THE GEOLOGY OF THE ENOSBURG AREA, VERMONT

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TABLE OF CONTENTS

List of Illustrations

Plates

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ABSTRACT

The area here described extends broadly from the Hinesburg thrust in the west, to the west flank of the Green Mountain anticlinorium to the east. The central feature is the Enosburg anticlinorium.

The Oak Hill succession in Vermont is re-defined as a result of recent work. To the west, the succession is miogeosynclinal in character, and easily classified. It changes eastward into a eugeosynclinal sequence difficult to classify stratigraphically.

The movement pattern as revealed by minor structures is displayed to tell its own story: In the Green Mountains, bedding schistosity (s_1) associated with down-dip minor folds indicates stretching, implying that updoming was the deforming mechanism of the anticlinorium. Farther west, the same schistosity becomes "axial plane" cleavage to folds which face west.

A later spaced cleavage (s_3) in the Green Mountains is steep and is associated with shear folding. The overall tectonic pattern seems to emphasize vertical movement as the primary cause of orogenic deformation.

Metamorphism is of chlorite grade throughout, but intensity indicated by grain size increases from west to east throughout the area.

Iron and lime, and possibly copper, have been exploited locally in the past. Gravel is of some limited local economic importance at the present time.

INTRODUCTION

Location

The present Bulletin deals with the geology of the Enosburg Falls quadrangle, Vermont, and adjacent areas in the St. Albans and Jay Peak quadrangles (see Fig. 1).

The Enosburg Falls quadrangle is located between latitudes 44°45′ N and the Canadian border, and longitudes 72°45′ and 73°00 W. The

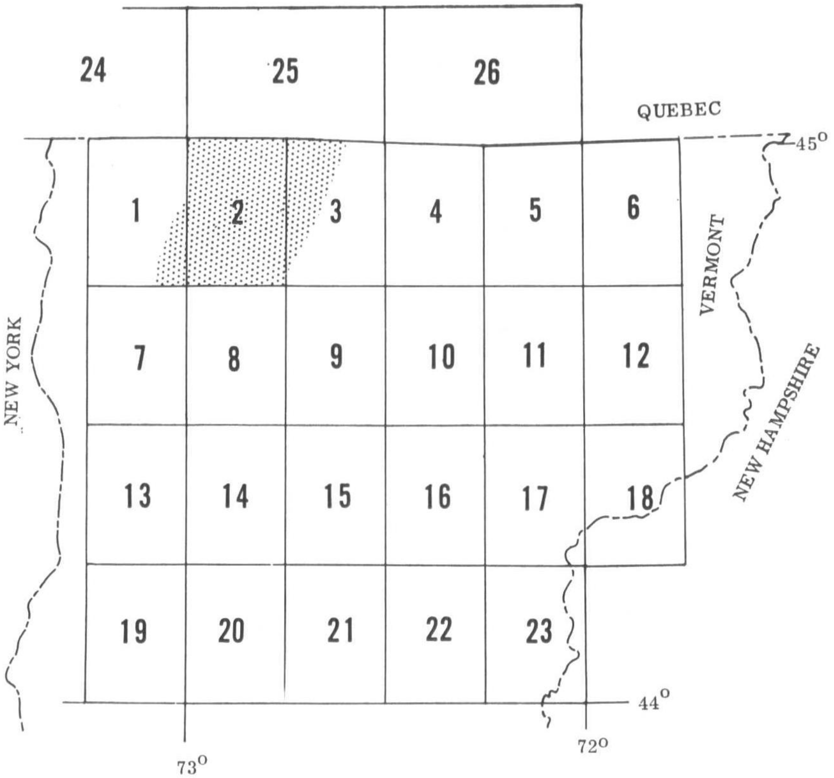


Figure 1. Key Map.

Names of quadrangles:

1. St. Albans
2. Enosburg Falls
3. Jay Peak
4. Irasburg
5. Memphremagog
6. Island Pond
7. Milton
8. Mount Mansfield
9. Hyde Park
10. Hardwick
11. Lyndonville
12. Burke
13. Burlington
14. Camels Hump
15. Montpelier
16. Plainfield
17. St. Johnsbury
18. Littleton (N. H.)
19. Middlebury
20. Lincoln Mountain
21. Barre
22. East Barre
23. Woodsville (N. H.)
24. Lacolle (Qué)
25. Sutton (Que)
26. Memphremagog (Qué)

Area of this bulletin is stippled

St. Albans quadrangle adjoins it on the west, the Jay Peak quadrangle on the east. Vermont State highways 105 and 36 traverse the area from west to east, and State highway 108 reaches Enosburg Falls from the south. The Central Vermont railroad maintains a freight service along

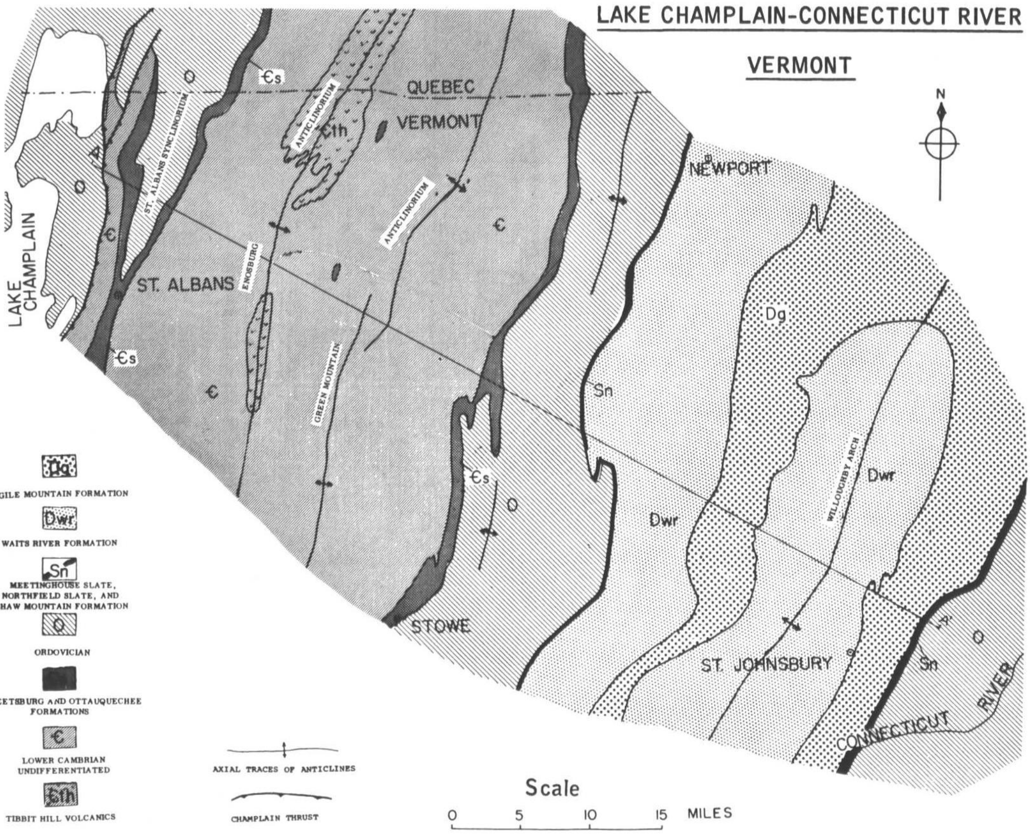


Figure 2. Regional geologic setting.

the Missisquoi valley, and the St. Johnsbury and Lamoille County railroad traverses the area from southeast to northwest. The old Montpelier to Montreal stage passed through Enosburg Center and Berkshire.

Geologic Setting

The Enosburg area is in the western foothills of the Green Mountains. Geologically it is part of a belt of low-grade metamorphic rocks of Cambrian and Ordovician age, which form the western margin of the Acadian orogenic zone of eastern North America. This belt is bounded to the west by the Champlain thrust and to the east by the Green Mountain anticlinorium (see Fig. 2). The rocks form part of a typical eugeosynclinal sequence; subdivision by time-rock units is difficult and hazardous, and rapid facies changes complicate correlations.

Previous Work

Adams (1846) and Hitchcock (1861) visited the area and commented on it, but can hardly be considered to have "worked" on it. Logan (1863) made a reconnaissance survey in adjacent Quebec. The first coherent, but rather general account was by Keith (1923; also 1932). Later Schuchert (1937) and Cady (1945) wrote on the rocks and structure of the area. Clark (1934 and 1936) mapped adjacent equivalent rocks in Quebec in great detail; his excellent map (unpublished) formed the basis of the first consistent subdivision of the rocks concerned. Booth (1950) published the first detailed map of the same rocks in Vermont, partly in the area covered by this Bulletin.

The extreme western margin of this Bulletin covers part of the eastern margin of the area mapped by Shaw (1958). Figure 3 shows the extent of mapping by Booth and Shaw.

Method of Study

Field mapping was carried out in the summers of 1957 and 1958. A few rapid checks were made in the summers of 1959 and 1960, while the author was principally engaged elsewhere.

The U. S. G. S. one-inch maps of the areas concerned, and aerial photographs, served as topographic bases. During the work it became clear that the local topographic maps are out of date and inaccurate in many details; as a result, a few discrepancies may have crept into the location of geologic contacts. In mountainous, forest covered terrain, location was by means of an aneroid barometer.

The present Bulletin had originally been intended to continue Booth's (1950) mapping toward the east. When it became clear that Booth's map required substantial revisions, the present survey was extended to include all of the area originally mapped by Booth (*loc.cit.*). This revision concentrated mainly on critical areas; in the intervening areas traverses were more widely spaced, and much of Booth's work there was used as a guide. The southern portion of the revision is being published in the Vermont Geological Survey Bulletin No. 26 on the Milton area (Stone and Dennis, in press).

Acknowledgments

The writer is indebted to Charles G. Doll, State Geologist, who suggested the work and gave constant support. Wallace M. Cady, of the U.S.G.S., gave much valuable help and advice, both in the field and with the manuscript. Field conferences with workers in adjacent areas always resulted in valuable contributions. Many days were spent with M. J. Rickard then of McGill University, whose approach to structural problems was most stimulating and helpful, with P. H. Osberg and with W. M. Cady, who have unraveled much of the complicated stratigraphy in adjacent areas, and with R. A. Christman who was the first to give close attention to the Tibbit Hill Volcanics.

Throughout the field work the writer has had occasion to appreciate the helpful attitude of the local residents toward his work. He would particularly mention Dr. L. Judd, and Mr. F. Tupper who has an intimate knowledge of the mineralogy of the area. The writer has also been assisted in his work by the cooperative attitude of the U. S. Border Patrol, and of the Canadian Border officials.

Rosemary Donica ably typed the manuscript and Pamela White helped with the drafting. The laboratory and office work was in a large part financed from a research grant by Texas Technological College.

Physiography

A gently undulating plain, the Champlain Lowland, extends from Lake Champlain to a west-facing escarpment comprising Bellevue Hill and Rice Hill. The area here described begins just west of that escarpment, which is a dominant feature aligned from southern Quebec as far as central Vermont. Booth (1950), following Clark, has named it the Oak Hill escarpment. Since it is somewhat discontinuous between the Canadian border and the name locality at Oak Hill near Dunham, Quebec, a Vermont name seems preferable here. Traditionally (Jacobs, 1950), the escarpment has been known as the Green Mountain Front. More commonly, however, the Green Mountains are held to begin some 12 miles farther east. Fairfield Hill escarpment (a local name) seems more appropriate. It is carved into the tough quartzose graywackes of the Cheshire Formation which are thrust over the younger rocks of the lowlands. A number of tectonic outliers or klippen to the west are witnesses of this thrusting. These klippen stand out in the topography as isolated hills, for they are predominantly composed of erosion-resistant Cheshire Graywacke.

East of the escarpment the Green Mountain foothills form a number of north to northeast trending low ridges. Superimposed streams, such as the Missisquoi and its larger tributaries, are evidence of an early peneplenation. Booth (1950, p. 1134) points out that this peneplain must have been drained toward the west. The Green Mountains proper begin with the Cold Hollow mountains, which rise to 2000-3000 ft. at the southeastern limit of the area. Since the region of high elevations and relief coincides with a zone of higher grade metamorphism, it seems possible that the relief of the Green Mountains was to a large extent determined by the higher erosion resistance of higher grade metamorphic rocks.

The whole of the area was glaciated, at least during the Wisconsin glacial epoch. Erratics and morainal debris are common. A number of striae orientations have been recorded (see Pl. 1). A sand pit some 1½ miles northeast of Greens Corners reveals recumbent folding due to a moving ice mass. Examples of roches moutonnées may be found in many parts of the area. The meandering of the Missisquoi is characteristic for the collector stream of a maturely dissected region. Entrenched meanders (e.g. at Sheldon Springs and Enosburg Falls) are evidence of post-glacial uplift.

STRATIGRAPHY

Introduction

The Enosburg area is underlain entirely by a eugeosynclinal type lower Paleozoic succession. Except for the Ordovician Morses Line Slate in the far northwest corner, all the rocks are Cambrian or older. These rocks have long resisted stratigraphic resolution, partly because of the paucity (indeed, near-absence) of fossils, partly because of the rapid facies changes and lensing typical of rocks deposited in a eugeosyncline.

Clark (1934) was the first to attack the problem in the only way that offered any promise: detailed outcrop mapping in an area in which relative abundance of outcrops and reasonably good individualization of units offered prospects for positive results. Clark chose a strip across the strike, adjoining the area here discussed immediately north of the Canadian border. The stratigraphic sequence which he established, known as the Oak Hill succession, is shown in Fig. 3.

Booth (1950) attempted to follow Clark's units southward into Vermont. His mapping was on a smaller scale, and there was some misunderstanding as to the identity of some of Clark's units. In the course of the present work it has been possible to resolve these difficulties and also to establish a link with Cady's work (1945; and in press) to the south.

All the rocks here described have undergone low-grade metamorphism and belong to the greenschist metamorphic facies. Recrystallization due

Zone	ST. ALBANS SYNCLINORIUM		OAK HILL SUCCESSION	"OAK HILL SUCCESSION IN VERMONT"	ENOSBURG ANTICLINORIUM	RICHFORD AND CAMBRIDGE SYNCLINES	BROME SYNCLINE (QUEBEC)	GREEN MOUNTAIN ANTICLINORIUM	
Sources	Howell 1938 Keith 1932 Schuchert 1937 Shaw 1958		Clark 1934 1936 Cook, unpubl. Rickard, unpubl.	Booth 1950	Dennis, this bulletin	Cady, in press Dennis, this bulletin	Rickard (unpubl.) Osberg (unpubl.)	Albee, 1957 Cady, in press Christman, 1959 Konig & Dennis, in press	
ORDO-VICIAN	Morses Line		Stanbridge		Morses Line			Missisquoi = Beauceville	
	Highgate							Stowe	
CAMBRIAN	Gorge								
	Woods Corners	Hungerford	Melbourne / Vail						
		Saxe Brook / Rock-ledge	Sweetsburg		Sweetsburg	Sweetsburg	Sweetsburg	Ottauquechee	
		Skeels Corners		Skeels Corners					
		Mill River							
		St. Albans							
	Rugg Brook		Scottsmore	Rugg Brook	Bridgeman Hill	(Bridgeman Hill)	Scottsmore		
	Parker		Oak Hill	Parker			Oak Hill		
	Dunham		Dunham	Dunham			Dunham		
	Gilman		Gilman	Gilman / "Jay Peak"	Cheshire	Underhill	Bonsecours	Camels Hump	Pinney Hollow
? CAMBRIAN			West Sutton	West Sutton	Fairfield Pond				
			White Brook	White Brook	White Brook	White Brook	White Brook		Hazens Notch
			Pinnacle	Pinnacle	Pinnacle	Pinnacle	Pinnacle		
			Call Mill						
			Tibbit Hill	Tibbit Hill	Tibbit Hill				Underhill

Figure 3. Correlation chart. Vertical line pattern: Equivalent units not exposed here.

to metamorphism will be more fully discussed in the chapter on metamorphism.

