THE GEOLOGY OF THE RUTLAND AREA, VERMONT

By

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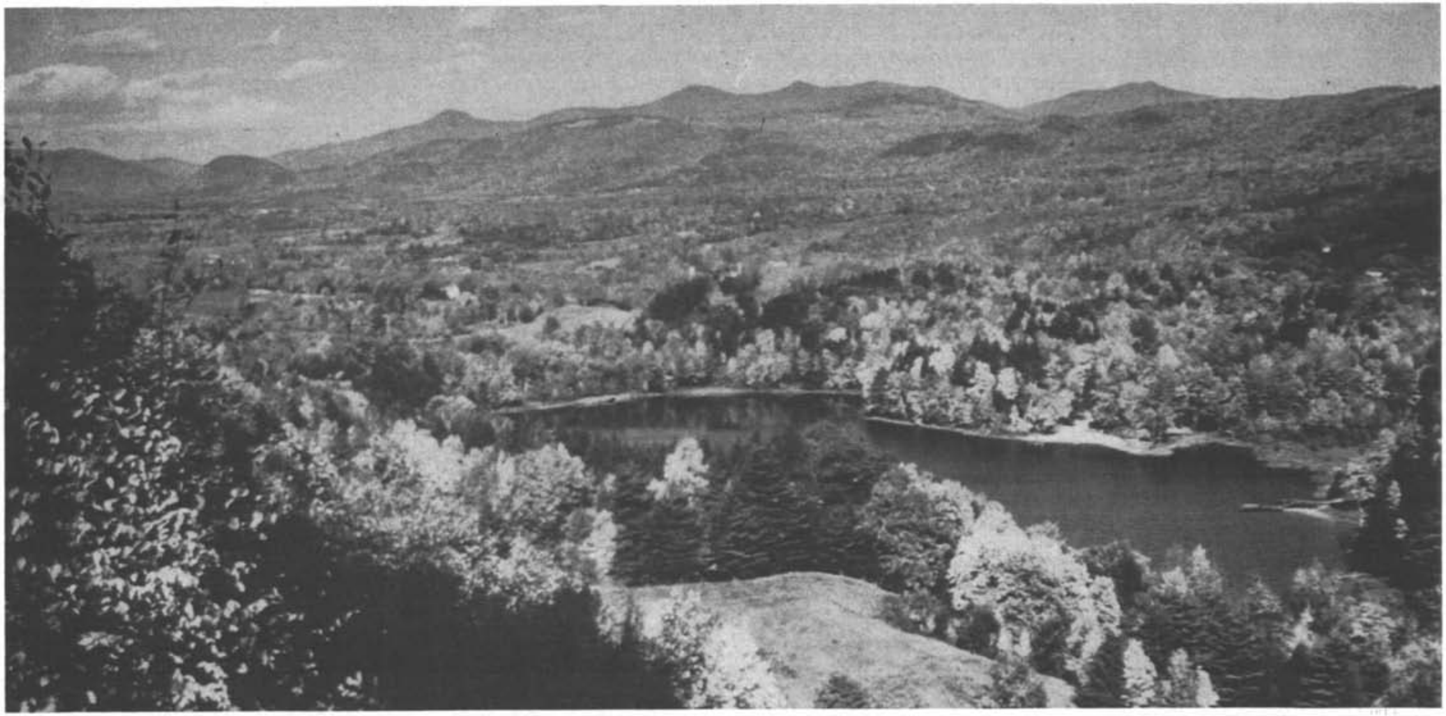

View from Elfin Lake, Wallingford, looking northward into the Rutland area. Picture is taken from the east slopes of the Pine Hill ridge looking northeastward across the Rutland valley to the Green Mountain Front and the high peaks of the main ridge. The highest mountains on the horizon are from left to right: Pico Peak, Mendon Peak, Killington Peak and Shrewsbury Peak. Photograph through the courtesy of Harold C. Todd, Fanwood, New Jersey.

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ABSTRACT

The Rutland area, in the Green Mountains of central Vermont, is underlain by a pre-Cambrian basement complex and flanking Cambro-Ordovician sequences arranged in the form of an anticlinorium.

About 10,000 feet of gneiss, schist, quartzite, marble and amphibolite are exposed in the core of the anticlinorium and lie unconformably beneath Lower Cambrian strata. These pre-Cambrian rocks resemble the Grenville of New York State and may be largely of sedimentary origin, as shown by abundant beds of marble, quartzite, graphitic schist and widespread compositional banding.

About 6,000 feet of dolomite, quartzite, phyllite and graywacke are included in the western sequence which ranges in age from Lower Cambrian to Middle Ordovician. Major unconformities occur at the base of the sequence and beneath Middle Ordovician black phyllite. The formations of the western sequence are correlated across the Pine Hill thrust with units described in west-central Vermont. The Mendon formation, the basal member of the western sequence and conformable beneath the *Olenellus*-bearing Cheshire quartzite, is correlated with rocks in Massachusetts and southern Quebec.

The eastern sequence of late pre-Cambrian to Middle Ordovician age includes about 17,000 feet of phyllite, phyllitic sandstone, dolomite, quartzite and the metamorphosed equivalents of mafic volcanic rocks. The sequence rests unconformably on the pre-Cambrian basement complex with the lowest unit, the Saltash formation, separated from the rest by a pronounced unconformity.

The eastern and western sequences are not in contact, but similarity of rock types suggests correlation of Mendon and Saltash formations, and correlation of black and green phyllites above the Middle Ordovician unconformity in the western sequence with the lower phyllite and metavolcanic units of the eastern sequence.

The pre-Cambrian basement complex shows the effects of two metamorphisms, the earlier producing garnet, diopside and actinolitic hornblende, the later partly altering these early minerals to chlorite, biotite and talc. The Cambro-Ordovician rocks show the effect of a regional metamorphism with a single isograd near the eastern edge of the area separating argillaceous rocks bearing chlorite and biotite to the west from those farther east bearing garnet. The later metamorphism of the pre-Cambrian probably occurred at the same time as the metamorphism of the Cambro-Ordovician rocks, as the younger mineral assemblages of the basement complex are similar to those in the younger mantling rocks.

Structurally the rocks of the Rutland area have the form of a gently north plunging anticlinorium and subsidiary anticline which are overturned to the west. Cambro-Ordovician primary structures and secondary rotational features are for the most part consistent with this pattern. Structural complications are abundant on the overturned west limbs of these major folds and include irregular fold plunge, thrust and normal faulting and locally more than one generation of folds. The overall structure points to the operation of a couple, with eastern rocks moving westward over western rocks. The important episodes of deformation and metamorphism probably occurred during the Middle Ordovician, and during the Taconic orogeny.

The basement complex consists of a central zone which may preserve intact earlier oblique pre-Cambrian trends, and a complicated thin marginal zone in which basement and flanking rocks have interacted during Paleozoic deformation. Structures of basement and mantle in the marginal zone indicate on the one hand that younger units may have locally developed trends paralleling older banding; and on the other that basement structures have been bent or dragged into near-conformity with the mantle. Major faulting appears to have been unimportant as a means of adjustment of the basement. No fault was found with appropriate geometry to have acted as a root zone for overthrust Taconic masses to the west of the area.

Minor structural features of the Cambro-Ordovician rocks are in a general way compatible with the major structure. This is shown by the agreement of rotational features (folds and rotated grains) with the mechanical requirements of larger scale flexural folding, and in linear features (mineral lineation and elongation, pebble elongation, boudinage, and intersections of planar features) by agreement either with the regional axis of folding or with the presumed direction of overall tectonic transport.

Compositional banding and principal rock foliation consistently parallel bedding. Slip cleavage is developed in micaceous rock types as a consequence of and during advanced stages of flexural slip parallel to foliation surfaces. Rotation and flattening parallel to foliation is expressed by internal spirals in albite and by external quartz overgrowth on albite, magnetite and pyrite. Crinkles, folds and intersections of foliation with fracture and slip cleavage are lineation which is parallel to the direction of regional fold axes. Quartz pod and pebble elongation, streaming, crenulation and rare folds are lineation subnormal to regional fold axes. Divergence from perpendicularity of these assumed contemporaneous lineation systems is explainable by examining the formation of doubly plunging folds. Pebbles are elongate in foliation surfaces, and apparently reflect flattening normal to foliation, rather than the usual deformation by uniform shear on foliation surfaces. Y-fabric, polygonal arches of mica and relics of lenticular and augen structure reflect widespread post-deformational recrystallization or annealing in the younger rocks.

INTRODUCTION

Location

The Rutland area covers about 230 square miles and includes the Rutland quadrangle (latitude 40°30′ to 43°45′, longitude 72°45′ to 73°00′) and portions of the Castleton, Wallingford and Woodstock quadrangles of the United States Geological Survey. The area is in central Vermont and extends from the marble quarries of West Rutland east over the main arch of the Green Mountains into the vicinity of Sherburne and Plymouth.

This area is traversed by several main highways: Routes 7 and 100 in a north-south direction, and Routes 4 and 103 in an east-west direction. About two thirds of the area is readily accessible by automobile during the drier months.

Regional Geologic Setting

The southern and central Green Mountains of Vermont are underlain

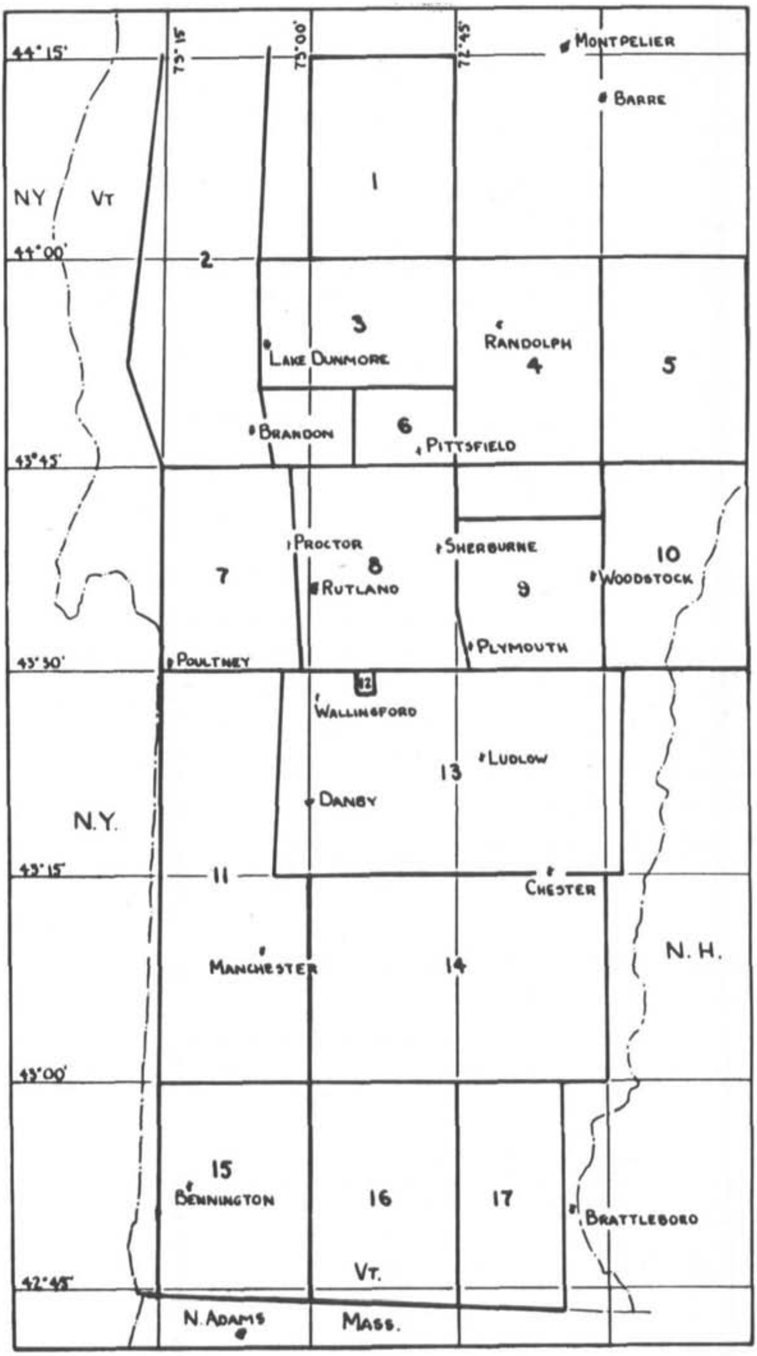

Figure 1. Index Map. (Key on opposite page)

KEY TO FIGURE 1

1. Lincoln Mountain Quadrangle, W. M. Cady, unpublished
2. West-Central Vermont, (Cady, 1945)
3. Rochester-East Middlebury Area, (P. H. Osberg, 1952)
4. Randolph Quadrangle, (White and Jahns, 1950)
5. Strafford Quadrangle, (Doll, 1944)
6. Pittsfield-Rochester Area, (H. Hawkes, 1941)
7. Castleton Quadrangle, (P. Fowler, 1950)
8. Rutland Area
9. Plymouth-Woodstock Area, P. Chang, unpublished
10. Hanover Quadrangle, J. Lyons, unpublished
11. Portions of Pawlet and Equinox Quadrangles, Dale (1894, 1904)
12. Cuttingsville Area, (Eggleston, 1918)
13. Portions of Wallingford and Ludlow Quadrangles, J. B. Thompson, in progress
14. Portions of Londonderry and Saxtons River Quadrangles, J. L. Rosenfeld, in progress
15. Bennington Quadrangle, J. MacFadyen, in progress
16. Wilmington Quadrangle, J. Skehan, unpublished
17. Brattleboro Quadrangle in part, G. MacDonald, in progress

by deformed and metamorphosed sedimentary, extrusive and plutonic igneous rocks ranging in age from pre-Cambrian to Ordovician. A central anticlinorium, a western synclinorium with thrust sheets and an eastern homocline are the major structural units. Prevailing structural trends are north-south and folds are generally overturned to the west. Metamorphism increases toward the east rather gradually from a low grade in and to the west of the anticlinorium to middle grade through the homocline. These general features of the section across central Vermont are markedly similar to those of sections to the north and south. Thus the crystalline pre-Cambrian core of the Green Mountains is matched within the Sutton Mountain of Quebec, the Berkshires of Massachusetts and the Blue Ridge of the southern Appalachians. Carbonate rocks and quartzites of the western limb of the Green Mountain anticlinorium appear with remarkable persistence from Quebec to Georgia. Finally, certain features within the highly metamorphosed southern Appalachian Piedmont suggest contemporaneity with the eastern band of metamorphic rocks of Vermont and Massachusetts.

At least three and possibly four Paleozoic disturbances have affected these rocks. One has occurred in mid-Ordovician and one at the close of the Ordovician; both Acadian (late Devonian) and Appalachian deformations have also left their imprint, but to an unknown degree.

In Vermont the western sequenee of autochthonous rocks consists of slightly metamorphosed limestone, dolomite and sandstone exposed in a synclinorium. The two limbs of the synclinorium rest unconformably upon the pre-Cambrian; the lowest member of the west limb is the Upper Cambrian Potsdam sandstone and that of the east limb is the Lower Cambrian Mendon formation. Fossils found throughout many of the units indicate a fairly complete section (excepting the Middle Cambrian) up to rocks of mid-Ordovician age. A large klippe reported within the southern portion of the synclinorium also contains rocks of Cambrian to Ordovician age. These latter rocks, however, are not of the same lithology as the autochthonous units but are predominantly slates. It has been suggested that their source was many miles to the east, in the central parts of the geosyncline far from the marginal facies upon which they have been thrust. The klippe underlies the Taconic Mountains and areas to the west in southern Vermont and western Massachusetts.

The pre-Cambrian rocks of the Green Mountain anticlinorium are similar to those within the Pre-Cambrian of the Adirondacks to the west. Paleozoic sedimentary rocks of the anticlinorium rest unconformably upon the pre-Cambrian in a large fold overturned to the west. The anticlinorium plunges gently northward in central Vermont and the northern limits of exposure of the pre-Cambrian rocks are complicated where long slivers of infolded younger rocks appear. The western limits of the pre-Cambrian rocks in the west limb of the anticlinorium are also rather complex for minor folding is irregular and faulting is common. The rocks of the Green Mountain anticlinorium show the effects of a low grade of metamorphism. The pre-Cambrian rocks in addition may have passed through an earlier and more intense stage of metamorphism.

The eastern sequence is, broadly speaking, a homocline in which rocks dip rather uniformly east off the core of the Green Mountain anticlinorium. Most of the rocks of the eastern sequence were originally argillaceous but also included volcanic rocks or their derivatives, sandstone and shaly limestone. The grade of metamorphism increases across this zone from the west where biotite occurs, to the east where diopside, staurolite and kyanite are fairly common. As would be expected, the age of the metamorphic rocks of the eastern sequence is uncertain, although a few poor fossils found throughout the length of the state indicate probably Middle Ordovician age and older.

The area of the present study is centered on the Green Mountain anticlinorium and includes several units of Cambro-Ordovician age and older.

The current major problems in Vermont geology concern the stratigraphy and structural setting of the Taconic Range, and the correlation of Paleozoic strata on the east flank of the Green Mountain anticlinorium with those on the west flank. The present study bears on both of these regional problems.

Previous Work

As early as 1861 a geologic map of Vermont had been completed. Edward Hitchcock (1861) and others had grasped the major elements of extremely complex structure and stratigraphy and presented an excellent starting point for further detailed work. The anticlinal nature of the Green Mountains was recognized largely through correlation of conglomeratic horizons in the latitude of Wallingford and Plymouth. Fossil discoveries soon thereafter by Wolff (1891), Foerste (1893) and

Dale (1892) in the carbonate beds of the Rutland and West Rutland valleys combined with Walcott's (1888) discoveries of organic remains in the Cheshire quartzite, dated the western mantling formations of the anticlinorium as Lower Cambrian through Ordovician. Whittle (1894 a,b) defined more fully the lowest western clastic formations and briefly described the basal units to the east of the range and the ancient rocks exposed in the core of the Green Mountains.

Some of the most significant early work was that of Dale (1892 to 1916). His studies included the slate belt of the Taconic area, the marble producing areas near Rutland, the Pine Hill-Danby Ridge and parts of the pre-Cambrian basement and eastern rocks. Dale discovered or clarified most of the structural units and correctly interpreted many of the structural details.

Perry (1928) developed the stratigraphy of the eastern sequence in nearly its present form, and correlated units in central Vermont with those in western Massachusetts and southern Quebec. At about the same time Foyles made several detailed studies of the western limb of the Green Mountain anticlinorium (1928, 1930) and attempted a far-reaching correlation of all the current stratigraphic units known in the region. Keith (1932) outlined the stratigraphy of the western sequence of Paleozoic rocks, and Bain (1931 to 1938) examined many of the units in some detail. Bain's studies of marble quarries near the Rutland area prompted the first description of the somewhat unique manner in which carbonate rocks were deformed.

In mapping a large area extending across the range near Mendon and north and south along the Plymouth-Pittsfield valley, Hawkes considered both the Taconic problem and the correlation of units across the Green Mountain anticlinorium (1940, 1941). He traced the units of Perry northward and suggested that the Taconic thrust fault was located within the western exposure of the pre-Cambrian basement rocks. Thompson (1950) and Chang (1950) continued work in the eastern sequence with detailed studies of the Ludlow and Woodstock-Plymouth areas respectively. Thompson clarified the stratigraphy and established criteria for the interpretation of most of the minor structural features in the area. He demonstrated for the first time that map units could be traced within the pre-Cambrian basement complex.

Cady (1945) established detailed stratigraphy of the western sequence and provided the basis for work to the south and in the Rutland area.

His mapping was continued to the south by Fowler (1950) and to the east by Osberg (1952) in areas immediately adjacent to this study.

At present, work is being conducted in nearby areas by P. H. Osberg, John MacFadyen, J. B. Thompson, J. L. Rosenfeld and E-an Zen.

Purpose of Study

The primary purpose of this study was a detailed investigation of the structural features of the fold belt and the examination, interpretation and correlation of structural features from the scale of major folding to that of mineral orientation and micro-fabric.

The structural geologic setting of the Rutland area posed many intriguing problems. It was hoped to clarify:

1. the broader structural and stratigraphic relations within the pre-Cambrian core rocks of the Green Mountain anticlinorium near Rutland. This might lead to significant formulation of the interaction of crystalline and mantling rocks during mountain building. Information was sought relating to possible root zones of the Taconic klippe.

2. the stratigraphic relations of the rocks immediately above the core gneisses, and

3. complex arrangements of units within the Pine Hill arch and in the Pittsford-Chittenden area.

Method of Study

About ten months of summer field work were carried out during 1950, 1951 and 1952. Petrographic studies and preparation of manuscript occupied a major portion of the winter season 1951–52 and the entire winter season 1952–53.

Two types of field map were used, the quadrangle maps of the U. S. Geological Survey (Scales 1:62,500 and 1:31,680) and aerial photographs (Scales: 1:31,680 and 1:25,000 approx.). Various combinations of topographic map, altimeter and photograph were used to plot field data, the formation boundaries being ultimately placed on the appropriate quadrangles in true horizontal position.

Poor exposure in the Rutland area is responsible for a few of the inconsistencies and the over-simplification of the geologic mapping. Easily two-thirds of the area underlain by carbonates in the Rutland valley is completely mantled by glacial and river debris. The situation in the main range, underlain by Pre-Cambrian, is scarcely improved by the addition of heavy forest cover.

Many important features of central Vermont stratigraphy have not yet been firmly established. Several of the stratigraphic problems of the Rutland area can only be solved by detailed mapping of the surrounding areas. This mapping is now in progress but the results are as yet unpublished; therefore the correlation and age designations presented in this paper are tentative.

Acknowledgment

Many people have given me inspiration, criticism and material assistance during the course of this study; though perhaps unnamed here, they should know my appreciation, without accepting responsibility for my errors or opinions.

The problem was suggested by Professors Harold W. Fairbairn of M.I.T. and James B. Thompson of Harvard University, who have since guided the study and provided constructive criticism.

The field work was supported by the Vermont Geological Survey. Professor Charles G. Doll, State Geologist, spent several days with me in the area. Through the support of the National Science Foundation the entire winter of 1952–1953 was devoted to the laboratory aspects of the problem. Much of the petrographic equipment used at M.I.T. is the property of Professor Harold Fairbairn.

Dr. Wallace M. Cady of the United States Geological Survey reviewed the manuscript and map and was extremely helpful in some of the most complex field problems. Professor Marland P. Billings and Mr. Gordon MacDonald, both of Harvard University, critically examined parts of the manuscript and map.

Four weeks of the field studies were made especially pleasant through the assistance of E-an Zen, Severo Ornstein and Alexander Nicholson.

Many of the geologists working in Vermont and at M.I.T. have been most helpful by discussion of various aspects of laboratory and field problems. (If any of their ideas have been incorporated with my own and seem uncredited, it is regrettable but probably inevitable when a group works closely together.) Among these are Gordon MacDonald, E-an Zen, John MacFadyen, Terrence Podolsky, Philip Osberg and John Rosenfeld.

My mother, Mrs. Frances Brace, helped to improve the literary style of the manuscript and throughout the work gave inspiration and material support.

Physiography

The Rutland area consists of a mountainous central and eastern belt and a narrow lowland strip to the west. A major divide extends north-south separating eastward drainage to the Connecticut River from that emptying to the west into Lake Champlain. Several lesser divides separate the headwaters of the east-flowing White, Ottauquechee and Black Rivers.

The north or northwest trends of the hills and valleys are conspicuous. The Pine Hill ridge just west of Rutland is a remarkably straight rise which extends from Proctor at the north to Danby at the south. The relief of the ridge is 800 to 1,000 feet above its companion depression the Rutland valley; the latter is part of a continuous lowland that extends into southern New England. A confusion of small hills of less than a thousand feet relief appear in the northwest corner of the area, giving way northward to the more orderly north-south ridges and valleys of the Brandon area.

East of the Rutland Valley the Green Mountain front rises 1,500 feet abruptly, to form a pronounced linear break in the topography traceable for the length of Vermont. Behind it to the east rise the highest points in the area reaching in Killington Peak an elevation of 4,200 feet above sea level. The ridges in this central area conform to no general trend, showing both northwest and west alignment. The high peaks, Killington, Pico, Blue Ridge, and others are part of the main ridge which extends with few breaks for the length of Vermont and includes Mansfield, Lincoln and Carmel to the north and Stratton and Haystack to the south.

Still further east the average elevation of the sharply dissected upland decreases and the trends of the ridges and valleys are again roughly north-south with about 1,500 feet relief. A deep narrow valley near the eastern margin of the area can be followed along a fairly straight course for nearly 20 miles; it includes the Plymouth Lakes, Woodford Reservoir (Timber Lake) and the villages of Pittsfield, Sherburne, West Bridgewater, and Plymouth Union.

The shape of the land surface conforms in a broad way to the structure and composition of the bedrock. Thus large areas of lowland, such as the Rutland Valley, are underlain by carbonate rocks; the more resistant quartzites and schists being exposed in ridges such as Pine Hill and the Green Mountain front. The conformity of the topography to the nature of the bedrock is hardly traceable in detail due to the effects of both glacial action and metamorphism. The erosional and depositional agencies of glaciation have indifferently carved the upland and have flooded the lower elevations with poorly sorted debris, altering drainage patterns and blanketing the lower mountain slopes. Metamorphism treated broadly has acted to equalize these primary differences of the rocks which guide the weathering processes.

STRATIGRAPHY AND GENETIC RELATIONS

General Statement

Unconformities separate the rocks of the Rutland area into a central basement complex, an eastern sequence (Table 2) and a western sequence (Table 1). Correlation of the eastern and western sequences is discussed following their description. Petrographic details of the rocks and definition of terms will be found in the Appendix.

The use of time designation in this report (Lower Cambrian, for example) is naturally less accurate than it might be in regions of less severe deformations. The emphasis in mapping therefore is placed on delimitation of *rock units* rather than *time units*. Lithologic similarity guided by the principles of facies change is used in correlation, and the major subdivisions in a succession are placed at surfaces of unconformity. The more accurate biozonal subdivision[1] is not possible in central Vermont.

Few features of the sedimentary record are preserved in the metasedimentary rocks of the Rutland area. Details of stratification and texture, minor structures and original lithology are largely masked in the formation of crystalline schist and marble. Although gross features may remain, the origin of these rocks often must be interpreted on the basis of approximate chemical composition. Here numerous difficulties arise. The chemistry of sediments is poorly known (Albee, 1952); the routine optical study of low grade metamorphic rocks yields at best a crude approximation of the chemical composition. The investigation of the origin of the metasedimentary rocks proved nevertheless to be extremely interesting and is reported in the paragraphs to follow.

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[1] A case in point is the question of the base of the Cambrian system. Wheeler has properly suggested that the base of the Olenellus biozone be used (1947); immediately the problem arises of dating the several hundred feet of rock conformable beneath the lowest fossil-bearing beds. These unfossiliferous beds rest with profound unconformity upon an older metamorphic complex. I consider this major unconformity to be the base of the Cambrian in the area, for it is entirely reasonable that the lack of fossils below the lowest *Olenellus* horizon may be explained as well by destruction as by non-formation.

Throughout this report the thicknesses of formations are approximate in as much as both the identity of primary features and the secondary thickening and thinning of strata are hard to determine.

Pre-Cambrian Basement Rocks

General Statement

Preliminary examination of about 100 square miles of pre-Cambrian exposure has been surprisingly fruitful. Several distinct rock types are sufficiently widespread to show the structural trends as well as the primary characteristics of the basement complex.

The Pre-Cambrian is divided into the Saltash and Wilcox formations and the Mount Holly complex (Whittle, 1894). The relative age of the Saltash and Wilcox formations is not known; but since their grade of metamorphism is lower and they rest unconformably upon the Mount Holly, they are considered younger than the Mount Holly. The Saltash grits, black phyllite and quartzite are closely related to the Eastern sequence and are described with that group. The Wilcox consists of over 3,000 feet of schists, phyllite, dolomite and gneiss. The rock types in the Mount Holly and the percentage of total area of pre-Cambrian which they underlie are as follows:

gneiss, 75 percent
quartzite and quartz schist, 20 percent
amphibolite and greenstone, 2 percent
marble and lime silicate, 2 percent
pegmatite, quartz-tourmaline veins, 1 percent

An early metamorphism has produced garnet in the argillaceous types and actinolitic hornblende in the rocks of basaltic composition. A second metamorphism has followed, probably at the time of regional metamorphism of the Cambro-Ordovician rocks, and has partly reduced the original minerals to those typical of the biotite or chlorite zones of metamorphism. Within the central part of the pre-Cambrian area the structural trends of the Mount Holly complex appear rather simple and persistent, with strata dipping for the most part gently to the northeast. At least 7,000 feet of strata has been mapped in this central region where the quartzite and interbedded marbles attest the sedimentary origin of the rock; within the large area underlain by gneiss, however, origin is obscure and interpretation must await further field study.

The Mount Holly complex and Wilcox formation are in contact with

Cambro-Ordovician strata at a major unconformity. The structural trends, lithology and metamorphism are so different in the two sequences that the surface without doubt represents the greatest break in the sedimentary record in the area. The Mount Holly and Wilcox are considered to be pre-Cambrian since the rocks above them are of Lower Cambrian age.

Relation to Cambro-Ordovician Strata

The Cambrian-pre-Cambrian boundary, while often difficult to locate in detail, emerges with regional mapping as a surface of profound unconformity. This surface has been located with some difficulty in the past since a certain amount of experience with the structural and metamorphic aspects of both units is necessary for differentiation.

