted porphyroblasts in strongly crenulated schist indicate a late development.

Alteration in varying degrees is common in the garnets, the replacing minerals quartz and sericite, and often chlorite, occurring at tufted ends of garnets parallel to the schistosity, in bleached border zones in which biotite, staurolite, and sulphides, in addition, are present, and in six-rayed fractures. In some cases replacement has become so complete that only phantom crystals remain. Finely disseminated carbon has rendered some crystals a dark gray. Garnet occurs also in quartz and pegmatite veins and in granites.

Andalusite—In the thin sections examined, andalusite appears to be confined to sedimentary rocks bordering intrusives. It appears here, in the high-grade zone, as a retrograde mineral which occurs in various stages of alteration to quartz and muscovite. Large porphyroblasts of the variety chiastolite, exhibiting symmetrically-arranged carbonaceous inclusions, are greatly altered to quartz and muscovite, as is also a strongly pleochroic colorless to rose-colored variety. Andalusite has formed apparently in a retrogressive metamorphic environment of the high-grade zone in argillaceous sediments, developing into the retrogressive condition with falling temperatures.

Staurolite—Staurolite occurs generally in the Westmore formation, and only at igneous contacts in the Barton River formation. It is found in association with andalusite, which is rare according to Harker (1932, p. 232). In this relationship, the staurolite is present as small porphyroblasts oriented to conform to a wavy schistosity which bends around the much larger andalusite porphyroblasts that are considerably retrogressed. In this instance, the staurolite formed while deformation was in progress and the andalusite at a later time of falling temperature. Many large porphyroblasts of staurolite, variously oriented, are greatly fractured, distorted and broken apart, which is suggestive of either a recurring period of deformation or their development during the waning stages of the main orogeny. The large porphyroblasts are profuse with inclusions of quartz and carbonaceous matter, the latter sometimes oriented parallel with the schistosity. Porphyroblasts that have suffered little or no distortion, possess borders relatively free from inclusions.

Sillimanite—Sillimanite is restricted to contact areas of high thermal metamorphism surrounding the plutons. In some places it occurs abundantly in sheaf-like bundles and as isolated, needle-like crystals in random orientation, in aluminous and highly quartzose sediments. It is late to form and has grown at the expense of biotite and muscovite. At the contact between a very fine-grained rock composed of quartz grains and abundant carbonaceous material with phantom porphyroblasts of muscovite visible only under crossed nicols, fan-shaped groups of slender sillimanite crystals extend into a bordering, medium-grained quartz-calcite rock.

Lime-silicate Minerals—This group of minerals is here represented by tremolite, actinolite, diopside, clinozoisite, some zoisite, wollastonite and scapolite, and possibly vesuvianite. They occur almost exclusively at igneous contacts in the calcareous rocks of the Barton River and

Westmore formations. The development of these minerals at igneous contacts is suggestive of the introduction of materials in addition to the activation of the original constituents of the sediments, such as the clay mineral impurities.

Tremolite ranges in size from fine grains to large euhedral crystals, and sometimes occurs as phantom crystals. It is often associated with quartz, calcite and diopside, in which relationship diopside has taken the place of tremolite in an environment of rising temperature. Large acicular porphyroblasts of tremolite and muscovite have, in some instances, been observed with brush-like groups of tremolite crystals attached to the ends of the rectangular muscovites. Actinolite accompanies tremolite in white, saccharoidal, highly siliceous, marbleized layers of the limestones at igneous contacts.

Amphibole—Amphibole is common in the rocks of the middle- and high-grade metamorphic zones. Colorless and yellow-green porphyroblasts are frequently found corroded by quartz and altered to chlorite and epidote, and occasionally clinozoisite, suggesting a change to a retrogressive metamorphic mineral assemblage. On the other hand, biotite partly altered to hornblende containing streaks of magnetite, indicates a progressive metamorphic environment. The amphiboles have a random orientation and, therefore, are considered to have formed when the period of dynamo-metamorphism had subsided.

HISTORICAL GEOLOGY

The Memphremagog area lies within the Magog eugeosynclinal belt (Kay, 1937, p. 290; 1942, p. 1642; 1948, p. 1332) which subsided deeply during early Paleozoic with accompanying volcanism, represented sparsely by tuffs in the immediate section. The paleogeography and sedimentation were strongly affected by deformation toward the west in the later Ordovician, culminating in the Taconian orogeny. Deep subsidence continued in a more restricted belt including the Memphremagog area through Silurian and into early Devonian time, closing with the severe folding, thrusting and plutonic invasions of the later Acadian or Shickshockian orogeny. The sediments are generally poorly sorted, dominantly argillaceous, with varying quantities of quartz and calcareous constituents. Deposition was in waters of varying depths having persistent source lands for terrigenous clastic sediments that filled the rather rapidly subsiding geosynclines. Rapidly changing lithologies in thin-bedded sections are frequent. Extensive conglomerates of rounded quartz pebbles in lower formations, and crossbedding, frequent in some units, indicate shallow water conditions at times. Most of the argillites are quartz arenaceous, and much of the quartzite has argillaceous constituents, suggesting that subsidence proceeded too rapidly for the debris to become well sorted by the currents that swept it in and passed over it. Notwithstanding effects of subsequent metamorphism, the rocks in many localities are sufficiently well preserved that fossils should be found if originally common; few have been recognized confidently.

The area was occupied by medial Ordovician seas when the oldest sediments in the area were deposited. The prevailingly silty muds laid on a gravelly base, becoming sandy in higher horizons are represented in the Cram Hill formation, suggesting progressive deepening, sinking exceeding accumulation. After an erosional interval, perhaps representing rise of a belt to the west, the seas advanced again over the area within medial Ordovician time, beginning the deposition of quartz gravels that with the succeeding sands interbedded with tuffaceous material and thin calcareous zones, constitute the Shaw Mountain formation. Crustal oscillations controlled the extent of erosion and deposition reflected in the variable thicknesses of this formation, particularly of the conglomerates.

Subsequently, with deepening water, dark, siliceous muds of the Northfield slates were deposited, becoming more calcareous toward the top and finally passing into the calcareous, somewhat sandy, muds of the Ayers Cliff formation. The Taconian orogeny may be represented in the conglomerate of sedimentary and igneous fragments at the base of the succeeding Barton River formation and the unconformity at its base. The succeeding beds are replete with abrupt changes in lithology involving calcareous muds to sands, denoting shifting currents in the shallow seas above the rapidly subsiding geosyncline. Similar conditions of sedimentation continued throughout the span of the Westmore formation, but with the fine terrigenous clastics becoming progressively more siliceous. Deposition of the Westmore formation may have extended into early Devonian time (Doll, 1943, p. 57), though evidence is tenuous. That this portion of the Magog trough was flooded by marine waters in Devonian time, is supported by the fossils in the Bernardston marble far to the south in Massachusetts (Whitfield, 1883).

Although secondary structures were imposed upon the strata during the earlier stages of the geosynclinal development, it was not until the late Devonian that the present structures, revealing a deformation of great intensity, were formed. This was the Acadian or Shickshockian disturbance. The formations were squeezed into strongly folded and faulted structures with northeast trends, the structures notably accentuated by the forceful invasions of the magmatic bodies which shoved the beds from the earlier acquired northeast trends to a perimetric orientation in proximity to the plutons. The final emplacement of the igneous bodies lagged behind the dynamic phase of the orogeny which produced the folded and faulted structures, but that they were actively rising during the period of deformation is shown by sills infolded with the sediments. The intrusions were definitely related to the orogeny and very possibly played an important part in the evolution of the *en echelon* reentrants, in which they occupy "cornerstone" positions. The thrust faults originated during the final stages of the folding and probably long before the intrusives had ceased to be active. A solitary basic dike affirms the occurrence of post-orogenic injections.

Metamorphism is intimately associated with the deformation and igneous intrusions. It is both deformational and thermal, the latter producing widespread recrystallization of high rank. Although recrystallization began early in the deformation, it probably had not established a zonal pattern until the arrival of the more advanced intrusive stage of the orogeny. This is substantiated by the proximal relationship of the high-grade metamorphic zone to the intrusives. Besides heat, the intrusives sent gases and liquids far out into the enclosing rocks, resulting in the development of a variety of new minerals in the easily altered argillaceous and calcareous sediments.