Pinnacle Formation

Name and Distribution

The name Pinnacle Formation here includes Clark's Tibbit Hill Schist, Pinnacle Graywacke, and Call Mill Slate (see Fig. 3). This reclassification became necessary because the Tibbit Hill was found interbedded with typical Pinnacle Graywacke near East Fairfield. The Call Mill Slate occurs very sporadically in Vermont, never in outcrops large or consistent enough to map, and is, therefore, also included in the Pinnacle Formation.

The rocks of the Pinnacle Formation were first described by Logan (1863, p. 246). Logan did not differentiate between the volcanics and the graywackes. He called the rocks below the White Brook (unnamed then) "chloritic, micaceous and epidotic rocks." He erroneously considered them to be at the core of a syncline.

The bottom of the Pinnacle as here defined is not exposed in the Enosburg area, but the formation unconformably overlies the Precambrian Mount Holly Complex in the Middlebury quadrangle (lower part of Mendon Formation, Osberg, 1952; Pinnacle Formation, Cady, personal communication, 1958). The top of the Pinnacle is, by implication (Clark, 1936) defined by the overlying White Brook Dolomite or its equivalents, where present. Definition by lithology of the Pinnacle itself is not practicable, because of the very great facies variations within the formation. Where the White Brook is absent, the top of the Pinnacle is taken to be at the stratigraphically highest appearance of the coarse graywacke facies.

In Clark's usage the type locality of a formation is not necessarily its name locality. Thus, the type Pinnacle, as defined by Clark, is in "lot one, range one, Dunham (Quebec)." The name is taken from Pinnacle Mountain near Abercorn (Quebec). The type locality Pinnacle is a coarse graywacke, the Pinnacle capping Pinnacle Mountain is essentially phyllitic (V. Booth, personal communication to Cady, 1947). These two lithologies, coarse graywacke and quartz-chlorite-sericite phyllite, make up the bulk of the metasedimentary rocks of the Pinnacle Formation in Vermont.

Graywacke

The graywackes of the Pinnacle Formation were first defined and given stratigraphic status by Keith (1932, p. 394) who called them Nickwaket Graywacke. The term graywacke is a controversial one. The original definition of *Grauwacke*, the German term from which graywacke is derived, is: "Sandstone made up of fragmental granite debris" (Lasius, 1789, quoted in A.G.I. Glossary of Geology, 1957).

The representative graywacke is traditionally the graywacke of Tanne in the Harz mountains[1]. Similar rocks make up a high percentage of the rocks in eugeosynclinal belts. Most geologists would recognize a typical graywacke, but when it comes to defining it, opinions differ widely; all seem to agree now (1961) that rocks of the Tanne type must be comprised within the meaning of the term.

There are a number of good recent discussions of graywackes (e.g. Krynine, 1948; Helmbold, 1952 and 1958; Pettijohn 1954). The writer does not feel called upon to renew the discussion, but would merely refer to the above papers, and to the second edition of Pettijohn's text (1959). Pettijohn suggests that a true graywacke should have some of the diagnostic features of a turbidite (graded bedding, lack of bedding planes, lack of or poor sorting). The graywacke facies of the Pinnacle is indeed poorly sorted, sometimes has graded bedding, and comprises at least one exposure of slump breccia. But the beds of the graywacke facies are often quite thick (up to 10 feet at least) while in the case of "classical" graywackes each bed is considered to be the product of one turbidity current, never more than 2–3 feet thick. Yet, the graywacke facies of the Pinnacle is a polymineralic microbreccia, with usually too much groundmass to be called an arkose. On balance, the arenitic facies of the Pinnacle would be considered a graywacke in both Pettijohn's and Krynine's classifications.

H. C. Cooke (personal communication, 1953) has followed the Pinnacle northeastward from Clark's type area, and found that it grades into a white metaquartzite south of the St. Francis river.

In the Enosburg Area the arenitic facies of the Pinnacle predominates between the Canadian border and the Missisquoi valley. South of the Missisquoi, the proportion of phyllitic interbeds (Underhill facies) increases considerably. Locally, quartz pebble and polygenous boulder conglomerates appear near the top of the formation. The boulders are irregularly spaced within the graywacke matrix, they appear to "float" in it. This is normally a good indication of turbidite sedimentation.

[1] Often incorrectly rendered as "Tanner graywacke"; "Tanner graywacke" would be the correct stratigraphic name in English, for inflexions of proper names must not be carried from one language into another.

The most striking lens of boulder conglomerate makes up a bald hill one mile due east of Bellevue Hill, just south of the St. Albans-Fairfield town line.

Petrography. The petrography of the graywacke and conglomerate facies of the Pinnacle is given in great detail by Booth (1950, p. 1142–1145).

Underhill Facies

Phyllitic interbeds in the Pinnacle are common and their proportion within the formation increases considerably from north to south and from west to east. In the adjacent Mount Mansfield quadrangle Cady and Christman (personal communication, 1960) have found it convenient to group all phyllitic rocks in a separate formation, the Underhill Formation. This is a magnafacies, as defined by Caster (1934). The connected problem of stratigraphic nomenclature will be discussed below with the Underhill Formation.

The chief minerals of the Underhill phyllites are quartz, sericite and chlorite, with accessory magnetite and albite. Small amounts of heavy minerals can be seen in many thin sections (zircon, rutile, apatite, tourmaline, sphene, epidote, as well as pyrite, muscovite, biotite and hornblende). Quartz grains are often present as phenoclasts, well-rounded and about 0.5mm in diameter. Occasionally lithic fragments can be observed, e.g. west of the bridge south of East Fairfield village. Here phyllitic fragments of all sizes are set in a silty phyllitic matrix (Fig. 4). This seems to be a slump breccia. Elsewhere lithic fragments usually escape identification, owing to the rather uniformly fine grain of the phyllites, which would reduce such fragments to the size of their constituent minerals. Booth has found a number of fragments of igneous origin (Booth, 1950, p. 1142). Quartz veins, often parallel to the bedding or schistosity, are common, and may represent both quartz segregations and recrystallized chert.

It seems that, in Krynine's classification of sedimentary rocks, these phyllites would rank as fine-grained graywackes.

Tibbit Hill Volcanics

The lower part of the Pinnacle Formation, as exposed in the Enosburg area, largely consists of a succession of dark green metavolcanics interbedded with the graywacke facies, and, more rarely, with a few beds of the phyllite. Originally Clark (1936) had mapped these volcanics as a separate formation, the Tibbit Hill Schists. The name is taken from

Figure 4. Large fragments in Pinnacle graywacke. East Fairfield.

Tibbit Hill, a little over a mile southwest of Brome Lake near Knowlton, Quebec. In the Sutton, Quebec, map area the outcrop of this formation is wide and unified, and appears to underlie the Pinnacle Graywacke at all known contacts. Hence Clark's classification of the Tibbit Hill as the oldest formation of the Oak Hill succession was perfectly justified. In the Enosburg area, however, the volcanics appear to diminish in volume from north to south, and at the same time they acquire intercalations of graywacke and phyllite which can be shown not to be pinched anticlines: One band of pillow lava south of East Fairfield has pillows whose tops uniformly face east. This band is both overlain and underlain by graywacke. Hence graywacke and volcanics are interbedded, making it preferable to combine both in one formation. In the Lincoln Mountain and Middlebury quadrangles (Cady and others, 1962) Pinnacle Gray-

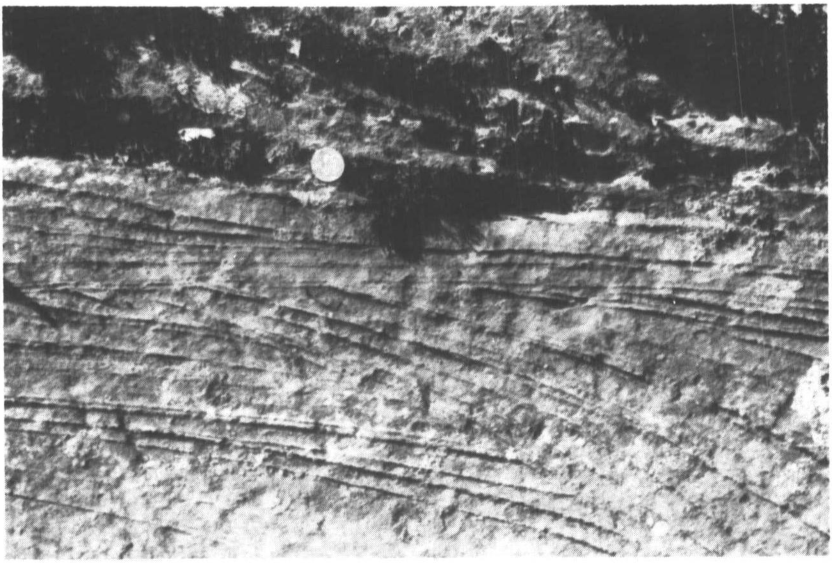


Figure 5. Cross-bedding in Tibbit Hill. West of Richford.

wacke directly overlies Precambrian; so that either the lowest Pinnacle is a graywacke, or the volcanics, while possibly at the bottom of the formation, die out toward the south.

Most outcrops of Tibbit Hill are very distinctive in the field. They are fairly massive, fine grained, usually dark green, sometimes vesicular and often characteristically pitted. The typical Tibbit Hill has no distinctive schistosity, but there are more schistose varieties which often resemble fine-grained graywacke or phyllite. These may be metamorphosed tuffs and/or water-deposited mafic volcanic detritus. In doubtful exposures the presence of epidote usually confirms basic igneous origin; that is, Tibbit Hill as against metasediment. There are a few unusual occurrences, such as pods of pistachio-green epidote-rich material near Ayers Hill, and pillow lavas south of East Fairfield (either side of the road south of the abandoned Lapland school). Lenses of calcite and calcite-filled amygdules are quite sporadic.

The eastern border facies of the Tibbit Hill Volcanics is partly detrital, and often the contact between volcanics and graywackes is uncertain. On the road from West Berkshire to Richford, at elevation 642, an outcrop of what looks like good characteristic Tibbit Hill Volcanics is, on closer examination, cross-bedded (see Fig. 5).

The common minerals that make up these metavolcanics are albite, epidote and chlorite. Occasional amphibole (actinolite?) becomes more abundant toward the south. Christman (1959) has confined the name Tibbit Hill to greenstones in which amphibole and/or feldspar can be recognized in the field. This was not the intent of Clark when he defined the Tibbit Hill; the abundance of amphibole appears to be a characteristic of metamorphic grade rather than composition. In the Enosburg area, all greenstones within the Pinnacle have been included in the Tibbit Hill Volcanics. The micropetrography of the Tibbit Hill has been so well described by Christman (1959) and also by Booth (1950), that it seems unnecessary to repeat these descriptions here.

Age

As stated above, the Pinnacle is known elsewhere to overlie the Precambrian Mount Holly complex. It has not yielded any fossils or radiometric ages. Since it is below the *Olenellus*-bearing Cheshire Formation, its age may be late Precambrian or what has sometimes been named "Eocambrian": conformably below the oldest known fossil-bearing rocks. On the geologic map of Vermont such rocks are shown as Cambrian (?).

Underhill Formation

Name and Distribution

The rocks of the Underhill Formation are mainly greenish quartz-chlorite-sericite phyllites lying stratigraphically between the Pinnacle and Cheshire formations, where present. Their type locality is in the township of Underhill, in the Mount Mansfield quadrangle (W. M. Cady, personal communication, 1960). The present writer would place the rocks of the type locality within the Underhill facies of the Pinnacle Formation, for they are clearly stratigraphically equivalent to rocks of the Pinnacle Formation in the Enosburg area, being below an excellent marker horizon, the White Brook dolomite and slate. However, the Underhill facies of the Pinnacle and the phyllites of the Underhill Formation are practically indistinguishable in the field, and so it is unavoidable, wherever the dividing White Brook dolomite and slate are absent, to map all rocks of Underhill facies as one unit. In the western outcrop belt Underhill rocks are well defined, between the White Brook Dolomite or coarse Pinnacle Graywacke below and the Cheshire Formation above. Rocks in this clearly defined interval are here recognized as a separate member within the Underhill Formation, the Fairfield Pond Member, which is thus a parvafacies (Caster, 1934).

Fairfield Pond Member

The Fairfield Pond Member includes the West Sutton Slate and the lower (phyllitic) part of the Gilman "Quartzite" of Clark (1936). In this report the name "Gilman" has been abandoned, because there has been some confusion as to its application.

As mentioned before, Clark distinguished between the *name locality* and the *type locality* of a unit. The name locality of his Gilman Quartzite, Gilman, Quebec, is actually underlain by green quartz-chlorite-sericite phyllite of Underhill type. The type locality for the Gilman, on the other hand, is the top of Oak Hill, Quebec (Clark, 1936, p. 144), where the rock is an impure quartzite which correlates with the Cheshire Formation of this report (see below). Since the quartzitic upper part of Clark's Gilman is readily distinguishable in the field from the underlying Fairfield Pond phyllite, and has indeed been recognized separately by Keith (1923) as Cheshire Quartzite, this report distinguishes the upper quartzitic part of Clark's Gilman as the *Cheshire Formation* and lower phyllitic part of Clark's Gilman as the *Fairfield Pond Member* of the Underhill Formation which also includes the West Sutton Slate.

In the eastern exposure belt, where the Cheshire is absent and the lower contact of Underhill type rocks is uncertain on the eastern limb of the Richford syncline, Fairfield Pond equivalents are mapped with possible older rocks as Underhill Formation, undivided. This same belt correlates, in part, with the western part of Clark's (1936) Sutton Schists, and Osberg's (manuscript map, 1958) and Rickard's (manuscript map, 1959) Bonsecours Formation in adjacent areas in Quebec. Farther north in the Warwick area of Quebec, H. C. Cooke and the writer (manuscript map, 1953) had mapped equivalent rocks as Gilman.

Cady (U.S.G.S., *in press*) and J. B. Thompson, Jr., have followed Fairfield Pond and Underhill type rocks southward and have established their probable partial stratigraphic equivalence to Perry's (1928) Pinney Hollow Formation, to which they bear a great resemblance in the field. Cady has found that Keith's (1932) Moosalamoo Phyllite is equivalent to the Fairfield Pond Member.

Booth (1950) was uncertain as to the application of Clark's names Gilman and West Sutton. Consequently he continued mapping Clark's Gilman, both quartzite and phyllite, southward as such, but near Sheldon he swung its lower contact west, so that, toward the southern part of his map, Booth's Gilman included little but the quartz-rich Gilman; in other words, the Cheshire. At the same time Booth inadvertently extended the application of Clark's West Sutton, including within it the phyllitic parts of the Gilman that he excluded from his new interpretation of that name, as well as considerable amounts of Pinnacle graywacke and phyllite, including intervening White Brook Dolomite. These facts must be borne in mind when referring to Booth's description of the area.