The lowest beds of the Cambro-Ordovician often closely resemble the lithology in the underlying Mount Holly, because the basal conglomerate and grits are in part composed of material derived from sources in the Mount Holly near the site of deposition. Fragments of blue quartz are typical in the Tyson and Nickwacket units, and can usually be related to beds of blue quartz or blue quartz bearing pegmatites in the Mount Holly. Frequently, however, these sources have themselves been strongly deformed and the blue quartz beds broken and strewn about in the foliation mimicking a true clastic texture. Locally fragments of vitreous quartzite apprcach the size of the outcrop in which they are seen; mapping in these areas may be rather uncertain. A peculiar occurrence of magnetite helps to identify the location of the unconformity while obscuring it in detail. Octahedra of this mineral appear on both sides of the contact for several feet within both units. With regard to the Mount Holly, this is probably due to prolonged chemical weathering of the ancient surface, with conversion of ferrous iron of the original rock into ferric iron, which appears upon metamorphism as magnetite.

As indicated on the geologic map (Plate 1) the Mount Holly units are often truncated at the unconformity beneath the Cambro-Ordovician. This feature is rarely observed in the limited field exposures, as the typical situation in outcrop is rather perfect parallelism of both sequences across their surface of contact. This parallelism is in part produced by younger structural elements, such as cleavage and folds in the Mount Holly, but to a larger extent is due to the actual rotation of the older compositional banding into apparent conformity with the surface of the contact.

The metamorphic characteristics usually offer the most reliable criteria for the separation of the Mount Holly complex and the Cambro-Ordovician sequences. The basement rocks have been elevated to the garnet zone of regional metamorphism and strongly recrystallized; they contain abundant pegmatite and quartz-tourmaline veinlets. Gneiss and coarse-grained marble are common and contrast strongly with the thick, medium- to fine-grained grit, phyllite and sugary fine-grained dolomite of the Tyson and Nickwacket units. In addition the basement rocks show the effect of a second metamorphism. Thus, higher grade minerals such as almandine, hornblende, tremolite and intermediate plagioclase have been altered to assemblages similar to those formed during the younger metamorphism. The grain size of the Mount Holly rocks is reduced to the characteristic gritty and phyllitic texture of those above the unconformity. The incomplete nature of the second metamorphism, however, imprints a rather distinct character in the basement rocks. They appear weathered, due to red iron oxides released in the alteration of the mafic minerals and to the chalky, greenish color of the saussuritized feldspar. The "dirty" appearance of the rocks and the occurrence of pegmatite and rock types peculiar to the Cambro-Ordovician is usually the basis for differentiation of these two sequences.

The unconformity is rarely observed in the Rutland area[1], but can usually be located within a hundred feet. It probably had little relief except where thick pre-Cambrian quartzite is now in contact with the Cambro-Ordovician.[2]

Map Units

Wilcox Formation: The name Wilcox is applied here to a group of schists, dolomite and gneiss about 3,000 feet thick, exposed in the western part of Mendon and Shrewsbury towns. This unit may best be seen on the slopes south of Cold River at Wilcox Hill and on the northwest slopes of Mendon Peak. The general synclinal nature of the Wilcox in the area exposed is suggested by minor structural features and is supported by the appearance of the Mount Holly type of lithology in apparent unconformity to the southeast. Unfortunately the details of the west limb of the syncline are not known due to poor exposure.

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[1] About 1.3 miles SSE of East Clarendon, 1250 feet elevation; just north of route 103 at Clarendon Gorge, Clarendon; 0.7 mile SE of North Chittenden, 1700 feet elevation; east slopes about 1 mile SE of North Sherburne.

[2] "Rabbit Ledge," 0.4 mile SE of Mendon Village; "Bald Mountain" 2.3 miles ENE of Pittsford and the summit of Pine Hill 1.3 miles ESE of Proctor.

Most of the Wilcox is green, white and black schist, enclosing thin dolomite beds several hundred feet above the base. About 500 feet of pegmatitic quartzose gneiss occur near the middle of the formation. The schists are feldspathic and dolomitic, show a fine banding and contain thin buff and blue quartzite beds. Chloritic schist is most abundant although graphitic and sericitic schist are conspicuous. Quartz pebbles are locally abundant near the base of the formation, but no basal conglomerate was observed. Above the middle gneiss member quartz and feldspar increase in amount, forming dark schistose grits which contrast with the lower strictly argillaceous types. Many of the rock types of the Wilcox resemble varieties found in the Tyson and Pinney Hollow formations, but the Wilcox is retained in the pre-Cambrian as it contains abundant evidence that it is more closely related in time to the Mount Holly. The coarser texture of the schist and dolomite attest the higher grade of metamorphism reached by these materials; the blue quartz beds and rare saussurite further suggest that this formation had undergone somewhat the same thermal history as the older basement rocks. It is quite probable though that the Wilcox formation is a relatively youthful member of the Pre-Cambrian.

Mount Holly Complex: Marbles, lime-silicate-bearing rocks: Limestone, dolomite and various lime-silicate-bearing rocks occur as thin variable units throughout the Mount Holly complex. They appear above the thick quartzite-schist unit north and east of Pico Peak and Blue Ridge Mountain, in two thin bands south of Killington Peak and north of Mendon and Shrewsbury Peaks, and again in two bands south of Comfois Hill and west of Saltash Mountain. Tremolite schist occurs in the Pre-Cambrian north of Blueberry Hill and west of Chittenden, and in scattered localities on Flat Rock and Boardman Hill and south of the Mill River. Some of these occurences have been described by Hitchcock (1861, p. 555), Whittle (1894, p. 414), Dale (1915, p. 20), Eggleston (1918), and Hawkes (1940, p. 47).

The thickness of these units ranges from 10 to 200 feet and the lithology varies from a pure dolomite to a complex lime-silicate rock. The fine-grained dolomite on the north slopes of Pico Peak, for example, is nearly identical with varieties found in the Mendon formation of Cambrian age, whereas in the same band near the Gifford Woods State Park in Sherburne the rock is coarsely crystalline and contains abundant tremolite, talc and epidote. This variation in mineralogy and texture is ascribed to differences both in original composition and in the chemistry of the surrounding rocks. Elsewhere, as at Northam[1] and south of the Mill River, the metamorphism has been of about the same intensity, but high lime and alumina in the original sediment is reflected by abundant epidote and zoisite. Graphite, phlogopite, sphene and quartz are constant accessories in all of the marbles and lime-silicate rocks; chlorite, albite and microcline appear rarely.

Typically the tremolitic rocks are a mesh of coarse colorless amphibole rods in a matrix of carbonate and fine micaceous material; talc appears in thin schistose partings. Structural features are vague, but show that these rocks have generally undergone intense deformation.

Amphibolite, greenstone: Amphibolite and greenstone occur sparingly in the southern part of the Mount Holly complex as units 5 to 50 feet thick conformable with, or cutting, the banding of the enclosing gneiss and quartzite. Coarse plagioclase-garnet amphibolite appears in great abundance on the south and west slopes of Robinson Hill and in a few outcrops south of the Mill River. The more typical epidote amphibolite occurs sparingly as thin discontinuous units in the vicinity of Comfois Hill and Northam. Close to the unconformity at Sherburne and north of Great Roaring Brook amphibolites are altered to chlorite, epidote, albite and carbonates and resemble the greenstones found in the Cambro-Ordovician Pinney Hollow formation. Amphibolite and greenstone are not sufficiently thick map units to warrant separate designation on the map and sections (Plates 1, 2).

The typical minerals are a greenish actinolitic hornblende, highly altered intermediate plagioclase, and biotite. Alteration of the mafic minerals has yielded a pale birefringent chlorite, and the feldspar in most cases is almost completely converted to albite containing a core of zoisite, epidote and calcite (Plate 4, Figure 2). Sphene, magnetite and quartz are the common accessories.

The amphibolites of the Mount Holly are usually massive,showing at best a weak lineation of the amphibole rods. The majority of the occurrences probably represent dikes and sills emplaced before the final episode of pre-Cambrian metamorphism. The thick masses of coarse amphibolite with associated mafic pegmatite at Robinson Hill may well represent a larger mass of gabbroic intrusive. At a single locality one mile northwest of Shrewsbury, large amounts of quartz with the amphi-

[1] North Shrewsbury

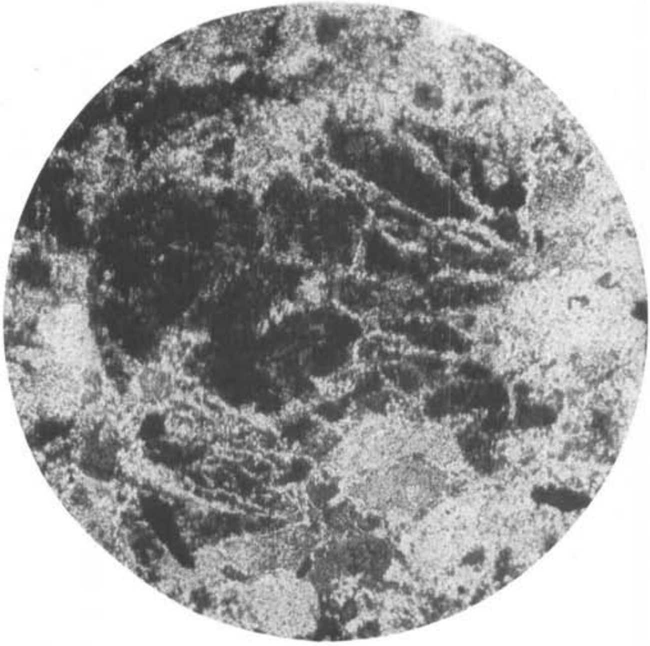

Figure 1. Garnet partly altered to chlorite and cut by veinlets of chlorite, sericite and quartz; from plagioclase-garnet gneiss, a half mile south of the summit of East Mountain, Mendon; crossed Nicols. x 37.

Figure 2. Amphibolite near summit of Robinson Hill, Shrewsbury. Andesine (gray clouded masses) is largely altered to albite, zoisite and quartz. Amphibole is partly replaced by chlorite; crossed Nicols. x 20.

PLATE 4, PRE-CAMBRIAN GNEISS AND AMPHIBOLITE.

bole and epidote suggests a tuffaceous origin for thin beds of amphibolite within grits, gneiss and schist.

Biotite-microcline gneiss: Biotite-microcline gneiss is one of the common rock types observed in the Mount Holly complex. The gneiss occurs both as large uniform masses devoid of other lithology, and as beds several feet thick within garnet gneiss, quartzite and limestone. Areas typical of the first variety are the south and west slopes of Mendon Peak between elevations 2,600 to 3,300 feet, the slopes east of Chittenden Reservoir to elevation 2,100 feet, and lower elevations south of the village of North Sherburne.

The gneiss is grey with a weak foliation due to dark schist septa which separate bands or augen of pink feldspar. The augen or bands are an eighth to a half inch thick and contain broken microcline grains, sodic plagioclase and abundant clear quartz. Whereas the microcline is surprisingly fresh, plagioclase commonly encloses abundant clinozoisite, epidote, calcite and quartz inclusions in the center of rounded grains. The thin dark schist layers contain sericite, biotite, chlorite and epidote. The whole rock is peppered with dark dust-like particles and stained with iron oxides.

The origin of the biotite-microcline gneiss is not known. The association of this rock with quartzite and marble suggests a sedimentary origin for part of it, but a volcanic or plutonic origin is more likely for most occurrences. The great extent of the gneiss bodies suggests, as does the mineralogy, that they represent sheared felsic plutons. Thin conformable beds of gneiss may have been felsic tuffs or intrusives (Oral suggestion of Thompson, 1952).

It is impossible in our present knowledge of the gneiss to assess the possibility of a metasomatic origin.

Quartzite, Schist: Thick vitreous quartzite occurs near the summits of nearly all of the mountains underlain by pre-Cambrian strata in the central core of the range and to the west and northwest in the smaller folds. This is extremely fortunate for it has been possible to unravel much of the structure of the core by tracing these resistant units. The most impressive quartzite bodies are those of Blue Ridge Mountain and Pico Peak, Rabbit Ledge (0.5 mile southeast of Mendon) and Bald Mountain (2.3 miles northeast of Pittsford). These formations are nearly pure quartz in vitreous, blue-white beds 1 to 20 feet thick. Bedding is marked by thin dark bands or schist layers, with coarse white quartz-tourmaline-muscovite pods. Within the 2,500-foot unit exposed on the east slopes of Blue Ridge Mountain massive quartzite beds are associated with quartz-sericite schist, chloritoid schists and thin gneiss.[1]

Schists are not conspicuous within the basement complex except in thin beds associated with quartzite. They are pink or white quartz-sericite-chlorite rocks and commonly contain several percent of minerals such as chloritoid, coarse graphite and epidote. The crinkling and faint compositional banding call to mind types of lithology common in the younger Tyson and Pinney Hollow formations; the pre-Cambrian schists, however, exhibit such distinguishing features as strongly undulose quartz grains and fairly coarse texture. Quartz-feldspar-tourmaline pegmatites commonly cut the pre-Cambrian schists as well as the other rocks of the basement complex.

Garnetiferous Quartzite: Garnetiferous quartzite is found at the summits of Killington, Little Killington and Mendon Peaks and in a band north and west of Saltash Mountain and south of the Mill River. These quartzites contain 60 to 90 percent quartz with garnet, chlorite, sericite and minor sodic plagioclase. The rock is usually red or green in color, in beds several feet thick and associated with plagioclase-garnet gneiss and lime-silicate rocks. The quartz is fractured, shows intense undulose extinction, with sutured or crushed grain boundaries. Garnets as much as a quarter inch in diameter are largely altered to chlorite, and are commonly broken and strewn out in the foliation. Micas and accessory sphene, epidote and opaque minerals form a fine mesh containing the quartz and garnet. The fragmental nature of these quartzites has often been misleading; Dale (1916), Whittle (1896) and Foyles (1930) have described conglomerate in the Mount Holly which in all probability are these impure quartzites. Some unknown deformational factor has hindered flowage of quartz, allowing it instead to fracture and mimic a true clastic texture.

Garnet gneiss, plagioclase gneiss: Gnarled dark greenish gneiss is common throughout the Mount Holly complex. A 2,500-foot band is associated with quartzite and limestone between Killington and Little Killington peaks; similar rock appears north of Saltash Mountain, south of North Sherburne and on the slope a mile north of Shrewsbury. Because of the heterogeneous character of the rock it is a singularly poor map unit. The mineralogy is usually obscured by the dark rusty alteration

[1] Whittle (1892) has described "Ottrelite" from these beds. I have found it only at the summit of Blue Ridge Mountain.

products although occasional chlorite pods a quarter of an inch or so in mean diameter mark the site of early garnets. The more feldspathic varieties resemble the cleanly banded microcline gneiss although plagioclase alteration has usually given the light colored bands a greenish tinge. Formation boundaries within the gneisses are not drawn on map or sections (Plates 1, 2) since it was not possible to subdivide them consistently in the field.

Plagioclase in the gneiss curiously resembles that found in basal Cambro-Ordovician beds. Albite-oligoclase is rounded and contains abundant inclusions of epidote, zoisite, calcite and sericite, recalling the typical porphyroblasts of the younger strata. Apparently the gneiss contained an early plagioclase more calcic than is now seen in the rock; its complete reconstitution rather contrasts the partial decay of the associated minerals such as garnet and biotite. Chlorite usually exceeds biotite in the gneisses, following the trend commonly observed that potash feldspar and biotite appear together as do plagioclase, garnet and chlorite. Titanium is in sphene and iron-stained pits suggest ankerite and pyrite. Garnet is filled with veinlets of chlorite and rarely biotite (Plate 4, Figure 1).

Miscellaneous: Peculiar gritty and schistose rocks occur in the Mount Holly near the contact with Cambro-Ordovician and Wilcox. Mineralogy and fabric are similar to that of the younger rock above the contact although mineralogic and structural relicts suggest that the rocks are merely disguised gneiss and schist of the basement complex. Garnet is absent, feldspar and micas rearranged and the banding lost in the devolopment of foliation parallel to that in the mantling sequences. The few vestiges of the initial character of the rock are the sheared pegmatite and thin but intact beds of bluish quartzite and grey gneiss. These grits and schist are the most troublesome units in the area since they weaken the lithologic distinction across the unconformity, and can seldom be correlated with the more typical Mount Holly members into which they grade.

Bright green chrome mica schist occurs on a small saddle on the north side of Round hill in Shrewsbury township. It is enclosed by grits and schist of the Mount Holly and has only been found near the unconformity with Cambrian rocks. A chromium-bearing white mica is 85 percent of the rock[1] with chlorite, quartz and accessories.

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[1] According to J. S. Diller and W. T. Schaller of the U. S. Geological Survey, the mica is fuchsite and the Cr_2O_3 content of the rock is 2.03 percent (Dale 1914, p. 42).

Genesis and comparison with the Grenville of New York

The Mount Holly and Wilcox formations have the following common features that suggest a sedimentary origin: (1) interbedded quartzite and limestone, (2) graphitic schist and (3) pronounced compositional banding; thick masses of microcline and plagioclase gneiss and the amphibolites are exceptions.

The Mount Holly complex and the Grenville[1] of New York have features in common although the most recent thermal histories of the two groups have without doubt been profoundly different. Both contain abundant feldspathic gneiss, quartzite and limestone; graphitic rocks are not uncommon and amphibolite is unimportant. Grenville marbles are much thicker, however; as much as 12,000 feet are reported (Miller, 1914, p. 8). The origin of the Grenville series is universally considered to be sedimentary, in fact Miller (1924, p. 31) has referred to the Grenville as "marine water laid rocks." The age of the Grenville is "probably Archean" (Miller, 1924, p. 34).

Western Sequence Of Paleozoic Rocks

General Statement

The western sequence consists of Cambrian and Ordovician rocks and is exposed from the western margin of the pre-Cambrian rocks to the Pine Hill ridge. It includes about 6,000 feet of sparsely fossiliferous dolomite, quartzite, limestone, graywacke and conglomerate (Table 1). The only major break is an angular unconformity beneath the Hortonville phyllite and marble.

With the exception of the Mendon formation the members of the western sequence are treated briefly, for they are poorly exposed in the Rutland area; reconnaissance studies outside of the area of detailed study was necessary to make separation of the units and correlation possible.

The effects of metamorphism appear to be about the same throughout, with biotite occuring in feldspathic argillaceous rocks and phlogopite in siliceous dolomite.

[1] Alling, H. L., 1927; Buddington, A. F., 1939; Buddington, A. F. and Whitcomb, L., 1941; Cannon, R. S., Jr., 1937; Dale, N. C., 1935; Krieger, M. H., 1937; Miller, W. J., 1914, 1917, 1924.

Table 1

Western Sequence

Age	*Formation*	*Lithology*	*Thickness*
M. Ord.	Hortonville slate	black phyllite, blue calcite marble	>300′
		UNCONFORMITY	
Ord.?		grey, white calcite marble	?
U. Camb.	Clarendon Springs Dolomite	dark grey limy dolomite	200′±
	Danby formation	thin-bedded grey dolomite with glassy cross-laminated quartzites	1000′±
L. Camb.	Winooski dolomite	pink, white, blue thin-bedded dolomite	600′±
	Monkton quartzite	varicolored, thin-bedded dolomite; thick sandstone, red, black, green schist	400–800′
	Dunham dolomite	thick-bedded grey, pink sandy dolomite; thick grey sandstone in middle	1700′±
	Cheshire quartzite	massive buff to white vitreous quartzite	1000′±
	Mendon formation	Moosalamoo member: dark quartzite, black fine-banded phyllites	500–800′
		Forestdale member; pink sandy dolomite	0–150′
		Nickwacket member: thick graywacke, quartzite, thin schist and conglomerate	25–800′
		UNCONFORMITY	
Pre-Cambrian	Wilcox formation	schist, dolomite, gneiss	3000′±
		UNCONFORMITY	
	Mount Holly complex	schist, gneiss, quartzite, amphibolite	>7000′

Lower Cambrian Formations

Mendon Formation: Clastic rocks comprising the Mendon formation occur at the base of the Cambrian beneath the *Olenellus*-bearing Cheshire quartzite. While the lower boundary has never been a point of mutual agreement, the formation in nearly its present form was first described by Whittle (1894a) as the Mendon series. He recognized a three-fold division and included units were later named by Keith (1932, p. 394) the Nickwacket graywacke, Forestdale marble and the Moosalamoo phyllite. The Mendon formation and Cheshire quartzite are shown diagrammatically in Figure 2.

The Nickwacket member consists of massive grey-green, medium- to fine-grained, quartz-rich rocks, rather varied in composition in which the principal structures are a weak foliation or slip cleavage and a rare limy or conglomeratic bed. Locally graded bedding is preserved within conglomeratic beds. The rock types of the Nickwacket member include quartz-sericite-microcline-albite-chlorite grit, quartz-sericite-chlorite schist, sericite-quartzite and quartz-sericite-chloritoid-chlorite schist. Biotite, ankerite and magnetite are present in small amounts, with accessory ilmenite, sphene, zircon, leucoxene and tourmaline. Evenly distributed quartz and feldspar pebbles as much as a quarter inch in mean diameter are very common. Conglomerates contain quartzite cobbles whose original diameter must have reached 10 inches. Quartz of pebbles and sand grains is blue and in thin section appears as strongly undulose quartzite fragments cemented by a recrystallized paste of clear equant quartz and mica. Feldspar pebbles are commonly of unaltered microcline broken across the grain. Tiny albite prophyroblasts are a rare constituent. Quartz, sericite, chloritoid and chlorite form a rather unique assemblage of minerals that occur locally in the schist both as lenslike masses and as the matrix for coarse quartz-conglomerate.[1] The dark green chloritoid shows distinctly in thin section as tabular porphyroblasts or poorly formed rosettes[2] (Plate 5).

The Forestdale member of the Mendon formation is a brown-weathering, sandy, fine-grained dolomite. Bedding though rare is shown by thin layers of dolomitic conglomerate or sandy schistose dolomite. The

1 About 1.5 miles north of Mendon on the west slopes of Blue Ridge Mountain; 1.6 miles east of Proctor at the north end of the Pine Hill, and in the short north-south portion of Clarendon Gorge on the Mill River 1.8 miles east-southeast of East Clarendon.

2 Chloritoid (ottrelite) has been described earlier by Whittle (1892); his locality is within the Pre-Cambrian, not as he believed, in the Mendon formation.

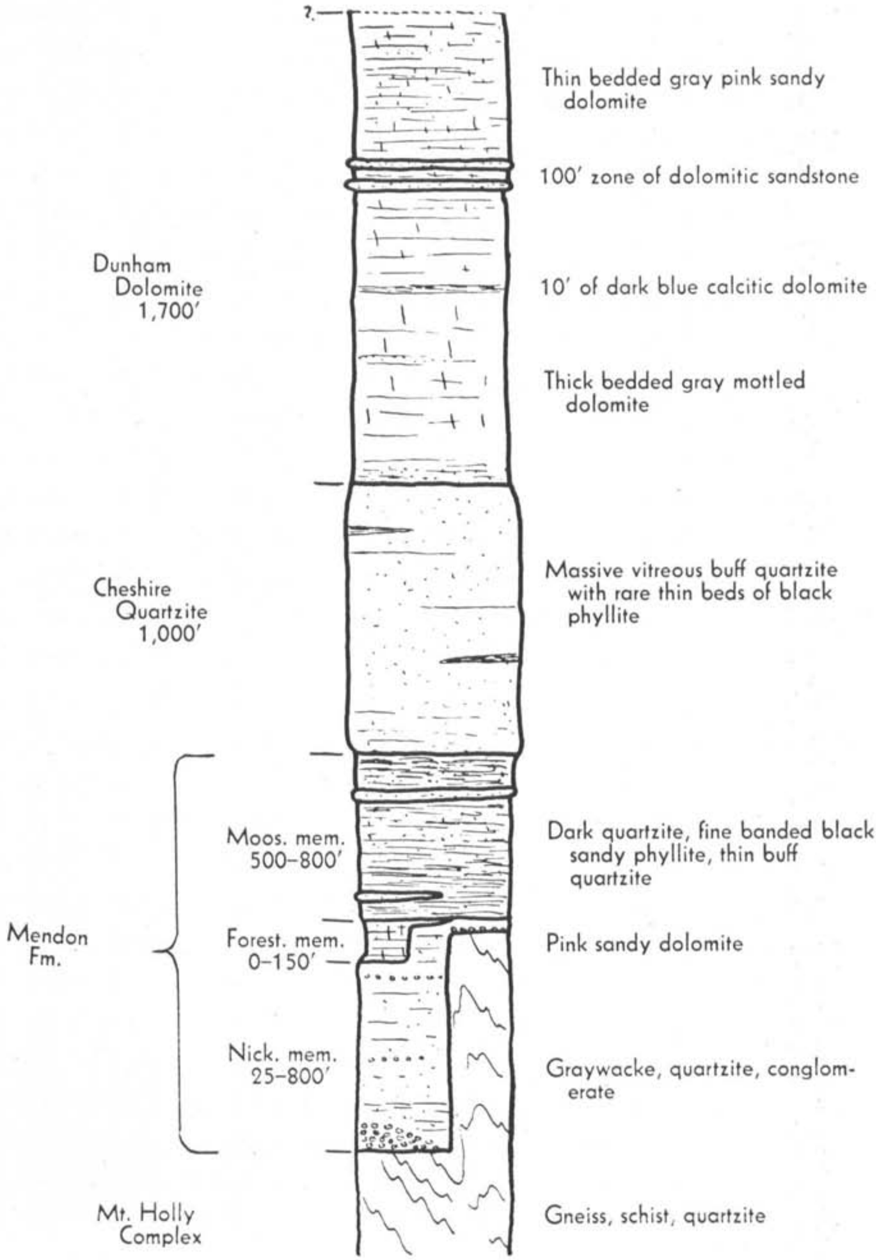

Figure 2. Columnar Section of Mendon Formation, Cheshire Quartzite and Dunham Dolomite.

Chloritoid rosettes in sericite-quartzite conglomerate matrix, North end of Pine Hill, Proctor; crossed Nicols. x 37.

PLATE 5, MENDON FORMATION.

mineral dolomite, in very tiny equant grains, forms 90 per cent of the typical rock. Quartz and microcline pebbles and sericite, calcite, quartz, albite and opaque minerals occur in small amounts.

The Moosalamoo member is composed of sandy phyllites and pebbly grey quartz-sericite-chlorite-albite schists, dark shades being typical. The minerals in these rocks are quartz, sericite, biotite or chlorite, microcline and albite. The dark shades are produced by scattered biotite and traces of graphite. Most of the Moosalamoo is very fine grained and the minerals are concentrated in delicately crinkled bands not over a millimeter thick. Slip cleavage is very common; indeed much of the abundant planar structure seen in outcrop proves in thin section to be slip cleavage.

The Mendon formation crops out along the Green Mountain front and also along the Pine Hill ridge. These belts of exposure join in the area of complex structure near Pittsford and Chittenden and extend north toward the East Middlebury and Lake Dunmore area. The Nickwacket member varies in thickness from 800 feet to less than 25 feet. Fairly uniform maximum thickness is maintained along the Green Mountain front through the Pittsford area but it is very thin and locally may be absent at and to the south of Pine Hill. The Forestdale member has a maximum thickness of about 150 feet in the valleys northeast of Pittsford, but south of Rutland it occurs only in a few places as discontinuous masses a few tens of feet thick. The Moosalamoo member maintains a thickness of 500 to 800 feet throughout.

The upper contact of the Mendon formation, between the Moosalamoo member and the overlying Cheshire quartzite appears at first sight to be sharp, but detail shows it to be more complicated. Rock types common to each unit are often found throughout the pair. Thus, vitreous quartzite beds occur at several places within and as low as the base of the Moosalamoo. In the present study the boundary between the formations is placed beneath the lowest bed of vitreous quartzite more than 25 feet thick.[1]

The age of the Mendon formation is considered to be Lower Cambrian. Correlation of the Mendon seems justified although no fossils were discovered within the Mendon formation of the Rutland area. Fragments of *Olenellus* were found by Walcott (1888) in a quartzite bed 100 feet above an unconformity with the pre-Cambrian near North Adams, Massachusetts. In company with J. B. Thompson and J. Skehan, the writer examined the units overlying the pre-Cambrian in that area and it was generally concluded that the fossiliferous horizon is near the base of a graywacke-quartzite unit resembling the lower part of the Mendon formation in central Vermont. The description of the area by Prindle and Knopf (1932) also places these Lower Cambrian fossils in a stratigraphic position equivalent to that of the lower part of the Nickwacket member. Thus the Nickwacket member of the Mendon formation is probably Lower Cambrian.

No evidence of unconformity between the Mendon formation and the overlying Cheshire quartzite was found in the Rutland area. The

[1] This is probably equivalent to the boundary used by Osberg, separation being made "where the quartzite beds and the graphitic quartz-muscovite schist beds each make up 50 percent of the strata." (Osberg, 1952, p. 30).

writer accepts this as additional support for the Lower Cambrian age of the Mendon formation inasmuch as the Cheshire of the area to the north contains Lower Cambrian fossils. This view, however, is not shared by several who have worked in the region (Whittle, 1896a, Keith, 1932, Fowler, 1950), who believe that Cheshire and Mendon are unconformable. One of the best examples of supposed unconformity has been recently examined by Osberg (1952, p. 38) who considers the surface to be a disconformity relating to a rather short break in deposition. No unconformity has been detected to the south of the Rutland area from Wallingford to Danby (J. B. Thompson oral communication). The areal pattern of the pre-Cambrian basement rocks and its relation to the pattern of the units of the western sequence implies that the unconformity at the base of the Mendon formation is probably the most pronounced in the region. This unconformity overshadows the disconformity between the Mendon and Cheshire referred to above and seems a much more appropriate base for the Cambrian than the first appearance of fossil remains. It is concluded then that the Mendon formation is of Lower Cambrian age.