Later Devonian time introduces a long span characterized chiefly by erosion and intermittent uplifts, the warping of the late Tertiary heralding the Great Ice Age or Pleistocene epoch. The glacial ice modified the earlier landscape, rasping material from the higher elevations and distributing it over the surface generally as till, but to greater thicknesses in the valleys. Much of these deposits was reworked by the meltwater liberated by the waning ice sheet and in the major valleys was submerged under the waters of marginal glacial lakes, of which Glacial Lake Memphremagog was the most extensive (Hitchcock, 1907–1908, p. 641).

Post-glacial history is principally one of uplift with continued erosion, in which time the major streams have incised themselves to form scenic gorges in places.

ECONOMIC GEOLOGY

Granite quarries and prospects at many places in the plutons bear witness to a time of extensive activity in this industry. The granite probably found local use mostly as foundations for bridges and houses. The Willey Granite Quarry in Derby appears to have been in production longest and, after a period of inactivity, is again in operation. For a time this quarry furnished stock for gravestones, as may be seen in the cemeteries in Newport. In the past, phyllite of whetstone grade from the upper part of the Westmore formation was a natural resource of some note. Several openings are located on the slopes north of Lake Willoughby and east of East Brownington School. The chief source of whetstone stock was the large quarry located on the westerly slope half a mile east of the junction of state highway route nos. 58 and 5A. The phyllite was fashioned into whetstones at a mill on the Willoughby River in Evansville about 3 miles west of the quarry. Phyllites have also found local use as foundation stone and flagging.

Gold and copper have been reported in the area (Richardson, 1907–1908, p. 289); the former in hardly more than a trace and the latter in small samples in combined form. The quest for these metals in the past is revealed by pits in pyritized rock and openings in quartz veins.

A generous supply of gravel is being exploited in the various towns for road metal and other uses.

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Plates 2 through 17

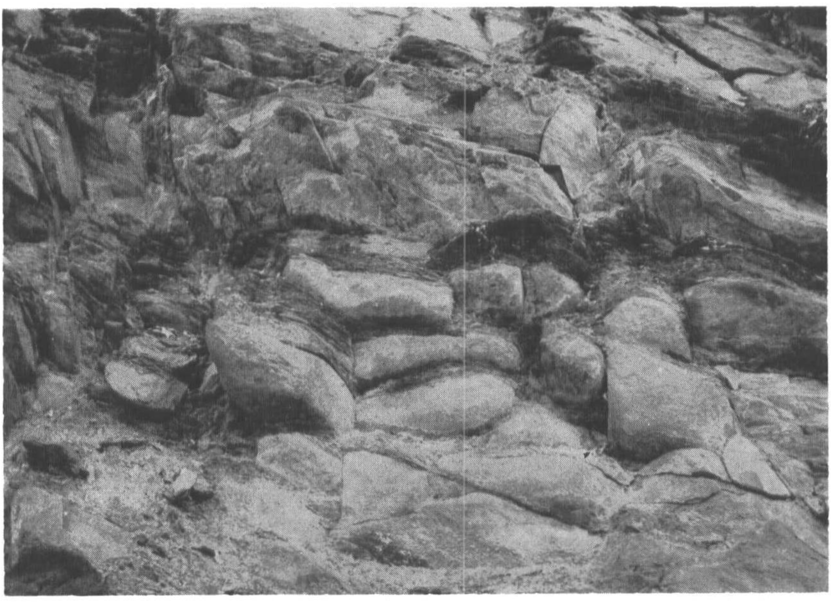


PLATE 2

Figure 1. Surface of an arenaceous limestone pock-marked by parallel solution cavities, bordering road at farmhouse, 1 mile southwest of Bemis School (Island Pond quadrangle).

Figure 2. Thick phyllite layer in escarpment of Black River south of Newport City, exhibiting spheroidal weathering.

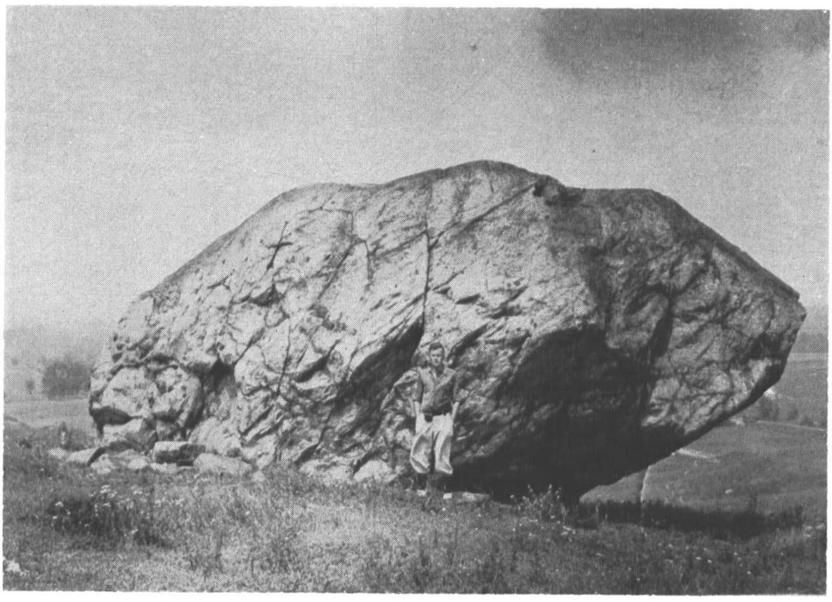


PLATE 2

Figure 3. "Tombstone" weathering of phyllite, ½ mile northwest of Brownington Village.

Figure 4. Glacial erratic of pillow lava belonging to the Bolton igneous group, in pasture 2 miles north of Orleans and still within sight of its home.

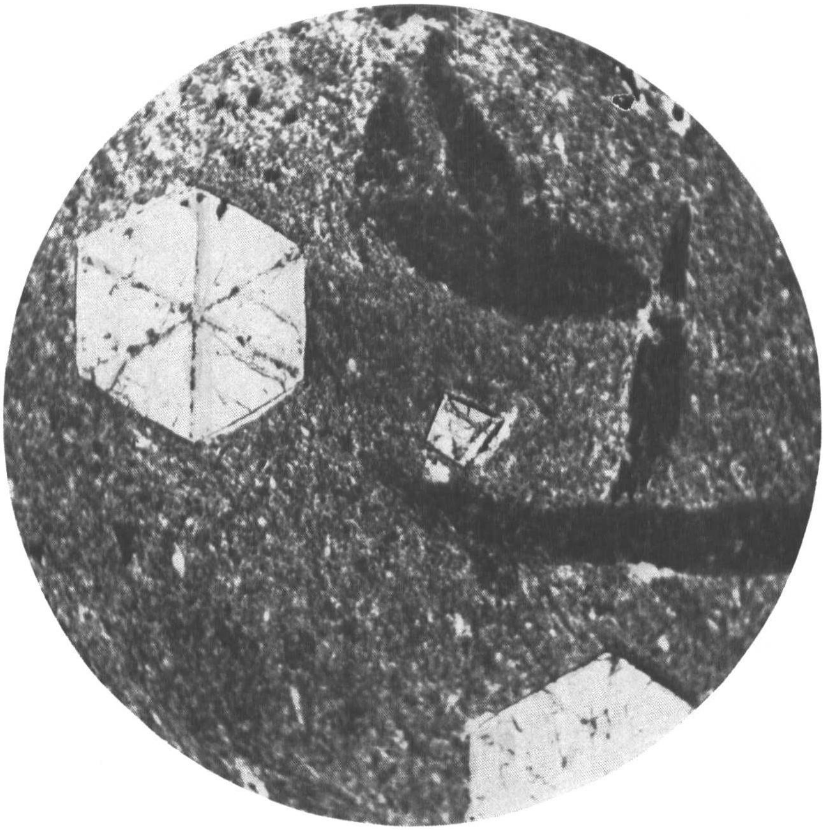


PLATE 3

Figure 1. Metacrysts of six-rayed garnets and bent and broken amphiboles in Barton River phyllite. (Vt. 366).* Nicols not crossed. (X 25).