White Brook Member

The White Brook is an excellent marker horizon. At its type locality, the headwaters of White Brook south of Sutton, Quebec, (Clark, 1936) it is a dolomite of varying light coloration, usually whitish to pinkish or cream. The weathered surface is very conspicuous: a dull buff to brown, often on a low erosion-resistant ridge. The erosion resistance is largely due to a high quartz grain content, as well as quartz pods and quartz stringers criss-crossing the dolomite. Bedding has never been observed in Vermont, though Clark (1936) mentions outcrops showing sand grains concentrated along bedding planes. In some outcrops there is an abundance of irregular pods of whitish marble, possibly of secondary origin. Undoubted White Brook Dolomite has not been seen south of the Lamoille river, but Cady (manuscript map, 1960) has correlated it with the Forestdale Member of the Underhill Formation in the Middlebury area. The Forestdale Marble (Keith, 1932, p. 394) occupies the same stratigraphic position as the White Brook. On the explanation of the above-mentioned map Cady has also implicitly correlated the White Brook with the Battell Member of the Underhill, Osberg's (1952) Battell Member of the Monastery Formation. The Battell (Osberg, 1952, p. 44) is a black graphitic quartz-muscovite schist with lenses of dolomitic marble. It occupies a stratigraphic position above the Pinnacle Formation which may well be equivalent to that of the White Brook. In the Enosburg, Sutton (Quebec) and Mount Mansfield areas black slate bands often mark the top of the Pinnacle in its eastern exposure belt. In at least one occurrence this slate is associated with dolomite that closely resembles the type White Brook: the best exposure is one mile south of the Bakersfield village cross-roads, west of Route 108, where the dolomite is associated with black slate and also with a greenstone-like chlorite schist band. Therefore, failing contradicting evidence, it seems reasonable to assume with Cady that the discontinuous black slate band at the eastern limit of the Pinnacle is indeed in or close to the same stratigraphic position as the White Brook. Hence the White Brook Member is held to include both the dolomite of Clark and the black slate and limestone of the Enosburg and adjacent areas.

In many places hematite from the overlying West Sutton Slate (see below) has become enriched at the top of the White Brook Dolomite, and at least two formerly exploited iron pits are known in the town of Sheldon (see Pl. 1): one is 1 mile east of the village, the other about 1.2 miles south of the village; both are marked on the geological map, Plate 1. It may not be coincidental that both are near the hinges of minor folds, though one is a syncline, the other an anticline.

A dolomite lens marked ?Ꞓuw on Plate 1, on the Canadian border NNW of Richford village, may be White Brook Dolomite. It, too, is associated with black slate. White Brook here would imply that the phyllite between this dolomite and the Tibbit Hill farther west is Pinnacle, *i.e.* the Underhill facies of the Pinnacle. Such an interpretation would not meet serious difficulties on the Quebec side of the border. Rickard (manuscript map, 1959) has mapped the same rocks as Bonsecours which, in this position, would be equivalent to Underhill. It is strictly a matter of local definition and interpretation, and this writer prefers to accept, as far as reasonably possible, the White Brook as a marker horizon, marking the bottom of the Underhill Formation.

An extensive petrographic description of the White Brook Dolomite appears in Booth (1950, p. 1145–1147).

West Sutton Slate

This member sporadically overlies the White Brook. The name and type locality are both in the vicinity of West Sutton, Quebec (Clark, 1936, p. 143). It is a well-cleaved slate, usually with a characteristic reddish tinge due to disseminated hematite. The member is continuously mappable in Quebec, starting from some distance north of the border, but, with an average thickness of less than 100 feet, it cannot appear on a one-inch map (the White Brook, of course, should also be omitted from a one-inch map, but it has been included in part on Plate 1 with locally somewhat exaggerated thickness because of its importance as a marker horizon). In Vermont the West Sutton is rarely exposed, and probably mostly absent. It nowhere in Vermont appears to exceed the thickness noted by Clark in Quebec. North of the abandoned hematite pit one mile east of Sheldon (see Pl. 1), there is a fairly wide exposure area of the West Sutton, in the crest of a fold.

Booth's (1950) interpretation of the West Sutton, as noted before, is erroneous. The bulk of his "West Sutton" (including the true West Sutton) is really Fairfield Pond, the remainder Pinnacle or Cheshire.

Bonsecours Facies

The eastern (Richford syncline) belt of Underhill rocks is very similar in appearance to the Fairfield Pond rocks of the west. However, instead of repeating the rather monotonous green phyllite with quartz lenses, the rocks in the Richford syncline have, in addition to that lithology, a number of bands of contrasting lithology. The eastern facies is here given the distinguishing name of *Bonsecours facies*, because of its distinctive development: On the west flank of the Richford Syncline the base is well established by the White Brook Dolomite. But the Peaked Hill Greenstone forms a rather shaky base on the east flank. The top is beneath the Bridgeman Hill or the Sweetsburg Formation, wherever they occur (cf. Pl. 1). The name is taken from Bonsecours, Quebec, Osberg's (personal communication, and manuscript map, 1959) type locality for his Bonsecours Formation, which correlates broadly with the Bonsecours facies of the Underhill Formation.

The phyllites are indistinguishable from those of the Fairfield Pond Member. They contain somewhat more segregation quartz bodies, and towards the east they tend to contain small (about 1mm diameter) albite porphyroblasts, easily identified as white spots on the fresh and weathered exposure surfaces. A characteristic occasional layering (lighter and darker greenish to yellowish layers) parallel to the schistosity is not true bedding, but, rather, bedding drawn out along the schistosity in highly exaggerated coalescing isoclinal folds. More competent beds sometimes transect these layers. The process by which these layers form has been called transposition (Knopf and Ingerson, 1938). Similar transposition layering has been observed in Ordovician rocks near Hardwick (Konig and Dennis, in press), and in Devonian rocks near Greensboro Bend (*ibidem*).

In the eastern limb of the Richford syncline the grain locally coarsens somewhat, and, toward the east sericite is gradually replaced in part by muscovite. More quartz appears to be present here, but this may be due to recrystallization of fine-grained groundmass quartz. Magnetite grains are locally abundant, often in perfect octahedra. In the extreme southeast corner of the Enosburg quadrangle, around Peaked Hill (named Peaked Mountain on the topographic map, possibly in accordance with earlier usage), a rusty weathering graphitic schist appears. On the 1:250,000 geologic map of Vermont this is shown as part of the

Hazens Notch Formation. This writer prefers to retain the Peaked Mountain Greenstone (see below) as a horizon marker, and thus include that part of the Hazens Notch facies which locally appears to the west of that greenstone, with the Underhill Formation; it is thus a "parvafacies" (Caster, 1934), rather than a "magnafacies," as on the 1:250,000 geologic map of Vermont. Where the Peaked Mountain Greenstone is absent, as farther south, the Hazens Notch includes rocks younger than those so named in the Enosburg area. A number of *graphitic schist and slate bands* (Ꞓug on Plate 1) are intercalated in the Bonsecours rocks. They seem to foreshadow an increase in the graphite content toward the east and downward in the section (Hazens Notch facies).

Greenstones

According to Tyrrell (1933), greenstones are "all the basic and ultrabasic igneous rocks characteristically found in fold-mountain ranges—spilites, basalts, gabbros, diabases, and serpentine,—which are generally regarded as having been erupted during the geosynclinal phase of mountain-building. . . . They have in general suffered a low-grade regional metamorphism." There are two main bands of greenstones in the Underhill Formation of the Richford syncline (cf. Pl. 1): a more westerly one, passing through the village of Bakersfield and named the Bakersfield Greenstone, Ꞓub; and a more easterly one, passing immediately east of Peaked Mountain and named the Peaked Mountain Greenstone (Ꞓup). The latter forms the eastern boundary of the area described in this Bulletin. Other outcrops of greenstone shown on Plate 1 may or may not be related to the Bakersfield and Peaked Mountain greenstones.

Characteristically these rocks are chlorite-albite-epidote greenstones. The chlorite gives them their dark green color. Sometimes more or less pronounced layering is produced by segregation of albite and chlorite, and also (much more rarely) by segregation of epidote. The layering often outlines minor folding. Foliation is purely compositional, and mechanically the rock has little tendency to break along the foliation. The exposed surface weathers rusty, and is often characteristically pitted, rather like the Tibbit Hill. Both the Bakersfield and the Peaked Mountain greenstones are in many places associated with a white well-crystallized marble. This association is very characteristic. A few lenses of marble in the southeast corner of the area (cf. Pl. 1.) may well take the place stratigraphically of a greenstone band. The northernmost exposure of the Bakersfield greenstone in the Jay Peak quadrangle is almost exclusively marble, with just enough chlorite schist stringers to confirm equivalence with the greenstone band aligned along the strike to the south.

Stratigraphic Relations of the Greenstones

The Bakersfield Greenstone lies consistently between the White Brook and the Sweetsburg, but never in contact with either. Osberg (personal communication, 1958; and manuscript map, 1959) has traced a similarly positioned greenstone with similar lithologic characteristics in the area northeast of Brome Lake and named it the Lawrenceville Greenstone. Both Osberg and Cady (personal communication, 1958) believe that field evidence is sufficient for a provisional correlation of the Lawrenceville Greenstone and the Belvidere Mountain Amphibolite (Albee, 1957), which separates the Ottauquechee Formation on the east flank of the Green Mountains from the underlying Hazens Notch Formation. Thus, that part of the Underhill between the Bakersfield Greenstone and the Sweetsburg Formation would correlate with the lower part of the Ottauquechee Formation, and the Sweetsburg Formation itself with the main (upper) part of the Ottauquechee. Since, in the western part of the Enosburg area, the Cheshire and Bridgeman Hill formations appear between the Underhill and the Sweetsburg, the lower part of the Ottauquechee may be partly or wholly stratigraphically equivalent to the Cheshire and Bridgeman Hill.

The *Peaked Mountain Greenstone* is lithologically very similar to the Bakersfield Greenstone. Moreover, since the Bakersfield Greenstone seems to be traceable to the eastern flank of the Green Mountains, one would expect it to appear on the east flank of the Richford syncline. Indeed, Osberg found his probably equivalent Lawrenceville Greenstone both on the east and west flanks of the corresponding syncline in the Brome Lake-Richmond area of Quebec. A difficulty arises because in the Sutton area (Rickard, personal communication, and manuscript map, 1959) greenstone exposures are so sporadic that it is difficult, if not impossible, to trace, with any assurance, greenstone bands from Vermont far into Quebec. On the geologic map of this bulletin (Pl. 1), it would be possible to correlate the band of Peaked Mountain Greenstone north of South Richford either with that near Stevens Mills or that near the Richford Country Club. Now, the exposure (abandoned copper prospect) near the Country Club is lithologically more like the Peaked Mountain Greenstone than the exposure near Stevens Mills (see under Metamorphism). But the latter is better aligned structurally.

The greenstone near the Country Club *could* be correlated with the Lawrenceville; that near Stevens Mills is stratigraphically too low, as shown on Rickard's (1959) manuscript map of the adjoining Sutton area in Quebec. The question cannot be resolved on the basis of presently available evidence, and any correlations indicated on published maps, including Plate 1 of this bulletin, are highly hypothetical.

In Quebec, Rickard (manuscript map, 1959) correlates the Country Club greenstone with Osberg's (manuscript map, 1959) Bolton Glen Formation, which is *below* the Lawrenceville Greenstone.

Cheshire Formation

Name and Distribution

As reported by Cady (1945), the Cheshire was first named by B. K. Emerson in 1892, on the unpublished Hawley sheet of the United States Geological Survey. The first published account of the Cheshire Quartzite is in Emerson (1917). The formation was named from Cheshire, Massachusetts, where it occurs as "a granular quartz rock of very massive habit, rather fine and even grain. . . . It is extensively used for making glass. In places it is very feldspathic. . . . The feldspar washed out of it forms small beds of very pure kaolin." (*ibid.*) This description of the Cheshire is also applicable to the formation in southern and central Vermont (Cady, 1945; Keith, 1923). In northwestern Vermont, however, there is a facies change to an impure quartzose subgraywacke, although more quartzitic portions occur in certain areas (notably on Oak Hill, Quebec). These rocks were included in Clark's (1934, 1936) Gilman Quartzite which, as was stated earlier, also included the Fairfield Pond Member of this bulletin.

Booth (1950) followed Clark's nomenclature, since Clark's area adjoined his. Shaw (1954) also followed Clark. Booth (*ibid.*, p. 1148) recognized four facies in his Gilman: "(1) a coarse to very coarse-grained sandstone or quartzite (mainly in the Milton quadrangle), (2) a fine-grained argillaceous siltstone, (3) a mottled argillaceous quartzite, and (4) a whitish relatively pure massive quartzite. M. J. Rickard (personal communication, 1959), distinguishes (1) greenish-gray phyllitic schist (siltstone), (2) dark gray siltstone—fine quartzite schist, (3) white-fawn schistose quartzite (top).

Lithology

In this bulletin the Cheshire Formation includes lithologies (3) and

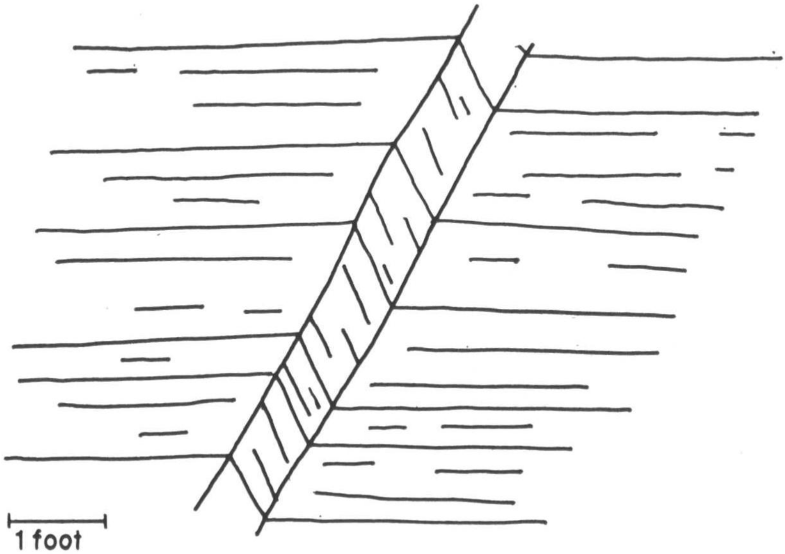


Figure 6. Kink band.

(4) of Booth, or lithology (3) of Rickard. The contact between the Fairfield Pond lithology and the Cheshire is gradational. The line was drawn above the frankly phyllitic dark gray to greenish phyllite, and below the rather characteristic mottled gray silty impure quartz schist of the Cheshire. Breaks that would be joints in the Cheshire are developed as "kink-bands" or "fold layers" (Fig. 6.) in the more phyllitic Fairfield Pond. Osberg (manuscript map, 1959) has recognized Cheshire almost as far north as the St. Francis river. North of the St. Francis river H. C. Cooke and the writer (manuscript map, 1953) mapped what is essentially Fairfield Pond lithology as Gilman.

A good detailed description of the Cheshire as developed in the Enosburg quadrangle is given by Booth (1950, p. 1149), where he refers to his lithologies (3) and (4).

Age

The Cheshire is the lowest fossil-dated formation in Vermont. Walcott (1888) found *Olenellus*, *Nothozoe*, and *Hyolithes*, near Bennington (reported by Cady, 1945). Clark (1936) found *Kutorgina* near Scottsmore, Quebec. Shaw (1954) found *Salterella* near St. Albans and *Hyolithes* near Sheldon Junction. Fossils are scarce and mostly poorly preserved, but the above finds establish the Cheshire as early Cambrian.