The Mendon formation extends from north to south in length of the State. Barker (1950) traced the upper and lower members southward to Bennington and Cady (oral communication, 1952), has traced them to northern Vermont and Quebec where Booth (1950) and Clark (1934, 1936) respectively have described units very similar to those in the Rutland area. The Pinnacle graywacke, the White Brook dolomite and the West Sutton slate of these northern areas are equivalent respectively to the Nickwacket, Forestdale and Moosalamoo members of the Mendon formation.

Cheshire quartzite: The name Cheshire was first applied by Emerson (1892) and has since been widely used for this conspicuous thick quartzite. The type locality is in Cheshire, Massachusetts.

The Cheshire is quite distinct as a rock type in the Rutland area and consists of massive buff to white vitreous quartzite. Primary features are rarely seen; bedding planes are ten to twenty feet apart and are seldom as conspicuous as the joint systems which commonly divide the rock. Although the formation is nearly pure quartzite, thin beds of graphitic black phyllite and sandy dolomite are included, the latter near the contact with the overlying Dunham. The quartz grains of the Cheshire range in diameter from a tenth to one millimeter, they are usually equant, and in thin section they commonly show weak undulose extinction. In spite of rather complete recrystallization, granularity is commonly apparent and distinguishes the Cheshire from similar massive quartzites within the pre-Cambrian. Shreds of sericite, grains of altered feldspar and miscellaneous opaque material accompany the predominant quartz in the Cheshire.

The Cheshire is exposed along the Pine Hill ridge, and on Blueberry Hill northeast of Rutland. The thickness where exposed and as inferred along the Green Mountain Front ranges from 800 to 1,200 feet. This compares favorably with the 1,000 feet reported by Cady (1945, p. 526) and Osberg (1952, p. 37) to the north. The quartzite grades into the overlying Dunham dolomite within a dozen or so feet[1] and the upper limit of the Cheshire is therefore clear.

Lower Cambrian fossils have been found in the Cheshire outside the area of this study. *Hyolithes, Nothosoe* and *Olenellus* were found by Walcott (1888, p. 285) near Bennington, Vermont; Wolcott and Seely (1910, p. 307) discovered *Hyolithes* near Lake Dunmore and Hitchcock found *Lingula* near Monkton, Vermont (1861, p. 356).

Dunham Dolomite, Monkton Quartzite and Winooski Dolomite: These formations of Lower Cambrian age have been fully described by Cady (1945, pp. 528–534) and his definitions and terminology are used here. The lowest formation, the Dunham, weathers to dark shades of buff, is generally siliceous and forms thick beds. The calcite content of the Dunham as well as that of the Monkton and Winooski is low thus helping to distinguish these formations from carbonate rocks and sandstones in the Upper Cambrian. The Monkton and Winooski are thinner bedded and the Monkton usually shows contrasting pink, blue and grey shades and contains conspicuous beds of quartzite and dolomitic sandstone.

The Dunham is rarely exposed in central Vermont. Within the Rutland area, however, over several square miles just northwest of the city of Rutland there is excellent opportunity for examination of this unit and its relation to the underlying Cheshire quartzite. The Dunham here includes about 1,700 feet of gently dipping strata with beds ranging in thickness from 6 to 24 inches. The rocks consist of grey, white or buff fine-grained dolomite, throughout which are tiny quartz grains, thin schistose partings, and irregular shaped knots of quartz and coarse-

[1] This contact is best seen near East Creek and the Pittsford road one to two miles north of Rutland on the east slopes of Pine Hill.

grained dolomite. Mottling is very common. Irregular shaped forms an inch or so wide contrast in color with the surrounding rock. Occasionally the bedding surfaces are sharply irregular in detail owing to stylolitic development of the thin iron-stained micaceous partings. The lower fifty feet of the Dunham is conspicuously rich in sand, has 2- to 4-inch beds, often of a pink or buff color. About 700 feet above the base a dark blue-grey limy dolomite occurs with characteristic platy weathering. Near this horizon the fossils are found in the vicinity of Rutland City (Wolff, 1891, Foerste, 1893) (*Salterella, Kutorgina*). Large amounts of quartz sand appear about 400 feet higher in the section as sandstones 50 feet or more thick. These are coarse grey dolomitic sandstones (Plate 6, Figure 1) made largely of blue quartz grains one half to one millimeter in diameter; these sandstones weather to large rounded outcrops and locally show a crude cross lamination. The sandy beds are unique although one- to two-foot beds of dolomitic sandstone are found through the dolomite above and below. Toward the upper part of the section thin bedding and pink color suggest a transition to the varicolored Monkton. At Chippenhook in the southwestern corner of the area upper Dunham also shows rusty coloration and includes beds of greenish schist and dolomitic sandstone 1 to 3 inches thick.

All the rocks considered here are referred to the Dunham dolomite (Cady, 1945, p. 528) because of their thick bedding and position above the Cheshire. It may be possible, however, that some of the upper 600 feet of this section includes portions of the overlying Monkton and Winooski formations. In the area just to the west Fowler had some difficulty in subdividing these formations (1950, p. 49). The 1,700 feet of dolomite suggested here for the Dunham dolomite compares favorably with the 1,700 to 2,000 feet of Cady but is much larger than the 1,000 feet of Osberg and the 800 feet described in the Wallingford-Danby area to the south (Billings et al, 1952, p. 39).

The Monkton quartzite of this area includes 400 to 800 feet of pink, white, blue and buff fine-grained, thin-bedded dolomite and sandstone and thin black and green schist. It is exposed in a single band in the southern part of the valley south of Rutland and is best seen along Route 7 south of the Mill River in Clarendon. The base of the Monkton is marked by a 10- to 20-foot bed of dolomitic sandstone quite similar to that in the upper middle part of the Dunham; in addition thin salmon-weathering quartzites 1 to 3 inches occur throughout the formation, with numerous rusty 1- to 3-foot beds of grey sandstone. Cross-lamina-

Figure 1. Thick-bedded dolomitic sandstone from middle part of Dunham dolomite. Note east-west joint normal to picture. About 1 mile WNW of power station on East River, Rutland.

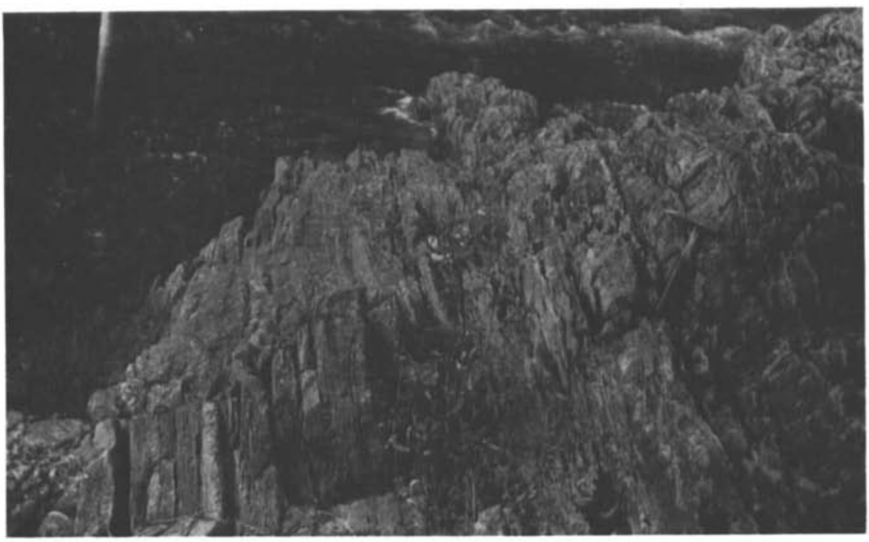

Figure 2. Thin-bedded, pink-grey dolomite with 3- to 5-inch white quartzite beds (lower left) from middle of Winooski dolomite. The trace of east-west jointing is visible in bed beneath hammer handle. Cold River at railroad bridge.

PLATE 6, DUNHAM AND WINOOSKI DOLOMITES.

tion is common within the sandy dolomites but rare in the sandstone and quartzite. Differentiation of the Monkton from Dunham and Winooski is based on the recurrent one- to two-foot beds of sandstone which are separated in the Monkton by less than 25 feet of dolomite and in the Winooski by greater thicknesses of dolomite.

Thin-bedded grey, buff, yellow and blue dolomites occurring between the Monkton sandstones and the Danby quartzites are included here in the Winooski dolomite (Cady, 1945, p. 532). They crop out in a band in the southern part of the area and are best seen in the lower beds of the Mill and Cold rivers and along Route 7 between them (Plate 6, Figure 2). Bedding throughout is 2 to 10 inches thick; the dolomite is buff or yellow below and grey to white above with increasing content of calcium carbonate upward. Cross-lamination in the dolomite and thin sandy yellow schist are fairly common. A thin band of grey chert fragments appears near the top of the formation in the Cold River, several hundred feet above a thin reddish intraformational conglomerate. The thickness of this unit is about 600 feet.

Upper Cambrian Formations

Danby Formation: Glassy quartzites and grey dolomite of the Danby formation (Cady, 1945, p. 535) appear in two discontinuous bands in the vicinity of Bald Mountain and East Clarendon where the formation is probably about 1,000 feet thick. Grey coloration again prevails throughout this lithology; the beds of dolomite are 3 to 20 inches thick and contain noticeable amounts of calcium carbonate. The quartzites of the Danby are most characteristic and occur as 3- to 6-foot beds, white, vitreous and often cross-laminated, separated by a dozen or so feet of dolomite. Thick and thin beds of dolomitic sandstone, so characteristic of the underlying formations, are rarely seen in the Danby. The best exposure of this formation is at the foot of Bald Mountain, 0.4 mile east of the railroad bridge in Cold River and east of Routes 103 and 7 in the fields north of Mill River. Just east of Pittsford in Furnace Brook the Danby appears with overlying dolomite and calcite marble in a narrow syncline

Clarendon Springs Dolomite: About 200 feet of limy dolomite appears between the Danby and the overlying calcite marbles (Keith, 1932, p. 397). It is exposed on the east side of the valley at North and East Clarendon and east of Pittsford both in Furnace Brook and in the pastures to the south. The color of the rock is predominantly dark grey, the bedding thin and the lime content low but conspicuous. The otherwise uniform nature of an outcrop is broken by thin seams of white calcite, tiny quartz knots and rare sandy beds and white limestone. Passage into either overlying or underlying units is gradational. At Bald Mountain, for example, separation of Danby, Clarendon Springs and calcite marble is not sharp; calcite marble appears between the highest quartzite beds of the Danby and sandy facies appear throughout the group.

Ordovician (?) Formations

Marble: A few feet of grey, or white calcite marble occur within the area examined. Both at Pittsford and Clarendon this unit is a crystalline calcite marble in beds 1 to 2 feet thick with dark grey or black partings. At the base of Bald Mountain it contains thin septa of graphitic schist.

Hortonville slate: Black phyllite crops out along the western edge of the Rutland area from Pine Hill southward, and is correlated with the Hortonville slate of Middle Ordovician age (Keith, 1932, p. 369). In the absence of associated dark fossiliferous limestones, it is extremely difficult to distinguish this unit from the Lower Cambrian Moosalamoo member of the Mendon formation. The Hortonville is similarly a black phyllite showing a fine banding, has large amounts of fine sand and numerous folded white quartz stringers. This unfortunate likeness has caused great confusion in the Pine Hill ridge as these two units have apparently been juxtaposed through faulting. At one or two places the limy nature of the Hortonville as opposed to the quartzose characteristics of the Moosalamoo has been utilized in defining the formations. Probably less than 300 feet of Hortonville phyllite occur in the Rutland area.

A large unconformity occurs beneath the mid-Ordovician Hortonville. At Chippenhook it rests directly upon Dunham dolomite and to the south it rests directly upon Pre-Cambrian (J. B. Thompson oral communication).

Genetic Relations

The western sequence resembles a foreland facies (Pettijohn, 1949, p. 454). Dolomite and limestone predominate, accompanied by small amounts of clean sandstone, shale and graywacke-conglomerate. The units are thinly bedded, the composition may vary abruptly from bed to bed and they are commonly cross-laminated. Much of the western sequence consists of calcarenites, and the regularity and purity of the sandstones suggest either intense first cycle cleansing or the reworking of earlier more normal sandstone types.

The lowest unit of the western sequence, the Mendon formation, has features which suggest that conditions during its deposition were somewhat different from those existing during the deposition of the overlying limestones and dolomite. Two rather different controls seem to have competed in the formation of the Nickwacket member, one producing quartzose conglomerate and aluminous schist, the other producing thick uniform masses of graywacke and chloritic schist with graded bedding. The first case implies concentration of material which had undergone rather complete chemical weathering and the second implies floods of detritus from a land mass denuded by severe mechanical weathering and erosion. When their position within the formation is considered these assemblages of rocks seem compatible. The quartzose conglomerate and aluminous schists form lenticular masses near the base of the Nickwacket and grade upward into the massive graywacke and chloritic schist. Though thin conglomerates occur toward the upper parts of the Nickwacket, they commonly reflect a more generous sampling of the source area and, in addition to quartz, show pebbles of gneiss, pegmatite and feldspar. It is inferred that the pre-Cambrian terrain must have undergone an extended period of deep chemical weathering in order to form at the opening of Cambrian time regolithic materials consisting of alkali-poor clays, thin quartz sands and scattered pockets of quartzite boulders. Concentration of this material must soon have been followed, however, by severe changes in the characteristics of the source area; poorly sorted clastic sediments next flooded the basin and covered or mixed with the thin residual deposits. It is interesting that in Quebec Clark (1936, p. 140) described a thin slate beneath the Pinnacle graywacke called the Call Mill slate and considers it in somewhat the same relation to the overlying Pinnacle as that suggested here for the aluminous schists within the Nickwacket.

The Forestdale member of the Mendon formation is believed to be of clastic origin, for the dolomite commonly contains grains of feldspar and quartz of various sizes and at one locality cross-lamination is abundant. The antecedents of the phyllites and dark quartzites of the Moosalamoo member were probably organic shale and sandstone. The common fine banding of the Moosalamoo, however, indicates a more stable environment of deposition than that accorded the members below it; abundant sodic and potassic feldspar and tourmaline suggest a continued period of strong erosion of the gneisses, the felsic intrusive rocks and the pegmatites of the source area.

The relationship of the environment of the Moosalamoo black phyllite to that of the clean quartz sands of the overlying Cheshire remains an intriguing puzzle. As mentioned above, the boundary between these widely different rocks is not sharp in a stratigraphic sense; the appearance of a thin unit of one rock type within the other, though always rather startling, is fairly common. Contrasting environments were obviously close to each other in space and time. Owing to the small grain size of the Moosalamoo, it is unlikely that the Cheshire is its "washed" equivalent. In all probability, the material making up the Moosalamoo was organized in somewhat restricted basins within an otherwise normal shallow water shelf environment capable of producing orthoquartzite.

The location of the source of the sediments is not obvious from study of the Rutland area. Booth, working with less metamorphosed materials believes that the source lies to the west (Booth, 1950, p. 1160). This conclusion is probably supported in the Rutland area by the thinning of the Nickwacket member of the Mendon formation to the west.

Eastern Sequence of Paleozoic and Late Pre-Cambrian Rocks

General Statement

The eastern sequence, flanking the basement rocks on the north and east, consists of metasedimentary rocks of predominantly argillaceous antecedents. They are exposed in the map area in a strip that extends from Plymouth north to Pittsfield. Their maximum thickness is about 15,000 feet. The effects of low-grade metamorphism are fairly uniform throughout and the rocks are now green, white and black phyllites, phyllitic sandstone, quartzite, dolomite, conglomerate and metamorphosed derivatives of mafic volcanic rocks. Minor structural complications abound although contacts are fairly regular, lithologic features monotonous and the units commonly rather thick. Two unconformities within the eastern sequence separate two younger successions from the Mount Holly complex of the Pre-Cambrian. The older succession is the Saltash formation considered to be late Pre-Cambrian; the younger, which includes rocks from the Tyson through the "Bethel" formations, is conformable outside the Rutland area with strata bearing Ordovician fossils (Billings, et al, 1952, p. 18) and is assigned to the Cambro-Ordovician.

TABLE 2

Eastern Sequence

Age	*Formation*	*Lithology*	*Thickness*
Cambro-Ordovician	"Bethel" Formation	green phyllite, sandy finely banded green sandstone	6000′ ?
	Ottauquechee Formation	black phyllite, grey, black quartzite	3500′±
	Pinney Hollow Formation	upper: green phyllite, finely banded green and grey sandstone middle: green, white phyllite, thin greenstone, black phyllite lower: white green phyllite, green albitic phyllite and grit, thin black phyllite	3500–4000′±
	Grahamville Formation	albite grit, dark sandstone, finely banded quartzose black phyllite; Plymouth member: thin white quartzite and grey dolomite	700–1500′
	Tyson Formation	variable green or grey grit, thin conglomerate, albite grit, or buff weathering pink dolomite	0–600′
		UNCONFORMITY	
Pre-C	Saltash Formation	C: vitreous grey quartzite	700′
		B: black graphitic phyllite and thin dolomite, limestone	3–500′
		A: massive white grit, sericite, quartzite, thin conglomerate	7–800′
		UNCONFORMITY	
	Mount Holly Complex	gneiss, schist, quartzite, amphibolite, marble	>7000′

Brace 1953	Osberg (1952)	Chang (1950)	Thompson 1950	Hawkes (1940)		Perry (1928)
"Bethel" Formation	Stowe Formation	Stowe Formation	Stowe Formation		Schist	
Ottauquechee Formation	Ottauquechee Formation	Ottauquechee Formation	Ottauquechee Formation	Ottauquechee Formation		
Pinney	Pinney Hollow Formation					
	Granville Formation	Pinney Hollow Formation	Pinney Hollow Formation	Pinney Hollow Formation	Series	Pinney Hollow Formation
Hollow Formation	Monas-					
Plymouth Member	Battel Member	Plymouth Mem.	Plymouth Mem.		Plymouth Union Ser.	
Grahamville Formation	tery Formation					
Tyson Formation	Tyson Member	Grahamville Formation	Tyson Formation			Older Cambrian rocks
Saltash Form A B C	Mt. Holly Complex	Tyson Fm.	Mt. Holly			Mendon Ser.
Mt. Holly Complex				Pico Ser.		

Table 3 Correlation Chart of Eastern Sequence

The formations mapped in the eastern sequence are listed in Table 2 and presented in Table 3 to show correlation with stratigraphic sections in nearby areas.

Younger Pre-Cambrian (?) Rocks

Saltash Formation: The name Saltash is proposed for grits, phyllite and quartzite of unknown age beneath the Tyson formation on the east flank of the Mount Holly complex. The formation is bounded above and below by pronounced unconformities, the most pronounced being against the underlying Pre-Cambrian. It is unconformable with overlying strata bearing Paleozoic fossils and is therefore provisionally included in the Pre-Cambrian. The Saltash is well exposed in all of the east-flowing brooks of the Ottauquechee and Black River valleys from Saltash Mountain to West Bridgewater.

Although deformation of the Saltash formation has been extreme, lithologic relationships are sufficiently clear so that a three-fold stratigraphic division can be made. The lower grit is termed member A, the middle graphitic phyllite and carbonate, member B, and the upper grey vitreous quartzite, member C. Just west of Woodward Reservoir these units are seen with minimum structural complication.

The basal Saltash member consists of green or white grits and schist, and sericite quartzite. Two types are conspicuously present, one bearing carbonates, white in color and quartz-rich, the other greenish with abundant albite porphyroblasts. Quartz and feldspar pebbles are conspicuous throughout but conglomerate beds are rare and usually encountered several hundred feet within the member. Member B is graphitic black phyllite with thin dolomite and limestone beds, separated from the lower grits by pyritic white sericite schist, quartzite and dolomitic sandstone beds. The phyllite has none of the banding so common elsewhere (as in the Moosalamoo or Grahamville black phyllites), but is a greasy lustrous rock with occasional pyrite euhedrons There is sufficient graphite to soil the hands. The upper part of the Saltash formation, member C, consists of massive grey vitreous quartzites, with a few beds of dolomite and thin-bedded graphitic quartzite separating it from the black phyllite. The quartzite resembles the Cheshire quartzite of the western sequence although its fine texture, grey color and occasional black partings are not characteristic of the Cheshire.

The lowest member has a thickness of 700 to 800 feet. The map pattern is deceiving in this regard for it expresses an unfortunate combination of nearly equal dip of strata and slope of ground which makes the unit appear thicker. The middle member is from 300 to 500 feet thick with at least 700 feet of the upper quartzite above this. Although the unconformity of the Saltash formation against the overlying Tyson formation is distinct, neither strong angular unconformity in outcrop is observed nor significant difference in metamorphism. On Route 100, for example, truncation of the various Saltash members is apparent. The quartzite is in contact with the Tyson near Woodward Reservoir, the black phyllite beneath it in Killington Brook and finally the grits south of Sherburne. Similarly the unconformity beneath the Saltash formation, although demonstrable through mapping, is not always obvious in the field. The relation to the vastly different rocks and structure of the underlying complex is similar to that of the Cambro-Ordovician.

A small body of Saltash occurs near the summit of Sherburne Pass (Plate 1). Within this tiny syncline only the basal grits of Member A have been observed; these are well exposed in the cliffs north of the main road (locally called Deer Leap). The rock consists of grey-green grits bearing albite porphyroblasts and biotite, and encloses highly folded grey vitreous quartzite beds 2 to 10 inches thick, frequent pebbly zones and rare thin dolomitic beds. Although unconformity with the Pre-Cambrian augen gneiss and dolomite is evident on the west side of this structure, the eastern boundary is indistinct due to a strong resemblance of the younger grits and quartzite to underlying rock.

Cambro-Ordovician Formations

Tyson Formation: The Tyson (Thompson, 1950, p. 31) is a thin discontinuous unit of grit, conglomerate, and dolomite appearing between the rocks of the Pre-Cambrian and the albite schists of the Grahamville formation.

In the Rutland area the Tyson formation is either a thin dolomite or a conspicuous graywacke-conglomerate unit. The dolomite is exposed on Route 100 from Plymouth Union to Black Pond, and the graywacke on the slopes east of North Sherburne and south of 2964 mountain[1] in Chittenden.

The Tyson dolomite is massive, fine grained, often brown weathering and contains in addition to dolomite several percent of quartz, sericite

[1] This designation refers to an unnamed mountain; the identifying number is the elevation as given on the Rutland quadrangle map, 1893.

and opaque minerals. In the vicinity of Plymouth Union an unusual variety contains sufficient specular hematite to justify mining during the last century. Bedding is rarely seen in the Tyson except in the basal foot or two of sandstone and conglomerate.

The arenaceous Tyson consists of grey-green quartz-rich albitic grit and schist, chlorite, graphite or sericite schist and conglomerate. Carbonate-rich white quartz-sericite grits, graphitic schist and pebbly chloritoid-bearing types occur within a narrow syncline in the north-central part of the area. These rocks commonly show a weak banding, good foliation and pronounced slip cleavage. Albite porphyroblasts less than a millimeter in size are often conspicuous as well as small quartz and feldspar pebbles. When feldspar is present biotite may be more abundant than chlorite. The Tyson exposed in the narrow central band has a somewhat different mineralogy probably due in part to original compositional differences and in part to differences in metamorphism. Thus ankerite and calcite are always present, chloritoid is fairly common and albite is rare. The chloritoid is found only in feldspar-free, white quartz-sericite schist as small weakly pleochroic tablets closely associated with chlorite.

The conglomerates of the Tyson are of interest although deformation has been so intense that the attention is drawn rather quickly from the few primary features to the host of well-developed secondary ones. Quartzite pebbles and cobbles predominate, with a few gneiss and pegmatite representatives. All have undergone extreme deformation and have the present form of paddles or cigars, 3 to 30 inches long with length 8 to 10 times the thickness. The dark grey-green graywacke matrix varies from 20 to 80 percent of the volume of the rock. It is not uncommon to find a 6-inch quartzite cobble suspended within massive graywacke made up of $\frac{1}{8}$-to $\frac{1}{4}$-inch grains.

The Tyson dolomite is 200 feet or less in thickness and the graywacke-conglomerate unit 400 to 600 feet, the graywacke being thickest to the northwest. In the Rutland area these two lithologies are not in contact; in the southern part of Plymouth, however, the dolomite overlies graywacke and conglomerate (Thompson, 1950, p. 32).

The coarse conglomerate of the Tyson occurs in a bed 20 to 100 feet thick in the northern part of the area near the base of the formation. It has been described as the Sherburne conglomerate (Richardson and Maynard, 1938, p. 89; Hawkes, 1940, p. 54) and assigned various modes of origin, none of which seem justified in view of the present mangled condition of the rock. It is probably near the stratigraphic position of the famous Dry Hill conglomerate of Plymouth (Hitchcock, 1861, p. 386; Thompson, 1950, p. 31).

The Tyson is considered to lie at the base of the Cambro-Ordovician sequence in central Vermont, resting unconformably on either the Pre-Cambrian Saltash formation or the Mount Holly complex. As shown in Table 3, however, this relationship has not always been accepted. The unconformable nature of the Saltash has not been recognized and it has in fact been included completely or in part with the Cambro-Ordovician (Chang, 1950, p. 9; Hawkes, 1940, p 45; Perry, 1928, p. 14). Additional difficulties arose when the dolomite facies of the Tyson was considered the same unit as dolomites found higher in the section, but repeated through faulting. During reconnaissance with the writer in 1950, Thompson clarified this point for the first time by suggesting that the major unconformity did lie beneath the lowest dolomite and above the Saltash. Detailed work which followed has borne out his opinion.

Grahamville Formation: The Grahamville formation (Thompson, 1950, p. 33) includes albite grit, dark sandy phyllite and quartzite-dolomite; these occur between the Tyson formation and the green schist of the Pinney Hollow formation. The upper dolomite-quartzite unit is called the Plymouth member. These rocks are well exposed in the lower slopes of the Ottauqueechee and Black River valleys south of West Bridgewater.

The base of the Grahamville is marked by persistant strata 200 to 400 feet thick, consisting of dark grey-green, massive albite grit. The mineralogy is sericite, albite, quartz, chlorite and biotite with accessory magnetite-ilmenite, tourmaline and epidote. The albite, one half to one millimeter in diameter shows distinct porphyroblastic habit and common evidence of rotation (See "Rotational Features"). Banding, foliation and slip cleavage are usually indistinct. The albite grit is succeeded by a variety of dark fine-banded sandstones, sandy graphitic phyllites and in the southern area buff, ankeritic sandstone. The sandy rocks appear lowest, just above the albite grit and give way upward to thinly banded phyllites. The mineralogy, similar to the lowest unit of the Grahamville is quartz, albite, sericite and biotite with minor chlorite and ankerite, and accessory magnetite-ilmenite, tourmaline and graphite. The dark color is due principally to the few per cent of biotite, as graphite is unimportant in quantity. The most conspicuous characteristics of this middle Grahamville lithology are the fine banding, lack of porphyroblasts and fine texture; the similarity of these rocks to the Moosalamoo member of the Mendon formation is rather striking. Atypical varieties include thin green phyllite and sandy, porphyroblastic red-green schists, and are more commonly found in the northern exposure of the Grahamville.

The Plymouth member includes grey dolomite, and white vitreous thinly bedded quartzite. The thickness may reach 100 feet in the vicinity of Plymouth but is typically 10 to 20 feet; its occurrence as short lenticular masses may be a primary characteristic or may be the result of large-scale boudinaging during later deformation. Though massive the grey dolomite locally appears brecciated; it is quite pure with only a few per cent of tiny quartz and albite grains, sericite shreds, and a trace of calcite and dustlike opaque minerals. Although the associated quartzite is quite similar to the Cheshire of the western sequence, it is less pure, containing several per cent of albite, microcline, sericite and ankerite.

The Grahamville, with slight variation in lithology can be traced the length of the map area, but the thickness changes from about 700 feet in Chittenden to 1500 feet at Plymouth; the middle group of dark sandstone and phyllite is responsible for this variation as the other units have nearly constant thickness.

Although the lower boundary of the Grahamville is distinct beneath the thick albitic grit, some difficulty is encountered in defining an upper boundary in this area. Within as much as 300 feet above the Plymouth member green phyllites of Pinney Hollow appear; above this thin black phyllites continue. Since it is nearly impossible to trace a surface separating black and green lithology an arbitrary boundary was chosen with the dividing surface placed beneath the lowest green bed fifty feet or more in thickness.[1] The top of the Plymouth member might be preferable as a more pronounced surface of discontinuity, but could not be used here due to its discontinuous nature.

The Pinney Hollow Formation: The Pinney Hollow formation (Perry, 1928, p. 24) consists of thick green phyllite and schist which appear between the black schist and quartzite of the Ottauquechee, and the

[1] Referring to Table 3 the Grahamville is equivalent in part to the Monastery of Osberg (1952, p. 42). The writer correlates the Grahamville with the Battel member and the lower portion of the Monastery exclusive of the Tyson member. The Plymouth member in the Rutland area, therefore, would be equivalent to the thin dolomitic lower parts of the Battel member of the Rochester area. The upper surface of the Battel black schist probably corresponds to the top of the Grahamville.

underlying Grahamville. Only the lower half of the formation was mapped south of Sherburne, but good exposure of the entire section is found in the hilly terrain in the northern part of Sherburne and Stockbridge.