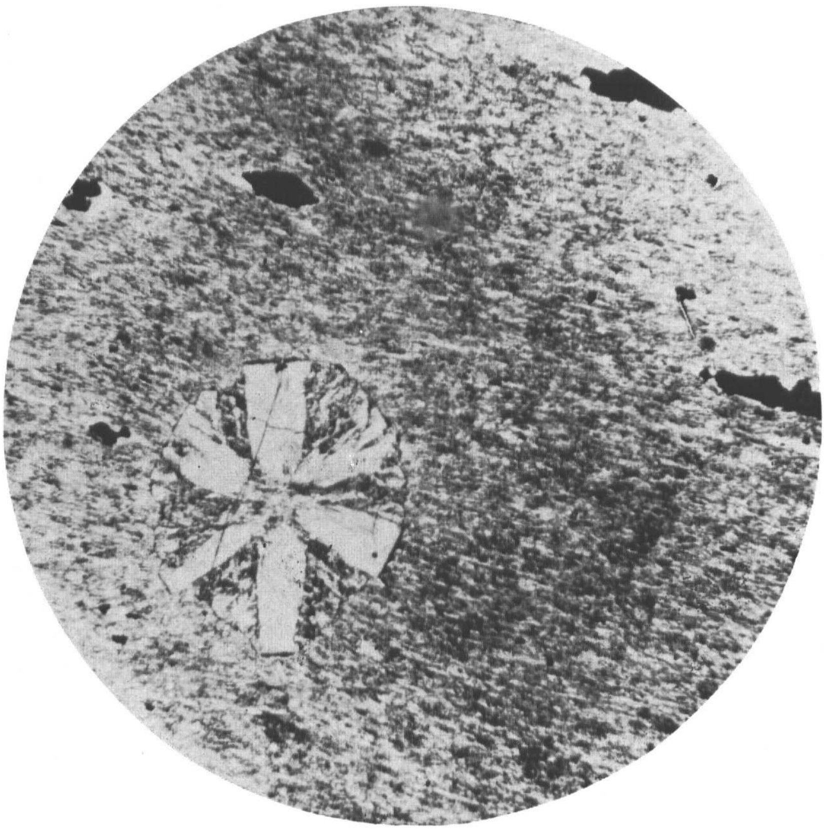


PLATE 3

Figure 2. Advanced stage in alteration of six-rayed garnet in Barton River phyllite. (Vt. 355).* Nicols not crossed (X 50).

*These accompanying numbers in the illustrations refer to slide collections at the University of Vermont.

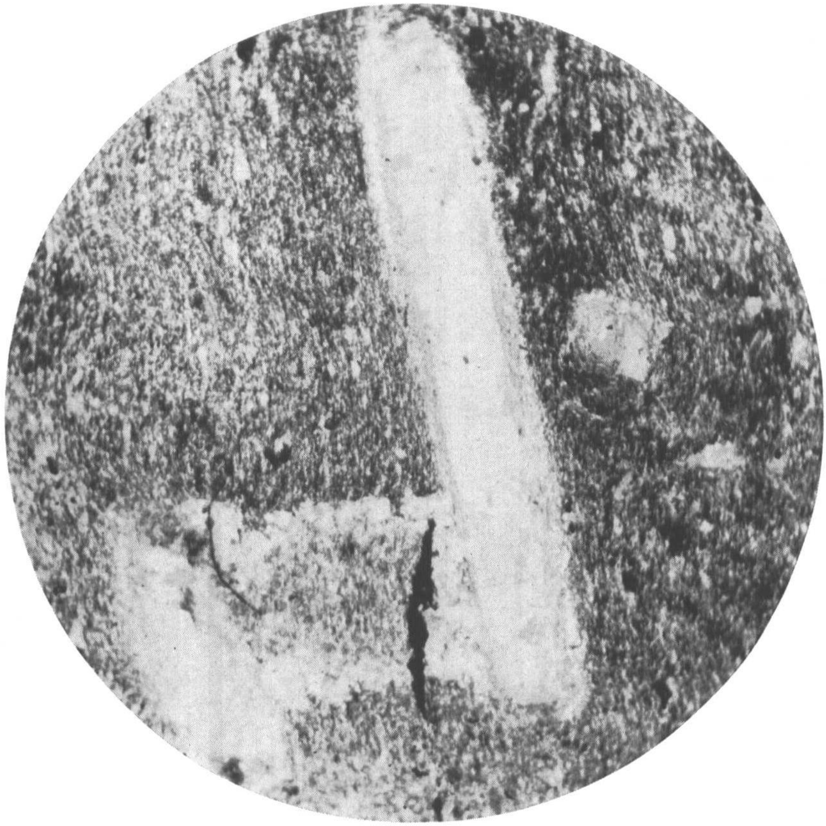


PLATE 4

Figure 1. Andalusite metacrysts in Barton River phyllite showing longitudinal and transverse sections, the latter considerably altered to a skeletal frame. (Vt. 421). Nicols not crossed. (X 24).

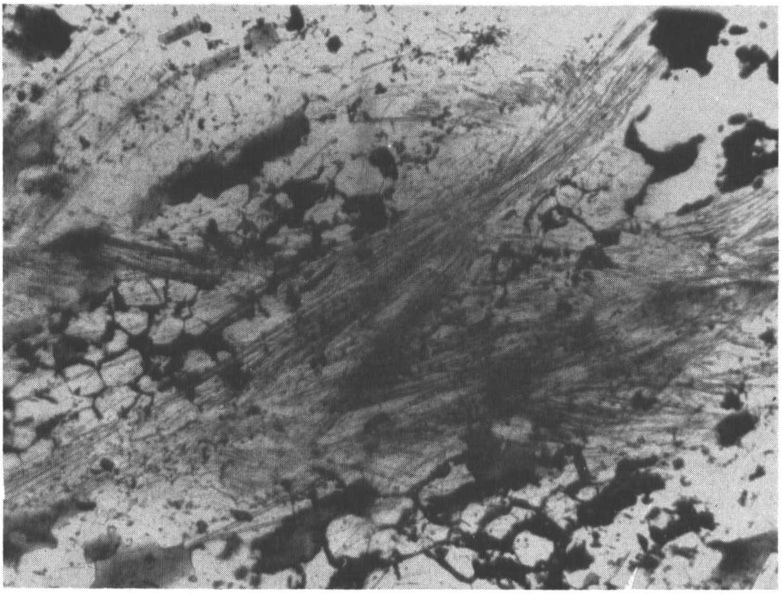


PLATE 4

Figure 2. Sheaf-like bundles and isolated acicular crystals of sillimanite in impure quartzite of the Barton River formation (Vt. 368). Nicols not crossed. (X 180).

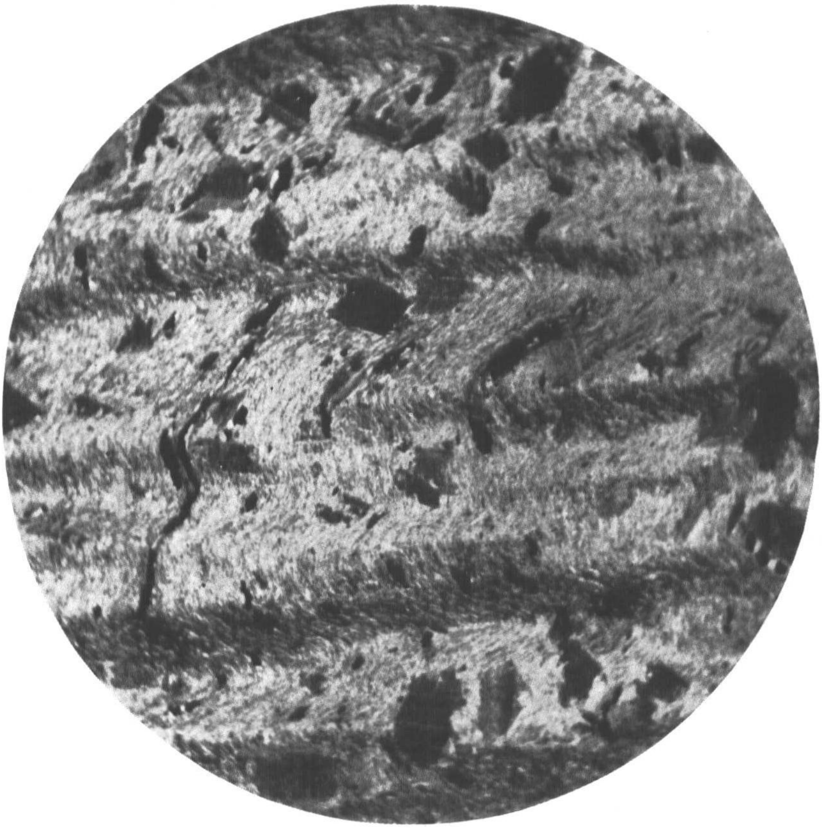


PLATE 5

Figure 1. Quartz-sericite-biotite schist of the Barton River formation showing mica and sulphides bent with the herringbone structure illustrating "slip cleavage." (Vt. 479). Nicols not crossed. (X 24).