Bridgeman Hill Formation

Name and Distribution

The Bridgeman Hill Formation of this report includes, as members, the Dunham Dolomite, the Oak Hill Slate and the Scottsmore Quartzite and conglomerate of Clark (1936). The formation forms a comparatively narrow outcrop band on the east side of the St. Albans synclinorium. It is stratigraphically equivalent to the Dunham Dolomite, Parker Slate, and Rugg Brook Dolomite of the west limb of the St. Albans synclinorium (Shaw, 1958). Following Clark (1934) the succession on the east limb of the synclinorium is often called the "Oak Hill sequence," that on the west limb the "Rosenberg sequence." Clark believed each to be in separate, but stratigraphically partially equivalent thrust slices. The present study has found no valid evidence for thrusting between the two sequences (other than the purely local Hinesburg thrust, which only affects pre-Dunham rocks, and Dunham Dolomite in part). Hence the only structure involved in the two sequences is the St. Albans synclinorium. The facies change from west to east within the synclinorium has been described and discussed by Shaw (1958). In the present report the Bridgeman Hill rocks have been grouped as one formation, because they are not separately mappable in the area covered. North of the international boundary this difficulty seems to persist to some extent (M. J. Rickard, oral communication, 1959). The name of the formation is taken from Bridgeman Hill, near Franklin, where a good section is well exposed. A new member, the Rice Hill Member, is here recognized.

Dunham Dolomite

The Dunham Dolomite (Clark, 1934) is the lowest member of the Bridgeman Hill Formation. It conformably overlies the Cheshire Formation. The name is derived from Dunham, Quebec, and the type locality is at Oak Hill, Quebec (Clark, 1936, p. 147).

Cady (1945) considerably extended the name Dunham to a number of correlatives then known in Vermont; this led him to reclassify adjoining formations. The Dunham of the west limb of the St. Albans synclinorium is much thicker and rather different from the Dunham of the east limb. An excellent description is given by Shaw (1958). The Dunham of the east limb crops out as a narrow (50–100 feet) band, rusty weathering and relatively erosion resistant owing to an appreciable quartz content as stringers and grains. Weathered outcrops are rather similar to those of the White Brook Dolomite. The fresh surface, however, is most often bluish gray to gray. In addition, the Dunham may be well-bedded and cleaved, particularly between Bridgeman Hill and the village of Franklin, and near where the St. Albans-Fairfield Pond road crosses the outcrop.

Rice Hill Member

This is a limy to dolomitic shale, locally grading into a shaly dolomite, which overlies the Dunham Dolomite. It is well exposed on the west slope of Rice Hill, and usually underlies the Hinesburg thrust. Near its top, bands of whitish to pinkish marble have been found east and northeast of Franklin village, and in the reentrant some 0.8 miles east-southeast of Greens Corners. The typical Rice Hill is a medium gray, brownish weathering fissile, reasonably tough dolomitic silt rock. Parts of it are quite sandy, others (especially east of East Highgate) are as dolomitic as the Dunham. It is evidently a transitional lithology between the Dunham and the Rugg Brook, and sporadically incorporates lithologies from each of those members. This makes separate mapping of the members of the Bridgeman Hill Formation very difficult. Even on Clark's (unpublished) outcrop map of the adjacent area in Quebec it is not possible to separate the members satisfactorily into stratigraphic units (M. J. Rickard, oral communication, 1959).

Oak Hill Slate (Parker Slate)

Clark (1936) described a tough rusty-flecked black slate from Oak Hill which intervenes between the Dunham and the Scottsmore (Rugg Brook in Vermont). It is equivalent to the Parker Slate of Keith, in which a rich Lower Cambrian fauna has been found. The last and most comprehensive description is by Shaw (1954).

True black Oak Hill Slate has only been observed north of the village of Franklin, in one small exposure, and in a few exposures northeast of Aldis Hill. However, it is possible that the Rice Hill member is partly or wholly equivalent to the Oak Hill-Parker. Shaw's map (1958) suggests that he may have had such a correlation in mind, but he did not apply it consistently. On the other hand, W. M. Cady (oral communication, 1960) believes that the rocks of the Rice Hill Member represent the lower part of the Rugg Brook Dolomite.

Rugg Brook Dolomite (Scottsmore Quartzite)

This member, which is at the top of the Bridgeman Hill Formation, is a sandy dolomite or dolomitic sandstone. The type locality (Schuchert, 1933) is on the western limb of the St. Albans synclinorium and, hence, is not suitable as a reference locality for the rather different eastern facies. A better reference locality is Oak Hill, where Clark (1936) described the correlative Scottsmore Quartzite.

The Rugg Brook or Scottsmore of the east limb is a rusty weathering, characteristically sandy dolomite. Occasional conglomerate lenses with variegated dolomite, limestone and slate pebbles occur. The horizon is not continuous, and part of it is undoubtedly cut out by the Hinesburg thrust. Where the cement is siliceous, the rock is correctly referred to as a quartzite. Evidence on the west limb of the synclinorium led Shaw to postulate an unconformity between the Parker and the Rugg Brook. The structural complexity and attenuation of the east limb make it impossible to extend this hypothesis to that limb. The Rugg Brook probably is Middle Cambrian, as suggested by its inclusion in the *Cedaria* zone (Shaw, 1954).

A thorough description of the Rugg Brook is by Howell (1939).

Richford Syncline Exposures

The presence of vestigial Bridgeman Hill Formation in the Richford syncline below synclinal outliers of Sweetsburg-like slate is inferred from correlation by similar lithologic association. Two exposures appear to be critical: one, of dense blue-gray dolomite (?Dunham) below black Sweetsburg-like slate on the bank of Bogue Branch brook near School No. 11. The other, of brown-weathering sandy dolomite (?Rugg Brook), also underlying Sweetsburg-type black slate, on a small hill close to the southern tip of the large island in the Missisquoi river west of Richford village.

Sweetsburg Formation

Description and Distribution

No type locality for this formation is named by Clark (1936) who first described it, but sections near Oak Hill are appropriate type sections. The Sweetsburg is a black, highly graphitic well-cleaved slate with very characteristic thin (1–5mm) whitish silty layers. Joint and weathered surfaces are often rusty brown, owing to the presence of pyrite in the slate. Two- to four-inch layers of a bluish quartzite and layers of a black graphitic limestone are occasionally present in Quebec, rarely in Vermont. Slightly dolomitic, more coarse-grained (silty) bands occasionally present have a tendency to weather a rusty brown. These bands are common in the upper part of the Sweetsburg, which was named Vail Slate by Clark (1936) and Hungerford Slate by Schuchert (1937).

On the western limb of the synclinorium the Sweetsburg has been divided into *Skeels Corners Slate* (Howell, 1939 and Shaw, 1958) and *Hungerford Slate* (Schuchert, 1937), separated by the intervening Rockledge Conglomerate (Schuchert, 1937). This subdivision is not practicable on the east limb owing to the absence or lack of outcrops of the Rockledge Conglomerate. An attempt at separation was, however, made by Shaw (1958), who also grouped all the Middle Cambrian formations of the west limb of the St. Albans synclinorium under the name *Woods Corners Group*. The absence of the lowest member of this group, the St. Albans Slate, on the east limb and a marked facies change make it inadvisable to carry the group name over to the east limb.

The rocks of the Sweetsburg Formation appear in synclinal outliers in the Richford-Brome syncline. The correlation of these black slates with the western Sweetsburg has presented some problems, since the Underhill in this area also contains black slates. Two lines of evidence have led to reasonably certain correlation: association with Bridgeman Hill type rocks (see above), and alignment along the tectonic trend with undoubted Sweetsburg in the core of the same syncline to the northeast (Osberg, manuscript map, 1959).

Age

Shaw (1958, P. 548) has shown that the *Cedaria* zone, to which the Sweetsburg must be assigned, is most probably wholly Middle Cambrian.

Morses Line Slate

Distribution and Age

No rocks of undisputed Upper Cambrian or Lower Ordovician age have been found in the area covered by this bulletin. The rocks overlying the Middle Cambrian Sweetsburg Formation belong to the Middle Ordovician Morses Line Slate (Shaw, 1951 and 1958). Its type locality is at the border station of Morses Line. The Morses Line Slate has been tentatively correlated with the Stanbridge and Beauceville formations of Quebec (Osberg and Cady, oral communication, 1959) and the Missisquoi Formation of the east limb of the Green Mountain anticlinorium

(Cady, written communication, 1960). On the east limb of the Green Mountain anticlinorium, the Stowe Formation intervenes between the Ottauquechee (=Sweetsburg in part) and the Missisquoi formations. On the east limb of the Sutton Mountains anticlinorium of Quebec, the "Caldwell" rocks intervene between the Sweetsburg and the Trentonian Beauceville. Thus, the "Caldwell"-Stowe could be either Upper Cambrian or Lower Ordovician or both. In Quebec, at least, the "Caldwell" conformably overlies the Sweetsburg; interbeds of Sweetsburg type slate in the lowest Caldwell graywackes make this a reasonable assumption. In Vermont the Missisquoi Formation locally begins with the Umbrella Hill Conglomerate. It therefore seems reasonable to place the "Caldwell" -Stowe in the Upper Cambrian, at least in part.

Lithology

The Morses Line Slate of the area covered by this bulletin has been thoroughly described by Shaw (1958, p. 553). Briefly, it is a light to medium gray, often banded and partly calcareous slate, with many intercalations of limestone lenses. It superficially resembles the probably equivalent Beauceville Slate of Quebec, but it lacks the latter's volcanics.

General Stratigraphic Relationships

Cady (1960) gives a very thorough account of the stratigraphic relations of the region surrounding the area of this bulletin. Briefly, Cady shows that there is an interfingering of graywacke-shale-greenstone assemblages from the east with carbonate-quartzite assemblages from the west (*ibidem*, p. 538-539). This is well illustrated by the area described in this bulletin. We are, therefore, in the transition zone between the miogeosyncline and the eugeosyncline of Cambrian time.

STRUCTURAL GEOLOGY

Terminology

Since there is insufficient agreement concerning the terminology of structural geology, and in order to avoid misunderstandings, some definitions and explanations of tectonic terms used in this bulletin may be useful.

Geosyncline: The terminology developed by Stille (1940) and Kay (1951) will be used.

Geotectonic cycle: (Stille, 1940, p. 13): Includes all events taking place during the development of a geosyncline, from the initiation of subsidence to final magmatic activity. Includes Stille's magmatological cycle.

Orogeny: Intensive deformation of the fabric of rocks, limited in time and space (cf. Stille, 1924, p. 11). The term fabric is used in its widest sense.

Primary orogeny: characterized by deformation, regional metamorphism and granitization; in eugeosynclinal zones. Unique within any one geotectonic cycle in any one place. "Orthotectonic" (de Sitter, 1956); "anatexitic" (de Cserna, 1961).

Secondary orogeny: (a) in the basement: Sharply outlined vertical movements, without important geochemical changes. (b) in the cover rocks: intensive deformation; often several phases within the same geotectonic cycle, more or less separated from primary orogeny both in time and in space. Secondary orogenic movements are usually triggered by basement movements.

The terms "primary" and "secondary" are here used as in Wilson (1954).

Where the criteria of primary or secondary orogeny as defined above are absent, the writer is reluctant to label "orogeny" any deformation or intrusion, of uncertain extent or intensity, as has been done in the past. It would be preferable to make use of the non-committal term "event," so as to eliminate any danger of unwittingly introducing a hypothesis in one's terminology.

Folding: This term is here used in its kinematic meaning exclusively. Unfortunately, "folding" is often used as a synonym for "orogeny." This connotation of the term is too wide, and in many instances, misleading.

Fabric: See under "Minor Structural Features".

Regional Tectonic Setting

The geographic application of the name "Appalachians" is often restricted to the mountain range extending from northern Alabama to just north of the Pennsylvania-New York State line. The geologist's Appalachians comprise the entire orogenic belt from Alabama to Newfoundland. In New England, five axial zones may be distinguished (Fig. 7.): 1 and 2, the eastern and the western schist zones; 3, the zone of domes and arches, mainly straddling the Connecticut River valley; 4, the plutonic core in the White Mountains; 5, the western zone of decollements. The latter is hardly developed at all in the area here considered: to the extent that it might exist it would comprise the somewhat dis-

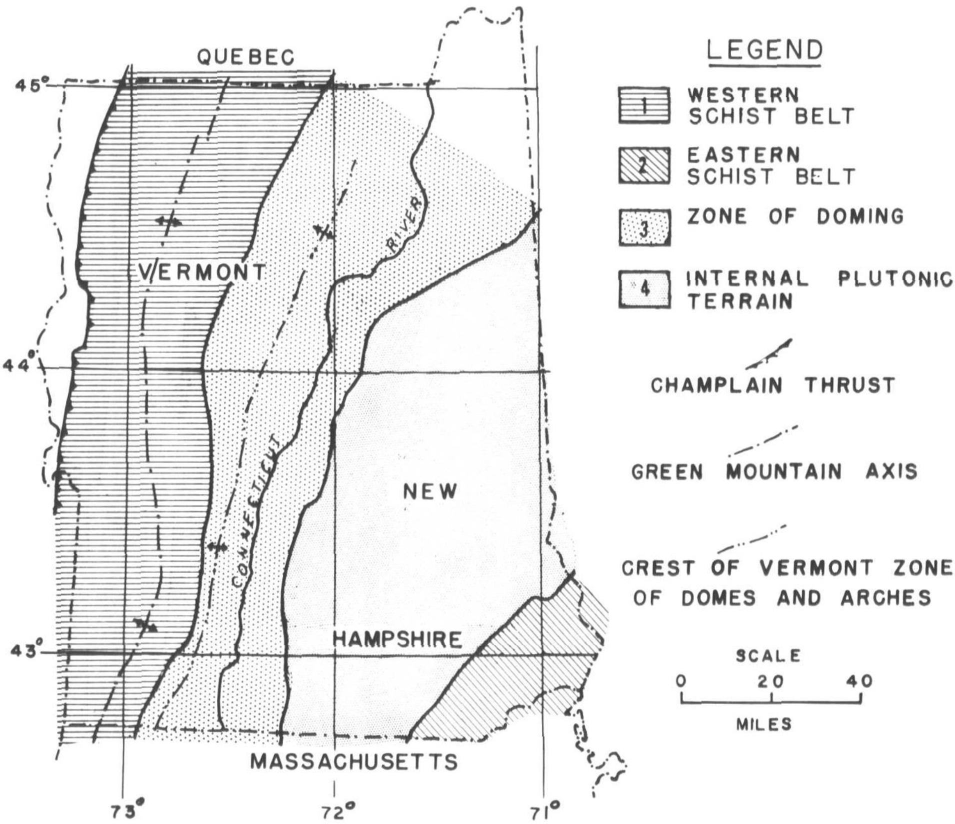


Figure 7. Major tectonic zones in New England.

turbed zone just west of the Champlain thrust. It corresponds to the folded and thrusted Appalachians farther south. Zones 1 - 4 are all within the primary orogenic zone. Secondary orogeny is principally confined to (or recognizable in) zone 5, and the westernmost part of zone 1.

It is noteworthy that the secondary orogenic zone is very differently developed in different parts of the Appalachian orogenic belt, while the primary zone in New England does not differ basically from that of the Piedmont (see E. Cloos, 1953; King, 1955). Radiometric age determinations by Hurley *et al.* (1958), Kulp (1959), Long *et al.* (1959) and others tend to confirm the essential unity of the primary orogenic zone of the Appalachians whose last principal orogenic event (Acadian orogeny) has been determined by these workers to fall within the age range of 320-365 million years, mainly 330-350 million years (Mississipian).