The Pinney Hollow is a thick monotonous unit consisting of pale green, grey and white phyllite with minor albitic schist and green fine-banded sandy phyllite. Folded stringers of white vitreous quartz are abundant throughout this and the enclosing formations; spaced two to ten inches they are pod-shaped elongate in the foliation from one half an inch to two inches thick. Discontinuous beds of greenstone, pebbly sandstone and graphitic black phyllite appear in the northern exposure of the Pinney Hollow.

A subdivision of the principal lithology based upon a combination of field and petrographic characteristics is possible. The lowest 1,000 feet of the Pinney Hollow contain albitic sericite-quartz-chlorite schist with minor green phyllite, sandy phyllite and graphitic schist. Albitic grits identical to those in the Grahamville occur in beds several feet thick. The middle 1,500 to 2,000 feet consist of very fine grained green phyllite, thin greenstone and at one locality a white pebbly quartzite.[1] Chloritoid and quarter-inch magnetite octahedrons are abundant; albite is uncommon. The rocks usually show several sets of cleavage and lineations. Near the top of the Pinney Hollow sandy fine-banded green phyllite occurs with occasional beds of black phyllite and sandstone; tiny albite porphyroblasts reappear and biotite is in sufficient quantity to give the rock a darker color than the types below.

Greenstone in the middle of the Pinney Hollow occurs in two bands, one north of Sherburne which was traced for three miles, and the other a mile in length traced southward from the northern quadrangle boundary along the east slopes above the Tweed River. The southern band is yellow-green, very fine grained and consists of epidote, chlorite, calcite, ankerite, quartz, sericite and plagioclase. Banding is weak and in places marked by a thin film of silver-black opaques. The boundaries of the greenstone are sharp, and parallel the foliation of the surrounding phyllites. The northern band contrasts with the southern in pronounced banding of mineral constituents, with a much coarser texture. Minerals include chlorite, sericite, ankerite, epidote, calcite, quartz and plagioclase. It is probable that both bands reflect low grade metamorphism of

[1] 1.8 miles NNW of Sherburne at elevation 2,100 feet.

mafic intrusive or volcanic rocks, with the present differences due to original composition and texture. The origin of the greenstones is not known from study of the Rutland area; lacking contrary observation, the writer will concur with others who have worked with these units elsewhere in Vermont and consider them to be of volcanic origin (Thompson, 1950, p. 38; Hawkes, 1940, p. 75).

The Pinney Hollow[1] maintains a thickness of 3,500 to 4,000 feet. Separation from the overlying Ottauquechee formation is at the top of the highest green phyllite beds. It is an easier boundary to map than the lower one, although the uppermost several hundred feet of the Pinney Hollow contain thin units similar to those found in the Ottauquechee.

Ottauquechee formation: The black phyllites and quartzite of the Ottauquechee formation (Perry, 1928, p. 27) strongly contrast with the green phyllites of the enclosing formations. Outcrop in the map area is limited to a short strip in the northeastern corner where the exposure is poor with extreme deformation of unusual form.

The formation consists of fine-grained black phyllites and black vitreous quartzites with rare thin beds of green sandy phyllite. The common minerals are quartz, sericite and chlorite with lesser biotite, garnet and graphite and typically show a great range of grain size. The Ottauquechee rocks might closely resemble black units elsewhere in the area such as Moosalamoo, Grahamville, and others were it not for the occurrence of 1- to 3-foot glassy black quartzite beds and finely banded *grey* sandstone which are particularly common low in the formation. The color of these quartz-rich lithologies, their purity and regularity are unique, at least at the grade of metamorphism shown in this area.[2]

The Ottauquechee is about 3,500 feet thick. Passing upward on the section the base is marked by the disappearance of green Pinney Hollow phyllite, the upper boundary by the reappearance of these types in the "Bethel" formation. Within the area studied the possibility could not be discarded that the "Bethel," as mapped to the east, actually repre-

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[1] In the northern exposures a conspicuous bed of black phyllite, perhaps 50 feet thick, occurs about 1,000 feet above the base of the formation. This black phyllite may be equivalent to the Granville formation of Osberg (1952, p. 53,) although it is thinner and contains no thin dolomite lenses. The Pinney Hollow of the Rutland area would then be equivalent to the Pinney Hollow, Granville and part of the Monastery formation of the Rochester area as indicated in Table 3.

[2] Excellent examples are found in the stone buildings of the Gifford Woods State Forest Park at Sherburne, Vermont.

sents a structural repetition of the Pinney Hollow; decision will no doubt be possible when adjoining areas are mapped in detail.

"Bethel" formation: The sole appearance of the "Bethel" formation (Richardson, 1924, p. 82) is in an area of intense local deformation, in the northeast corner of the Rutland area. Excellent exposures appear in Stockbridge, in the vicinity of Sable Mountain and in the brooks to the east.

The lithology resembles that of the Pinney Hollow formation, except that the rocks consistently appear less phyllitic and with faint to distinct banding. Thus the principal minerals are again quartz, sericite and chlorite with a few percent of albite or garnet and accessory magnetite-ilmenite, tourmaline, zircon and epidote. Banding is about half a millimeter thick with a great size range of quartz and mica grains (0.01 to 0.4 millimeters).

Question arises as to the proper subdivision of the unit mapped here as the "Bethel." About 6,000 feet of rock is exposed with no horizons detected, whereas Chang (1950, p. 21) and Thompson (1950, p. 40) to the south and Osberg (1952, p. 65) to the north have placed the 1,000- to 2,000-foot Stowe formation above the Ottauquechee. The Stowe is described as very similar to Pinney Hollow and separable from the overlying Moretown formation which contains thinly banded, so-called "pinstripe" quartzite.

The next zone reported higher in the section is the dark graphite or biotite-bearing schist considered to be the base of the Missisquoi formation (Richardson, 1924, p. 101) or the base of the Whetstone Hill member of the Moretown formation (Thompson, 1950, p. 41). In the Rutland area neither the abrupt appearance of "pinstripe" quartzite nor black schist was detected.

The writer has therefore followed Chang's procedure in the northern part of the Woodstock area (1950, p. 23) and applies the ill-defined term "Bethel" to the entire unit, realizing that it may include the Stowe and portions of the Moretown formations in an unknown structural relationship.

Genetic Relations

The metasedimentary rocks of the eastern sequence exclusive of the Saltash formation are probably eugeosynclinal or can be considered "geosynclinal facies" (Pettijohn, 1949, p. 444). Thick-bedded chloritic and graphitic shale, shaly impure sandstones, graywacke and inter-

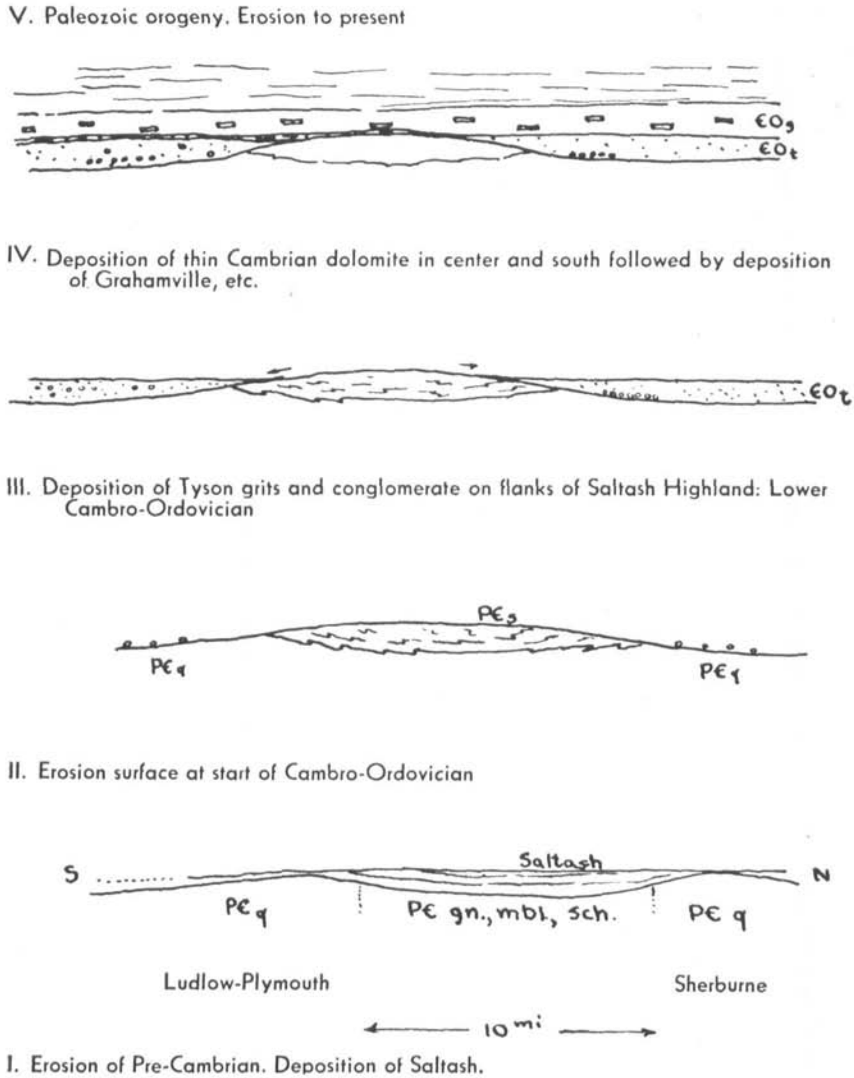

Figure 3. Diagram Illustrating Distribution of Tyson and Saltash Formations between Sherburne and Ludlow, Vermont.

bedded volcanic rocks are their sedimentary antecedents with subordinate carbonate and clean sandstone units. Rare primary features include graded bedding. Certain features, however, are atypical of the purely geosynclinal facies (Idem p. 443). Conglomerate cobbles are mostly quartzite. Beds of pure quartzite and dolomite appear in the lower part of the section and a considerable portion of the shale representatives appear to have rather a high aluminum-alkali ratio. Although possibly present in the guise of a phyllite or schist, graywacke is not conspicuous in the sequence, which is dominated by black and green argillaceous types.

The Saltash formation is quite similar to the Mendon formation. The sequence grits-black phyllite-quartzite is alike in the two cases and the general environment for deposition appears to have been similar. There are differences in lithology, of course, such as the presence of carbonate through much of the Saltash and the finer texture and shaly impurities of the upper Saltash quartzite, but these might be due to deposition of the Saltash in deeper water. Few of the eugeosynclinal features of the overlying Cambro-Ordovician rocks are found in the Saltash.

The present arrangement of Tyson and Saltash formations may have been controlled by the nature of the basement complex. Various stages in the hypothetical development of the present relations are sketched (Figure 3). In the North Sherburne and the Plymouth-Ludlow areas (Thompson, 1950, p. 19) abundant Pre-Cambrian quartzite appears in the basement complex. The coarse quartzite conglomerate is similarly distributed in the overlying Tyson. Between these two areas the Tyson is a dolomite and the Saltash appears beneath the Tyson. Assuming that the Saltash was metamorphosed before the deposition of the Tyson and has resisted erosion (as indeed the Nickwacket does today) explanation of many of these curious features as possible. Stage I of Figure 3 indicates the probable shape of the basin in which Saltash materials were deposited. The resistant Pre-Cambrian quartzite underlay elevated tracts in contrast to the less resistant gneiss, schist and marbles. Sands and clays filled the depressions and possibly covered the Pre-Cambrian monadnocks. Stage II shows position of an erosion surface following the deformation and metamorphism of the Saltash formation, the higher elevations now being underlain by the younger massive grits and quartzite. First deposition of the Tyson, Stage III, would naturally have occurred at lower elevations marginal to the Saltash. Conglomerate formation was localized near these areas and consisted of a mixture of the coarse debris overlying the quartzite, and transported matter. Deposition of the Tyson grits continued until the floor of the basin was nearly level, but without completely covering the Saltash surface. Deposition of carbonate followed that of the grits, finally covering the Saltash surface and extended a short distance southward to rest on the grits. Dark shale and sandstone were then deposited in thick regular units above the dolomite. Thus it is possible to explain the restricted occurrence of the Saltash and Tyson dolomite between the areas of conglomerate deposition. The coarse cobbles and boulders of the Tyson conglomerates would be of local derivation; their matrix transported from an active source area perhaps at some distance.

Some indication of the variation in the Tyson is shown by comparison of sections exposed in the Ottauquechee and Black River valleys with those found in the narrow syncline in the central part of the area. On the south and west slopes of 2964 hill in Chittenden, black schist first appears within micaceous rock which has lost much of the coarse gritty features common at Ludlow. Within the syncline further west black schist is even more abundant as a thin bed near the base of the formation and again near the top. The Tyson here is quartz-rich schist with some chloritoid. Graywacke-like material then diminishes to the west, being replaced by shale and clean micaceous sandstone. This relationship suggests an eastern source of material for the Tyson, and possibly for the overlying shales and volcanic rocks as well.

Correlation of Eastern and Western Sequences

The Pre-Cambrian basement complex is flanked on each side by Cambro-Ordovician rocks and their correlation is of prime importance. While studies in the area mapped have provided no unequivocal solution to the problem, certain features present possibilities of correlation. These possibilities can doubtless be fully evaluated when further studies have been made in the northern and southern extremities of the Green Mountains and in the Taconic Mountains to the west.

Two currently favored schemes of correlation are shown diagrammatically in Figure 4 (see also Billings, et al, 1952, p. 18). The older idea is that the western dolomite and quartzite, and eastern shale and volcanic rocks were contemporaneous, the former approaching a platform facies, the latter a geosynclinal accumulation (Sketch I). According to this scheme the Dunham dolomite is, for example, considered equivalent to the Pinney Hollow phyllite and volcanics. Large tracts of slate in the

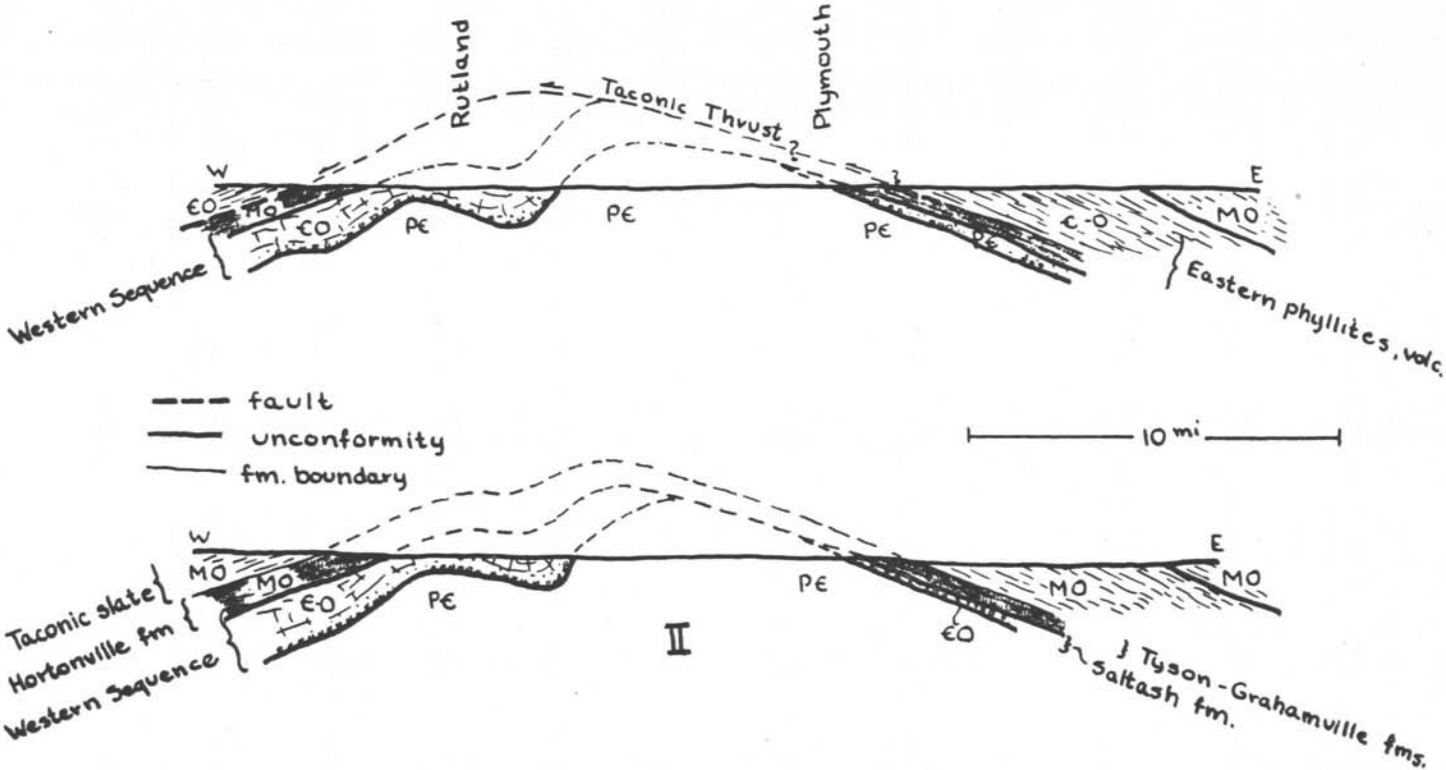

Figure 4. Correlation of Eastern and Western Sequences
(Sketched from Billings, et al, 1952)

Taconic Mountains (Fowler, 1950, p. 64) are believed to be members of the eastern section thrust far to the west to rest as a klippe upon contemporaneous platform facies. The approximate distance separating these facies during their deposition should be considered. In the northern part of the Rutland area they are now less than ten miles apart and, although the original distance was indeed greater, studies of the basement complex indicate that the original distance may not have been more than twice the present distance.

The other scheme (Thompson, in Billings, et al, 1952, p. 18) suggests that the eastern shale and volcanic rocks do not extend in age back to Lower Cambrian but are represented in part by the Hortonville slate and certain Taconic slates to the west both of Middle Ordovician age (Sketch II). A regional unconformity lies beneath the Hortonville and in this correlation it is made equivalent to the unconformity separating Tyson from the underlying Saltash formation and Mount Holly complex. The Tyson and Grahamville, being then roughly equivalent to the fossiliferous Hortonville, would be of Middle Ordovician age. A further implication is that most of the eastern Taconic shale is autochthonous, not equivalent in age to underlying carbonate rocks, but resting unconformably upon them. Several observations made in the Rutland area support this correlation. Tyson-Grahamville rock types appear to change westward to types above the major mid-Ordovician unconformity west of Rutland; Tyson rocks become more schistose and contain conspicuous black phyllite and dolomite. To the east of the Pre-Cambrian, Lower Cambrian units of the western sequence may be represented in part by the Saltash formation which appears briefly between Tyson and Mount Holly. The striking lithologic similarity of the Saltash to the Mendon and Cheshire formations was marked by Perry (1928, p. 14) and has been described above. Eastern representatives of carbonate formations above the Cheshire would, according to this hypothesis, have been eroded in pre-Hortonville time; at least they do not appear in the map area.

Although the correlation of the lower part of the eastern sequence with mid-Ordovician rocks removes the rapid facies change otherwise necessary, another question is raised, for Midde Ordovician fossils have also been found about 20,000 feet above the base of the Tyson (Currier and Jahns, 1941). Thus the deposition of 20,000 feet of sediments and volcanic rocks is apparently compressed into a rather short interval of time. The variety of lithologic features further requires many rapid changes in the environment. In this regard Smith (1953) has shown that in northern Venezuela complex igneous intrusion, two episodes of metamorphism and the deposition of 10,000 feet of sediments must all have occurred during a small part of the Cretaceous. Perhaps our demands on Middle Ordovician time are not unreasonable.

Unstratified Rocks

Unstratified rocks: A few thin mafic and felsic dikes and sills cut the rocks of the Rutland area. These intrusives are older than the major deformation and metamorphism which have affected the Cambro-Ordovician strata, but show a slight mineralogic readjustment. The most abundant intrusive is lamprophyre; the labradoritic plagioclase and brown hornblende have altered to clinozoisite, sericite, carbonates and chlorite. Elsewhere serpentine and chlorite have formed from pyroxene. Bostonite dikes[1] are silicified, and their feldspar is wholly altered to sericite.

Small intrusive bodies that are related to the Cuttingsville syenite stock south of the Rutland area, occur a few miles west of Shrewsbury. No additional study of this group of rocks was made except to verify certain details of Eggleston's work (Eggleston, 1918). A conspicuous variety of the rock is a dark breccia exposed along the railroad tracks at the quadrangle boundary, and 2.3 miles WNW of Shrewsbury. Trachytic and andesitic porphyry contain small rotated angular fragments of gneiss, amphibolite and quartzite, the host of rocks of the vicinity; at the southern exposure the porphyry is peripheral to medium-grained white syenite and diorite. These intrusive rocks are correlated with the White Mountain magma series of Mississippian (?) age (Billings, 1934).

METAMORPHISM

General Statement

Discussion of the thermal history of the rocks of the Rutland area naturally follows a two-fold division. Broad regional metamorphism has affected the Cambro-Ordovician rocks, whereas the basement complex is characterized by a two-fold metamorphism; analysis of the mineralogy and interpretation of the metamorphism are therefore considered sep-

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1 Clarendon Gorge of the Mill River, 0.7 miles ESE of East Clarendon.

arately for each group. The younger metamorphism of the basement rocks was probably produced by the same agencies which produced a regional metamorphism of the Cambro-Ordovician strata.

Thorough analysis of metamorphism in the map area presents difficulties at this time. The rocks are predominantly argillaceous, have minerals typical of low rank metamorphism and show fine texture throughout. It is therefore particularly difficult to determine the distribution and variation of chemical elements throughout the area. Key elements like sodium, potassium, aluminum, iron and magnesium are bound in the ubiquitous micaceous minerals "sericite" and chlorite, the crystal-chemical nature of which precludes an accurate determination of chemical composition through the customary optical examination. Chemical analyses of chlorite, sericite and chloritoid are needed, but are ruled out for the present by the great difficulty of separating the minerals from the fine-grained grits and phyllite.

Metamorphism of the Cambro-Ordovician Rocks

The minerals found in the Cambro-Ordovician rocks are quartz, microcline, albite, sericite, chlorite, biotite, chloritoid, carbonates (calcite and ankerite-dolomite), garnet, epidote, zoisite and magnetite. Accessories are tourmaline, ilmenite, sphene, zircon, pyrite, graphite and apatite. Compositionally the rocks fall into three groups: argillaceous rocks with various ratios of alkali-alumina, impure dolomites and mafic volcanics. All but the volcanics have a wide range of distribution through the mapped area. Quartzites and graphitic schist, while compositionally unique, have behaved as one of the first two groups above.

The mineral assemblages of these rocks are shown in Figure 5 using triangular diagrams. Rocks which fall upon a join or within a field on the diagrams may contain two or three minerals, respectively. Additional minerals containing iron and magnesium may be present in a rock, since we are dealing in effect with at least a four or five component system. Ferrous iron and magnesium cannot be considered as substituting completely for one another in such minerals as chlorite, chloritoid and garnet. The AKF diagram illustrates the mutual exclusiveness of microcline and chloritoid, and microcline and chlorite. This is based on consistent observation.

A single isograd has been detected in the Rutland area, located at the first appearance of almandine garnet, roughly east of a line joining East Mountain, Sherburne and Sable Mountain, Stockbridge. West of

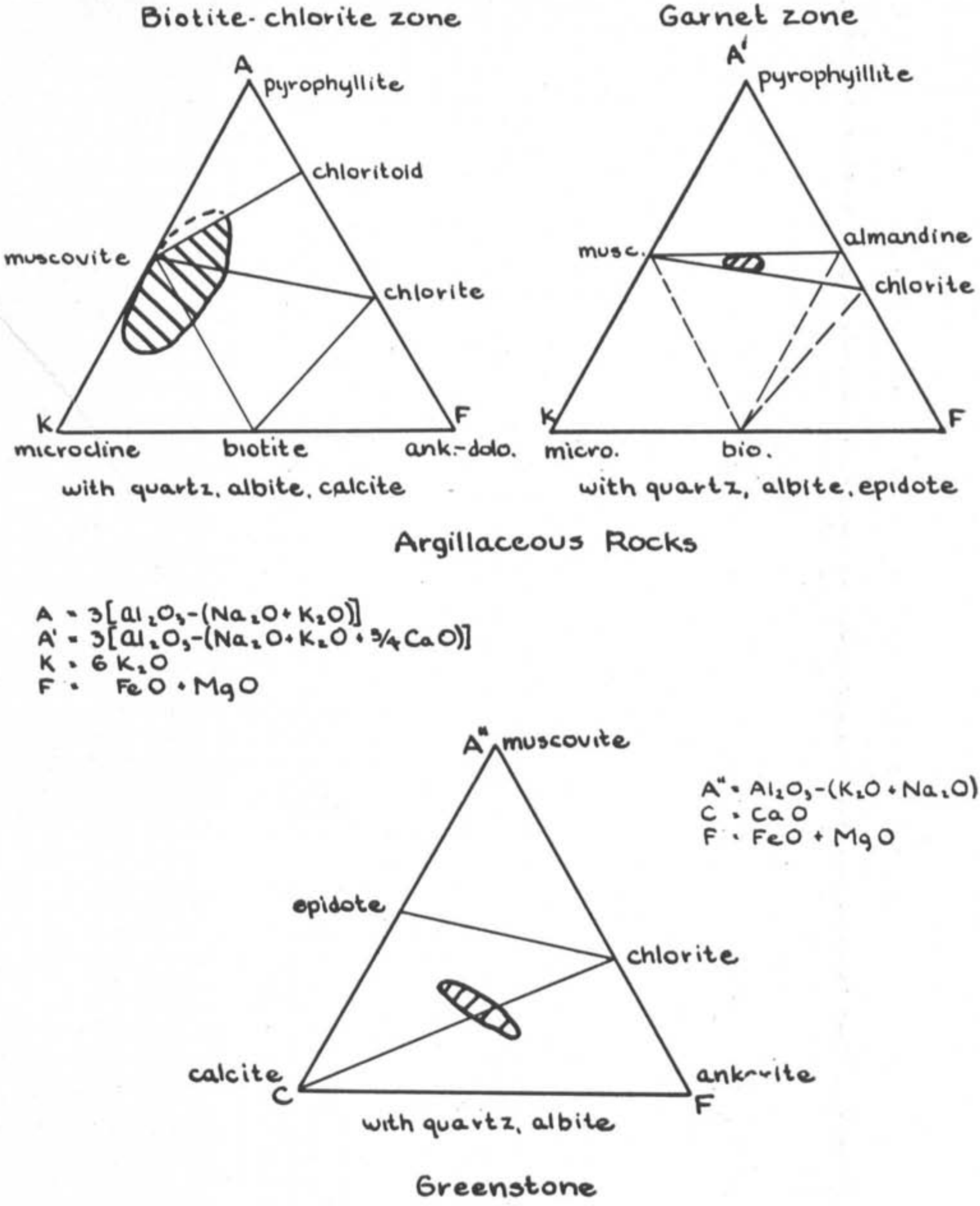

Figure 5. Mineral Assemblages of the Cambro-Ordovician Rocks illustrated in Triangular Diagrams (after Thompson, unpublished lecture notes).

this line garnet is absent in assemblages typical of the greenschist facies of Eskola (as described by Turner, 1948, p. 93–98). Thus, sericite-chlorite, chlorite-biotite, and chloritoid-sericite-chlorite schist, and quartzose dolomite occur with chlorite-epidote-calcite greenstone. The higher grade rocks east of this line contain sericite-chlorite-garnet. The occurrence of biotite does not appear to indicate a higher grade of metamorphism than that of the chloritic rocks. Biotite and chlorite form together in rocks with a relatively high potash-alumina ratio; biotite alone forms when the rock has even more potash and contains microcline. Throughout the area biotite and biotite-free rocks appear closely related in outcrop, with biotite in the microcline-bearing varieties. Discounting the effects of local deformation, texture is fairly uniform in all of the Cambro-Ordovician rocks and, except for the appearance of garnet, there is little to indicate rising grade of metamorphism to the east. Faint suggestion is given by the increase in abundance of accessory epidote over carbonates in the argillaceous varieties. To the east of the Rutland area (Chang, 1950; Thompson, 1950) garnet is abundant in argillaceous rocks, and mafic volcanics contain actinolite and epidote.

Certain features of the fabric and composition of the Cambro-Ordovician rocks indicate a fairly complete attainment of equilibrium during metamorphism. Again discounting the effect of local deformation, the range of grain size of a mineral species in a rock is quite small, zoning of minerals is absent and there are no gross violations of the phase rule i.e., never two varieties of a mineral species present together, such as two chlorites or two plagioclases. On the other hand, rare bent mica and possible twinned plagioclase indicate that the processes of annealing were not universally completed.

Metamorphism of the Basement Rocks

The Mount Holly complex has been affected by at least two episodes of metamorphism, an earlier period of high-temperature and a later low-temperature metamorphism similar in effect to that which left its mark in the Cambro-Ordovician rocks. Two problems exist in the analysis of these metamorphic rocks: (1) determination of the initial assemblages, and (2) interpretation of the effect and completeness of the younger metamorphism. The effects of the later episode of metamorphism are similar to those of retrograde metamorphism[1] but the term is probably not applicable here as the two periods of metamorphism are separated by interval of erosion and deposition.