PLATE 5

Figure 2. Barton River quartz-mica schist with herringbone structure illustrating "slip cleavage" and a rod of sulphide bordered by a halo of secondary biotite. Strut-like appearance of sulphide and curvilinear groundmass at its right end, indicate a late development of the sulphide. (Vt. 483). Nicols not crossed. (X 23).

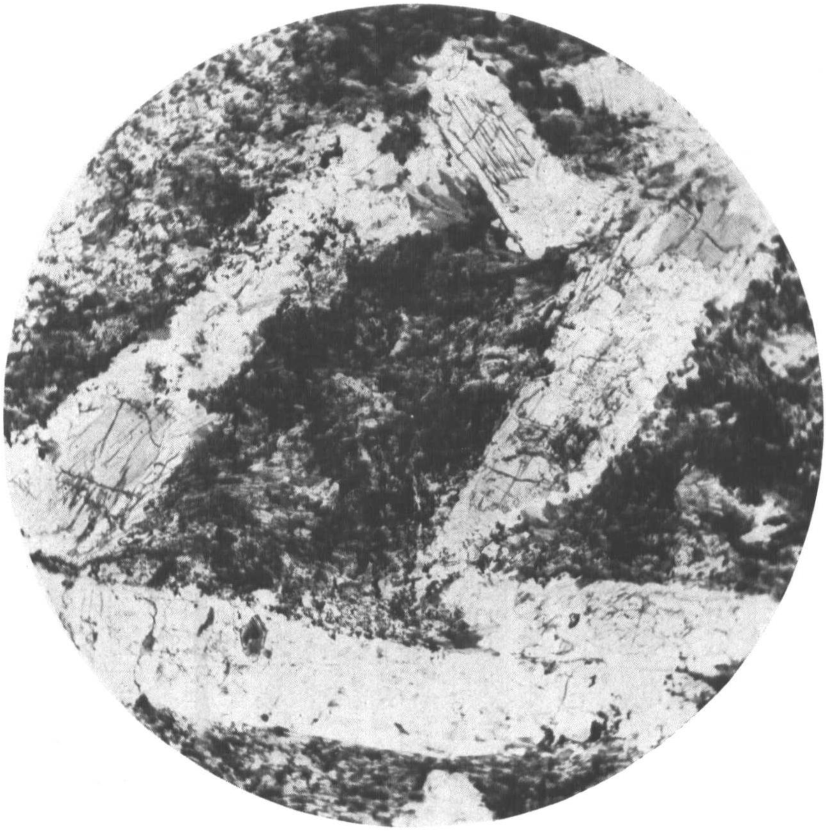


PLATE 6

Figure 1. Pleochroic, partly rose-colored (gray areas in the porphyroblasts) andalusite porphyroblasts. Andalusite schist of the Barton River formation. (Vt. 422). Nicols not crossed. (X 25).

PLATE 6

Figure 2. Porphyroblasts of chiastolite displaying geometrically-arranged carbonaceous inclusions. Barton River formation. (Vt. 412). Nicols not crossed. (X 24).

PLATE 7

Figure 1. Quartz-mica schist of the Westmore formation displaying parallel augen composed of biotite and scattered muscovite. (Vt. 464). Nicols not crossed. (X 37).

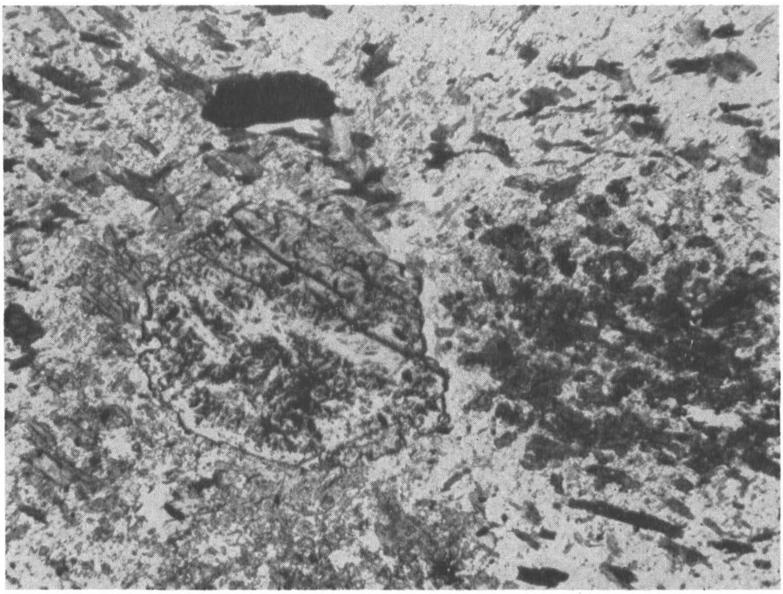


PLATE 7

Figure 2. Garnet-mica schist; corroded amphibole showing rhombic cross sections of cleavage. Garnet porphyroblast with corroded border and numerous inclusions. Westmore formation. (Vt. 465). Nicols not crossed. (X 51).

PLATE 8

Figure 1. Staurolite porphyroblast dark with carbon inclusions. Staurolite schist, Westmore formation. (Vt. 489). Nicols not crossed. (X 24).

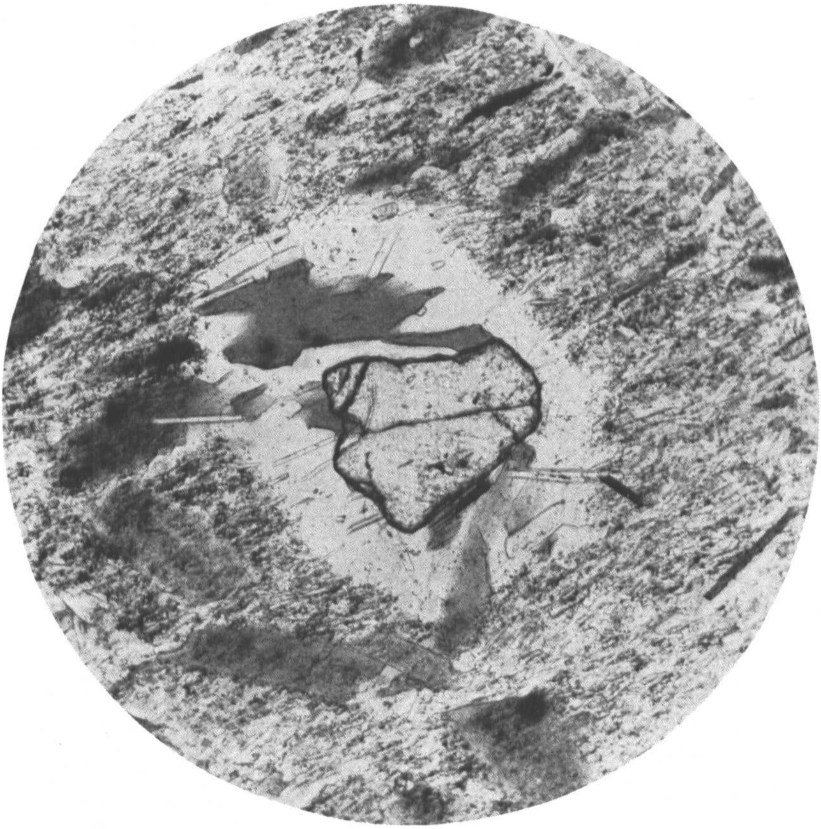
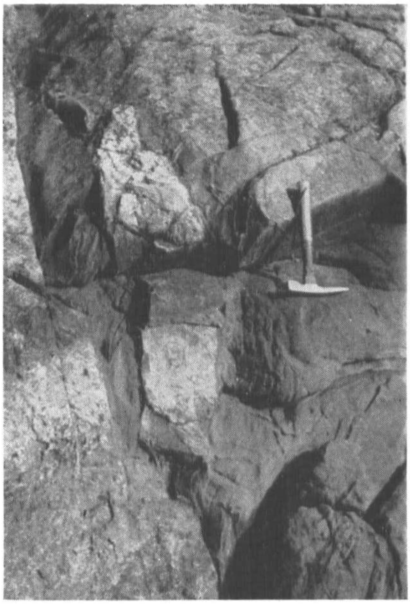


PLATE 8

Figure 2. Garnet porphyroblast surrounded by bleached zone of quartz, muscovite and biotite. Mica schist of Westmore formation. (Vt. 489a). Nicols not crossed. (X 84).