Major Structural Features

INTRODUCTION

The major structures of the area are shown in Fig. 2. From west to east, they comprise the St. Albans synclinorium, the Hinesburg thrust, the Enosburg anticlinorium, the Cambridge syncline and the Richford syncline. The last two synclines appear to be *en échelon*. A rather restricted anticline appears between them where the southern extremity of the Richford syncline and the northern extremity of the Cambridge syncline overlap, near Bakersfield (Pl. 1.).

ST. ALBANS SYNCLINORIUM

(Shaw, 1958). The trough of this feature is just outside the area; the Ordovician Morses Line Slate outlines its core. Its east limb is the west limb of the Enosburg anticlinorium. A break in this limb is developed as the Hinesburg thrust.

HINESBURG THRUST

Along the Fairfield Hill escarpment, the quartzites and graywackes of the Cheshire Formation overlie the Bridgeman Hill Formation. This may be due to overturning of the beds along the east limb of the St. Albans synclinorium in one of its minor folds, or to thrusting. A thrust should be postulated only when there is good field evidence for it. Such evidence is restricted to a few localities. Away from these localities faulting is difficult to confirm. Since the fault is at the contact between Cheshire Quartzite and Dunham Dolomite, and since these units are also found both in normal and reversed stratigraphic succession, it is usually impossible to tell whether, at any particular locality, the contact does indeed represent a fault. For simplicity's sake the contact concerned is drawn through as a thrust on Plate 1 and Figure 2. This is the Hinesburg thrust.

The first mention of the Hinesburg thrust is by Keith (1932, p. 364), in the vicinity of Hinesburg, Vermont. On his Fig. 1, Keith continued this same thrust into the St. Albans quadrangle. Cady (1945) first mapped the Hinesburg thrust in detail. He also recognized a thrust north of the Winooski river, but believed that this was Clark's (1934) Oak Hill thrust which apparently passed under the Hinesburg thrust toward the south (Cady, 1945, p. 566 and Pl.10.). Since then Cady (oral communication, 1960) has mapped the critical area in detail, and is satisfied that the type Hinesburg thrust is the same as that here described which runs along the base of the Fairfield Hill escarpment north of the Winooski river.

Within the area under consideration, evidence for thrusting exists in the reentrant east of Greens Corners and in the klippe area near Franklin village. At the north end of the Greens Corners reentrant the Cheshire overlies the Bridgeman Hill with apparent conformity, but at the south end the Bridgeman Hill beds strike into the Cheshire contact at a large angle. Near Franklin the two klippen are evidence that the Cheshire-Bridgeman Hill contact to their east is a thrust, unless the thrust plane has here passed to within the Cheshire. There is no evidence for this. Between these two localities the Cheshire of the escarpment directly overlies the higher Rice Hill Member rather than the lower Dunham Dolomite member of the Bridgeman Hill Formation; this suggests that the Dunham—present less than ½ mile to the west—has here been cut out by thrusting. The inlier strip of Cheshire west of the escarpment appears to be in an anticline. The only indication that this might be so (other than pure map interpretation) is furnished by the periclinal shape of a Dunham outcrop next to the Central Vermont railroad track about 1/3 mile west of the Fairfield-Swanton town line. Another possible interpretation is repetition by a steeply dipping fault; but there is no evidence at all for this.

In the Milton quadrangle, the Hinesburg thrust of this report was recognized by Booth (1950) and by Stone (1951). In the St. Albans area it was recognized by Shaw (1958) after Booth. In these areas an alignment of tectonic outliers in the foreland is a good indication of thrusting along the Fairfield Hill escarpment to the east. In the Milton area, cross-cutting relationships around Colchester Pond are satisfactory direct evidence of thrusting at the (topographically) lower contact of the Cheshire. Remnants of Dunham Dolomite along this contact in the Colchester Pond area, as well as in the Bridgeman Hill klippe, suggest that the thrust represents the drawnout forelimb of an overturned fold.

Booth (1950) and, after him, Stone (1951) were not certain about the correlation of this thrust with the Hinesburg thrust and they used local names for thrusts inferred to exist north of the Winooski river. Booth's and Stone's Brigham Hill thrust, (Milton quadrangle), shown on their geologic maps, can be traced north of Essex Junction until it crosses Indian Brook at the latitude of Colchester. The present writer specifically checked its postulated continuation northeast of that point, but could find no evidence for it. Rather, it here seemed to merge into the Dunham-Cheshire contact, and to continue from there into the St. Albans quadrangle where shown on Plate 1. Most of what Booth (1950) and Shaw (1958) called the Fairfield Pond thrust is, therefore, the northward continuation of the Hinesburg thrust, with its location revised in several places.

The projection at depth of the thrust plane on the cross-sections of Plate 1 is conjectural. The relationships as mapped make it entirely possible that some or all of the Hinesburg thrust sheet, and the klippen of Cheshire on the St. Albans syncline, became detached and slid from the crest of the rising Enosburg anticlinorium, particularly the Georgia Mountain anticline.

To the north, in southern Quebec, Rickard (written communication, 1960) has found evidence of thrusting at the same contact. Moreover, Rickard suggests that this thrust passes upward in the stratigraphic succession toward the north, and most likely becomes Clark's (1934) Oak Hill thrust near Dunham, Quebec.

Enosburg Anticlinorium[1]

East of the Hinesburg thrust successively older rocks of the Oak Hill succession crop out in a rather irregular fashion, as brought out on Plate 1. Even Plate 1 is simplified, and field relations in detail are far more complicated. Because of facies changes on the one hand, and lithologic similarities between succeeding formations on the other, the drawing of contacts has been exceedingly difficult. Only the White Brook Dolomite provides a reliable marker horizon for the delineation of structures, and even this horizon is discontinuous. Moreover, where it does outline the structure, it shows that the folding is very complicated in detail and that a great number of minor folds on all scales exist within the Enosburg anticlinorium. These minor folds are overturned toward the west on the west flank of the anticlinorium, but rapidly pass into a series of upright folds with near-vertical axial planes toward the core of the anticlinorium.

[1] The Enosburg anticlinorium had previously been named Fletcher anticline (Christman, 1959). The name change is by agreement with Christman. Cady (1960) called it the Enosburg Falls anticline. Since extensive subsidiary folding considerably complicates the anticlinal structure (see Pl. 1.), the present writer prefers to call it an anticlinorium, separated from the Green Mountain anticlinorium by the Richford syncline. Cady (1960) included the Enosburg Falls anticline in the Green Mountain anticlinorium.

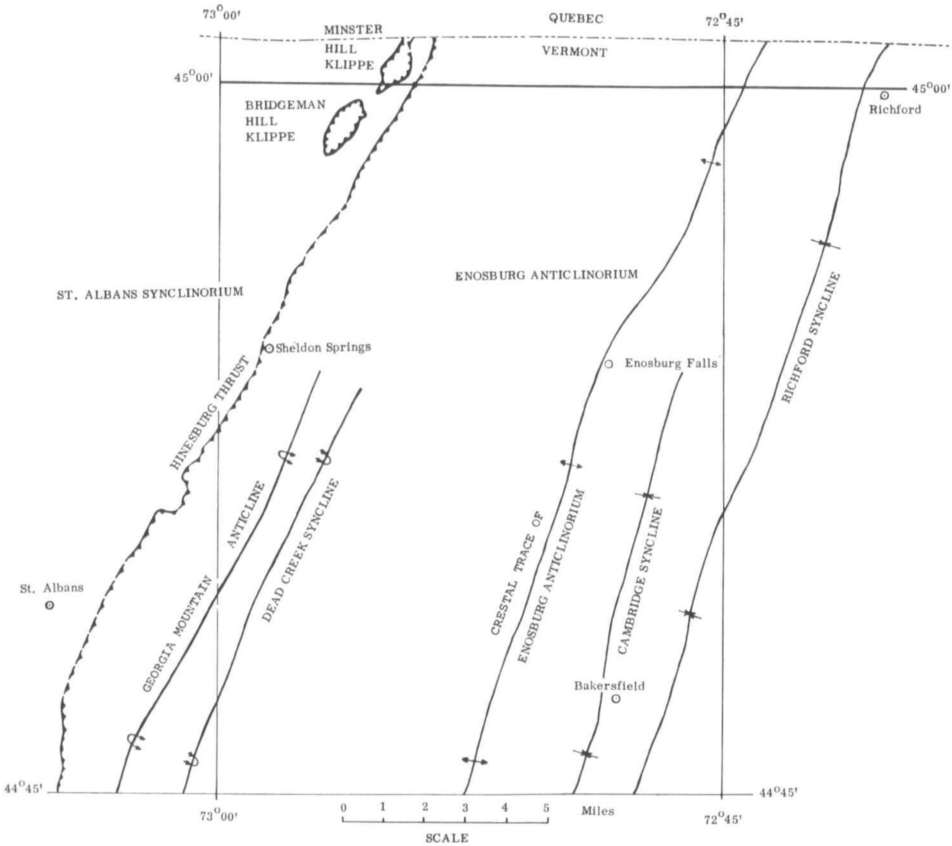


Figure 8. Major structural features of the Enosburgh Falls area.

The competent coarse graywackes are generally in open, undulating folds while the less competent phyllites are in tight isoclinal folds. Only two subsidiary folds could be outlined with any degree of certitude: the Georgia Mountain anticline and the Dead Creek syncline (see Fig. 8, and Pl. 1).

The east flank of the Enosburg anticlinorium cannot immediately and obviously be interpreted as such. All rocks above the Tibbit Hill exhibit appreciable facies change from west to east. The presence of Sweetsburg in two comparatively small outliers to the east (see Pl. 1) suggests a sequential relationship eastward from the main outcrops of Tibbit Hill. Originally Clark (1934) had postulated a thrust, the Brome thrust, along the east limb of the Enosburg anticlinorium. There is no evidence for such a thrust in the Enosburg area, or, in fact, anywhere to the south (Cady, oral communication, 1957). In Quebec, too, recent work by Rickard has shown that the structural relationships there can now be explained without assuming the presence of a thrust in the position assigned to the Brome thrust.

Cambridge Syncline

Originally all the black slates associated with carbonate rocks east of the Enosburg anticlinorium in the Mount Mansfield quadrangle had been assumed to be Sweetsburg equivalents (Cady, oral communication, 1957; Christman, 1958; Dennis, 1961). These workers named the syncline, of which these rocks were postulated to occupy the core, the Cambridge syncline. The Sweetsburg outcrops of the Enosburg Falls and Jay Peak quadrangles were then correlated with the Cambridge syncline rocks (which extend into the Enosburg area up to the village of Bakersfield) and the whole alignment of "Sweetsburg" outcrops was then named the Cambridge syncline (Cady, 1960). Later work by Cady, as explained in the chapter on stratigraphy, has placed the outcrops of black slate south of Bakersfield lower in the succession, most probably equivalent to the White Brook Dolomite. Thus, two synclines must be distinguished (on purely stratigraphic evidence): one contains the black slates and carbonate rocks south of Bakersfield and properly retains the name of its type locality, the Cambridge syncline; the second contains the outcrops of true Sweetsburg slate aligned along the core; it trends through Richford and Brome, and is here called the Richford syncline. In Quebec, Rickard (written communication, 1960) refers to the Brome syncline.

Richford Syncline

The Richford syncline is mainly occupied by rocks of the Underhill Formation (Bonsecours facies). At its core there are a number of outliers of Sweetsburg, and these constitute the best evidence of synclinal relationship. The west limb of the syncline is quite steep, and retains the style of the Enosburg anticlinorium. On its east limb dips gradually flatten toward the Green Mountain anticlinorium; a subsidiary anticline of this structure, the Cold Hollow anticline, is immediately east of the area here described. It has not been possible, so far, to correlate the rocks below the Underhill on each side of the Richford syncline. Cady (written communication, 1960) has suggested that the coarse chlorite-albite gneisses of the Green Mountains may be equivalent to the coarse graywackes of the Pinnacle.

Minor Structural Features

TERMINOLOGY

Cleavage—The term *cleavage* as applied to rocks refers to all types of secondary planar fabric elements (other than coarse schistosity) which impart mechanical anisotropy to the rock without apparent loss of cohesion. *Continuous cleavage* is the result of continuous parallelism of platy minerals throughout the rock, while in *spaced cleavage*[1] planes of discontinuity are spaced at finite intervals, however small. Cleavage planes are thus potential planes of parting. Any actual fracturing along cleavage planes is subsequent and incidental.

Foliation—Following Fairbairn (1949, p. 5) any planar anisotropy in a rock due to the presence of parallel planar fabric elements is a *foliation*. Foliation may be the result of sedimentary processes, of recrystallization (continuous cleavage, schistosity) or of mechanical deformation (all kinds of cleavage).

Schistosity—is foliation by dimensional parallelism of platy minerals, or by orientation of rod-shaped minerals along planes. In the case of fine-grained platy minerals, schistosity is synonymous with continuous cleavage.

Coordinates a, b, c are used to describe the symmetry of rock fabric at a point or within a fabric domain. The most prominent foliation is selected as the plane of a and b, or the *ab plane*. The most prominent symmetry plane perpendicular to *ab* is selected as the *ac* plane. Thus, a, b, and c are each fully determined. In addition, any set of planes which introduces a mechanical anisotropy into the fabric are known as *s* planes. Usually, stratification is labelled *ss*, and subsequently formed *s*-planes s_1, s_2, s_3, etc., in order of formation. Theoretically, this definition includes fracture planes. It is becoming customary, however, to restrict the designation "*s*" to closely spaced planar discontinuities such as foliations, cleavages and schistosities.

The *kinematic* axial cross is used to describe symmetry of *movement*. The plane of movement or gliding is the kinematic *ab* plane, the gliding direction is a, and maximum deformation is in the *ac* plane, which is the

[1] Chidester (1962, p.17) proposes classification of cleavages into "spaced" and "continuous." This non-genetic classification has great merit. Chidester uses "schistosity" as a synonym for cleavage as here defined, but agrees (oral communication, 1963) that "cleavage" may be preferable.

only plane of deformation in biaxial deformation. The normal to *ac* will then be *b*. In biaxial deformation there is no movement along *b*. *b* may also be an axis of rotation (*B*).

Inventory of Minor Structural Features

Three sets of *s*-surfaces have been observed: bedding planes are referred to as *ss*, secondary foliations as s_1 and s_3. s_2 does not extend into the area under consideration; it is a local cleavage associated with doming farther east and has been described by the writer in the Lyndonville area (Dennis, 1956 and 1961). In addition, there are linear features related to each set of *s*-surfaces. All these features will be described, grouped within the major structures of the area from west to east.

St. Albans Synclinorium: In this area s_1 is an east-dipping continuous cleavage. It is "axial plane" cleavage to west-facing folds, and it guides a number of minor thrusts (see Fig. 9).

Since s_1 in the kippen and thrust sheets of Cheshire Quartzite appears to be parallel to the Hinesburg thrust plane, it may also have had a role in guiding some of the major thrusts. The thrust shown in Fig. 9 is guided by s_1 planes.

In pelites s_1 is a true schistosity; in dolomites and coarser clastic rocks which contain a high percentage of granular crystals, the latter are not always dimensionally oriented; hence s_1 may be expressed only by the micaceous minerals present, forming discrete cleavage planes of rather irregular development (Fig. 10): Born's (1929) *Rauhschieferung*, here named *spaced schistosity*.