Earlier Metamorphism

The Mount Holly complex may have initially attained a medium to high grade of metamorphism. Almandine garnet and tremolite are abundant throughout the rocks and garnet-plagioclase amphibolites are common. The amphibolites are coarse, banded units containing actinolitic hornblende[2], albite with abundant zoisite and epidote inclusions and large almandine garnets. Assuming an original intermediate plagioclase, this assemblage is characteristic of the amphibolite facies (Turner, 1948, p. 86). The amphibolites in the southern part of the map area are associated with biotite-microcline and garnet-plagioclase gneiss, quartzite, and tremolitic marbles. Whereas the highest grade index mineral of the impure dolomites is tremolite, Thompson reports diopside from areas to the south (1950, p. 21). Chloritoid is the only aluminous silicate mineral found in these rocks. In all probability the Mount Holly complex has been elevated at least to the garnet zone of regional metamorphism.

The Grenville of New York is similarly composed of gneiss, marble, quartzite and amphibolite, but now contains sillimanite, diopside, wollastonite and augite in rocks of suitable composition. The present differences between Grenville and Mount Holly are, of course, profound; it is problematical whether this was always the case. A higher grade of metamorphism might be expected in Grenville from the presence of numerous plutons, which are either absent or unrecognizable in the core of the Green Mountains.

The Wilcox and Saltash formations have mineralogic features similar to the Cambro-Ordovician rocks. The Saltash gives no indication of an earlier metamorphism of higher grade than that now recorded. The Wilcox, except for a few thin beds of coarse quartzose gneiss and dolomitic marble, differs from the Paleozoic rocks only in its slightly coarser texture.

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[1] Becke, 1909; Kieslinger, 1928; Knopf, 1931, Harker, 1932, p. 344.

[2] A plagioclase-garnet amphibolite from Robinson Hill, Shrewsbury has 60 percent amphibole with the following properties: (-) $2V=72°$, $Z \wedge C=21°$, $n_x=1.663$, $n_z=1.683$. Absorption: $Z \gg Y > X$, Z: dark blue-green, Y: greenish brown, X: yellowish green. Amphibole is rimmed and partly replaced by chlorite (Plate 4, Figure 2).

Later Metamorphism

The Mount Holly rocks have been described previously in this paper; it remains now to summarize those features of mineralogy and fabric which bear upon their younger metamorphism. In the field, the "diseased look" of the rocks is characteristic; in thin section this dirty red and green chalky coloration can be traced to the products of alteration of earlier minerals. These early minerals indicate at least a garnet zone of metamorphism; their destruction has produced a mineralogy characteristic of the chlorite-biotite zone. Almandine garnet is rimmed and cut by veinlets of biotite and chlorite containing a dust of black opaques. Actinolitic hornblende is rimmed and partially replaced by chlorite. An early, more calcic plagioclase is suggested by albite grains filled with inclusions of zoisite, epidote and calcite. Tremolite is intergrown with talc. Chloritoid is not traceable to an earlier mineral; in fact in the Blue Ridge locality, chloritoid has the form of rosettes and has probably developed independently of staurolite or other aluminous silicate forebears. The rocks to this point contain physical relics of an earlier metamorphism of higher grade than the latest. It is tempting to consider further the possibility of rocks in the basement complex in which minerals have been completely altered. Such may be the grits and greenstones found near the unconformity with Cambro-Ordovician, although mapping does not as yet provide rigorous proof of this origin. In the vicinity of Saltash and Smith Peak plagioclase amphibolites are abundant in the Mount Holly; these rocks may be represented by chlorite-epidote-albite-calcite greenstone found at several places near the unconformity.

The mineralogic changes listed above lead in nearly all cases to the formation of hydrous minerals of the greenschist facies of metamorphism. The rare formation of later carbonate minerals is noteworthy. It is of interest to consider the relative persistance of early minerals judging from their appearance as relics. These can be listed roughly in order of decreasing persistance: biotite, hornblende, tremolite, garnet and intermediate plagioclase. While it is difficult with present knowledge to explain this order, numerous possibilities come to mind: differing mechanical strengths of minerals, differing thermal barriers in the change of minerals to their lower temperature offspring, and the possibility that certain minerals simply have a wide range of stability (biotite-microcline gneiss is throughout the least changed of the "ancient" rocks as its range of stability extends from the biotite-chlorite zone upward to the point at which biotite breaks down).

While the compositional banding and folds of the Mount Holly remain relatively intact in the center of the complex, a host of microscopic features attest to crushing and granulation. Garnet and microcline are shattered with trains of fragments strewn in the foliation, mica is often bent and a single mineral commonly appears in a great range of grain size. Unlike the quartz in the Cambro-Ordovician rocks, quartz in the Mount Holly has extremely variable extinction, shows crushed and sutured grain boundaries and is fractured[1] and veined with micaceous paste. Lime-silicate rocks are reduced to a pulp of tremolite and sericite fibers. There is, however, a sensible limit to this grain size reduction, for rocks are seldom reduced to phyllitic texture. Lacking first hand knowledge of Alpine counterparts, the writer hesitates to apply the terms here, "phyllonite," "augen schist" or "mylonitic gneiss" (Knopf, 1932, p. 10). Lenticules and augen of broken minerals are common and crystallographic orientation of quartz in these forms is conspicuous; however, it remains to be demonstrated that the present compositional banding, at least in the center of the complex, is of mylonitic origin. The writer considers this to be primary banding as it is often traceable for many miles—as for example, a thin bed of marble. However, the parallelism of structural elements near the unconformity may indeed be due to ultra-mylonitization or "tectonic unmixing" (Kieslinger, 1928) which has developed a secondary banding parallel to the unconformity and the planes of movement in the overlying younger rocks.

In summary, the rocks of the Pre-Cambrian basement complex show the effects of two metamorphisms. The earlier metamorphism which elevated the basement rocks at least to the garnet zone of regional metamorphism is separated from the younger episode by a period of erosion followed by deposition of the Cambro-Ordovician rocks. The younger metamorphism of the basement rocks probably occurred during the prograde metamorphism of the Cambro-Ordovician units. There is throughout a definite trend in the basement rocks favoring assemblages which are mineralogically identical and, therefore, in equilibrium with rocks in the Cambro-Ordovician. Mineralogic changes in the basement are probably due to the application of a temperature lower than that which formed the older and higher grade assemblages. Volatile con-

[1] Fractures have irregular orientation generally, but in a few cases are parallel to the undulatory banding, and hence the c-axis of quartz.

stituents, largely water, necessary to produce hydrous minerals and promote recrystallization were of course essential. There is evidence that water was fairly abundant: inasmuch as mafic volcanic rocks and garnet-plagioclase rocks have been almost completely altered to hydrous minerals and in so doing have not made serious inroads into the supply of volatiles present, at least no dessication is apparent in the immediately surrounding rocks.

Miscellaneous

Metasomatism: Metasomatism does not appear to have been important in the metamorphism of Cambro-Ordovician rocks; its role in the formation of the basement gneiss is not known. Albite and tourmaline, often considered metasomatic, are locally abundant in the Tyson-Grahamville and Moosalamoo. A rough calculation of the sodium or boron content of these rocks, however, gives values which do not exceed the amounts in comparable sediments (Pettijohn, 1949, p. 271, 285).

Thin conformable lenses of quartz and quartz-feldspar-carbonates are abundant in the Cambro-Ordovician and are generally related mineralogically to the immediately adjacent rocks. Thus, quartz pods occur in quartz-rich schist and phyllite, quartz-microcline-albite lenses in grits and feldspathic schist, and dolomite-quartz veinlets in siliceous dolomite. These lenses are probably due to a leaching of mobile material during metamorphism, with deposition in boudinage openings or within the rock foliation. This phenomenon probably illustrates the solution principle of metamorphic differentiation (as discussed by Turner, 1948, p. 144) and implies a very local metasomatism, or migration of material. Greenstones are conspicuously banded and the unusual composition of some bands (chlorite-epidote and dolomite-ankerite) may indicate migration of material other than volatiles for a distance of a few inches.

Volatiles such as water and carbon dioxide have plainly been involved in reactions throughout the area. The free movement of volatiles may be regarded as the most important phase of metasomatism in the area.

STRUCTURAL GEOLOGY

General Statement

The Rutland area is part of a complex fold belt. As a consequence of several episodes of metamorphism and deformation, a profusion of structural elements have been produced in the rocks, from the scale of the Green Mountain anticlinorium to that of microscopic grain orientation. Final analysis of the regional structures must await the completion of detailed mapping in the fold belt. The wealth of micro-structural data is as yet untapped and in this study features somewhere between these extremes are analyzed. Major structures will first be considered, then structures too small to be shown on the geologic map (Plate 1), and finally the relationship of deformation to metamorphism. Reference is made throughout to the geologic map, cross sections (Plate 2) and the tectonic map (Plate 3).

Several problems of general importance are clearly outlined in the structural geology of this central Vermont area:

(a) the analysis of the interaction of basement and mantle during deformation, in order to formulate the manner in which basement structure controlled the development of younger trends,

(b) the correlation of minor structures with major structure to determine the overall movement pattern of the Paleozoic deformation, and

(c) the search for a root zone for the Taconic klippe, a large detached mass that has been postulated to lie to the west.

These problems will be reviewed following description and analysis of major and minor structures of the Rutland area.

Major Structural Features

Structurally the Rutland area consists of the Green Mountain anticlinorium, Rutland syncline, Pine Hill anticline and the Pittsford-Chittenden structural complex (Figure 6). Regionally, the central anticlinorium, consisting of a Pre-Cambrian basement complex and flanking Cambro-Ordovician metasediments is traceable over the central and southern parts of the State (Billings, et al, 1952). The Pine Hill and Rutland folds are simply minor complications on the west limb of the Green Mountain anticlinorium. The Pine Hill anticline extends about 15 miles, from the vicinity of Proctor south to Danby. The Pittsford-Chittenden structural complex is a local feature; structures to the north and south trend fairly uniformly north. The rocks on the east flank of the Pre-Cambrian basement trend north and are the lowest units of a homocline extending eastward to the Connecticut River.

The Green Mountain Anticlinorium

The anticlinal nature of the Green Mountains was recognized by Adams (1846, p. 167) and has since been described by Keith (1932, p. 404), Cady (1945, p. 564), Thompson (1950, p. 86) and Osberg (1952,

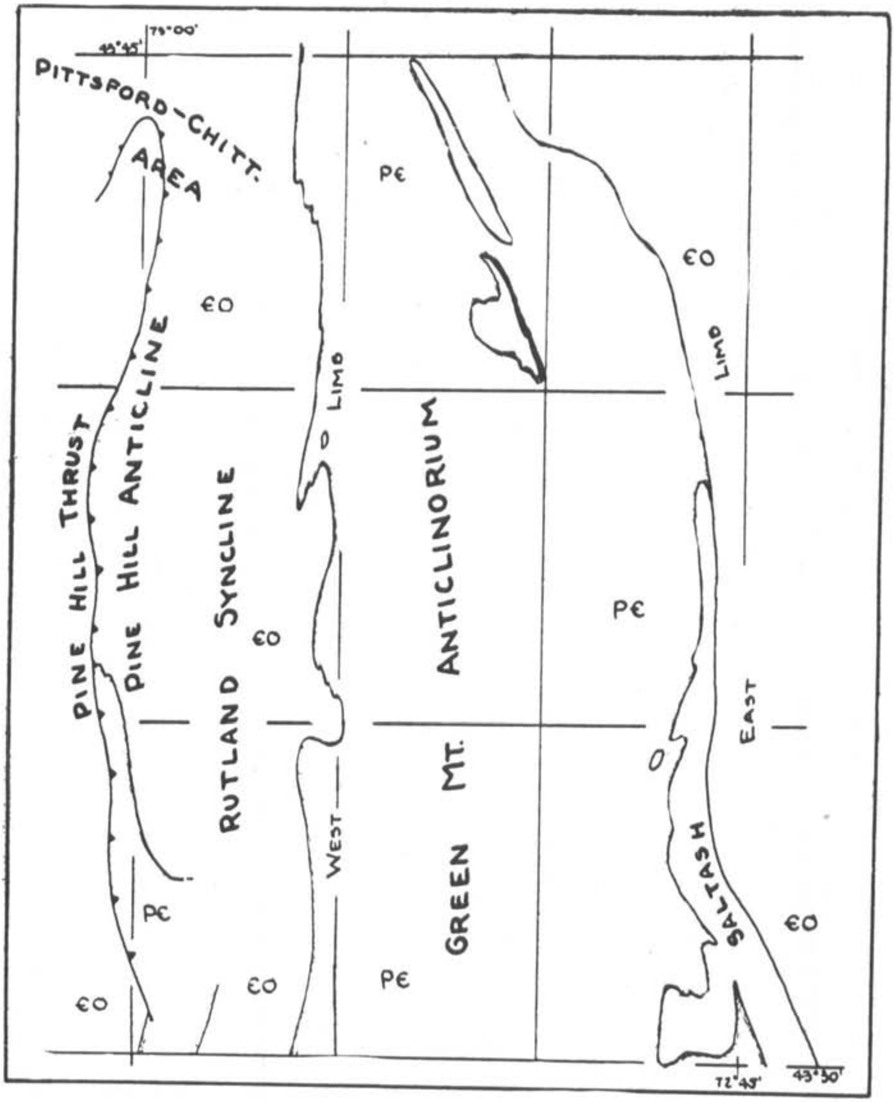

Figure 6. Major Structural Units (sketched from Plate 1, not to scale).

p. 75). Structural and metamorphic considerations offer the strongest proof of an anticlinal structure: a central metamorphic complex is bounded on either side by younger metamorphic rocks and lithologic units in the complex are sharply truncated at this boundary. Minor folds on the flanks of the anticlinorium have a consistent pattern which is typical of anticlinal structure (Billings, 1942, p. 77). The Green Mountain anticlinorium at the latitude of Rutland has a gentle northerly plunge. This is the consistent attitude of minor folds of the east limb, but not of the west limb where fold plunge varies rapidly along the strike. Nearly all of the folds, large and small, in the Rutland area are overturned to the west; in the central anticlinorium, this is expressed in a steep to overturned west limb, and moderate to steeply dipping east limb.[1]

Structure of the Pre-Cambrian Basement Rocks. Poor exposure, the unknown origin of much of the basement lithology and the complex metamorphic effects do not lend assurance to the interpretations which follow. This discussion of basement structures and their role in guiding the structural development of the younger rocks is of a preliminary nature.

In the center of the basement complex remote from the zone of interaction of basement and flanking rocks, the rocks strike northwest and dip gently northeast. No criteria have been found to indicate the relative ages of the 10,000 feet or more of rocks and it is doubtful whether superposition is reliable. From Shrewsbury southward the trends in the basement are about normal to those in the north and it is impossible to assess this either as a local feature developed during the folding of the younger Wilcox, or as a fundamental structural trend in the Mount Holly. In the central part of the basement complex younger structural elements (slip cleavage, folds, and other features) parallel to those in the Cambro-Ordovician rocks are rare and appear only in schist beds near the summits of Pico and Little Killington peaks.

Structurally the Grenville of New York State (Miller, 1916, p. 588–597) is remarkably similar to the basement complex of the Green Mountains. Although locally complicated by intrusion banding in the Grenville has an easterly strike and gentle southerly dip with no ap-

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[1] In general discussion the terms "gentle," "moderate" and "steep" are used for values of dip of strata and plunge of linear elements of 0 to 30, 30 to 60 and 60 to 90 degrees, respectively. Similarly, values of strike are generalized as NNE, NW, etc. in the northern quadrant of the compass.

parent major folding and faulting. As in the Mount Holly, rock foliation parallels compositional banding. Miller has described the development of Grenville structure by load metamorphism (*op. cit.* p. 597) accompanied by moderate doming and warping of the strata by intrusion and has decided that "lateral compression" has played no part in the metamorphism of the rocks. The efficacy of load metamorphism has been severely challenged (Fairbairn, 1949, p. 165; Turner, 1948a, p. 297) and seems unlikely as a mechanism for the development of the basement structures of the Mount Holly. Folds, rodding, streaming, and boudinage are abundant and attest to the existence of shearing movements but unfortunately it is not yet possible to synthesize this data.

Interaction of basement and Cambro-Ordovician can be analysed in several ways: (1) readjustment of basement trends as a result of arching and the formation of an anticlinorium, (2) the control of Cambro-Ordovician deposition by the trends in the Mount Holly and (3) the control exerted by variations in basement competency on the pattern of movements during deformation of the Cambro-Ordovician. As it is often impossible in a given Cambro-Ordovician structure to separate the effects of control during deposition (2) and control during later deformation (3), both aspects are grouped below as the influence of older structures on the development of the younger structures.

The Influence of Older Structures on the Development of the Younger Structures. The importance of Pre-Cambrian structural trends in guiding the geometrical development of the Green Mountain anticlinorium is doubtful; this regional structure trends north whereas basement structure both in Vermont and New York trends northwest to northeast. The details of this geometry may owe much to control by the basement. Certain of these features in the Rutland area can be discussed.

The control exerted on the *deposition* of the Saltash and the dolomite facies of the Tyson has been discussed. Review of the structural characteristics of the Saltash formation suggests control of the *deformation* of these younger units by the buttress-like Pre-Cambrian quartzites.

Unique structure of the Saltash suggests an unusual tectonic development. Two systems of structural elements are present, one a schistosity, slip cleavage and folding parallel in attitude to those in the overlying Cambro-Ordovician strata, and an older and more obscure system of folding and lineation. The older folds plunge gently eastward in the direction of the dip of the bedding and are parallel to a strong mineral lineation or streaming, and to the direction of elongation of stretched cobbles. Fold plunges (Plate 3) diverge slightly in direction of plunge. The folds are asymmetric; the axial planes approach the attitude of the bedding. Excluding the younger structural system as features impressed upon the Saltash during the deformation of the Tyson and overlying Cambro-Ordovician, there are several explanations for the present arrangement of the Saltash units and their structure. One is that the Saltash is one limb of an isoclinal fold whose eastward plunging axis parallels the axes of its minor folds; generation of such a structure would require either the action of a horizontal couple, or rotation of the major fold after construction by a vertical couple. This possibility is unlikely; such movements would not be at all consistent with the general pattern of movement throughout Vermont which requires a vertical couple with eastern rocks moving westward over western rocks. In addition no imprint of such bizarre movements appears in the basement immediately beneath the Saltash. Finally, the pattern of the minor folds cannot be integrated into a consistent pattern of shear. It is therefore doubtful that the Saltash forms the limb of an eastward plunging major fold.

The hypothesis considered here is that the minor folds in the Saltash are *a*-folds (literature summarized by Cloos, 1946, p. 26; Fairbairn, 1949, p. 222) that have developed during a deformation consistent in geometry with that elsewhere in the Paleozoic rocks. The pronounced and unique linear structure which marks the Saltash reflects the distribution and competency of rocks in the basement. Whereas the dominant movement during the deformation was in an east-west direction (maximum elongation and *a* fabric axis east-west) resistant masses in the basement interferred with a uniform movement of layered units westward up the eastern limb of the anticlinorium. Converging movement (Cloos, 1946, p. 28; Balk, 1936, p. 739) took place causing, at the level now exposed through erosion, a limited shortening normal to the direction of maximum elongation of the Saltash. The quartzite masses north and south[1] of the Saltash acting as buttresses, have not moved toward one another, but have in effect produced a north-south shortening by acting as constrictions in the overall westward flow of plastic rock masses. In response to this subsidiary north-south shortening, numerous small folds developed within the Saltash, with axes trending easterly.[2]

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[1] The northern quartzites are shown on Plate 1 and the southern quartzites are described by Thompson (1950 p. 19) and are located just off the map area south of Saltash mountain.

[2] From studies in the Bersdalen quadrangle, Norway, in which lineation and fold axes have a varying angular relationship to the principal direction of movement, Kvale sug-

Instead of the usual development of bedding foliation and folds with axes normal to maximum elongation (folds in *b* fabric axis), the pervading structure is linear in *a*. This may suggest some parallelism with linear flow structure in cylindrical plutons (Balk, 1937, p. 86) and salt domes (Balk, 1949). These counterparts have certainly experienced constriction in two directions normal to the direction of maximum movement and have developed as a consequence folds or mineral lineation in *a*. The linear structure of the Saltash further suggests a similarity to the features common at thrust faults[1] which include folds, rodding and streaming in *a*. The *a*-folds in the Saltash are overturned and even isoclinal and therefore the situation here may depart somewhat from that at a thrust zone, where *a*-folds are not commonly asymmetrical (Balk, 1936; Strand, 1944); furthermore, it is not often clear that shortening normal to the direction of thrusting has been the cause of the *a*-folds at thrusts, although this is strongly suggested for the Saltash.

In summary then it is suggested that the unique linear pattern of the Saltash is related to irregularities in basement competency which have guided the deforming movements; the deformation is somewhat similar to that in salt domes and cylindrical plutons, and in a few respects to that in thrust zones.

The synclines of Wilcox and Tyson have trends remarkably similar to that of the basement. Whether this is depositional or entirely deformational control by the basement is not known, but it is clear that the Paleozoic and late Pre-Cambrian movements which are responsible for this geometry must have been influenced greatly by basement structures. During deformation of the Wilcox and Tyson, movement probably occurred parallel to the banding in the adjacent Mount Holly to produce synclinal structures essentially parallel to the banding. Although the Mount Holly presumably had already been strongly recrystallized, great physical anisotropism would still have existed, for example, in beds of quartzite within gneiss; it seems reasonable that surfaces of banding might have acted as surfaces of slip in the zone of most intense strain near the contact with younger sequences.

Basement control of Cambro-Ordovician structure is particularly

gests that: "a solid body of rock . . . will tend to influence the direction of the linear elements in the less solid surrounding rocks in such a way that the lineation or axes of folds or both these structures become more or less parallel to the boundary of the solid body" (1947, p. 248).

[1] Summary by Fairbairn, 1949, p. 219–222; and Cloos, 1946, p. 26; Balk, 1937, p. 738, 743; Heim, 1878, p. 82, 202; Strand, 1944, p. 22, 25.

apparent where Pre-Cambrian quartzite and quartz schist occur near the unconformity. At Pittsford and Chittenden complex faulting and folding of Pre-Cambrian quartzite and gneiss and of the Mendon formation occur where the Green Mountain front is offset several miles to the west. The reason for this major offset is as yet unknown, though it is probable that it is related to the local occurrence of large amounts of resistant material in the basement. The peculiar east and northeast trends of Cambrian rocks west of Chittenden can be related to the similar trends of the thick quartzites present there in the basement: for example, at Bald Mountain, North Chittenden and Chittenden these quartzites strike northeasterly. At Pine Hill and Boardman Hill the situation is less clear although again Pre-Cambrian quartzite is conspicuous within the core of this subsidiary anticline. At East Mountain and Bald Mountain, Mendon, south and north plunging folds are outlined in the Mendon formation. Both of these younger folds follow to some extent the basement structure and the syncline at East Mountain roughly traces the outline of the thick quartzite beneath it, with minor fold axes nearly parallel to the trend of the quartzite. The southern fold on Bald Mountain follows even more perfectly the northwesterly trend of the basement, outlined by the Wilcox syncline.

In summary, many details of the Cambro-Ordovician configuration can plainly be related to basement control. The larger scale patterns, however, are considerably harder to relate. A major cause of this interaction apparently is movement along compositional banding of the basement which produces trends in the younger structures parallel to those of the banding. The movement responsible, although of Paleozoic age and generally oblique to trends in the basement, would resolve near the unconformity to directions within the banding of the basement.

Balk describes a mechanism by which younger rocks develop a pseudoconformity with elements in the basement which appears similar to that suggested here (Balk, 1937, p. 732–737, fig. 27, 28). Thrust faults cut the basement gneisses and unconformable younger rocks and through movement along the faults and by intense folding the surface of unconformity is rotated into an attitude near that of the fault. A "fracture cleavage" parallel to the fault develops on both sides of it and by guiding later recrystallization completely masks the original unconformable relation. It is thus a means of developing structural elements parallel to thrust faults and it is further apparent that these faults (judging from his sketches) are subparallel to the banding in the basement gneisses.

The effect then, would be similar to that suggested here, in that a conformity and parallelism of trend appears between younger structure and the banding in the basement.

From studies of complex structural relations in the Bergsdalen quadrangle in Norway, Kvale believes that "the directions of older structures may influence the directions of structures formed by a second deformation with the result that the old directions are followed or that the new directions are compromises between the old directions and the theoretical new directions." (1947, p. 249).

Readjustments in the Basement Caused by Younger Folding. Just as the basement trends have locally guided the development of structural patterns in the Cambro-Ordovician, the arching and formation of an anticlinorium has affected the basement structure. Theoretically the basement might yield in several ways. The change from a flat surface to a strongly arched surface might cause multiple fracturing and widespread granulation of the underlying material. Or if conditions were such that the material could act in a completely plastic manner, the basement might readjust its surface through some sort of irregular flow. It is possible that a basement of uniform competency might adjust its form by movement along shear planes developed parallel to its upper surface. Finally, the bulk of the basement might structurally remain more or less intact with all adjustments taking place by intense fracture and flowage close to the margin. The Green Mountain anticlinorium apparently demonstrates the last of these mechanisms. Few major faults were detected and these show only minor rotation of the units involved. Large-scale flowage or the development of surfaces of shear parallel to the contact are similarly absent, except within several thousand feet of the contact.

Before examining the marginal zone in detail certain aspects of the formation of a standard flexure fold (Billings, 1942, p. 87–98) are discussed. A segment of a fold limb rotates under the influence of several couples. In Figure 7, such a fold is sketched; the surface shown might be considered the unconformity between the basement and flanking strata. Segments on each limb of the fold rotate relative to an initial horizontal surface; first, due simply to the change in geometry of the surface; second, due to the influence of crestward moving overlying layers (expressed in minor folds); and third, due commonly to the effect of breaking and thrusting of the overturned limb. The directions of these rotations are not the same and cause irregularities on the limbs

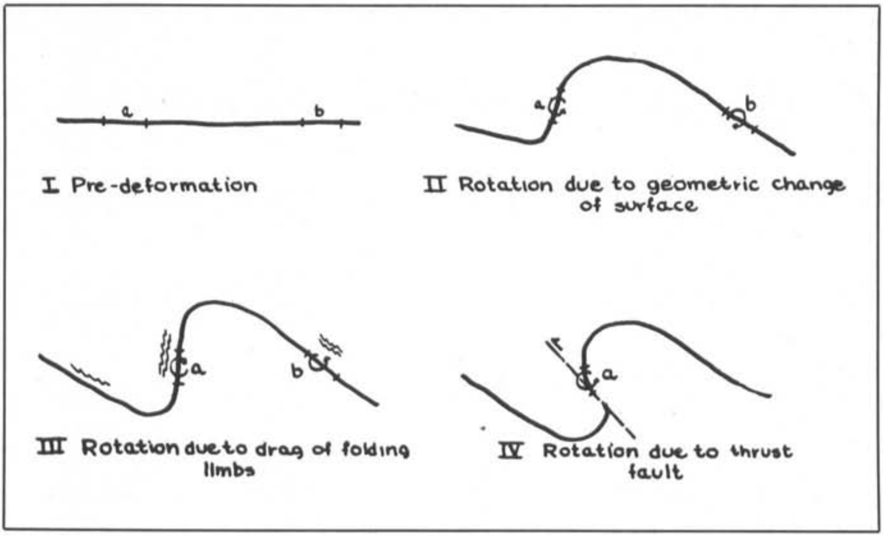


Figure 7. Rotation of Segments of a Fold Limb during the Formation of a Flexure Fold.

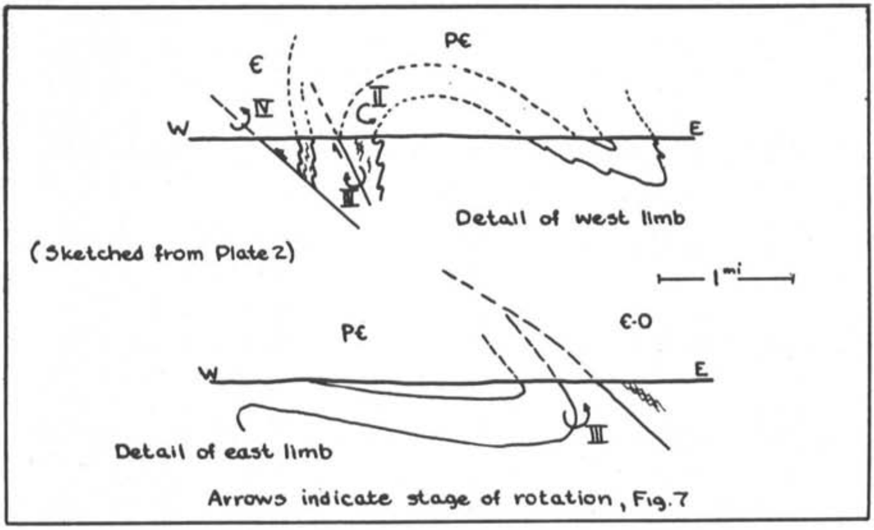


Figure 8. Detail of Basement Margin on west and east sides of the Green Mountain Anticlinorium showing Typical Bending of Older Structure.

of folds. Minor complications are usually more abundant on the overturned, or "drag" limb of a fold, than on the normal limb. This is immediately apparent since movement along thrusts on the drag limb reverse the shear sense acting on that limb (see Figure 7, IV).