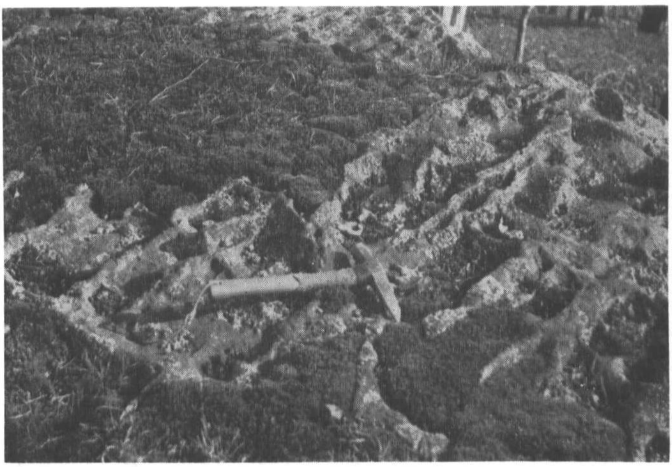


PLATE 9

Figure 1. Shore of Lake Memphremagog at Holbrook Bay. Granite inclusions in basic rock of the Bolton igneous group.

Figure 2. Bordering brook just off west side of road southwest of Maxfield Light. Reticular weathered surface of metadiorite. Bolton igneous group.

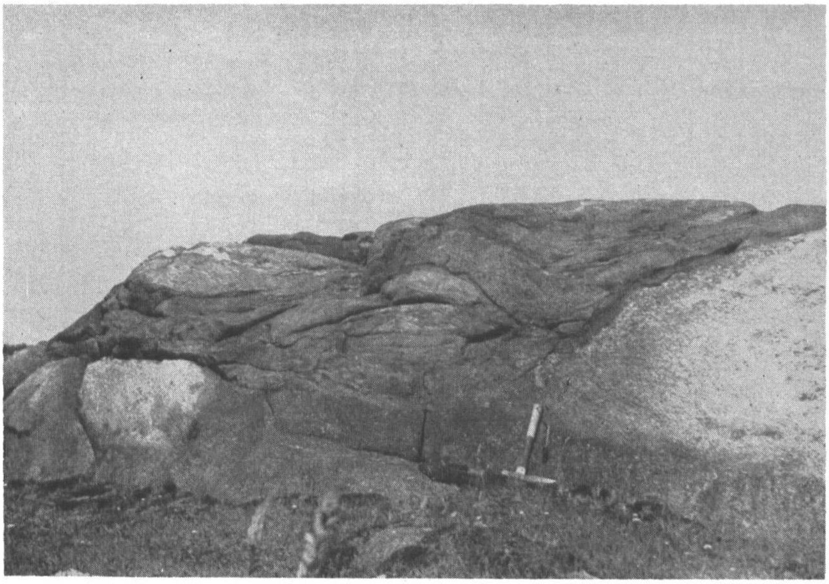


PLATE 9

Figure 3. Northeast slope of Sugar Hill, in pasture at sharp bend in road. Metamorphosed sediments of Barton River formation lying upon granite in synclinal depression.

Figure 4. ¾ mile east of Sugar Hill. Tabular inclusion of sediment in granite hogback. Note divergent fractures at the sharp corners of the xenolith.

PLATE 10

Figure 1. Steep part of west slope of Sugar Hill, opposite elevation 1226. Limestone of Barton River formation dipping beneath granite cliff.

Figure 2. North side of road near brook, ½ mile southeast of Cobb Pond. Gap in sediments (point of hammer) filled by flowage of coarse-grained granite.

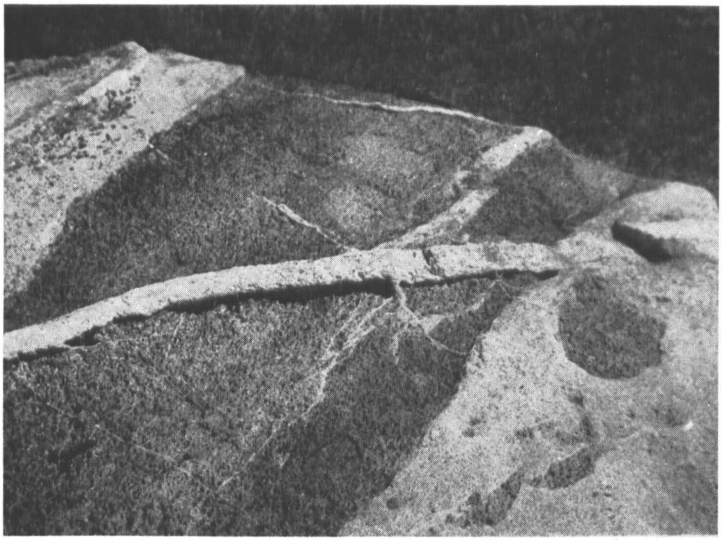


PLATE 10

Figure 3. In pasture on side of road opposite location of Fig. 2. Resistant metamorphosed shell and hollow interior of an inclusion.

Figure 4. Southwest slope of Wilcox Hill, ¾ mile northeast of Morgan. Age relations of two granites cut by pegmatite dike.

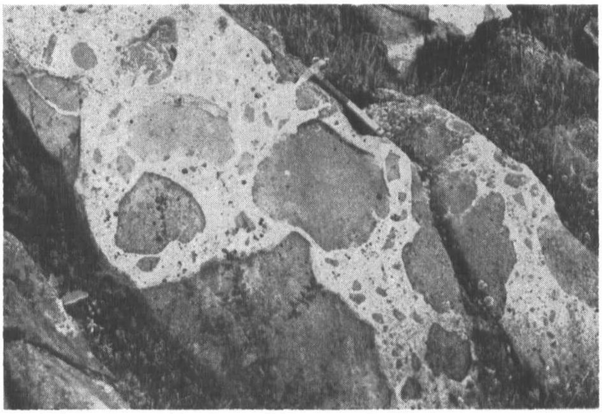
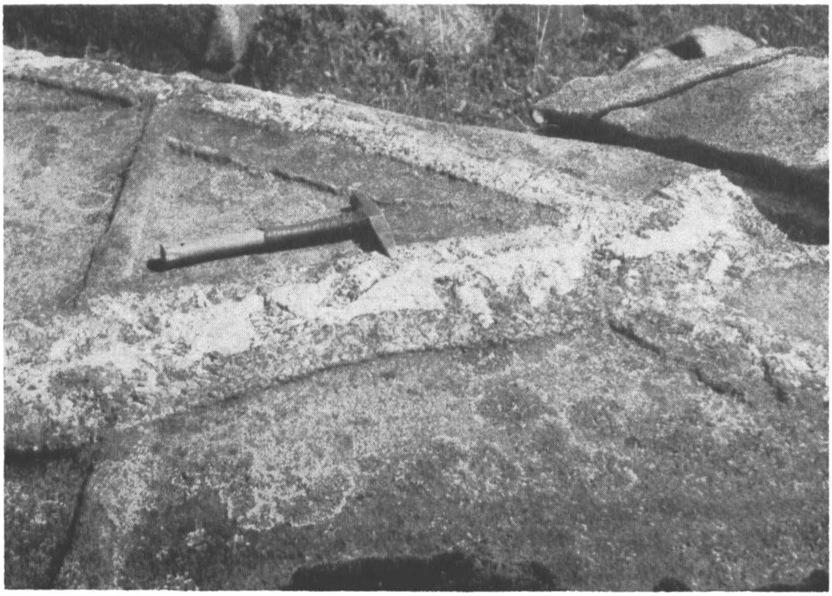


PLATE 11

Figure 1. West slope of 1600-foot hill, 1 mile southeast of West Charleston. Brecciated dark granite.