Linear structures related to s_1 are cleavage traces on bedding, and fold axes; both are gently plunging or horizontal lineations parallel to *b*. The later secondary foliation, s_3, is a spaced cleavage and almost uniformly steeply dipping, generally at 85° toward the east. It is only weakly and sporadically developed in the St. Albans synclinorium. No folds on s_3 have been observed in the field, west of the Richford syncline. However, evidence in the Middlebury synclinorium to the south (G. Crosby, oral communication, 1961), indicates that the synclinoria along the Middlebury-St. Albans trend are related to the s_3 deformation, and that the folds on s_1 belong to an independent earlier deformation phase: For in the Middlebury area s_1 folds trend at a large angle (50°) to the trend of the synclinorium, while s_3 folds are parallel to it.

Enosburg anticlinorium: Here s_1 is usually near-vertical. In the pelitic chlorite phyllites it is a continuous schistosity, in the coarse graywackes a spaced schistosity; see Fig. 10. In the phyllites a char-

Figure 9a. Local thrusting controlled by cleavage planes. In Dunham marble north of Franklin village.

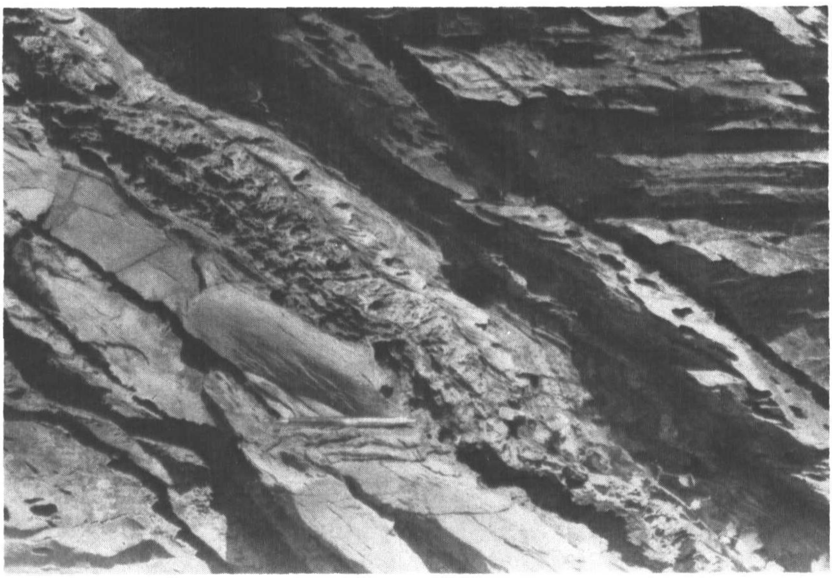


Figure 9b. Sub-horizontal axial plane cleavage in Sweetsburg Formation south of Greens Corners.

Figure 9. Development of s_1 near Hinesburg thrust.

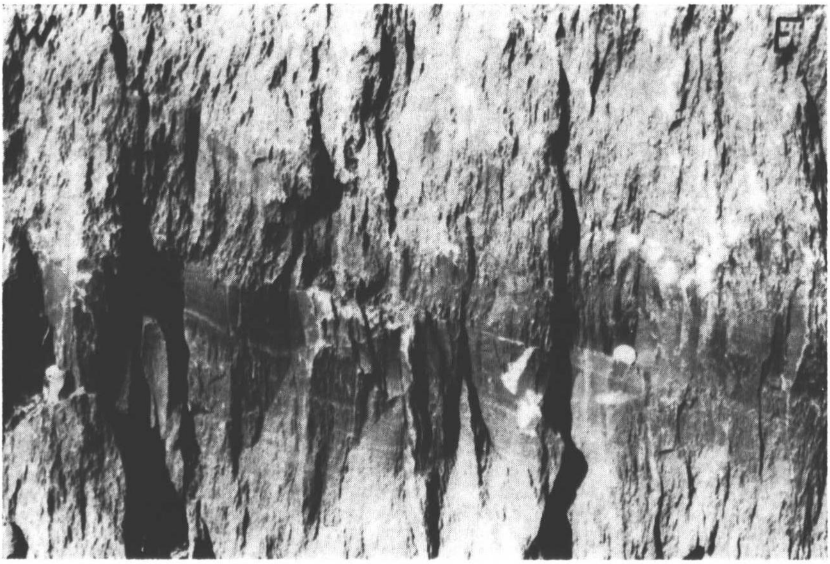


Figure 10. Spaced schistosity: s_1 in Pinnacle graywacke. Near West Berkshire.

acteristic banding parallel to s_1 must not be confused with bedding. It is caused by the drawing-out of the bedding inhomogeneities along the schistosity surfaces. This is quite clear wherever true bedding cuts across the schistosity. Quartz lenses are often aligned parallel to s_1.

The linear structures comprise gently plunging or horizontal fold axes and s-surface intersections. In contrast to the overturned folds of the St. Albans synclinorium, the folds of the Enosburg anticlinorium are gently undulating wherever true bedding planes can be determined. The steeply dipping transposition layering in the phyllites can be misleading in that respect. Naturally, there are also steeply dipping fold limbs in true bedding. In addition, steeply plunging intrafoliate folds are formed on s_1 surfaces as axial planes.

Cambridge and Richford synclines: Here s_1 changes from an essentially "axial plane" cleavage on the west limb to a bedding schistosity on the east limb (see Fig. 11). The probable identity of these two sets of surfaces was first recognized by Rickard (oral communication, 1958), in the adjoining Sutton map area. The bedding schistosity is axial plane cleavage to related minor folds. These usually face down-dip. In the Richford syncline many more steeply plunging minor folds appear. These are also related to s_1. Angles of plunge range from horizontal to vertical, and

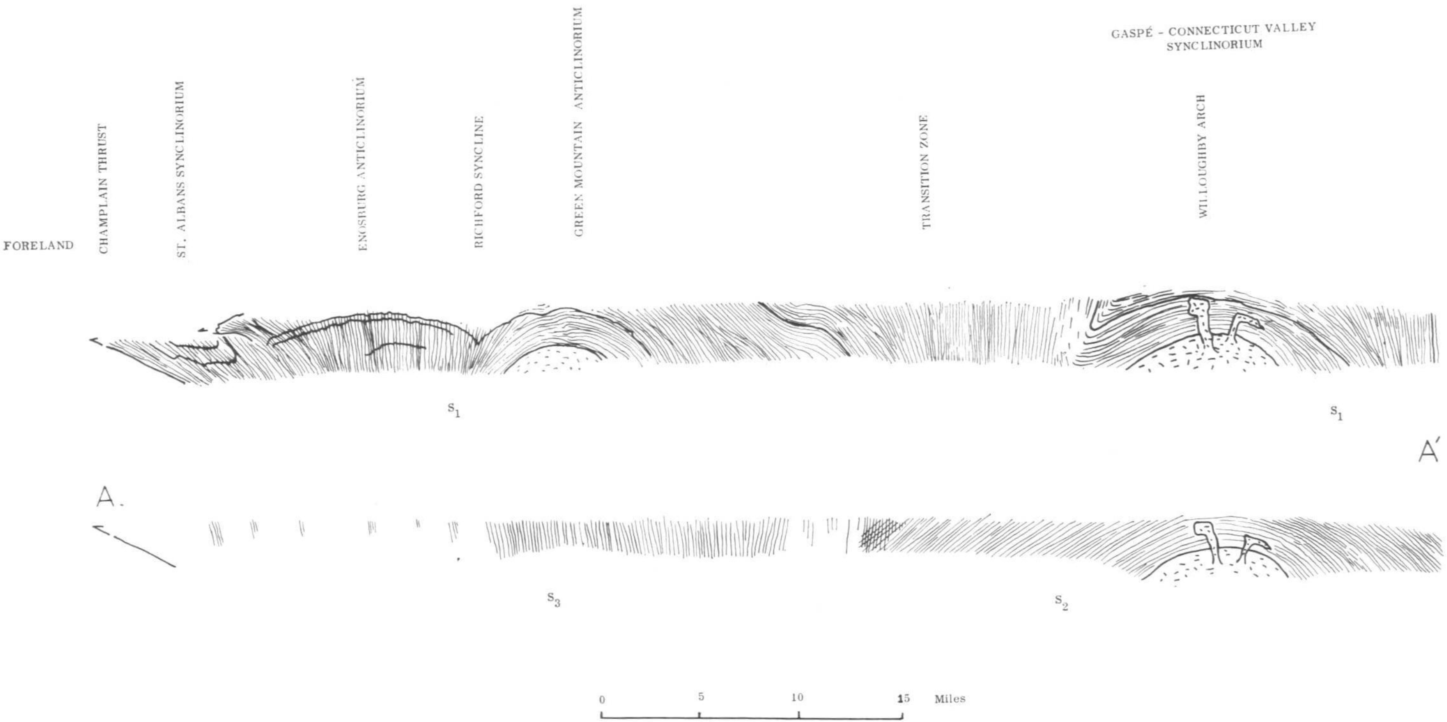


Figure 11. Cross-sections showing cleavage patterns, between Lake Champlain and the Connecticut River (Line AA′ in Fig. 2). Upper section: s_1 (fine lines), and schematic attitude of ss (heavy lines). Lower section: s_2 (in the east) and s_3 (in the west), intersecting near the center. Scale as in Fig. 2.

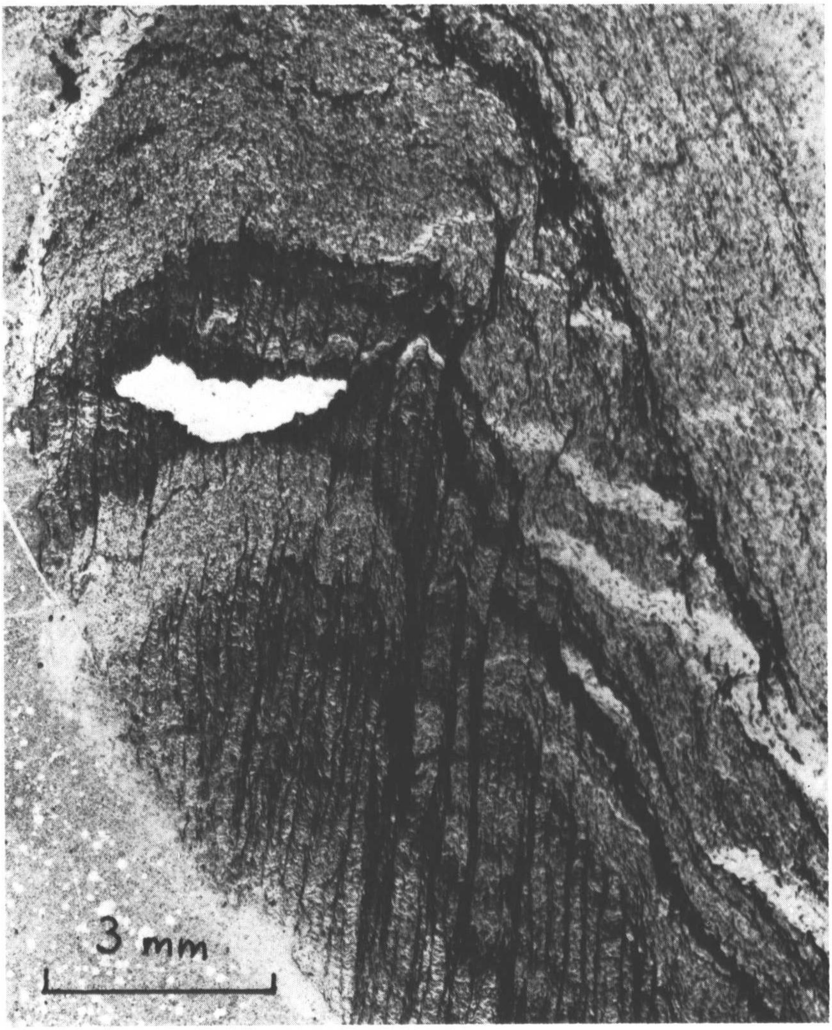


Figure 12. Photomicrograph of Underhill phyllite showing s_3 traces.

greatly divergent plunges may be observed within the same exposure. All are oriented within s_1, however. On the east flank of the Richford syncline s_3 becomes prominent. Its attitude throughout is very steep, normally with an easterly dip near 85°. This is Dale's (1896) slip cleavage. It is a spaced axial plane cleavage to upright chevron-type minor folds (see Fig. 12).

The Richford syncline is the transition zone between the tectonic style of the Enosburg anticlinorium and that of the Green Mountain anticlinorium. It occupies a comparatively narrow zone between these two major features.

Green Mountain Anticlinorium: Only a part of the west flank of this feature is within the area of this bulletin.

In this zone, s_1 is a bedding schistosity, except where it transects the hinge regions of related minor folds. Such minor folds usually plunge down-dip. Similarly, quartz rods are often oriented within s_1 surfaces, plunging down-dip. Chidester (1953) and Albee (1957) have interpreted these quartz rods as vestigial hinges of minor folds in quartz stringers and veinlets. Such quartz layers would probably have been along bedding planes, but some of them might have been cross-cutting veinlets; for relative compression perpendicular to *ab* folds all other surfaces along *B*-axes parallel to their respective intersections with *ab*. Osberg (1952) and Brace (1953), working in central Vermont, have described conglomerate pebbles elongated along the dip of s_1 surfaces; that is, perpendicular to *b*.

The later cleavage, s_3, is a spaced cleavage characteristic for the Green Mountains, and is often informally referred to as "Green Mountain cleavage." As in the Richford Syncline, it is typically developed as slip cleavage (Fig. 12). The related folds are open chevron-type folds with horizontal or gently dipping axes which are aligned parallel to the main Green Mountain arch.

Movement Pattern

Theoretical Premises

All secondary foliation planes in the area are regarded as *ab* surfaces (cf. Goguel, 1945). This hypothesis is confirmed by the consistent observation that all secondary foliations in the area are parallel to the axial planes of related minor folds. The direction perpendicular to fold axes, and *a* lineations where observable are taken to indicate the direction of flow along *ab*. These simple premises lead to a movement picture for each of the secondary foliations.

Interpretation

s_1: West of the Richford syncline s_1 is an axial plane cleavage and appears to express upward escape by flowage along *ab* planes. Flattening of the dip toward the Champlain thrust may represent "overflow" against the foreland, parallel to the marginal thrusting.