Over a distance of several thousand feet stratigraphically beneath the unconformity, Pre-Cambrian structures have been bent into apparent conformity with younger rocks. The direction of this bending gives the movement sense affecting this portion of the fold limbs, and is most clearly shown at the eastern margin south of Sherburne, to the west at Bald and East mountains, and in the syncline of Wilcox east of Clarendon (Plate 1). Typical relations at the east and west margin of the basement are shown in Figure 8.

The bending of basement structure on the west limb is in a counterclockwise sense, looking north (Figure 8). The west limb of the Wilcox syncline has been rotated about 90 degrees; its peculiar map pattern reflects the preservation of original drag folds consistent with its synclinal structure. Rotation of this same sense and degree appear likely for the thick quartzite just south of Mendon. In addition to this movement, it is possible to detect the effect of another couple within the basement which shows as a fault, with west side up relative to east side, cutting the west limb of the Wilcox syncline; its movement sense is the same as that producing drag folds in the overlying Cambrian. Still another fault appears to the west, with unknown effect on the basement, but reflecting the normal tendency of a drag limb to break with intense folding. Bending and faulting of the Wilcox syncline probably occurred at about the same time inasmuch as the couples involved exist at the same time. The final thrusting may mark the latest deformational event in the formation of this west limb of the anticlinorium.

Bending of the basement structures on the east limb was similarly in a counterclockwise sense, looking north. Several thousand feet of quartzite, gneiss and marble exposed from Pico to Killingon peaks are bent into near-conformity with the younger rocks which dip moderately northeast. Comparing the two limbs of the anticlinorium, the important changes in the geometry of the basement have apparently been produced by a couple of the same sense—eastern rocks moving westward over western rocks.

This mode of deformation of the basement, rotation of older structures into near-conformity with the younger is described elsewhere.[1]

In summary, the Green Mountain anticlinorium consists (1) of a central zone which seems structurally intact and may represent part of an ancient fold belt, and (2) of complex thin marginal zones in which basement and flanking rocks have interacted during the formation of this mid-Paleozoic structure. Chemical changes of the basement have probably been more extensive than structural changes during the most recent episode of mountain building in the region. The effects of the younger metamorphism is extensive whereas the development of conformable basement structure is limited to the margins of the complex.

East Limb of the Green Mountain Anticlinorium. Tyson and younger units of the east limb of the anticlinorium have fairly uniform structure. Large-scale folding is absent, beds strike northeasterly and dip moderately east, and minor structures have regular and consistent attitude. Minor structures are very abundant in the argillaceous lithology which is receptive to folding and the development of foliation surfaces. Although bedding is uncommonly seen in the phyllites and schists, its parallelism with the dominant rock foliation is consistent. Quartz veinlets and paddle-shaped masses are formed in this bedding foliation, which was apparently a surface of movement during deformation. Slip cleavage is often identified and its intersection with bedding foliation produces a crinkling typically parallel to the axes of minor folds. Folds up to a few feet in wave length are abundant; axial planes are nearly vertical, and axes plunge gently northward. Streaming, rodding and pebble elongation plunge steeply to the northeast at an angle of 50 to 70 degrees with fold axes in the plane of the foliation. Elongation in this direction of chlorite, chloritoid and quartz is weak and observable only in thin section. Albite, magnetite and pyrite often indicate rotation during and following their formation. The movement sense as recorded by the rotational features of the east limb is typically up east.

North and east of North Sherburne, folds of several hundred feet

[1] In the Alps Heim observed that schistosity of basement rocks of the north end of the Aar massif had been bent and dragged into near-conformity near the contact within mantling autochthonous strata (Heim, 1921, p. 157). In similar fashion too, the normal foliation-banding relations of the massif are disturbed near the margin of the massif with the appearance of various cleavages. In Dutchess county, New York, warping of the basement structure again results in near-conformity with mantling rocks, although in this case the surface of separation is a thrust fault (Balk, 1937, p. 739, 741). In the Rutland area bedding planes are surfaces of movement and the effects produced by movement might be expected to be similar to those produced by thrust faulting on a surface roughly parallel to bedding.

wave length appear, causing a greater outcrop width of formations. Two miles north of the village overall bedding trends WNW, but is obscured by the development of NNW trending slip cleavage and the more northerly minor fold limbs. Near Sable Mountain the dominant structure is linear, plunging steeply eastward. Folds with subhorizontal axes are replaced here by steep folds with no consistent movement sense. Since the fold axes make a large angle with regional fold axes the folds are considered *a*-folds, originating in subsidiary horizontal movements related in some way to the development of the regional bend (in plan) of formation boundaries in the region (map: Billings, et al, 1952).

Two steeply east-dipping thrust faults southeast of North Sherburne assist folds in offsetting the pattern of bedding to the northwest. The faults are probably of several hundred feet displacement and are not associated with any unique development of minor structures; folds and steep plunging lineation are abundant throughout the region.

Synclines of Tyson extend from North Pond south to the Deer Leap cliffs on Route 4 several miles west of Sherburne. The northern of the two synclinal masses has regular bedding foliation and minor folds with movement sense consistent with synclinal structure. South of South Pond fold axes have a more nearly WNW attitude with widespread variation in the direction of gentle plunge. The parallelism with basement trends is noteworthy, as is the control of map pattern by topography.

A faulted area occurs east of Woodward Reservoir and Plymouth. Aside from verification of Chang's description (Chang, 1950, p. 51), the writer made no detailed studies of this area. The existence of a moderately east dipping fault rather than a fold is suggested by minor features. Drag folds are not consistent with pure folding, and the attitude of foliation and fold axes changes abruptly near the eastern contact of Pinney Hollow and Grahamville.

West Limb of the Green Mountain Anticlinorium and Pittsford-Chittenden Structural Complex. Mendon and Cheshire formations of the west limb of the anticlinorium and in the Pittsford-Chittenden area reflect complex movement although lithologically the rocks are less suited for the development of minor structures than the phyllites of the east limb. Bedding of the west limb is vertical to overturned east, folds trend northward and plunge gently either north or south and fold axial planes dip moderately eastward. Large and small folds have a consistent up west movement sense, although the varying direction of plunge renders the map pattern irregular. Uncommon lineation of stretched pebbles, streaming and mineral elongation plunges steeply eastward at an angle of 60 to 90 degrees with the subhorizontal fold axes. Broad, open folds are characteristic of beds of quartzite and grit, but occur together with tightly folded schist and dolomite. From Bald Mountain southward through Clarendon the succession of the west limb is broken by a thrust fault, with the resistant sandstones and grits of the Mendon formation displaced westward over Ordovician carbonate rocks. The fault probably dips moderately to steeply east and with a dip slip of at least 7,000 feet.

In the Pittsford-Chittenden area the arrangement of the younger units is clearly the most complex of any of the area. Although knowledge of the northern extension of the structure is necessary for final analysis, it is possible at this point to grasp the essence of the structure. The area is located where the Green Mountain front is offset westward several miles, due possibly to the control exerted during folding by thick masses of eastward trending basement quartzite. While it is impossible to conceive of the irregular movements which have produced the details of the Pittsford-Chittenden complex, its larger elements can be related to the usual east over west shearing movement. Thus the structure consists of three south to southwest plunging anticlines strongly overturned to the west and complicated by irregularity of basement competency and possibly by a later crossfolding on a more southerly axis. Local thinning of the basal Cambrian clastics, as well as faulting, further complicates the picture. The westernmost fault is sub-horizontal and has served to displace quartzite and sandstone of the Mendon and Cheshire westward over carbonates. This may be the northern extension of the Pine Hill thrust. Another fault cuts the structure a mile or so northwest of Chittenden. Its existence is inferred by the disappearance of lowest Mendon and the absence of minor structural features compatible with a purely folded structure. Displacement on this fault is undoubtedly small with probably a few hundred feet dip slip. The easternmost fault, an inferred normal fault, is somewhat anomalous here, departing as it does from the usual thrusting movement on faults of the west limb. Evidence for its existence is well shown in the vicinity of Chittenden. A half mile south and a mile and a half north of the village Cambrian Moosalamoo and Pre-Cambrian biotite-microcline gneiss are a few tens of feet apart. The fold sense in the younger rock again precludes a purely folded structure with a local absence of basal Cambrian grits. A west dipping thrust, or east dipping normal fault between Moosalamoo and gneiss could produce the pattern found in the rocks; in the absence of additional evidence the latter is chosen, for it is consistent with the general tectonic movement of the area. Faulting could have occurred parallel to steep east dipping slip cleavage (abundant throughout the area) under the influence of an east over west couple. Counterclockwise rotation of blocks and fault between would bring older rocks up on the west giving the configuration seen at Chittenden. This is the "antithetic faulting" of Cloos (1928), applied in a somewhat different sense.

The Pine Hill Anticline. The Pine Hill structure is a complex anticline with the western or drag limb, broken by a major thrust, the Pine Hill thrust. This structure was recognized first by Wolff (1891) and carefully mapped by Dale (1892, 1894). Fossil discoveries to the east and west of Pine Hill suggested a thrust fault in the structure, and continuity of units above the fault was shown with those of the Green Mountain front to the east. Dale also hinted at the presence of Pre-Cambrian gneiss in the core of the structure (1892). The major failing of this early work was confusion of lithologically similar black phyllites of Lower Cambrian and of Middle Ordovician ages. Fowler (1950) described the structure of the folded carbonates west of the Pine Hill thrust, but had little success in delimiting the units to the east of the fault.

Pre-Cambrian and Mendon formations appear above the Pine Hill thrust in tight folds strongly overturned to the west; axial planes dip 30 to 50 degrees east above the fault, which probably dips 40 to 50 degrees east. Beneath the Cheshire quartzite minor folding is extreme with rapid changes in attitude of sub-horizontal folds; uncommon linear structure, folds, pebble elongation, rodding plunge gently easterly. The Cheshire itself is hardly deformed, but simply dips gently eastward beneath the dolomite of the Rutland Valley. The actual location of the Pine Hill thrust can seldom be determined exactly, for the rocks on either side of the fault are unfortunately very similar. Thus, the black sandstones and phyllite of the Moosalamoo often rest upon the black phyllite and marble of the Hortonville. The writer has placed much reliance on the quartz-rich, fine-banded nature of the Moosalamoo in attempting the subdivision. Since the Hortonville is seldom conformable with older units in the area (being at the base of a major unconformity), it appears simply as a wedge beneath the Pine Hill thrust. South of Chippenhook and Clarendon the Pine Hill thrust dies out, but the anticlinal structure of the ridge is disturbed by a new break. This fault is on the east side of the structure and its existence is principally supported by work to the south in the Wallingford and Pawlet quadrangles (Thompson, oral communication, 1952). This will be discussed further below.

The displacement along the Pine Hill thrust increases to the north. At Chippenhook the dip slip is negligible; at Rutland it may reach 4000 to 6000 feet, depending upon the relations beneath the Hortonville unconformity (Section E, Plate 2). Farther north in the Pittsford area, displacement along the thrust (assumed a continuation of the Pine Hill thrust) must be of the order of several miles. The formation of the Pine Hill thrust apparently postdated the deposition of the Hortonville and therefore the pre-Hortonville folding and erosion of Cambro-Ordovician rocks. The base of the Hortonville, as defined by fossiliferous limestone, is preserved only west of the thrust; hence, the Hortonville has also been affected by the faulting. Pre-Hortonville erosion in the Rutland area extends to Cheshire and younger formations and Bucher (1951) has described north of Peekskill, New York, a limestone and slate presumably of the same age as Hortonville which rest directly upon Pre-Cambrian. The possibility of correlation of this major mid-Ordovician unconformity with an unconformity east of the Green Mountains has been discussed elsewhere in this report.

The Rutland Syncline. Cambro-Ordovician carbonates and sandstones of the Rutland valley have the general form of a south plunging syncline with parts of the east and west limbs disturbed by faulting. North of Rutland gentle dips are south and east, the beds are gently warped and, aside from jointing and bedding planes, are structureless. Southward the situation changes. All the unusual features of carbonate deformation appear as the beds are faulted and folded into tight synclines. Dips are variable and change constantly along and across the strike; fold plunge is gentle north or south, but axial planes dip east and west and often have no consistent local movement sense. Wave length of fold seems to vary with thickness of bed, thick units broadly arched and occasionally broken, thin beds of sandstone and dolomite intricately faulted and dragged. Mapping was necessarily done by lithologic correlation and the use of rare cross-lamination, for secondary structural features, are quite unreliable.

Between Mill and Cold rivers synclinal structure is quite consistent with the regional pattern of deformation; a syncline with axial plane dipping steeply east is a reasonable counterpart for the neighboring anticline in Bald Mountain. The fault which disturbs the east limb of the structure must have been a late feature of the deformation and not directly related to folding, since it cuts the fold at an angle to the axis.

In the vicinity of Clarendon, all semblance of order disappears in the structure of the carbonates. The entire section exposed in the Mill River dips moderately to steeply west, fold axial planes dip west, whereas cross-lamination indicates that the top of the sequence lies to the east. This indicates a fold overturned to the east, which would be quite incompatible with its regional surroundings. A possible process of fold generation might be the "flowage folding" of Bain (1931). Flowage folds are produced by a gravity sliding of material into fold troughs and are probably similar to the penecontemporaneous folds formed in sloping beds of hydroplastic sediments (Shrock, 1948, p. 264–267). These folds in forming on the west limb of the Rutland syncline would in effect have axial planes dipping west. But considering the scale of the structure in the vicinity of Clarendon and the fair degree of consistency of its minor structures, the writer hesitates to apply this mechanism of flowage folding. In quarries it has been uncommonly observed in folds up to a foot in wave length; further, the penetration of this sort of movement does not appear to be extensive, judging from interformational counterparts (Shrock, 1947, p. 267), in which only the upper surface of a unit is plicated. It is much more likely that the formation of this anomalous syncline is related to a fault to the west. The fault has raised western rocks relative to those to the east and could be a west dipping thrust. The movement, west over east, although itself regionally anomalous, might have locally arched the west limb of the Rutland syncline into a west dipping attitude, preserving the otherwise normal minor structures. This seems to be the most acceptable hypothesis, but may no doubt be revised when studies to the south throw further light upon the rather unusual tectonics of this area.

Major Structure Summarized, Age of the Deformation. The basement structural complex of the Green Mountain anticlinorium consists of a central zone which may preserve intact earlier oblique deformation trends, and a complicated thin marginal zone in which basement and flanking rocks have interacted during Paleozoic deformation. Structures of basement and mantle in the marginal zone (about a mile thick stratigraphically) indicate on the one hand that younger units may locally develop structural trends paralleling older banding, and on the other that basement structures have been bent or dragged into near-conformity with the mantle. Major faulting appears to have been unimportant as a means of adjustment of the basement. Significantly, no fault was found with appropriate geometry to have acted as a root zone for overthrust Taconic masses to the west of the area.

The interpretation that the principal basement deformation occurred in a marginal zone, suggests the corollary that the ancient core of the Green Mountains underwent only a limited amount of shortening in an east-west direction.

Cambro-Ordovician structures of the Rutland area are for the most part conformable and their structural pattern is simple, consisting of a major anticlinorium and a subsidiary anticline overturned to the west and plunging gently north. Primary structures and secondary rotational features on both limbs of the folds are consistent with this pattern. Structural irregularities are more abundant on the west limb of the folds than on the east limbs, a situation to be anticipated on the "drag limbs" of folds. Slip cleavage and the axial planes of folds describe an incomplete fan across the area, being nearly vertical to the east and moderately to gently east dipping to the west. The overall structure points to the operation of a couple, with eastern rocks moving westward over western rocks.

Minor folding in a direction transverse to the range makes the Saltash structural unit anomalous in the map area. Stratigraphically the unit is an unconformable wedge lying between Cambro-Ordovician and Pre-Cambrian. In spite of having undergone deformation at an earlier period (either Pre-Cambrian or Ordovician) it seems possible to relate its anomalous structure with overall movements similar to those which deformed the overlying strata (presumably Taconic).

Until the stratigraphic relations are better known, it is impossible to date the deformation accurately or to evaluate the different Lower Paleozoic deformational episodes in the area. Regionally, metamorphism or deformation or both occurred within the Middle Ordovician, between Middle Ordovician and Upper Silurian (Taconic orogeny), in Middle or Late Devonian (Acadian orogeny) and late in the Paleozoic (Appalachian orogeny) (Billings, et al, 1952). Undeformed plutonic rocks correlated with those of Mississippian (?) age preclude a severe Appalachian deformation in the map area. In the absence of rocks younger than Middle Ordovician and older than the Devonian, the effect of the Acadian disturbance is not known. There is abundant evidence for Middle Ordovician and Taconic orogeny, however, just to the west of the Rutland area. To the east it is certain only that there were two periods of deformation and metamorphism separated in time by deep erosion and extensive deposition and that these episodes occurred within the late Pre-Cambrian to Middle Ordovician. The degree of metamorphism attained and the principal movement patterns recorded are remarkably similar.

Minor Structural Features

General Statement

Dale has written: ". . . the characteristic features of . . . mountain-making movements are often quite as truly shown by single ledges, or even hand specimens, as by whole mountain sides." (Dale, 1896, p. 549). Features of outcrop and hand specimen are, in fact, the only data available in central Vermont from which the geologist synthesizes major structure and the formation of the mountain range. Critical examination of these features is of great importance, therefore, for the light that they throw upon the mechanics of rock deformation and the manner in which they are integrable into a consistent movement pattern within major structural units.

Structural details of the Rutland area include planar, linear and rotational features, and details of rock fabric. It is convenient to refer the geometry and kinematics of these features to a coordinate system similar to that in common use (Fairbairn, 1949, p. 6; Cloos, 1946, p. 5). The axes a and b are in the movement plane; c is normal to it. b is parallel and a at a large angle to fold axes. Both a and b axes are defined by linear elements in the rocks; b is a rotation axis and a is a direction of elongation or stretching.

In the application of the strain ellipsoid (Billings 1952, p. 106; Fairbairn 1949, p. 199) the axes A, B, and C are used for the long, intermediate and short axes, respectively.

Planar Features

Bedding, Compositional Banding, Foliation. Compositional and textural banding is widespread in the rocks of the Rutland area and can usually be assigned to primary origin; it almost invariably parallels rock foliation. Banding and foliation are cut by slip cleavage, fracture cleavage and joints.

Bands differing in mineralogy or texture in the Cambro-Ordovician rocks have been interpreted as bedding. In rare cases they parallel undisputed primary features such as cross-laminated and graded beds and when mapped define the structure of the area. The scale of banding ranges from that of the delicately laminated black sandstones (less than a millimeter) to that of the Nickwacket greywacke and Cheshire quartzite in which dolomitic or schistose layers are tens of feet thick. Interband differences in mineralogy have doubtless been accentuated during metamorphism and deformation by such processes as Eskola's "concretion principle" (Turner, 1948, p. 139) or tectonic unmixing. Banding in the Pre-Cambrian remote from the margins of the complex may also be primary, since it is traceable for considerable distances through complex structures; but the general validity of this assumption awaits further understanding of the origin of many of the Pre-Cambrian rocks. In the zone of most intense deformation near the margins of the complex, compositional banding often parallels the surface of unconformity; here it is likely that processes other than primary deposition have influenced the character of the layering.

As a consequence of folding, planar features such as banding and foliation are offset; it is therefore necessary in mapping to ascertain the "trend of foliation," which is a surface estimated to be the mean of all of the folded surfaces visible. This mean surface may actually cross the strikes of minor fold limbs at a large angle where folding is most intense. Accurate "trend" determination obviously depends upon the three-dimensional extent of outcrop exposed to view.

Nearly all the rocks part along subparallel surfaces or folia which are conditioned by the parallel arrangement of compositional or textural bands and platy or elongate mineral shapes. This foliation is parallel to banding throughout, even on the noses of folds, and during folding has apparently served as a slip surface. Distinct linear elements such as crinkling and microfolding in *b*, slickensiding, streaming and mineral elongation in *a*, are usually visible in the foliation. Pebbles and quartz veinlets are elongate in the foliation through stretching, or fragmentation. Foliation appears at once as a surface of shear displacement, a bedding foliation (Mead, 1940, p. 1009) and a surface containing the direction of maximum elongation. Nowhere can foliation be identified as an axial plane cleavage or "flow cleavage" (Idem, p. 1010).

The origin of the foliation, or ability to part parallel to banding, is to a large degree a property developed early in the history of the rock. However, the coarse micas and lenticular mineral shapes developed parallel to banding and contributing greatly to the planar weakness in the rocks are unquestionably a product of metamorphism and attendant deformation. The actual mode of grain growth and orientation in foliation is not known as similar results have not been achieved as yet by synthetic means. There has apparently been no tendency for platy minerals to grow preferentially in parallelism with the axial planes of folds, normal to "the direction of the resultant of compressional stresses in the rock mass" (Mead, 1940, p. 1010). Platy and elongate grains other than biotite and chloritoid porphyroblasts, have only extended themselves in surfaces of shearing movement.

Slip Cleavage. Slip cleavage (Dale, 1896, p. 560) is a widely developed planar feature and consists of spaced surfaces of parting or incipient parting formed sub-parallel to the limbs of small folds (Plate 7, 8). It is usually parallel to fold axial planes and intersects the foliation in a line parallel with *b*. Slip cleavage is apparently restricted to micaceous rocks and is not developed in dolomites or quartzites (Plate 7, Figure 1). Recognition and use of slip cleavage to indicate movement sense are greatly enhanced by banded structure in the rock (Plate 7, Figure 2). The spacing of slip cleavage is somewhat variable even within a hand specimen, ranging from a fraction of a millimeter to 10 inches. Slip cleavage is often discontinuous even in wholly schistose rocks, with individual surfaces randomly becoming indistinct and reappearing offset. But regardless of its variable intersection with foliation and its discontinuity, parallelism with axial planes of folds is retained. Rarely slip cleavage is itself gently folded or, as in such complex situations as the margins of the basement complex, bears an irregular relation to folds of an early generation. In phyllites of the west limb of the anticlinorium slip cleavage is unusually developed, and due to its very close spacing and the lack of compositional banding, is the principal structural element in the rock. Elsewhere in homogeneous phyllites it is entirely possible that more than a single slip cleavage system has been formed and later rendered indistinguishable from structure mapped as foliation.

Slip cleavage is not related to the folds in the basement complex. It reaches maximum development in the younger rocks in the center and western limb of the anticlinorium. Following the trends of fold axial planes slip cleavage converges downward across the area, being nearly vertical to the east and dipping moderately eastward to the west (Plate 3).

All stages in the development of slip cleavage surfaces can be observed.

Figure 1. Slip cleavage in schistose bands of a quartz-sericite-biotite-chlorite schist of the Mendon formation. Light-colored quartz bands are not offset by the cleavage (vertical) which is about parallel to fold axial planes. (uncrossed Nicols) East Mountain, Rutland. x 10.

Figure 2. Incipient slip cleavage in quartz-sericite-chlorite-graphite schist. Note discontinuity through general parallelism of (vertical) cleavage surfaces. Movement sense indicated here is up left. Tyson formation, South Pond, Chittenden. (uncrossed Nicols) x 10.

PLATE 7, SLIP CLEAVAGE.

Figure 1. Detail of fold in banded quartz-sericite-biotite-chlorite schist. Slip cleavage is vertical. Recrystallized sericite and biotite (dark grains) form an arch growing parallel to bent foliation surfaces. Mendon formation. East Mountain, Rutland. (uncrossed Nicols) x 20.

Figure 2. Advanced stage of slip cleavage development in sericite-chlorite-quartz phyllite of the Mendon formation. Cleavage is accentuated by growth of opaque minerals and by iron oxide staining. Overall movement sense is not clear here. Pittsford. (uncrossed Nicols) x 20.

PLATE 8, SLIP CLEAVAGE.

In an initial stage (Plate 7, Figure 2) folding forms the characteristic drags on the limbs of larger folds; although no surfaces of discontinuity have as yet weakened the rock, the eye is nevertheless attracted to the axial plane symmetry in the herringbone pattern. With advancement of the folding the tiny drags are more acutely bent, and limbs are strongly thinned and rotated toward parallelism with the fold axial plane. Further thinning and stretching of drag folds causes actual surfaces to appear in the lining up of fold limbs. These surfaces seem to favor the growth of micaceous and opaque minerals, which accentuate the slip cleavage and render it a more noticeable surface (Plate 8, Figure 2).

Slip cleavage belongs to an advanced stage of the deformational history as it is seldom deformed. Also the associated acute drag folding might be expected in the later stages of flexural folding when the maximum amount of slip along foliation surfaces has been achieved, and tight, angular small folds develop instead of broader and simpler ones (Mead, 1940, p. 1021). With the formation of numerous fold crests where once there was one in the same length of bed, total linear slip is reduced along increments of the bed.

Structural features apparently similar to slip cleavage have been given various names:—*fracture cleavage* (Leith, 1905, p. 119; Broughton, 1946, p. 18; Balk, 1936, p. 706; Kvale, 1947, p. 19), *false cleavage* (Harker, 1932, p. 157), *transposition cleavage* (Weiss, 1949, p. 12), and *ausweichungsclivage* (Heim, 1878, p. 53). Not only are names different, but various modes of origin are assigned to slip cleavage (1) as a consequence of, or closely related to folding (Thompson, 1950, p. 118; White, 1949, p. 590; Knopf, 1931, p. 16; Dale, 1896, p. 567; Osberg, 1950, p. 85; Harker, 1932, p. 158), (2) due to fracturing parallel to surfaces inclined to AB plane of strain ellipsoid and unrelated to existant mineral arrangement (Leith, 1905, p. 126; Mead, 1940, p. 1010), or (3) as a fracture system unrelated to flexure folding but possibly to shear folding, micro-thrusting, normal faulting (Balk 1936, p. 706; Hawkes 1940, p. 124; Broughton 1946, p. 8; Schmidt 1932, p. 89). Apparently all of these names and origins refer to the same feature discussed here as slip cleavage. Differences in interpretation are due on the one hand to attempts to refer it to a cleavage classification developed in an area where slip cleavage does not exist, and on the other hand to the rather ambiguous movement sense given by the cleavage and related folds which led in many cases to divorcing slip cleavage formation from folding movements.

White has carefully reviewed the subject of slip cleavage in recent years (1949, p. 588) and clearly showed that it is not a fracture cleavage in the original genetic sense of the term (Leith 1905 p. 120). In the first place the efficacy of slip cleavage as a plane of weakness in the rock depends upon the subparallel arrangement of platy minerals in the limbs of tiny folds; indeed, the rock often breaks in the general direction of these fold limbs before an actual cleavage surface has formed. Secondly, fracture cleavage is referred to a different orientation in a strained rock than that which slip cleavage occupies. Fracture cleavage develops along surfaces inclined at an appreciable angle to the AB strain ellipsoid plane (Mead, 1940, p. 1010) and accepting the axial plane of folds as representing this ellipsoidal plane, slip cleavage is not inclined to it, but parallels it consistently. Finally, in eastern Vermont, White has been able to trace slip cleavage into a structure called schistosity which is similar to flow cleavage in that all or a majority of the micaceous elements in the rock are parallel (Idem, p. 591). If slip cleavage is truly equivalent to fracture cleavage it should show a distinct angular relationship to a flow cleavage (Leith, 1905, p. 125). Another interesting possibility of the close relationship between slip and flow cleavage is suggested by Dale in the discovery that the "slatey cleavage" of a shale from Bennington, Vermont, actually consisted of slip cleavage surfaces spaced about 360 to the inch (Dale, 1896, p. 563–564). In conclusion it seems unwise and inaccurate to apply the term "fracture cleavage" to the phenomena described here and named by Dale "slip cleavage" since neither the definitions nor the genetic implication of these terms have much in common and in some respects slip cleavage seems more closely allied to structure called flow cleavage.

In undertaking next the matter of movement sense associated with slip cleavage, the variety of genetic interpretations afforded this structure are to some degree made compatible. Primarily there are a number of patterns which can be included in the general term slip cleavage. These are shown in Figure 9 in relation to the cross sectional form of the Green Mountain anticlinorium. The patterns are all similar in that parallel surfaces arise as a result of the subparallelism of tiny fold limbs; the surfaces are slip cleavage and they parallel fold axial planes. The difference in the patterns can be traced to alignment of west or east limbs of the minor folds, or to alignment of both limbs. Considering as an example the possibilities at position A on the major fold, one pattern is formed if west limbs are thinned and brought into sub-parallel arrangement, another if the east limbs are so affected. At the crest of major folds both limbs may become aligned (position B), forming "Totfalten" (Ampferer, 1938). The factors which affect the ultimate development of a particular pattern are not entirely clear, but probably include the geometrical relationship of early foliation, the ultimate position of the fold axial plane, and various mechanical properties of the layered units involved.

The formation of all slip cleavage patterns is believed due to the action of couples with rotation sense consistent with the geometry of the major fold. Thus in Figure 9 both the patterns as position A result from the normal up west shear sense of layered units on the west limb of the anticlinorium. The "Totfalten" at the crest of the fold have no shear sense but reflect the customary shortening in a direction normal to axial planes (C of the strain ellipsoid). Again at D the peculiar geometric combination of inclined foliation and vertical axial planes gives in the gentle undulations no definite shear sense.