Figure 2. Escarpment along the Black River south of Newport City, above elevation 692. Granite sill folded with sediments. Barton River formation.



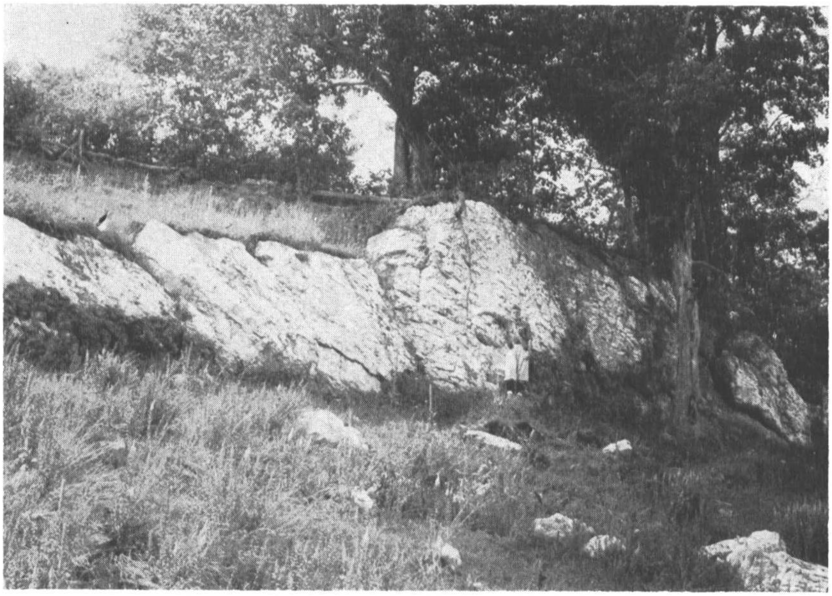


PLATE 11

Figure 3. Granite knob on northwest side of Barton Mountain, ⅞ mile southeast of Baird School. Two generations of pegmatite and an earlier aplite vein showing displacements.

Figure 4. On east slope, ¾ mile east of elevation 1003 in southern part of west rectangle, south of Coventry Center School. Milky quartz vein in Barton River formation. Man is 6 feet tall.

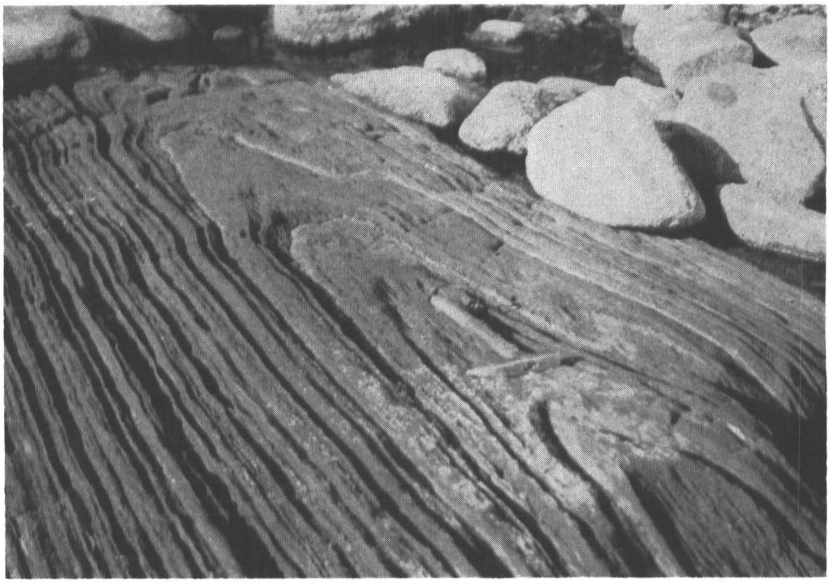


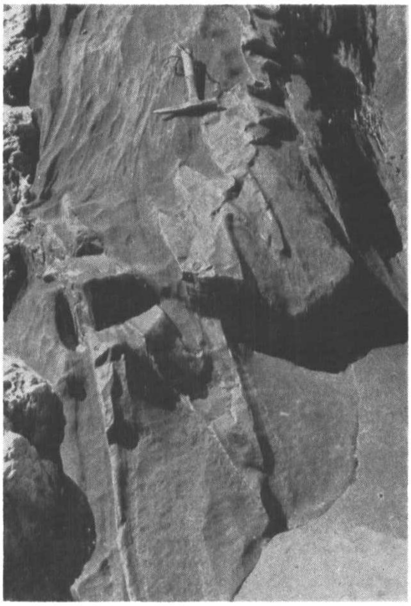
PLATE 12

Figure 1. Low hill ½ mile directly west of Echo Pond. Mass of quartz causing curve in strata of Westmore formation.

Figure 2. Cove Island in Lake Memphremagog. Minor fold in Ayers Cliff formation plunging northeast.

PLATE 12

Figure 3. South of Coventry Center School, $\frac{5}{8}$ mile northeast of elevation 1003. Fold in Barton River slates plunging northeasterly.



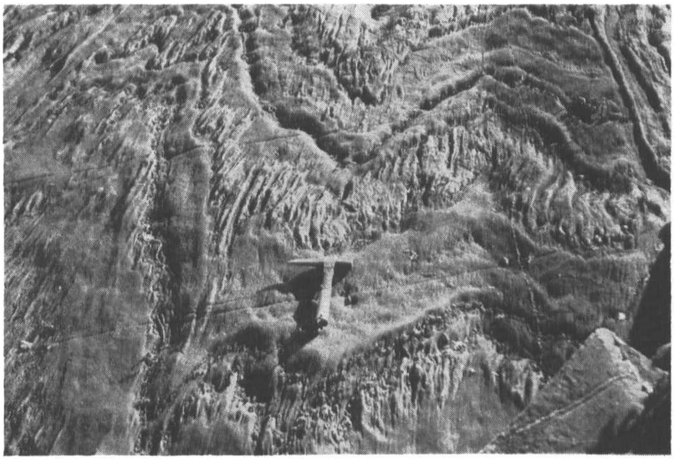


PLATE 13

Figure 1. Barton River ½ mile north of Country Club. Sharp drag-fold of phyllite layer in limestone. Barton River formation.

Figure 2. Shore of Lake Memphremagog, Lake Park. Fold in limestone traversed by closely-spaced shear planes. Ayers Cliff formation.

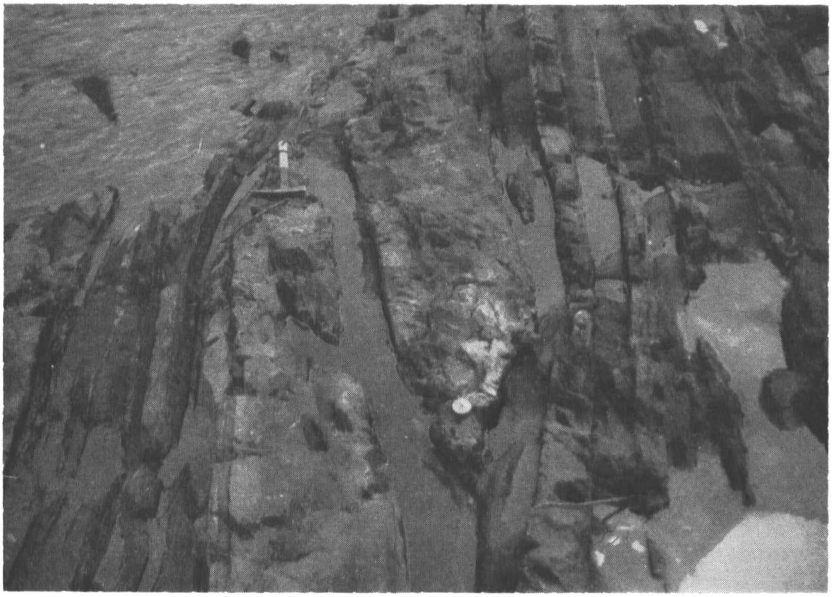
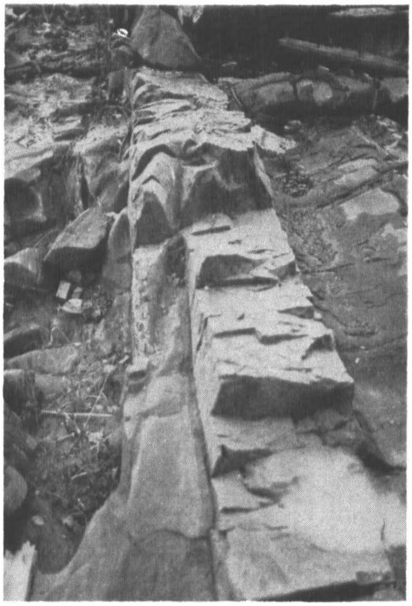


PLATE 13

Figure 3. Shore of Lake Memphremagog at Eagle Point. Marginal fractures and faults at border of granite pluton.