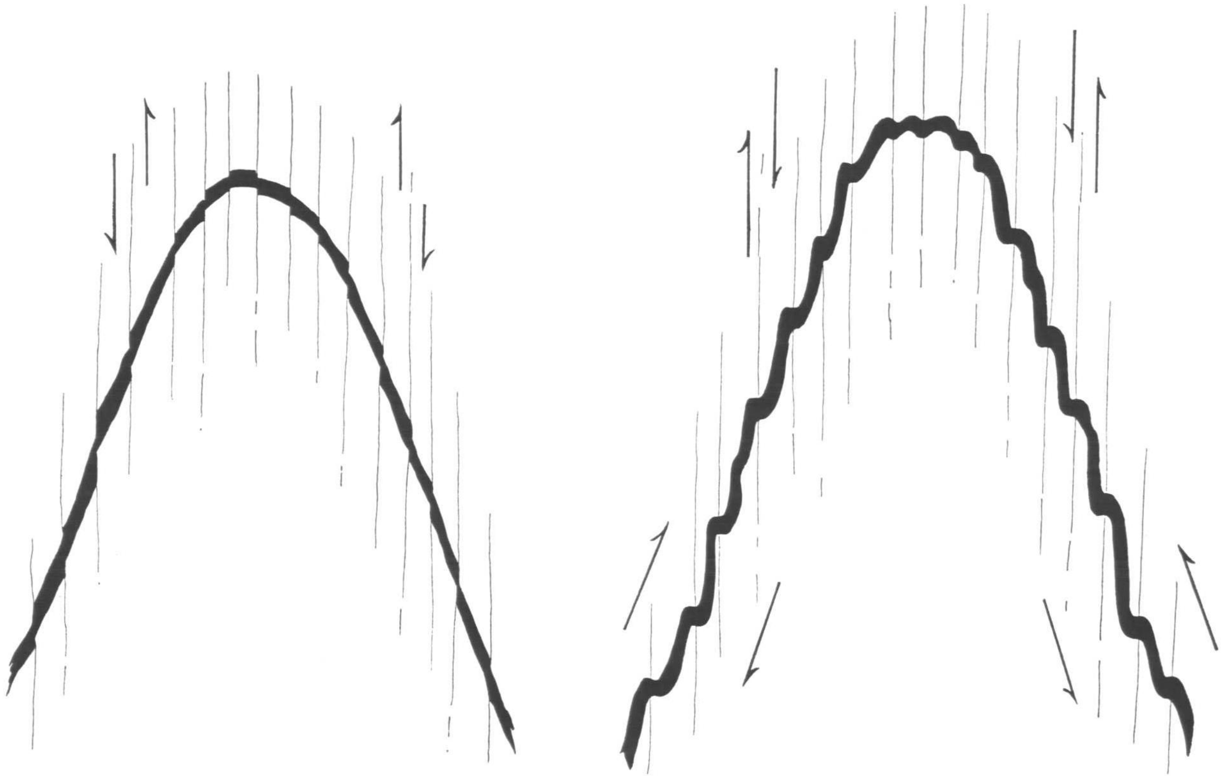
Hoeppener (1956, p. 254) has observed that it may be possible that the s_1 type of foliation referred to s_1 here only forms in rocks whose volume decreases during deformation. Hence the west-facing folds bordering the Champlain thrust may represent pre-cleavage folding due to gravity flowage away from the rising arches of the east (Enosburg or Green Mountain anticlinoria). s_1 would have formed, during subsequent crystallization of the sericite (i.e. contraction) at the time of the first phase of metamorphism. The question arises here why such post-folding sericite would crystallize parallel to the axial surfaces of the major folds. Conceivably, this might happen because the folding itself had formed a fabric with controlling planes for later continuous cleavage.

The pattern of s_1 in the Green Mountains arch is characteristic of a deeper tectonic style than prevails to the west. The bedding schistosity which envelops the arch may have been produced by stretching in *ab*. Updoming of that arch might account for such stretching. The observed linear structures in *a* (see above) tend to confirm this picture. The literature abounds with descriptions of structures in which minor fold axes are parallel to stretching as evidenced by elongated conglomerate pebbles. The mechanism has been described by E. Cloos (1946).

s_2 does not extend into the area under consideration. It is a regionally restricted cleavage associated with a zone of doming (Willoughby arch, Dennis, 1956) east of the Green Mountains. In an earlier communication (Dennis, 1958), the writer had provisionally correlated the two slip cleavages here referred to as s_2 and s_3. Later detailed work (Dennis, 1961; Konig and Dennis, in press) suggests that the slip cleavage associated with the Willoughby arch pre-dates that of the Green Mountains.

s_3 is always steeply dipping, rarely more than 5° from the vertical, and associated with shear folding. This implies a considerable vertical component of movement. There is no convincing evidence for lateral shortening, but that does not necessarily preclude it.

In the classical example of South Mountain, Maryland, E. Cloos (1947) has demonstrated that irrotational flow will produce lateral shortening. Nevertheless he recognized that a certain amount of rotational deformation (shear folding) had taken place there. Now, it would seem that s_3 in the Green Mountains is dominantly rotational. In the deformation plan of Fig. 13, originally circular designs in the plane of the cross-sections would have been elongated in simple shear, i.e. the long axis of the resulting ellipse would be at an acute angle to the cleavage traces. This is implicit in the deformation plan of the bedding, and is confirmed by occasional local faulting along s_3. Such a movement plan



a. b.

Figure 13. Diagrammatic cross-sections showing shear sense in cleavage. *a:* Normal shear sense; true shear folding. *b:* "Reverse" shear sense; hypothetical influence of drag. Probably due to flattening.

may be explained by active vertically directed differential movement or by passive escape resulting from lateral compression. Carey (1954; and oral communication, 1959) has shown that gneissic folds in the basement can be generated by laminar flow without lateral shortening. Folding in phyllites might take place in a similar manner, but in that case flow would be along cleavage planes; this was already assumed by Washburne (1940). Such shear folds are formed by heterogeneous differential movement along the cleavage planes. It is well to remember that these folds are only a consequence of heterogeneous movement. In the case of homogeneous movement along cleavage planes, no folds would form.

Fig. 14, a and b are stereograms of poles of s_1 and s_3. In the northwestern part of the Jay Peak quadrangle the point maximum of s_3 confirms field observations. The distribution pattern of s_1 and ss cannot be evaluated without more detailed work.

Throughout the area under discussion, therefore, both s_1 and s_3 deformation could be explained by either vertical primary movement or by lateral shortening, or both. However, no direct evidence favoring lateral shortening of the upper crust was observed. This does not, of course, exclude the possibility that some such shortening might have taken place.

General Conclusions

Theoretical Principles

The movement pattern developed in the last pages must now be interpreted dynamically. In the course of the last quarter of a century a number of geologists have begun to recognize that continuous rock deformation must be treated as flow (cf. *int.al.* Wegmann, 1931; Hubbert, 1937; Washburne, 1940; Carey, 1954; Bucher, 1956). Reiner has summarized the rheological basis of rock deformation: "As a first approximation it may be assumed that rock deformation is much like that of a Maxwell body. This is a material which reacts instantaneously like an elastic solid, when acted upon by a force. Therefore, . . . elastic stresses are first transmitted throughout the body at acoustical speed and elastic strains are produced. On top of these strains, an increasing deformation resulting from viscous flow is produced as long as the force is acting. In this process the directed elastic stress is continuously transformed into isotropic stress through relaxation. The time of relaxation is that time during which the elastic stress is reduced to its e^{th} part ($e = 2.71\ldots\ldots$). Jeffreys (The Earth, Cambridge 1929) calculated that

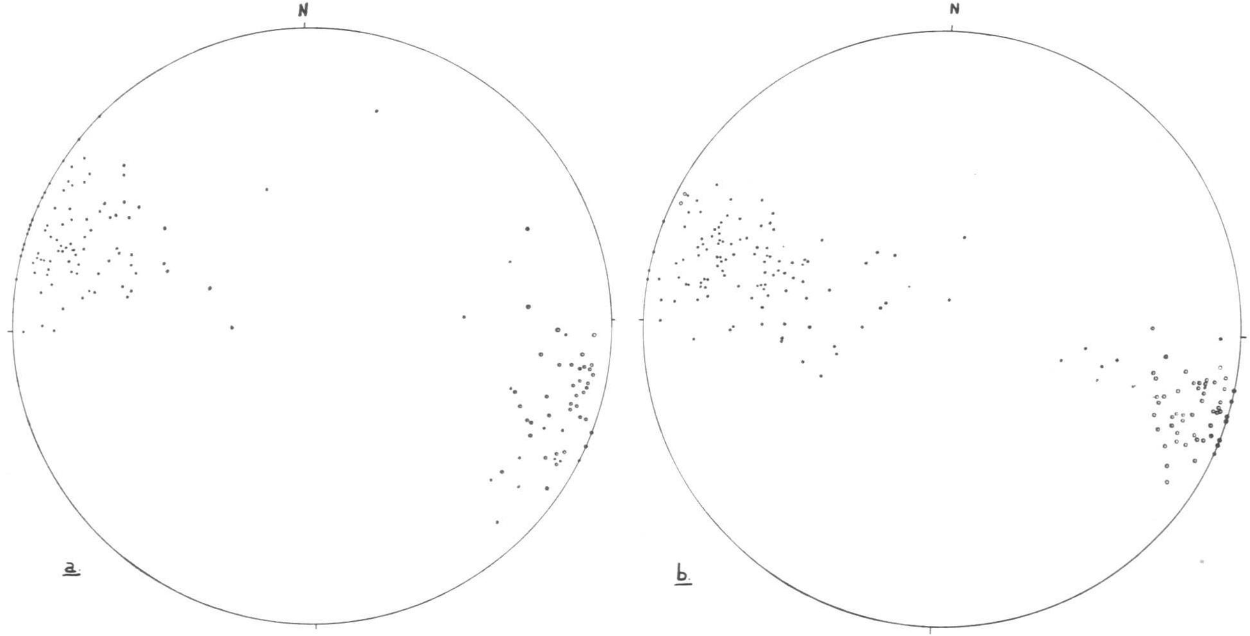


Figure 14. Stereograms (point diagrams) of cleavage poles in the northwestern part of the Jay Peak quadrangle. Points: *ss* and s_1. Circles: s_3. *a:* West of Missisquoi River and North Branch; *b:* East of Missisquoi River and North Branch.

the time of relaxation of the rock shell of the earth is of the order of 10^8 seconds = about three years." (Reiner, written communication, 1960).

Externally induced stresses, therefore, cannot add up significantly within a continuously deforming rock body in the course of time: tectonic movements are too slow, elastic stresses relax too fast by comparison. Only an insignificant part of the deformation can have been induced by elastic stress. For almost all of the deformation process the rocks react by flow to gradients in the existing internal stress field. The movement pattern of the known orogenic zones cannot, therefore, have arisen through continuing elastic deformation. Likewise, current observations do not support the hypothesis (Washburne, 1940) that elastic stresses may have been externally applied in pulses. *Differential flow can only be produced by continuously acting body forces.* This might be accomplished in three ways: (1) direct action of gravity; (2) indirect action of gravity; (3) body forces originating within the rock body.

Examples for (1) are furnished by the many instances of gravity tectonics, for (2) by the folding mechanism postulated by Beloussov (1959), and for (3) by crystal transformations, such as take place in the course of feldspathization and which may cause diapirism (cf. Wegmann, 1930). (2) and (3) may cause escape movements which, if directed between narrow "shores," could hardly be distinguished geometrically from lateral shortening. Steep to vertical flow in basement rocks has been described by many workers, e.g. Wegmann (1930), Eskola (1949), Haller (1956). Carey (1954) has shown that in such cases lateral shortening is not necessary, and that basement deformation patterns are more readily explained by laminar flow than by any other mechanism. Diapirism, too, does not require a "vise," but merely "shores," i.e. a less mobile frame.

Application to Observational Data

The movement pattern of s_1, as deduced above, may be explained as follows: In the Green Mountains, relative rise of a zone in the basement with consequent stretching of the mantling schists. Near the western border of the orogenic zone, escape by flow upward and over the Champlain foreland.

A dynamic interpretation of s_3 is more difficult. Deformation exclusively by horizontal compression seems unlikely, since rotational gliding along sub-vertical planes needs an independent vertical movement component. Also, s_3 is developed over a comparatively long cross-section, from west of the Hinesburg thrust to at least Hardwick, Vermont. The uniformly steep attitude of this set of surfaces and its somewhat patchy occurrence would seem to indicate that at least an important part of the deformation may have been by vertical application of the resultant of deforming body forces; in this case probably due to volume change during recrystallization.

Thus, in the Enosburg area, observational material interpreted according to simple mechanical principles suggests that primary vertical differential movement may have an important role in orogenic deformation.

METAMORPHISM

The whole of the area is in the chlorite zone of metamorphism. A number of features indicate increasing intensity of metamorphism from west to east, although chlorite is stable throughout. To the west, quartz grains are distinctly clastic. Undular extinction is not uncommon. Quartz grain boundaries are usually somewhat recrystallized and sometimes sutured. The groundmass of the more pelitic rocks is very fine grained. Going east, in the central portion of the area, quartz grains increase in size.

In the Richford syncline and west of it, quartz grains have fully recrystallized. Muscovite tends to replace sericite, and even chlorite. Albite porphyroblasts appear selectively in some of the more fine-grained lithologies. They are untwinned. Trails of opaque minerals outline microscopic folding within the albites.

The greenstones of the Tibbit Hill Volcanics and the Underhill Formation consist mainly of chlorite, albite, and epidote. Accessory calcite can become fairly abundant, especially where the greenstones are associated with marble lenses. Toward the south, amphibole and plagioclase porphyroblasts appear in some of the Tibbit Hill bands. Toward the east, especially in the Underhill greenstones, biotite occasionally replaces chlorite. But little biotite was observed in the metasedimentary rocks.

Magnetite is an accessory mineral throughout, in both metasediments and volcanics. It is most common in the finer grained phyllites of the eastern part of the area, where it often occurs as well-developed octahedra.

In the extreme northeastern part of the area, the grain size increases to about ¼ inch, and the rocks have the appearance of gneisses. Black equigranular porphyroblasts in hand specimens turn out to be, in thin section, albite with abundant carbon inclusions. This rock is common along the Green Mountain arch, and is more fully described by Cady *et al.* (1963, p. 11–12).

Metamorphism has also resulted in quartz segregations in the phyllitic rocks. These segregations increase in abundance toward the east. They are mostly in irregular veins and pods, whose orientation tends to be mainly controlled by s_1.

ECONOMIC GEOLOGY

Iron: Hematite concentrations, leached from the West Sutton slate, have been exploited in small pits near Sheldon and St. Rocks in the past (see locations on Plate 1). These occurrences are of no economic value to-day.

Copper: A small chalcopyrite occurrence associated with the greenstone near the Richford golf course attracted some prospecting activity in the past. A small adit remains. Little is known about this working; in particular, there is no information as to whether the occurrence has ever been of economic importance.

Gravel: A number of gravel pits are operating in the glacio-fluvial deposits. Most are shown on the U. S. Geological Survey topographic quadrangle sheets which serve as base for Plate 1. These gravel pits serve local needs only at the present time, but the fluvio-glacial deposits of the area should be surveyed for reserves that might permit more extensive operations.

Slate: The West Sutton Slate is of good roofing quality. Small size of occurrences, and limited demand make future exploitation unlikely.

Water: At the beginning of the century, Sheldon Springs was well-known as a spa. Water supply for domestic use presents no problem, and the Missisquoi river water is suitable for limited industrial use.

Lime: Two marble lenses in the Underhill Formation, in the eastern part of the town of Bakersfield (see Plate 1) have been used as local sources of lime.

Talc: A small lens of talc in the Hazens Notch Formation is exposed in a brook on the west flank of the Cold Hollow Mountains (see Plate 1). It is of no economic interest.