Now it is widely held that slip cleavage has arisen through microfaulting, with related dragging of foliation surfaces into parallelism with slip surfaces. The likelihood of such a mechanism seems obvious from the patterns, Figure 9 and Plate 8, Figure 2. Thus the forms shown in position A of Figure 9 might be due respectively to faulting in a thrust or normal sense, parallel to axial planes and with uniform small dip slip. Extreme flowage and thinning of tiny fold limbs produces an apparent offset of beds and renders a faulting hypothesis extremely attractive. However, in attempting to apply this hypothesis throughout a folded area, it is often necessary to create an intricate movement pattern which not only must vary in shear sense locally as slip cleavage patterns change but is often a complete reversal of shear sense as shown in competent folded beds and the position of the slip-cleaved rock in major folds. Thus, faced with the patterns at C, Figure 9, on the east limb of the Green Mountain anticlinorium Hawkes had to devise a complex scheme of movement in which some segments moved up east and others up west along the cleavage surfaces although the shear sense indicated by all rotational features is up east (Hawkes 1940, p. 142).

As mentioned earlier, it is difficult to understand the persistent agreement of slip cleavage with fold axial planes and axes if the cleavage is indeed a fracture phenomenon presumably later and unrelated to folding. It furthermore does not apparently affect competent beds (Plate 7, Figure 1) as fracturing doubtless should. A final interesting discovery

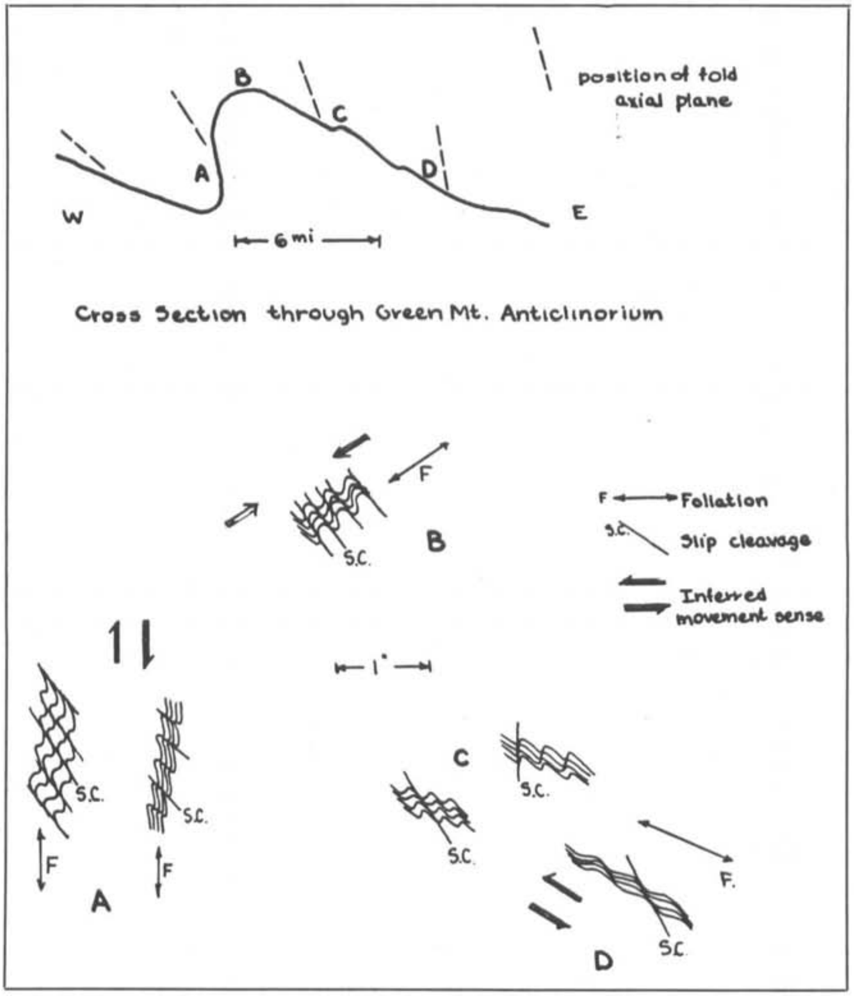


Figure 9. Patterns of Slip Cleavage.

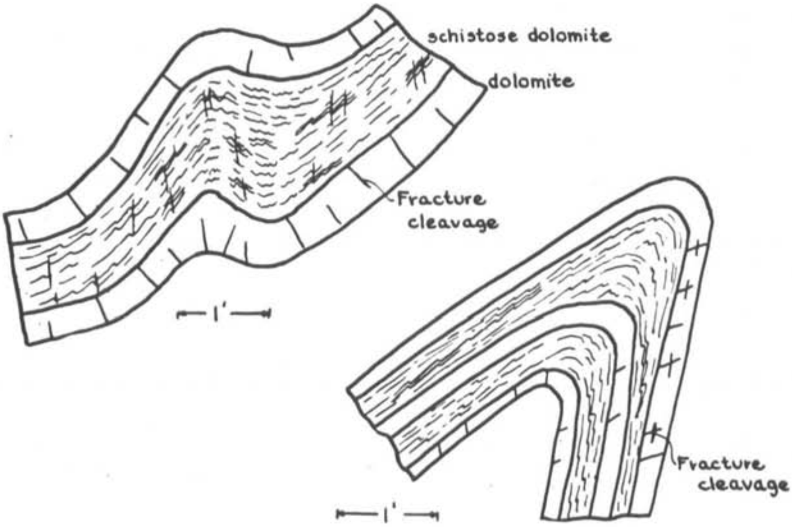
throws some light on this rather ambiguous matter of slip cleavage shear sense. Porphyroblasts of magnetite and albite (Plates 10 and 11) enclosed in cleavage bounded segments have a rotation sense, as indicated by external quartz pressure shadows, which agrees with a folding rather than a faulting development of the slip cleavage. Contemporary formation of the pressure shadows and slip cleavage is assumed since the rolled grains are often located in the part of the foliation which approaches parallelism with the cleavage plane. The grains would have suffered a reverse rolling if fault slip had produced the structure.

In summary, slip cleavage is believed to form as a result of minor folding in an advanced state of flexural deformation. While it is neither fracture nor flow cleavage in the strict definition, parallelism with fold axial planes and apparent transition into flow cleavage suggests that there are genetic controls common to both slip cleavage and flow cleavage.

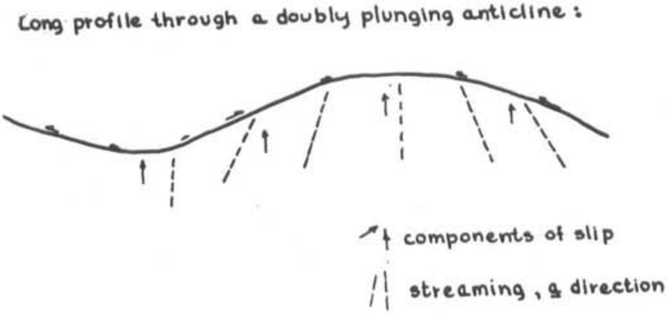
The development of slip cleavage is not only related to the local pattern of deformation but to rock type as well. The most important lithologic requirements for the formation of slip cleavage appear to be thin layering and mechanical properties such that slippage can occur parallel to the surfaces of the layers. Thus micaceous rocks such as phyllite and schist often show excellent slip cleavage whereas it is universally absent in thick-bedded sandstone and dolomite. Slip cleavage is not always developed in relatively incompetent rocks, for example, in a dolomite bed within a thick quartzite unit. Relative competence is apparently less important for the formation of slip cleavage than is micacous layering.

In summary, slip cleavage is believed to be the direct consequence of minor folding of micaceous beds during a somewhat advanced stage of the deformation. The existence of tiny similar folds is essential to the formation of slip cleavage; they exist without the cleavage and are by themselves completely in accord with the symmetry of the regional fold pattern. Slip cleavage, on the other hand, has never been seen in unfolded rock and it would be unwarranted to assume for it a post-folding origin in view of its persistent parallelism with fold axial planes. This parallelism and apparent transition into flow cleavage suggests that there are genetic controls common to both slip cleavage and flow cleavage.

Fracture Cleavage. Fracture cleavage (Mead, 1940, p. 1010) is not common in the rocks of the Rutland area, being found principally in mixed carbonate-sandstone sequences of the Rutland syncline. Typically, relatively competent beds have a fracture cleavage system in which rupture surfaces are spaced a few inches apart and oriented both parallel and nearly normal to bedding, Figure 10 (a). Competent lithology includes quartzite, sandstone, thick-bedded units and dolomite with re-



(a) Attenuation of Fold Limbs of Incompetent Beds, Fracture Cleavage developed in Competent Beds. Sketched from outcrop, Clarendon.



(b) Development of Streaming Subnormal to Fold Axis in Doubly Plunging Fold.

Figure 10. Folding

spect to dolomite, limestone, thin bedded units and schist respectively (See also: Dale, 1896, p. 551; Balk, 1936, p. 718).

Fracturing on a small scale often has the spatial orientation of fracture cleavage. Albite porphyroblasts (Plate 9, Figure 2) enclose planar inclusions at a large angle to bedding and it is possible that this arrangement is due to collection of material foreign to the growing porphyroblast into small-scale fracture cleavage surfaces. In the gneiss of the Mount Holly complex large altered garnets are often cut by distinct fracture systems with the approximate orientation of fracture cleavage.

Joints. No systematic study of joints was attempted here as it seemed to offer little help in analyzing the movement pattern of the area. The most consistent jointing occurs in the carbonate and sandstone rocks of the Rutland syncline and consists of surfaces spaced several feet or more, orientated subnormal to trend of bedding and fold axes. These are the *ac* joints or cross fractures (Fairbairn 1949, p. 156) and apparently form in response to tension or elongation parallel to fold axes. They are conspicuously formed in relatively competent beds in any lithologic succession, as for example in the massive dolomite and quartzite of the Rutland syncline, and in the quartzite and conglomerate of the basement and younger argillaceous sequences.

Due to the abundance of planar surfaces at a large angle to bedding (h01) surfaces with respect to fabric axes, the existence of *bc* joints could not be determined. Elsewhere in this type of fold belt they are apparently common (Thompson, 1950, p. 107; Osberg, 1952, p. 94; Broughton, 1946, p. 14).

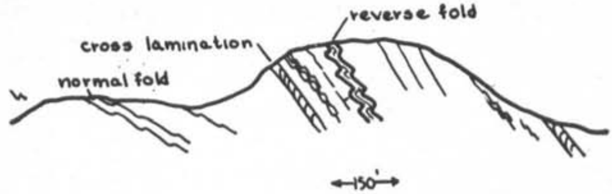
Rotational Features

Folds. Folds are one of the most obvious and widely used structural elements, ranging in useful scale down to microscopic dimensions. The essential elements of folds are the axis, axial plane, overall movement sense, wave length, and amplitude of fold and attenuation of limbs. Folds and crinkles (small-scale folds) generally are symmetric and similar (Billings 1942, p. 50) in the area with attitude of axial plane and movement sense consistent with their location on the flanks of larger structures. Up west is the movement sense on the west limb, and up east on the east limb of the Green Mountain anticlinorium and Pine Hill anticline. Axial planes converge downward across the area, dipping moderately east in the western part of the area and nearly vertical to the east. This is the "normal anticlinorium" (Idem p. 51). Isoclinal folding is uncommon in the Cambro-Ordovician but frequently observed in the basement complex where it is usually impossible to ascertain the movement sense. Small recumbent folds are found in the drag limbs of folds in the Pine Hill anticline (Plate 2). In fold crests minor folding may have no consistent asymmetry. This has been termed incompetent folding and is probably due to the complex readjustments which are forced upon incompetent beds in the folding of a sequence of mixed lithology.

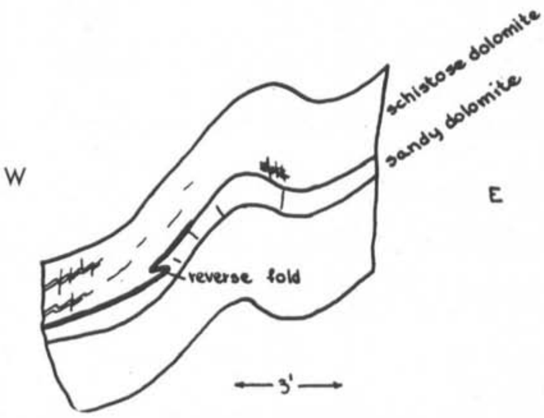
Many have observed that differences in fold geometry appear related to differing lithology, and hence to relative competency (For example: Heim, 1878, p. 40; Dale 1896, p. 564; Osberg, 1952, p. 79; Thompson, 1950, p. 94). In a folded sequence of mixed lithology, units which are relatively thick-bedded or free of micaceous material describe simple open folds of larger wave length than do the thin-bedded schistose units which have tiny compressed folds of an intricate pattern. The thin-bedded or micaceous beds can apparently change shape more readily, for their close folding shows a distinct thickening and thinning at crest and limb respectively of major folds; more massive beds, on the other hand, show only slight change in thickness (Figure 10, a).

Whereas most of the folds in the area, regardless of size, have axial lines which agree with the regional gentle north fold plunge, there are important exceptions. In the northeast corner of the area, and within the Saltash structural unit folds plunge moderately to steeply east at a large angle to the regional folds. In the first case these folds (approximately in *a*) have been ascribed to a second deformation with axis of maximum shortening normal to that of the major deformation (Hawkes, 1940, p. 118). This explanation is not yet supported by adequate field work. In the Saltash unit, however, an alternate suggestion for the evolution of the prevalent down-dip folds was advanced. It is believed that converging movements in a north-south direction were accompanied by the normal east-over-west passage of rocks on the east limb of the anticlinorium. Limited shortening in the B axis of the ellipsoid could be traced to the presence of marginal resistant masses in the basement immediately underlying and directly affecting the Saltash. Lack of confirmatory evidence in the basement structure appears to preclude a second deformation in this case with strain axes normal to the regional deformation.

Much of the microscopic "streaming" observed at a large angle to fold axes in the foliation surfaces appears to be a corrugation or gentle undulation of the micaceous surfaces (Plate 12). These are in a sense *a* folds. In their most conspicuous development the folds sheath the surfaces of stretched pebbles and quartz rods in which their form is symmetrical about the *ac* plane. These down-dip corrugations are again



(a) Folded Dolomite and Sandstone of Monkton Quartzite showing Bedding, Normal and Reverse Folds and Cross Lamination. Large outcrop 1¼ miles south of Clarendon Village.



(b) Drag Fold in Monkton Dolomite showing Small Normal and Reverse folds. Mill River at route 7, Clarendon.

Figure 11. Reverse Folds in Dolomite-Sandstone.

believed due to movement parallel to the axis or corrugation, or *a;* as poles to the mica in the surfaces would describe an incomplete *a*-axis girdle this would be in a sense movement normal to a girdle. It seems clear that the girdle has not arisen through rolling about the *a* axis, but rather as a consequence of the mechanical roughness of the slip surfaces in which projecting granular materials produced a slickensiding corrugation in a micaceous folia, in one direction of slip. It is furthermore held by some that subsidiary folds normal to principal folds (as found here) are entirely compatible with a single act of deformation (Kvale, 1947, p. 248; Strand, 1944, p. 22, 25).

In order to explain certain anomalies of fold pattern, such as those found locally in the Rutland area, Bain has proposed a theory of flowage folding (Bain, 1931) in which analogy was made to the movements of hot tar from the crest of a road. Or, as he writes, ". . . . due to load on the sides of . . . rising areas slumping into . . . sinking ones." (p. 525). In spite of the shallow source for such movements, he ascribes the resulting deformation to the "zone of flowage" (p. 521, 523) implying, in the application to marble, the importance of processes other than hydroplasticity. The need for such an ingenious mechanism arises in attempting to explain minor folds whose movement sense is not consistent with their position in a folded structure. Examples are shown in Figure 11. A fold pattern opposite to that of the normal pattern of drag folds is understandable if the action of slumping into synclinal troughs is assumed.

It is unfortunate that reversed folds have been seen only in carbonate rocks, where consistency of structural features is not the rule. Examination of these structures has not as yet provided an explanation of their formation which is satisfactory in all respects.

The hypothesis of flowage folding is tempting but its acceptance introduces serious difficulties. Fundamentally the deforming movements in flowage folding are the direct reverse of those which have unquestionably existed as the result of the arching of a layered sequence. Therefore the formation of flowage folds and normal drag folds (now adjacent in the rocks) must have taken place at different times in the structural history. Flowage folds might have originated late in the period of sedimentation (penecontemporaneous deformation, Shrock, 1948, p. 467) or during lapses in or following the principal tectonic events which have produced the major structural pattern. The early origin appears unlikely since flowage folds are sometimes found in highly deformed marble, and express a trough-ward yielding even in small folds obviously occurring later and not related to early history. Late formation is indicated, as Bain suggests, but it seems very unlikely however that gravity slumping would be a sufficiently powerful means of deforming rocks already rendered dense and crystalline. As yet no refolding of either reverse or normal drag folds by the other has been observed; such a relationship might further define the casual movements. At this point the origin of reverse drag folds remains as one of the unsolved and perplexing features of carbonate deformation.

Rotated Porphyroblasts and Pebbles. Porphyroblasts and suitably

Figure 1. Pyrite porphyroblast from black phyllite; pressure shadows of quartz indicate counterclockwise rotation; Grahamville formation, Sherburne. (crossed Nicols) x 20.

Figure 2. Albite porphyroblasts in quartz-sericite-chlorite schist. Planar inclusions of quartz and sericite are aligned nearly normal to foliation. Grahamville formation, W. Bridgewater. (crossed Nicols) x 10.

PLATE 9, PORPHYROBLASTS.

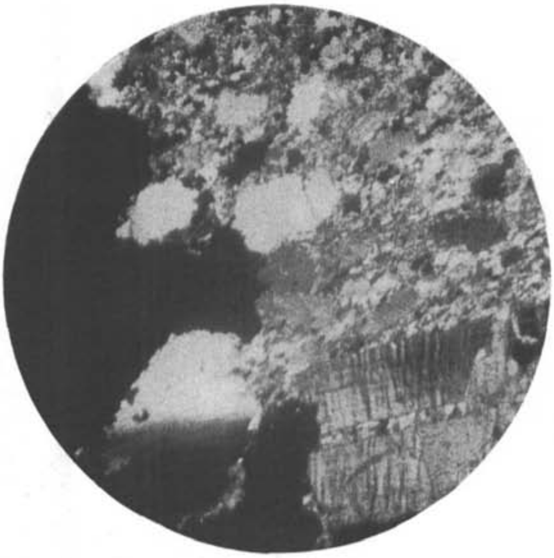


Figure 1. Broken microcline pebble from Nickwacket grit of the Mendon formation. Upper fragment has been moved to the right with respect to the lower. Section is approximately normal to foliation and fold axes. Coxe Mountain, Pittsford. (crossed Nicols) x 10.

Figure 2. Spirals of chlorite, tourmaline and opaque minerals within albite porphyroblasts in upper Pinney Hollow schist; section normal to fold axes. Rotation of about 90 degrees has taken place during growth of the albite. The typical relationship of slip cleavage and porphyroblast rotation is also shown. Sherburne. (crossed Nicols) x 37.

PLATE 10, ROTATION SENSE FROM PEBBLES AND PORPHYROBLASTS.

fractured pebbles commonly preserve certain components of the deforming movements of the enclosing rocks, while providing fundamental information about the time relation of metamorphism and deformation. Their significance has been noted by Mügge (1930), Heim (1921), Dale (1902), Becke (1924), Schmidt (1918) and many others.

In thin section albite, magnetite and pyrite often show trailing streamers of ribbon quartz, the arrangement indicating a rotational movement (Plate 9, Figure 1, Plate 11, Figure 2). Albite furthermore may show an S-shaped internal arrangement of inclusions which is assumed to indicate a rotation coincident with growth, (Mügge, 1930, p. 32) the nature of the internal arrangement indicating the rotation sense (Plate 10, Figure 2, Plate 11, Figure 2). In rare cases fractured pebbles indicate the rotation sense of the fracturing movements (Plate 10, Figure 1). None of these rotated materials exceeds a millimeter in diameter, and hence are only visible in polished section or thin section. The rotation axis is approximately *b* or the fold axes.

Albite with round or elliptical cross section very commonly encloses quartz, sericite, chlorite, graphite and ilmenite inclusions arranged in planes within the albite grain (Plate 9, Figure 2). These planes are at a large angle to bedding foliation and appear to be orientated uniformly throughout the rock as (h01) surfaces with reference to the fabric axes *a*, *b*, and *c*. The inclusion planes pass intact through (010) twin boundaries and in a single instance cross the boundary of adjacent albite grains. These inclusion planes define the intraformational rotation of some of the albite porphyroblasts, with the innermost segment of the surviving surface at angles up to 90 degrees with the outermost. The radius of curvature typically decreases outward but this could reflect changes in either of two independent functions, the velocity of shear, and the rate of growth of the porphyroblast. Rotation during growth is occasionally followed by rotation in which the albite acts exactly as pyrite and magnetite, and develops delicate exterior pressure shadows of granular quartz and chlorite (Plate 11, Figure 2) similar to those described by Mügge (1930, p. 48).

It was discovered that the rotation sense of porphyroblasts was not entirely reliable. Thus a case was noted in which internal and external features indicated an opposite rotation (Plate 10, Figure 2), another in which adjacent grains had differing internal spirals and finally, an example of quartz growth which had replaced part of a folded schist (Plate 10, Figure 1) retaining traces of S-shapes similar to those assigned

Figure. 1 Quartz replacing folded mica in sericite-chlorite schist. Saltash formation, Sherburne Village. (crossed Nicols) x 62.

Figure 2. Albite porphyroblast with external quartz pressure shadows and internal spirals of quartz inclusions. Movement sense is not in agreement. Grahamville formation, Sherburne. (crossed Nicols) x 10.

PLATE 11, PORPHYROBLASTS.

to the completely different mode of origin in the albites. In the majority of the rolled structures, however, rotation-sense agreed with that indicated by minor folds and map pattern. Caution should nevertheless be used in the application of these features and the possibility of origin of internal spirals by replacement borne in mind.

Certain aspects of rotated porphyroblasts have fundamental significance in rock deformation and metamorphism. The fact that grains in a deforming rock do apparently rotate about *b* is significant (Fairbairn, 1936). In addition the fact of growth during shearing movements precludes the possibility that porphyroblasts form only in rocks by recrystallization under compressive stress normal to bedding, or "after flowage has ceased" (Leith, 1905, p. 182) (Leith and Mead, 1915, p. 111, 180). Finally it is clear that deformation and grain growth are processes of considerable duration,[1] and should in no sense of the word be considered catastrophic (pointed out by Becke, 1924, p. 198).

Linear Features

Linear elements consist of (1) elongate pebble and quartz pods, mineral lineation, (2) streaming and corrugation subnormal to fold axes, and folds, crinkles, boudinage and (3) the intersections of foliation with fracture cleavage and slip cleavage in a direction parallel to fold axes. These features are shown in Figure 12 and described in some detail below.

Streaming, Mineral Lineation, Corrugation Subnormal to Fold Axes. Distinctly visible lineation commonly occurs on foliation surfaces in a direction sub-parallel to the dip (Plates 12, 13). It consists of tiny ridge-and-furrow surface irregularity, strewn out mineral fragments and less commonly elongate biotite, chlorite and opaques; dark streaks in the foliation resolve microscopically to broken, scattered dark mica flakes and the slickensided surface texture to elongate sericite flakes in tiny undulating discontinuous folds. This linear structure parallels the direction of elongation of stretched pebbles and is approximately normal to the rupture line in boudinage and to the axis of rotation of rolled porphyroblasts. It apparently represents the *a* direction or direction of tectonic transport in these rocks. This lineation is rarely repeated in

[1] A complex case illustrates this: Innermost straight inclusions in an albite grain suggest early growth without deformation; an outer spiral signifies later rotation during growth. Cessation of growth with continued rotation is next seen in external pressure shadows on the albite grain, and final slight movements are reflected in the mild straining induced in the quartz of the pressure shadows.

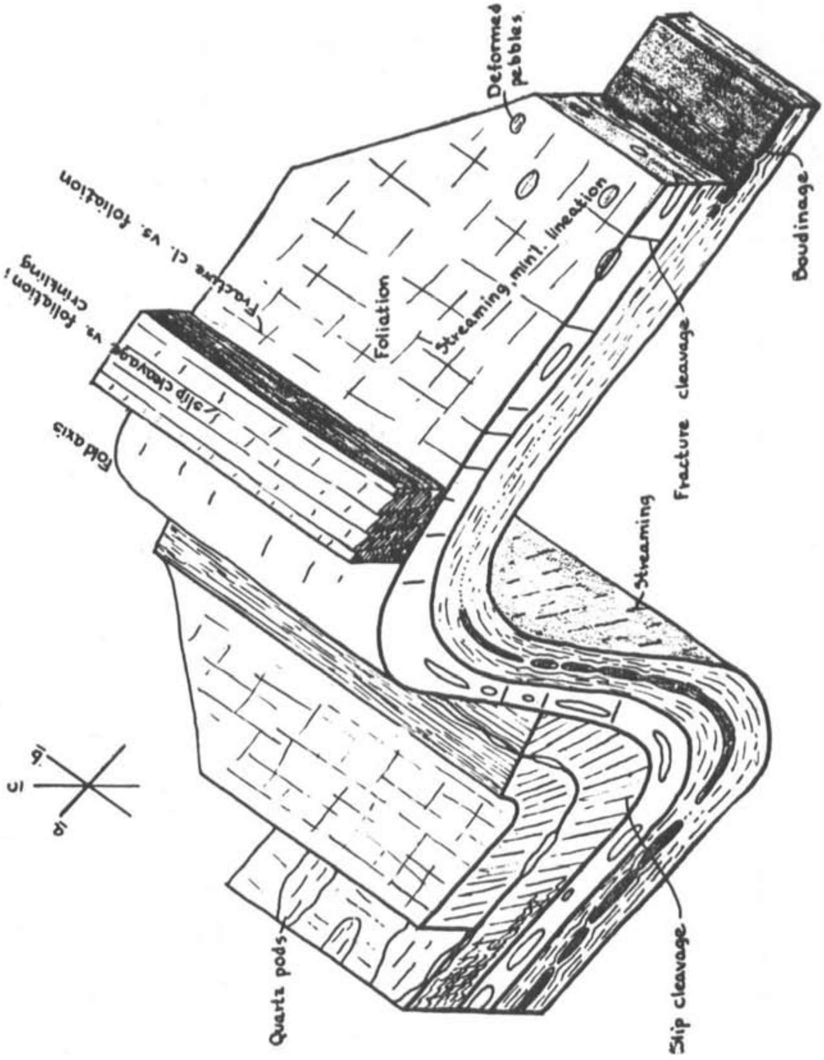


Figure 12. Block Diagram illustrating Minor Structural Features.

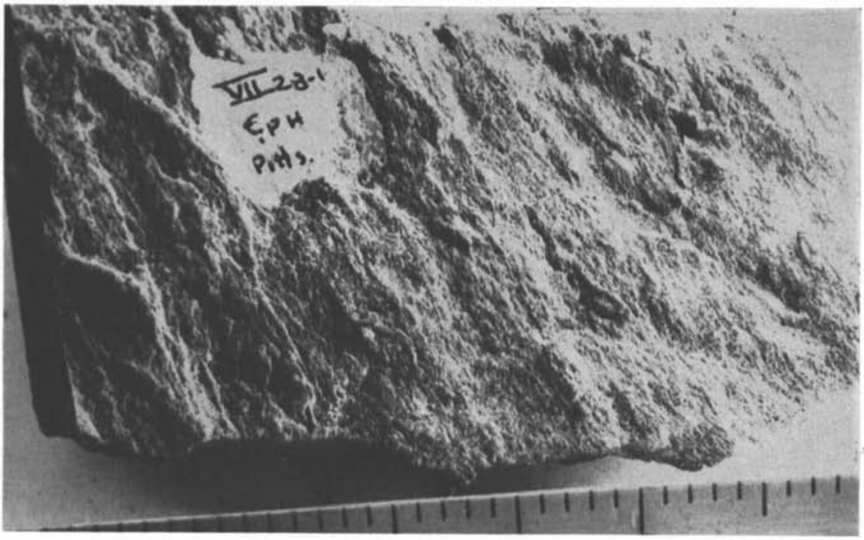


Figure 1. Lineation in the foliation plane in Pinney Hollow phyllite. Small folds (*b*) incline gently to the right; mica elongation (*a*) inclines steeply to the right. Pittsfield.

Figure 2. Lineation in the foliation plane of a stretched conglomerate from the Tyson formation, Dry Hill, Plymouth. The axes of crinkles are horizontal; streaming, crenulation incline to the left in a direction parallel to the elongation of the stretched pebbles.

PLATE 12, LINEATION.

Figure 1. Side view of small fold in Pre-Cambrian quartzite from Mendon. The folda xis parallels the upper left-sloping surface of the specimen, and streaming is nearly vertical. On both limbs, streaming $\wedge$ fold axis is 60°.

Figure 2. Grit, saltash formation, Sherburne. Cut section on upper surface of specimen is foliation. The lineation visible is due to parallel arrangement of chlorite and opaque mineral groups, and sericite elongation. Folds nearby are abnormal and have axes parallel to this lineation.

PLATE 13, LINEATION.

more than one direction in the foliation, suggesting that at least during the stage of deformation when movement was recorded this movement was undirectional.