Figure 4. Clyde River at lower bridge, West Charleston. Longitudinally displaced bed of phyllite. Barton River formation.



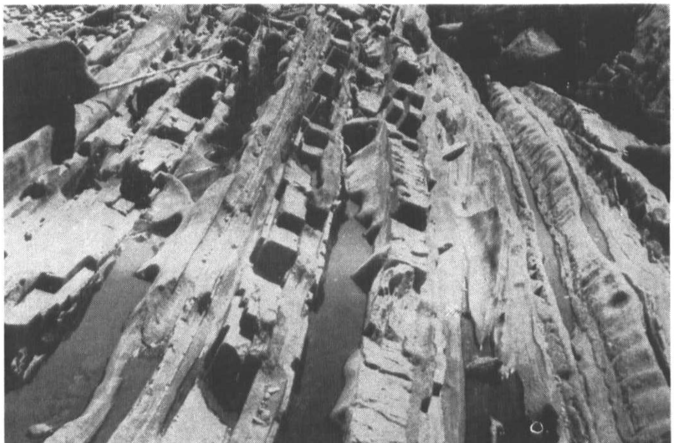


PLATE 14

Figure 1. Clyde River at lower bridge in West Charleston. Diagonal wedge-shaped fracture in phyllite layer filled with adjacent limestone. Barton River formation.

Figure 2. Below dam at Clyde Pond. Typical bedding in Barton River formation.

PLATE 14

Figure 3. West slope of hill south of Breezy Hill, ½ mile northeast of elevation 960. Bedding appears as folded fractures in slates. Barton River formation.

Figure 4. Southwest corner of Memphremagog quadrangle, 3½ miles west of Barton Village. Bedding disclosed by open fracture and solution holes. Slates in Barton River formation.

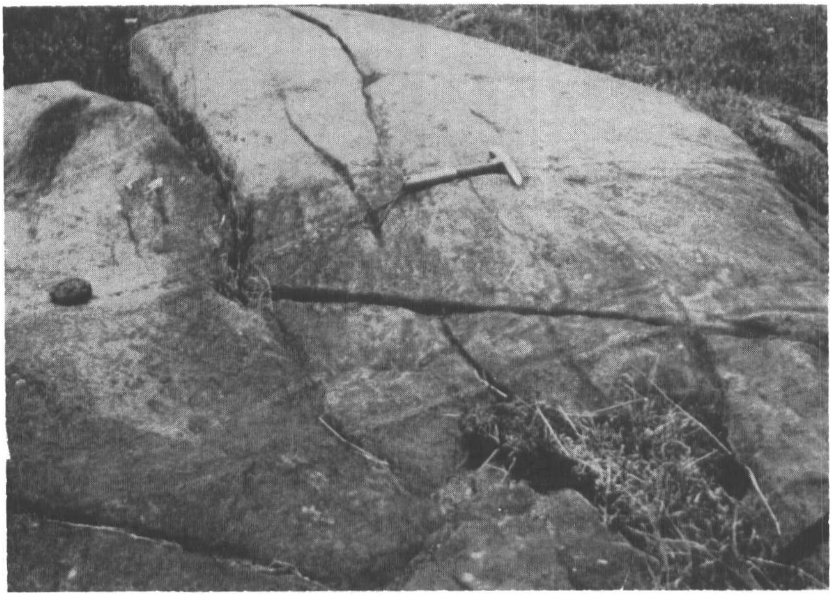
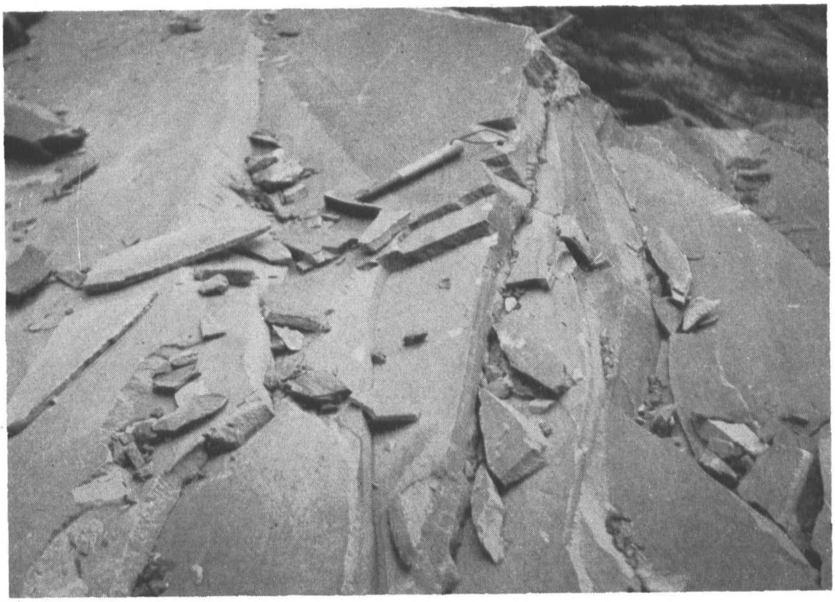


PLATE 15

Figure 1. Northwest end of Stony Hill, ¾ mile north of Brighton School. Bedding delineated by parallel grooves in slate. Hammer on highly contorted slate within 5 feet of granite contact.

Figure 2. Clyde River Valley, ½ mile north of Toad Pond. Bedding in amphibolite revealed by beaded divisions between beds.



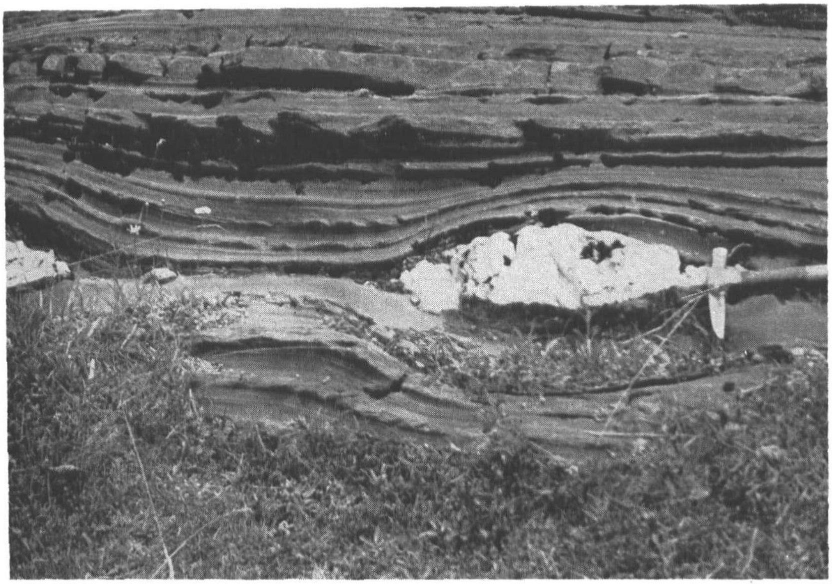


PLATE 15

Figure 3. Scythstone quarry, ¾ mile southeast of East Brownington School. Smooth bedding in phyllite of whetstone quality. Westmore formation.

Figure 4. Beside brook off Highway No. 5, 1 mile east of Newport City. Boudins of quartz in limestone of Barton River formation.

PLATE 16

Figure 1. Coventry Station. Bedding-cleavage relations in interbedded slates and limestones of Barton River formation.

Figure 2. Indian Point, ¼ mile south of elevation 721. Steeply-dipping bedding and practically horizontal cleavage in slates. Barton River formation.