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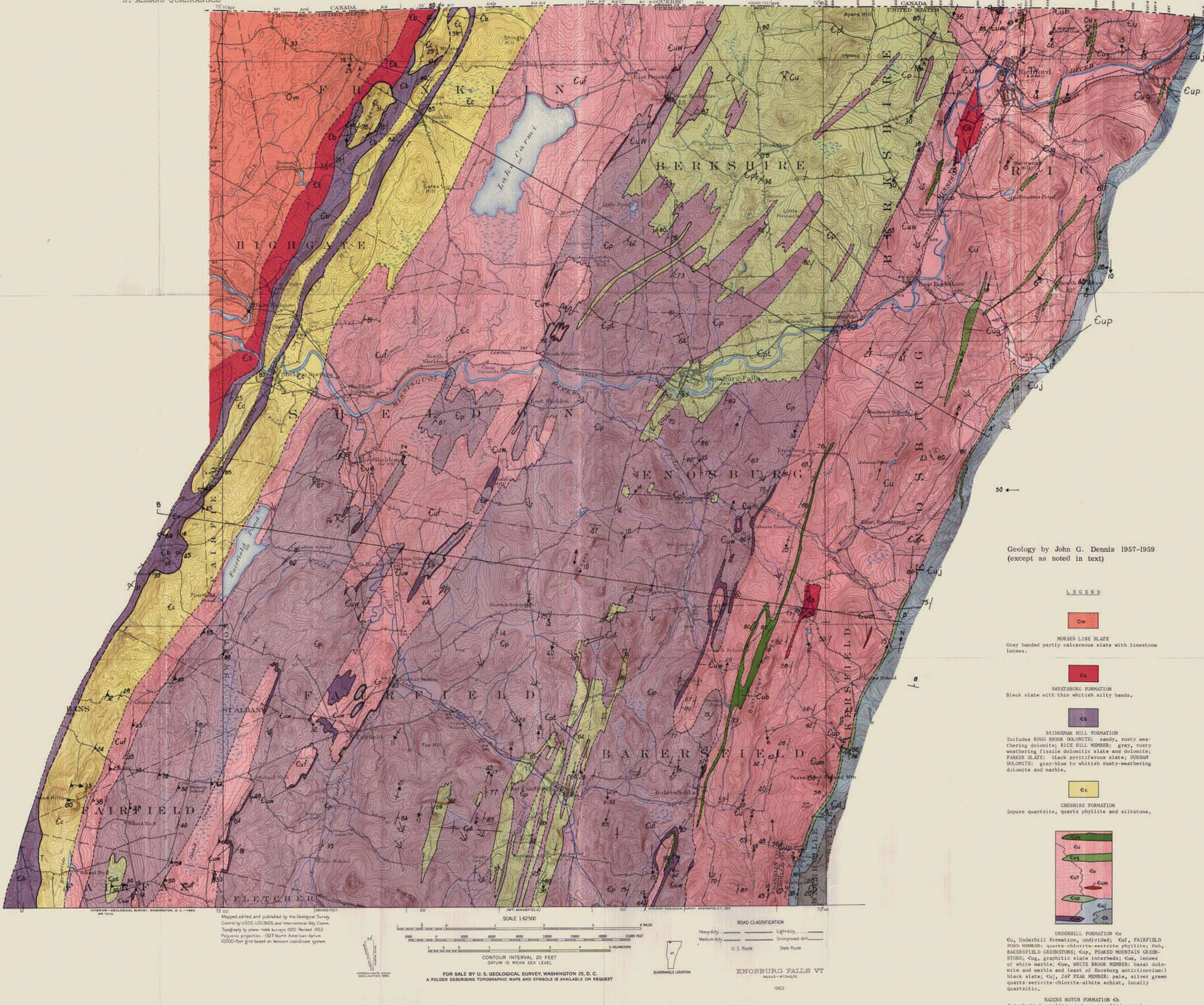
PLATE I



Geology by John G. Dennis 1957-1959
(except as noted in text)

LEGEND

Om

MORSES LINE SLATE
Gray banded partly calcareous slate with limestone lenses.

Ꞓs

SWEETSBURG FORMATION
Black slate with thin whitish silty bands.

Ꞓb

BRIDGEMAN HILL FORMATION
Includes RUGG BROOK DOLOMITE: sandy, rusty weathering dolomite; RICE HILL MEMBER: gray, rusty weathering fissile dolomitic slate and dolomite; PARKER SLATE: black pyritiferous slate; DUNHAM DOLOMITE: gray-blue to whitish rusty-weathering dolomite and marble.

Ꞓc

CHESHIRE FORMATION
Impure quartzite, quartz phyllite and siltstone.

UNDERHILL FORMATION Ꞓu
Ꞓu, Underhill Formation, undivided; Ꞓuf, FAIRFIELD POND MEMBER: quartz-chlorite-sericite phyllite; Ꞓub, BAKERSFIELD GREENSTONE; Ꞓup, PEAKED MOUNTAIN GREENSTONE; Ꞓug, graphitic slate interbeds; Ꞓum, lenses of white marble; Ꞓuw, WHITE BROOK MEMBER: basal dolomite and marble and (east of Enosburg anticlinorium:) black slate; Ꞓuj, JAP PEAK MEMBER: pale, silver green quartz-sericite-chlorite-albite schist, locally quartzitic.

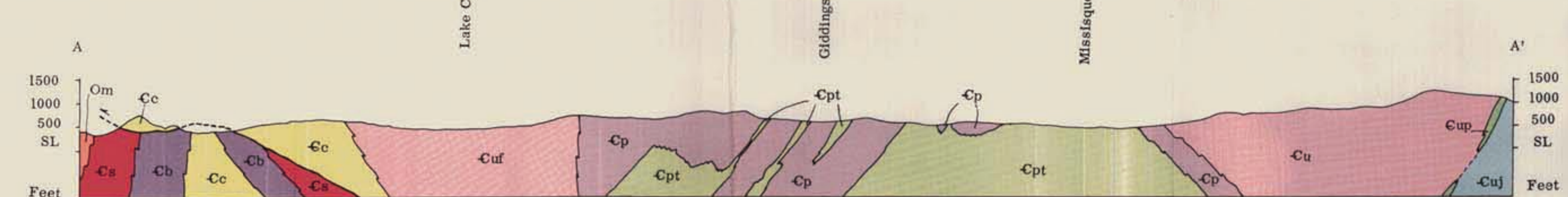
HAZENS NOTCH FORMATION Ꞓh
Interbedded graphitic and non-graphitic quartz-sericite-albite-chlorite schist; grades to quartzite.

Ꞓp
Ꞓpt

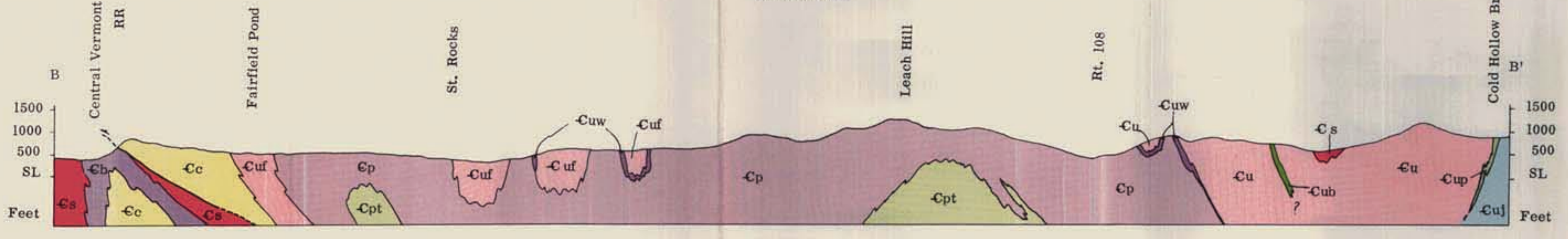
PINNACLE FORMATION
Ꞓp, Coarse greenish-gray graywacke and quartz-chlorite-sericite phyllite; Ꞓpt, Tibbit Hill volcanics and greenstone.

MAP SYMBOLS

Well-defined contact
Inferred contact
Dip and strike of beds
Strike of vertical beds
Dip and strike of overturned beds
Dip and strike of cleavage (St. Albans quadrangle only)
Plunge and trend of fold axes
Trend of glacial striations
Abandoned mine or prospect



SECTION AA'

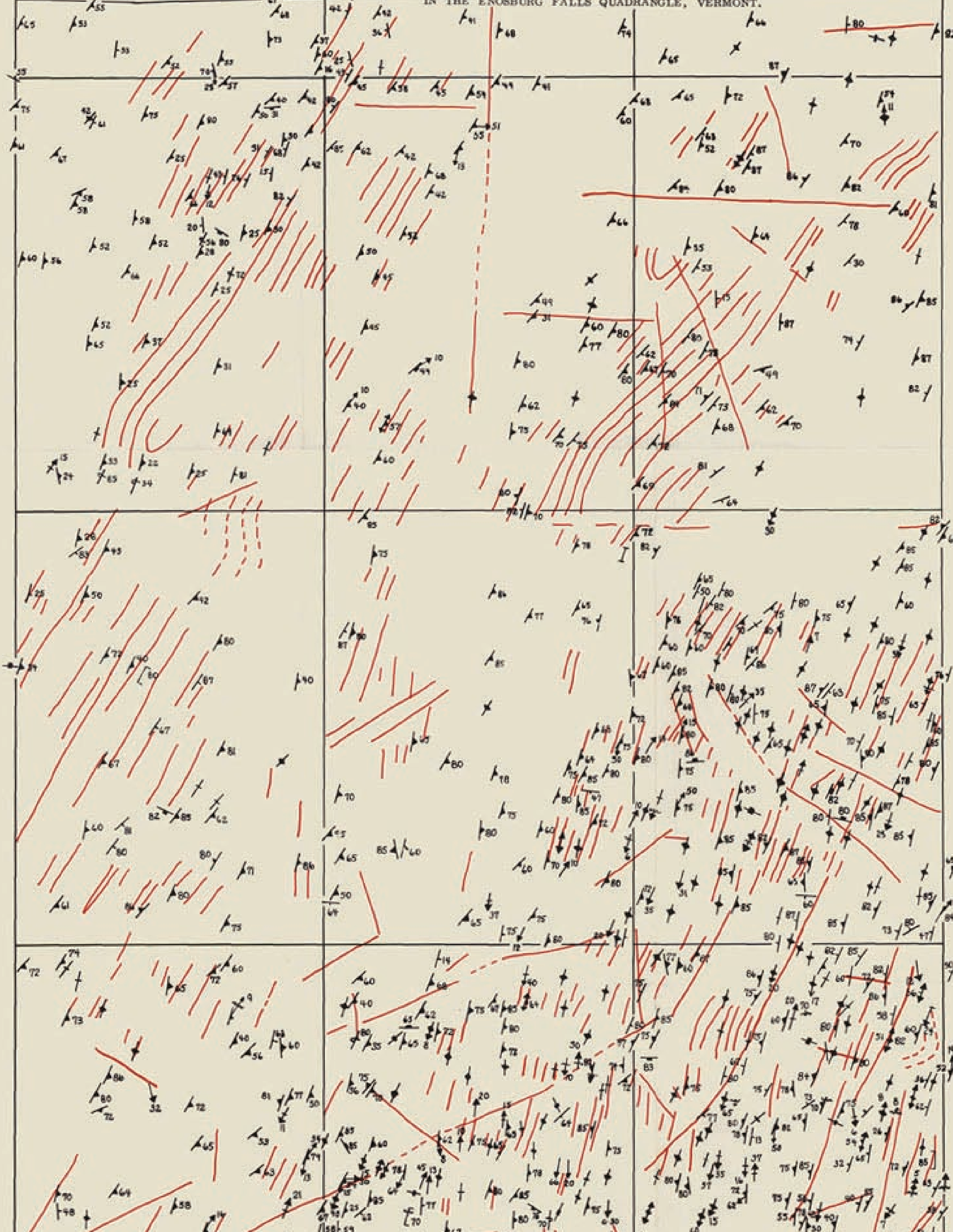


SECTION BB'

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 23)
Published 1964

PLATE 2. Legend on Plate 3

MINOR STRUCTURES AND AIR PHOTOGRAPH LINEARS
IN THE ENOSBURG FALLS QUADRANGLE, VERMONT.



VERMONT GEOLOGICAL SURVEY

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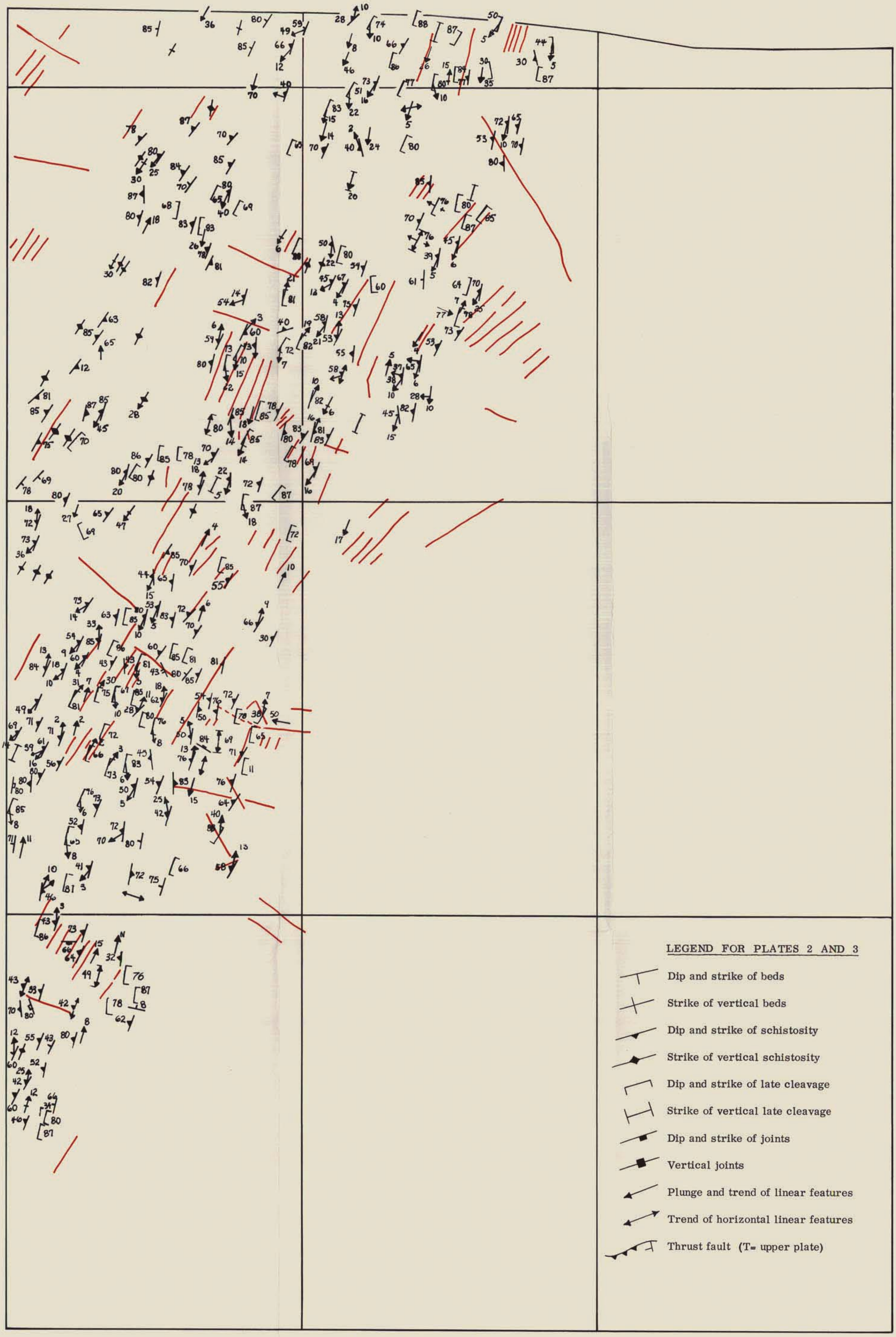
Published 1964

PLATE 3.

MINOR STRUCTURES AND AIR PHOTOGRAPH LINEARS IN A NORTHWESTERN PART OF THE JAY PEAK QUADRANGLE, VERMONT.



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