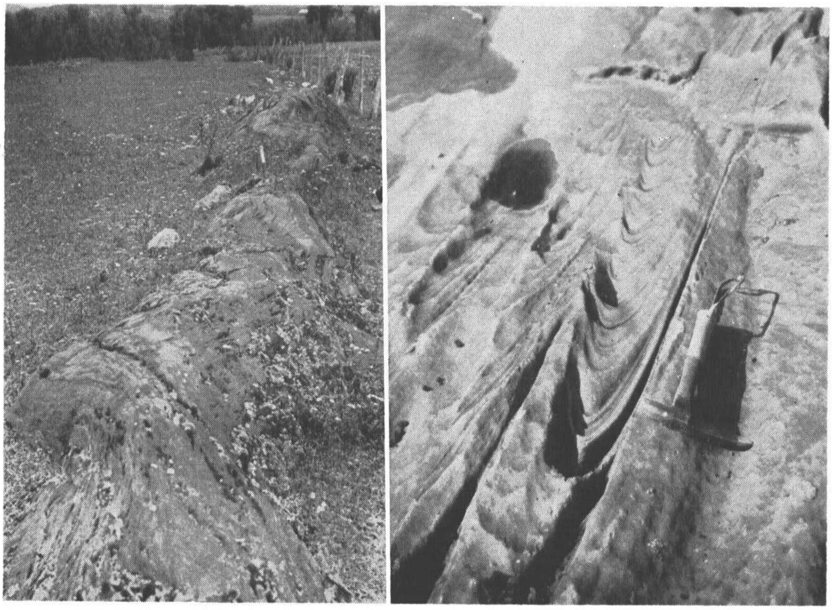


PLATE 16

Figure 3. North side of road, 5/8 mile south of Day School. Rolls in beds of limestone. Barton River formation.

Figure 4. At lower bridge on Clyde River at West Charleston. Flow fold in limestone of Barton River formation.

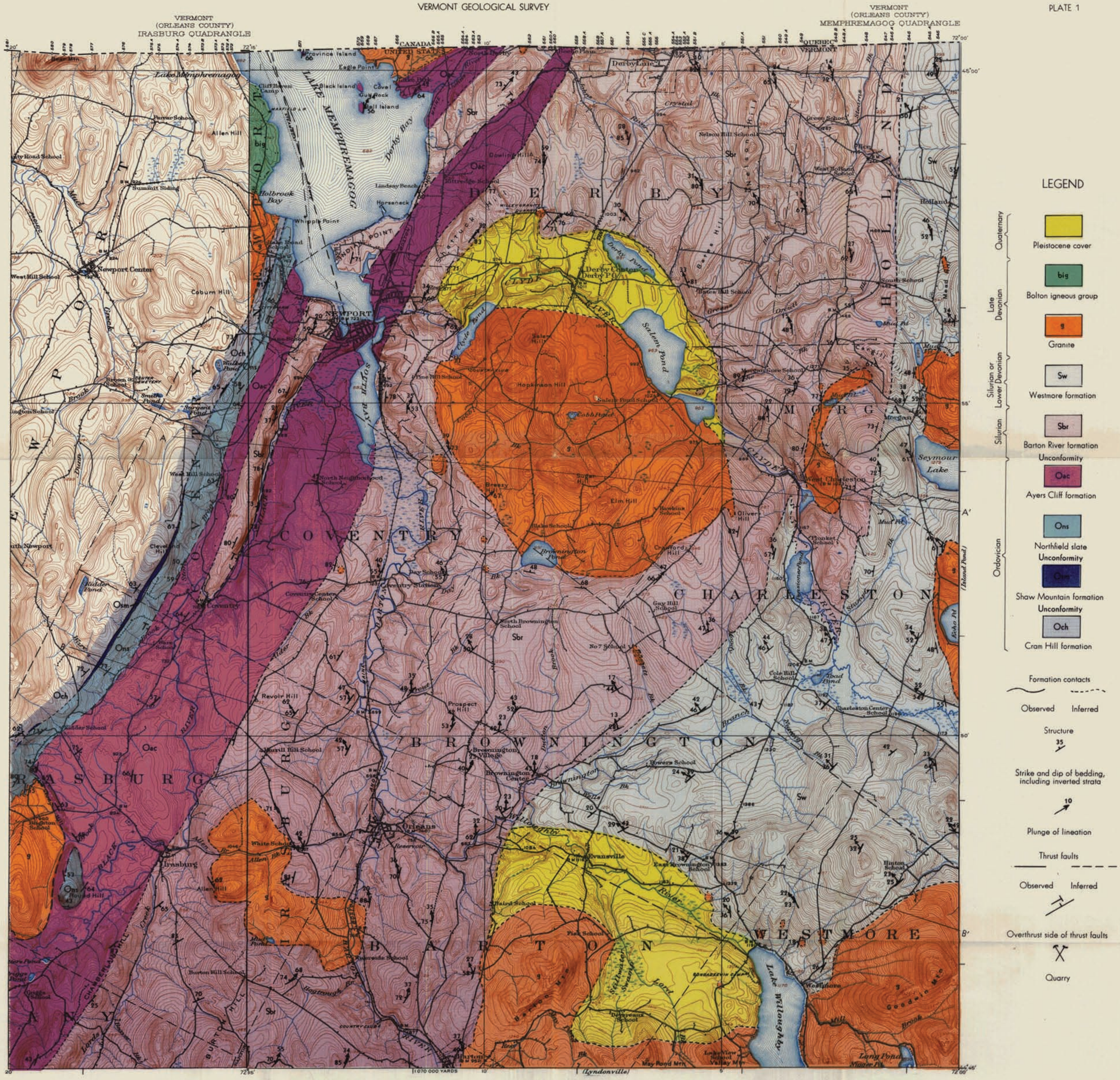
PLATE 17

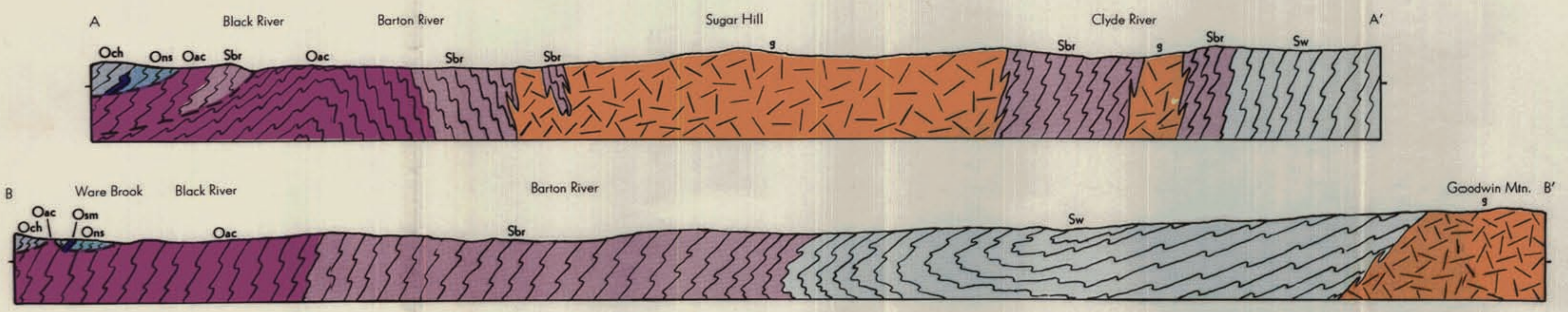
Figure 1. Shore of Lake Memphremagog at Lake Park. Intricate folds of thin slate layers in limestone of Ayers Cliff formation.

Figure 2. Abandoned road metal quarry beside Highway No. 5, 2⅛ miles northeast of Derby P. O. Limestone cores of synclinal folds overturned to northeast. Barton River formation.

PLATE 17

Figure 3. In pasture southeast of Orcutt Brook and ¾ mile northwest of Morgan Gore School. Sedimentary "inclusion" in limestone. Barton River formation.





GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE IRASBURG-MEMPHREMAGOG AREA, VERMONT

Topography by Hersey Munroe, Olinus Smith, G.E.Sisson, and F.H.Sargent
Control in part by International Boundary Commission, and U.S.Coast and Geodetic Survey
Surveyed in 1923

Polyconic projection. North American datum
5000 yard grid based upon U.S. zone system, A

Contour interval 20 feet
Datum is mean sea level

Geology by Charles G. Doll
Published in 1951