Fold axes and subnormal lineation commonly occur together in foliation surfaces, and their angular relationship could be easily determined. The two directions are rarely perpendicular, but intersect at angles of 55 to 90 degrees (see also Thompson, 1950, p. 91; Osberg, 1952, p. 90). Since these directions have been assumed to represent the *b* and *a* fabric axes, this relationship is rather startling as *a* is always considered perpendicular to *b* (Fairbairn, 1949, p. 6; Cloos, 1946, p. 5). The explanation is that flexural folding of a doubly plunging anticline requires a limited amount of slip parallel to its axis (Figure 10, b). Considering segments of a limb on such a structure, the differential movement in folding would consist not only of the customary slippage normal to the fold axis, but slight slip parallel to it, with upper beds moving in the direction of the longitudinal crest of the fold. The combination of these components gives an overall slip direction inclined to the fold axis; variation from perpendicularity with the axis is in the direction of the fold plunge, its magnitude dependent upon the relative intensities of folding parallel and normal to the structure. Thus in eastern part of the map area fold axes plunge gently north and streaming plunges moderately northeast with the angle of intersection in the foliation about 60 degrees. This is in agreement with the regional north plunge of the Green Mountain anticlinorium. Elsewhere in small folds with south or very small plunge, streaming plunges southerly or is normal to fold axes respectively.[1] In conclusion, the critical fabric axis *a* (Fairbairn, 1949, p. 5) should not be assumed to be perpendicular to plunging fold axes. As indicated here *a* may quite validly be inclined to these axes, which is indeed the case with streaming and mineral lineation in the Rutland area.

Boudinage. Boudinage (Cloos, 1946, p. 17) occurs in a few places in the basement complex and in the carbonates of the Rutland syncline, but neither environment permits accurate determination of the local

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[1] Kvale has described widespread lineation, apparently a streaming and mineral elongation which is inclined to fold axes at angles other than 90 degrees. The writer is unable to agree with Kvale (1947, p. 54) that lineation not normal to fold axes must be older than the folds. This possibility could be readily examined as follows: Paper is placed on the surface of a fold and linear elements traced from both limbs of the fold. When the paper is unfolded to a flat surface the relation of traced segments could be examined. These were never observed to fall on a straight line, as would be expected if the lineation had been an earlier feature which had undergone later folding.

movement picture. The general tendency is for the rupture line of the boudin to parallel fold axes, which would indicate stretching subnormal to fold axes. As Thompson points out (1950, p. 111, 120) boudinage is a tension phenomenon with the relatively competent, more brittle beds yielding to tension failure, permitting the less competent surrounding material to flow into the opening. He has utilized this principle in establishing a scale of competency for rocks in the Ludlow area.

Pebble and Quartz Pod Elongation, Rodding. Pebble and quartz pod elongation together with rodding are linear elements in foliation surfaces and are oriented subnormal to fold axes and parallel to streaming.

Rodding is the term applied to rods, pencils and other column-shaped structures. These are generally composed of folded thin quartzite beds or veinlets of quartz and are peculiar to the areas of strong down-dip folding (discussed above under "Folds"). Rodding develops there from the intersection of various planar features with folds in a direction parallel to the fold axes.

Conspicuous thin quartz veinlets and folded quartz pods are abundant throughout the argillaceous eastern sequence and in the Moosalamoo and Hortonville phyllites. These discontinuous, somewhat shapeless masses parallel the foliation and are occasionally elongate in a direction subnormal to fold axes. The pods may represent dismembered thin quartzite beds, but based on comparable quartzite deformation elsewhere boudinage should be evident. In all probability the pods and lenses formed as such during metamorphism and deformation (See "Metasomatism").

Pebbles[1] from all of the basal Cambro-Ordovician units are deformed into ellipsoids. The long and intermediate axes of the shapes are invariably in the foliation surface with the long axis subnormal to fold axes and parallel to streaming (Plate 12, Figure 2). Some idea of the initial condition of these pebbles is given by Osberg from his discovery of relatively undeformed samples in the trough of a syncline in the East Middlebury area (1952, p. 90). These have a shape index which is similar to that of beach and glacial gravels (Pettijohn, 1949, Table 49, p. 201). The present shape of larger pebbles is approximately that of

[1] The term "pebble" denotes all distinctly clastic grains larger than a quarter inch. A diametric classification might be preferable following the practice of sedimentologists but, as the materials are all deformed, mean diameter is not always an obvious feature. As the term "pebble" seems to be commonly used elsewhere in the description of deformed conglomerates, this usage is accepted here.

ellipsoids with ratios of major and minor ages ranging from 2 to 6. In a strict sense variance from ellipsoidal shape appears in lenticular cross sections, pointed ends and ends of different curvature. These ellipsoidal pebbles down to an inch in mean diameter lie with major and mean axes parallel to banding and foliation even in the noses of folds. At North Sherburne and Bear Mountain, Wallingford, tight folding has produced canoe- and S-shaped forms from the stretched grains. The grit matrix of the conglomerates is often strongly crinkled with development of slip cleavage; this axial plane cleavage nowhere parallels the elongation direction of the pebbles but is truncated against the pebble surface. Smaller pebbles commonly show less deformation than large ones, but even those visible in thin section have lenticular shape with major-minor axial ratio of about 1.3. Microcline grains up to half an inch in size are often highly angular and are broken and strewn in the foliation, contrasting with the quartz deformation in showing none of the typical undulose strain phenomena (Plate 10, Figure 1).

Neither time nor exposure permitted the statistical study of pebble dimensions in the area, nor was it possible to investigate thoroughly the relation of deformation to grain size and composition of pebble. The quartz and mica fabric of deformed conglomerates will be described elsewhere by the writer.

In view of the significance of stretching in the study of rock deformation, it is of some value to compare the meagre data from this area with that from certain other localities. Pebbles (Oftedahl, 1949, p. 477; Thompson, 1950, p. 104) and fossils (Heim, 1878, p. 61) are commonly flattened or elongate in bedding or in a foliation parallel to bedding. Yet this is not universal for pebbles (Fairbairn, 1936) and ooids (Cloos, 1947) may have major and intermediate axes in an axial plane foliation. The orientation of the long axis is even more variable than the planar relationships. In contrast to pebbles from the Rutland area Strand (1944, p. 17), Fairbairn (1936) and Kvale (1947, p. 28) report pebble elongation parallel to fold axis, or fabric *b*. Elongation in *a* is described in Vermont by Thompson and Osberg (1952, p. 90), in Norway (Oftedahl), in Maryland (Cloos) and in the Alps (Heim). It is clear then that conglomerate deformation of the Rutland area is easily matched elsewhere with regard to the orientation of pebble ellipsoids. The explanation of this pattern of deformation remains.

The observation that the pebbles lie with long and intermediate axes in the bedding plane may have considerable significance. In a medium undergoing change of shape by simultaneous slip on parallel surfaces, a spherical body should theoretically change to an ellipsoid with long axis inclined to the slip surfaces. This is, of course, the concept of the strain ellipsoid, and it will be recalled that the AB ellipsoid planes should parallel axial planes of folds (Billings, 1942, Figures 183, 184, 189). Now in the stretched pebbles an approximation to ellipsoidal shape is produced from forms which were originally more nearly spherical. The "AB plane" of the pebbles is invariably inclined to axial planes and is parallel to foliation which has elsewhere served as a surface of shearing movement. Elongation of pebbles has taken place *parallel* to the slip surfaces, the movement on which accounts for nearly all of the other structural features of the area. This elongation parallel to slip surfaces in the surrounding rock might be explained by a combination of slip and rotation, or by the assumption of a deforming mechanism for the pebbles different from that of the enclosing incompetent rock. In explaining a Norwegian occurrence similar to that found in the Rutland area, Oftedahl (1948, p. 484) believes that ellipsoidal form is due to homogeneous slip (in the manner of the strain ellipsoid analogy) followed by rotation of the ellipsoidal pebble within the rock matrix into parallelism with the slip surfaces. Now rotation of this sort should leave its mark. The pebbles of the Rutland area lie in a foliate matrix with weak banding of granular and micaceous minerals and this banding in its parallelism with distinct quartzose and limy beds is considered primary. The banding is sufficiently distinct so that its course can be traced through delicate folds and slip cleavage. The pebbles do not appear to have locally disturbed these delicate surfaces in the process of becoming elongate parallel with them although rotation would have been of the order of 30 degrees. Therefore Oftedahl's hypothesis of slip and subsequent rotation is not considered applicable. A mechanism other than homogeneous slip on parallel surfaces must have been a ruling factor in the development of pebble elongation here. It is tentatively suggested that flattening normal to foliation has been important.[1] It may be possible to trace this to some purely mechanical process or to a load recrystallization, either of which has produced maximum shortening normal to foliation, and maximum elongation in the foliation in the direction of maximum slip in the surrounding rocks.

[1] The non-rotational quartz overgrowths on magnetite and pyrite from this study and elsewhere may be recalled. These must be considered due to flattening, and an indication that such movements can occur in these rocks.

General Considerations of the Fabric

Certain features of the fabric of the rocks of the Rutland area deserve mention in that they can often be used to relate deformation to metamorphism. In general these processes must have been simultaneous although in detail they were probably of somewhat intermittent character and of varying degrees of intensity aerially and in time.

A feature termed "Y-fabric" universally characterizes the fabric of the Cambro-Ordovician and marginal parts of the Pre-Cambrian basement complex. It consists of grain boundary intersections at about 120 degrees in quartz and carbonates, a Y pattern being formed at such intersections of three grains. From the manner in which it occurs Y-fabric is regarded as a recrystallization phenomenon. It is least well developed in the massive flat-lying dolomite of the Rutland syncline where recrystallization appears to have advanced slightly beyond the secondary enlargement stage. In the basement quartzite Y-fabric is restricted to thin intergrain zones in which clear equant quartz contrasts with angular undulose fragments of original quartzite; here crushing was probably followed by annealing of the crushed intergrain quartz. Y-fabric is widespread in the younger rocks containing quartz and carbonates. The quartz matrix of intensely deformed conglomerate, and the dolomite in tight folds rarely show the anticipated features of strain such as lamellae, twinning, variable extinction, variable grain size and cataclastic structure. In this regard the contrast with Pre-Cambrian rocks may be traced to the fact that the younger materials were originally finer grained and would consequently have a greater tendency to recrystallize than their counterparts with texture already coarser than that which the younger rocks would develop. In regarding Y-fabric as the product of annealing, as suggested by theoretical studies of metals (Harker and Parker, 1934), it is clear that this annealing must have continued beyond the final stages of deformation. This suggestion that metamorphism (here regarded as recrystallization) as a general rule extended beyond the important movement phase in the rocks agrees, of course, with the occurrence of undeformed chloritoid and biotite porphyroblasts.

Weiss uses the term "polygonal arches" (1949, Plate 2) to describe another feature appearing here (Plate 8, Figure 1). These are tiny folds which are defined by tabular grains of mica. Recrystallization has apparently erased bending and other traces of grain irregularity which undoubtedly existed as a result of the tight folding. The arches are apparently associated with slip cleavage (writer's interpretation of Weiss, idem; Ch'ih, 1950, Plate 2, Figure 6) but give no indication of mica recrystallization parallel to this axial plane feature.

Although recrystallization has apparently succeeded the major stage of deformation and produced "annealed fabric," lenticules and augen testify to earlier crushing and granulation (Knopf, 1931, p. 16, Harker, 1932, p. 167). Lenticules and augen-structure are widespread in the basement gneiss but are poorly preserved in Cambro-Ordovician grits and quartzite. These structures range from microscopic size to an inch in maximum dimension. Their formation seems to require a predominantly granular lithology, with a few per cent of micaceous minerals. Quartz and feldspar fragments typically form the core of the lenticule or augen composed of equigranular small grains of the same meatrial. Thin films of mica and opaques separate the lenticules. With the increase of mica in the rock these structures are seldom formed. Their significance again lies in the indication of thorough recrystallization following the principal period of deformation.

Summary of Minor Structural Features

Minor structural features are in a general way compatible with the major structure. This compatibility is expressed in the agreement of rotational features with the mechanical requirements of larger scale flexural folding, and in linear features by agreement with either the regional axis of folding, or with the presumed direction of overall tectonic transport.

Compositional banding and the principal rock foliation consistently parallel bedding although these planar features may have been accentuated during deformation and metamorphism. Slip cleavage, fracture cleavage and joints relate to an advanced stage of metamorphism.

Major folding has produced subsidiary minor folds which are asymmetric, overturned or recumbent. At an advanced stage of the folding of argillaceous rocks slip cleavage was developed parallel to fold axial planes. The movement sense associated with slip cleavage is often ambiguous but is believed to be always compatible with the movement on larger associated fold limbs. Flow cleavage was not detected, but the name might readily be applied to certain varieties of slip cleavage.

Rotation and flattening is locally expressed by internal spirals in albite, and by external overgrowth of quartz and chlorite on albite, magnetite and pyrite. The reliability of these rotational features is somewhat weakened by local disagreement of internal and external patterns.

There are two principal systems of lineation. Crinkles and folds and the intersections of foliation with fracture and slip cleavage are parallel to the direction of regional fold axes. Elongation of pebbles, streaming and crenulation are in a direction subnormal to fold axes. The regional movement pattern indicates that the latter is the *a* fabric axis. Divergence from perpendicularity of these assumed contemporaneous lineation systems is explainable by examination of the mechanism of doubly plunging folds.

Quartzite pebbles and rods are elongate subnormal to fold axes with mean and major axes in the foliation (slip surface in the surrounding rocks). The deformation is ascribed to flattening in the foliation rather than to uniform shear on surfaces parallel to foliation.

Y-fabric, polygonal arches of mica, and relics of lenticular and augen structure reflect widespread post-deformational recrystallization or annealing in the younger rocks. With the exception of chloritoid and biotite porphyroblasts grain growth has taken place parallel to foliation even in the limbs of small folds. Growth in fold crests may produce larger grains than on the limbs, but even here new mica grains retain parallelism with bent foliation.

In general terminology the deformation is characterized by flowage rather than rupture. Observable folds of all magnitudes have thickened crests and attenuated limbs; stretching and bending is common whereas microfracturing and faulting is rare. It is impossible, however, to assess the thoroughness of annealing in erasing early ruptural strain.

Slip on surfaces of foliation appears to be the fundamental mechanism by which most of the rocks acquired their present structural features. This slip has produced folds and a variety of linear features subnormal to them in all but the most competent strata. Conglomerate and quartzite have more strongly resisted bedding plane shearing, but appear to have yielded to flattening normal to bedding.

Time Relationship of Deformation and Metamorphism

The basement complex of the Green Mountains has undergone an episode of deformation and metamorphism clearly younger than the period of folding and intrusion which produced its overall trend and mineralogic character. In all probability the younger deformation and metamorphism occurred simultaneously, although mineralogic changes appear to be more widespread than structural changes.

In spite of the occurrence of at least two orogenies of significantly differing ages, the recorded mineralogical and structural changes conform to a single general pattern. Likely reasons for this situation are that the younger episode either completely erased all earlier mineralogic and structural features, or that both episodes were entirely similar in nature of metamorphism and in movement pattern. From structural considerations the latter is more probable.

Deformation and metamorphism of Cambro-Ordovician rocks were nearly simultaneous processes, but it has been frequently emphasized that in a detailed sense the processes have been intermittent, overlapping and of considerable duration. Larger structures have been folded, refolded and faulted. Porphyroblasts have grown, been rotated intermittently, with rotational deformation often succeeded by flattening. Early porphyroblasts have been disrupted while the youngest grew independently of rock structure. Neither metamorphism nor deformation therefore can be considered catastrophic or widely separated in time.

CONCLUDING STATEMENT

In view of the current opinion that the Taconic rocks were thrust westward to their present location the writer sought to discover evidence of large-scale thrusting in the Rutland area. This was found neither within the younger rocks nor within the basement. Deformation by thrusting was generally not important in the area.

Inconclusive evidence prevents a definite correlation of Cambro-Ordovician sequences on the east and west limb of the Green Mountain anticlinorium; this is a matter which bears indirectly on the origin of the Taconic range. However, it is suggested that the Tyson and younger formations are of Ordovician age and that the Cambrian is solely represented in the eastern part of the area by the thin unconformable wedge of Saltash formation.

Preliminary study of the Pre-Cambrian basement proved remarkably fruitful. At least 7,000 feet of gneiss, schist, quartzite, amphibolite and marble of an older metamorphic complex show the effects of a second younger metamorphism similar to that which affected the Cambro-Ordovician rocks. Early structural trends are preserved within the center of the anticlinal core of the Pre-Cambrian and are separated from those of the younger rocks by a complex marginal zone. Interaction of core and mantle in this zone has, on the one hand, developed younger trends parallel to older banding and, on the other, a near-conformity of older trends with the mantle contact. Relatively resistant units in the basement have distinctly guided the structural development of the younger rocks, producing trends apparently anomalous with the Cambro-Ordovician movement pattern.

Minor structures conform with the local movement pattern in larger scale folds, which themselves are consistent with the anticlinal nature of the Green Mountains and Pine Hill anticline. The development of these features is ascribed to the action of a couple, eastern rocks moving westward over western rocks.

APPENDIX

Rock Nomenclature

Field terminology is used since it is simply descriptive and should convey to the reader the field characteristics which led to the subdivision of the map units.

foliation: that property of a rock which permits it to split along parallel surfaces whose position is controlled by the orientation of platy or micaceous grains.

banding: layering of material of varying composition or grain size.

schistose: having good foliation as a consequence of abundant mica or other platy grains.

granulose: having a granular fabric, generally lacking good foliation, often poor in mica.

schist: foliate rock made up of minerals visible in hand specimen and consisting of at least 30 percent mica.

phyllite: a fine-grained schist, minerals being unidentifiable in hand specimen.

grit: a granulose rock consisting mostly of quartz and feldspar.

marble: a carbonate rock showing crystallinity in the hand specimen.

gneiss: a medium or coarsely grained feldspathic rock with distinct banding.

augen gneiss: a gneiss with feldspar in "augen" or eyes a quarter inch or greater in smallest dimension.

greenstone: a granulose rock, typically greenish, containing epidote, chlorite and carbonates as principal minerals.

amphibolite: a dark weakly schistose rock containing at least 40 percent dark amphibole and varying amounts of plagioclase and garnet.

graywacke: a granulose thick-bedded rock predominantly of quartz and feldspar with more than 15 percent of camafic minerals such as chlorite, biotite, ankerite or calcite.

lime silicate rock: a granulite, medium- to coarse-grained, composed of carbonates and more than 25 percent of phlogopite, tremolite, talc, epidote, zoisite or grossularite.

Mineralogy

"Sericite" is abundant in nearly all of the rocks of the basement and flanking Cambro-Ordovician. Although this material may be principally muscovite, it is probably not free of the other white micaceous minerals, paragonite, pyrophyllite or phlogopite. Sericite is commonly red-stained or greenish, with optic angle of 20 to 30 degrees. Chlorite, shows great variety in thin section but few trends could be discerned in the association of a variety of chlorite with particular lithologic types. Chloritoid also varies in color and pleochroism in the different occurrences. While typically pleochroic (moderate to strong green-blue-yellow) chloritoid from Clarendon Gorge is nearly colorless. Winchell's tables (1951, p. 262) of composition-optical properties have been used exclusively in the determination of feldspar; extinction angles, optic angles and approximate refringence are the optical properties used. The ratio of calcite to dolomite was determined by a copper nitrate stain (Rodgers, 1940); accurate determination was not made of the remaining minerals; garnet, biotite, epidote and accessories.

Mendon and Cheshire Formations

1. Road near Sanitorium, S end of Coxe Mountain, Pittsford.
2. E slope Pine Hill elevation 900 feet; pebbles of quartzite up to 1.0 millimeter in diameter.
3. E end of Clarendon Gorge of Mill River, Clarendon; chloritoid as tiny colorless laths; "misc." is largely red iron oxides from the weathering of ankerite.
4. NW side of East Mountain, Mendon, elevation 1040 feet.
5. SW end of Long Hill, Pittsford; banding 1–2 millimeters thick.
6. Ridge NE of Chippenhook in Clarendon, elevation 1410 feet; banding 1–2 millimeters thick.
7. SW end of Long Hill, Pittsford.
8. Average of 10 samples of dolomite, Pine Hill and Green Mountain Front.
9. Average of 2 samples NW of Rutland; "Miscellaneous" in samples 4.5 and 6 refers to dark dust of unknown mineral(s); elsewhere it refers to sphene, zircon, apatite and red iron oxides.

Saltash and Tyson Formations

1. 1.2 miles S of Sherburne, 1650 feet elevation.
2. 0.4 mile SE of the summit of Smith Peak, 2700 feet elevation; albite in rounded porphyroblasts 1 millimeter.
3. Sherburne Village.
4. E SE of North Sherburne, 2,000 feet elevation.
5. Falls on Sherburne-North Sherburne road.

6. 1470 feet elevation, fork Townsend Brook, Chittenden; the base of the Tyson; miscellaneous includes black dust of unknown minerals.
7. Average of four dolomites, West Bridgewater to Plymouth Union.
8. 1 mile SE of South Pond Chittenden.

Grahamville and Pinney Hollow

1. Lower E slope 0.5 mile north of north end of Woodward Reservoir; albite as porphyroblasts 0.7 mm in diameter.
2. Dolomite of Plymouth member, average of 6.
3. Quartzite of Plymouth member, average of 3.
4. Lower E slope 3.2 mile S of Sherburne.
5. Lower albitic schist; 2.3 miles NNW of Sherburne, 1580 feet elevation; albite as round porphyroblasts up to 1.0 Millimeter in diameter.
6. Middle chloritoid-bearing phyllite; in brook 1.4 miles N of Sherburne, mod. pleoc chloritoid in tablets up to 0.2 millimeter in length.
7. Upper banded sandy schist; summit of East Mountain, Sherburne.
8. Lower black phyllite; 1.3 miles N of Sherburne in brook.
9. Greenstone from south band; 2½ miles N of Sherburne at elevation 1800 feet in brook.

Pre-Cambrian

1. 0.8 mile ESE of Northam, 2500 feet elevation; plagioclase, An_{30}, greenish-brown hornblende N_z about 1.680, $Z \wedge C$ 19 degrees.
2. 0.5 mile SE of North Pond, Chittenden, 2500 feet elevation; plagioclase An_{20} yellow-green hornblende, $Z \wedge C$ is 25 degrees.
3. On road 0.7 mile WSW of Chittenden Village; amphibole is colorless tremolite.
4. 0.9 mile W of Shrewsbury Peak, 3300 feet elevation; amphibole is colorless tremolite, epidote is low iron "clinozoisite," plagioclase: albite.
5. 1920 feet elevation, S slope Wilcox Hill, Shrewsbury; albite.
6. Ridge W of Mendon Peak, elevation 3240 feet; plagioclase is albite with central inclusions of clinozoisite.
7. S slope of Shrewsbury Peak, 2800 feet elevation; plagioclase is An_{10}.
8. 0.6 mile W of Comfois Hill, 1660 feet elevation; low iron epidote, plagioclase is An_{12}.
9. W summit of Mendon Peak; garnet and quartz in pebbly forms 2 millimeters in diameter.
10. Summit of Blue Ridge Mountain; chloritoid has moderate pleochroism, is twinned, in rosettes 5 millimeters.

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	schis-tose grit	*pebbly seri-cite-qtzite*	*pebbly seri-cite-qtzite*	*pebbly seri-cite-qtzite*	*dark qtz-ite*	*dark phyl-lite-qtzite*	*black phyl-lite*	*dolo-mite*	*quartz-ite*	
	1	2	3	4	5	6	7	8	9	
quartz	42	57	53	83	32	13	10	5	97	
albite	6				6	16	4	1	2	
microcline					1	6		tr	1	
sericite	36	29	28	15	36	48	55	2	tr	
chlorite	3	10	1				27			
biotite	1				17	16		tr	tr	
chloritoid			7							
ankerite	8	1	9		tr		tr	89		
calcite	4				1			2		
mag-ilm.	1		tr	1						
epidote	tr					tr				
tourmaline		2	1		tr	tr	1			
graphite					2	1	2			
misc.	tr	tr	1	tr	2	4	3	tr	tr	
TEXTURE	0.06	0.06	0.06	0.1	0.04	0.06	0.05	0.05	0.25	millimeter
STRUCTURE	sch.	gr.	gr.	gr.	gr.	phyl.	phyl.	gr.	gr.	

sch: schistose
gr: granulose
phyl: phyllitic

1–4 Nickwacket member
5–7 Moosalamoo member
8 Forestdale member
9 Cheshire quartzite

ESTIMATED MODES OF THE MENDON AND CHESHIRE FORMATIONS

	Limy grit 1	*albite grit*	*biotite grit*	*pebbly grit*	*Banded albite grit* 5	*black schist* 6	*dolomite* 7	*schistose grit* 8	*sericite graphite* 9
quartz	62	15	72	72	64	24	2	48	69
sericite	8	32	5	12	15	25	1	15	20
albite	15	32	15	6	12	30		9	
chlorite	7	15	4	7	5	15		10	2
biotite			2	3	2				
chloritoid									7
ankerite	1						96	9	1
calcite	6						1	7	
epidote	tr	tr	2						
mag.-ilm.		1	tr	1	1			1	
graphite						2			tr
misc.	tr	tr	tr	tr	tr	3	tr	tr	1
TEXTURE	0.06	0.05– 1.0	0.06– 0.5	0.08	0.08– 0.6	0.1– 0.3	0.06	0.08– 0.2	0.07 mm. 0.7
STRUCTURE	gr.	gr.	gr.	gr.	sch.	sch.	gr.	sch.	gr.

gr.: granulose 1–3: Saltash
sch.: schistose 4–9: Tyson

ESTIMATED MODES OF THE SALTASH AND TYSON FORMATIONS

	albite schist 1	*dolo-mite* 2	*quartz-ite* 3	*fine banded schist* 4	*albite schist* 5	*white phyllite* 6	*banded sandy schist* 7	*black schist* 8	*green-stone* 9
Quartz	10	5	80	68	42	20	75	72	15?
sericite	57	1	7	20	30	55	12	25	
albite	22	1	5	3	18		5		5?
chlorite	7				8	15	8		5
microcline			3						
biotite	1			7			tr	2	8
chloritoid						7			
dolomite-ank.		92	3						tr
epidote				tr					45
calcite		tr							20
mag.-ilm.	1		tr	1	1	tr			1
graphite								1	
misc.	tr	tr	tr	tr	tr	tr	tr	tr	tr
TEXTURE	0.08	0.04	0.1	0.06	0.1	00.50	0.70	0.80	0.05 mm.
STRUCTURE	sch.	gr.	gr.	sch.	sch.	phyl.	sch.	sch.	gr.

sch.: schistose
gr.: granulose
phyl.: phyllitic

1–4 are Grahamville formation
5–9 are Pinney Hollow formation

ESTIMATED MODES OF THE GRAHAMVILLE AND PINNEY HOLLOW FORMATIONS

	garnet-amphib. 1	*amphib.* 2	*tremo-lite grit* 3	*lime-sil. grit* 4	*Wilcox schist* 5	*micro-cline gneiss* 6	*plagio-clase gneiss* 7	*garnet gneiss* 8	*garnet quartz-ite* 9	*chlori-toid schist* 10
quartz	tr	10	tr	3	70	25	30	48	60	5
microcline						45				
plagioclase	25	20		5	10	15	40	25		
sericite				25	15	8	8	7	15	75
chlorite	5	20		30	4		7	3	15	
chloritoid										15
biotite	5	5				4	7	2		
talc			2							
garnet	5							7	7	
amphibole	60	40	96	15						
epidote	tr	tr	tr	20		2	7	5	tr	tr
sphene	tr	2	tr	tr	tr			tr		
graphite				tr					1	
magnetite-ilm.	tr	tr			tr	tr		tr		3
tourmaline		tr	tr		tr		tr			tr
TEXTURE	0.4	0.2–1.0	0.05–1.0	0.08–0.3	0.08	0.05–10.0	0.10–0.3	0.08–1.0	0.06	0.03–5.0 mm.
STRUCTURE	gr.	gr.	gr.	gr.	sch.	gn.	gn.	gn.	gr.	sch.

gr.: granulose
gn.: gneissose
sch.: schistose

ESTIMATED MODES OF THE PRE-CAMBRIAN

EXPLANATION

METAMORPHIC ROCKS

Middle Ordovician — Oh: Hortonville Slate. Black phyllite, blue calcite marble

unconformity

Ordovician (?) — O?: Gray, white calcite marble

Upper Cambrian — Ɛcs: Clarendon Springs Dolomite. Dark gray limy dolomite

Ɛda: Danby Quartzite. Thin bedded gray dolomite with glassy cross-laminated quartzites

Lower Cambrian — Ɛw: Winooski Dolomite. Pink, white, blue thin bedded dolomite

Ɛmo: Monkton Quartzite. Varicolored thin bedded dolomite; thick sandstone; red, black, green schist

Ɛdu: Dunham Dolomite. Thick bedded gray, pink sandy dolomite

Ɛc: Cheshire Quartzite. Massive buff vitreous quartzite

Ɛmm, Ɛmn, Ɛmf: Mendon Formation

Ɛmm: Moosalamoo member: dark quartzite, black fine banded phyllite

Ɛmf: Forestdale member: pink sandy dolomite

Ɛmn: Nickwacket member: thick graywacke, quartzite, thin schist, conglomerate

unconformity

Pre-Cambrian — pƐwc, pƐwb, pƐwa: Wilcox Formation

pƐwa: white, gray, green, black schist, thin buff dolomite

pƐwb: gneiss

unconformity

UNSTRATIFIED ROCKS

Miss. (?) — ib: Intrusive breccia, trachytic, andesitic porphyry

STRUCTURAL SYMBOLS

Contact, dashed where uncertain, gradational

Concealed contact

Thrust fault; T, upper plate, dashed where approximate

Reverse fault; D, downthrown block, U, upthrown block. Question marks indicate uncertainty as to existence of fault

Compositional banding

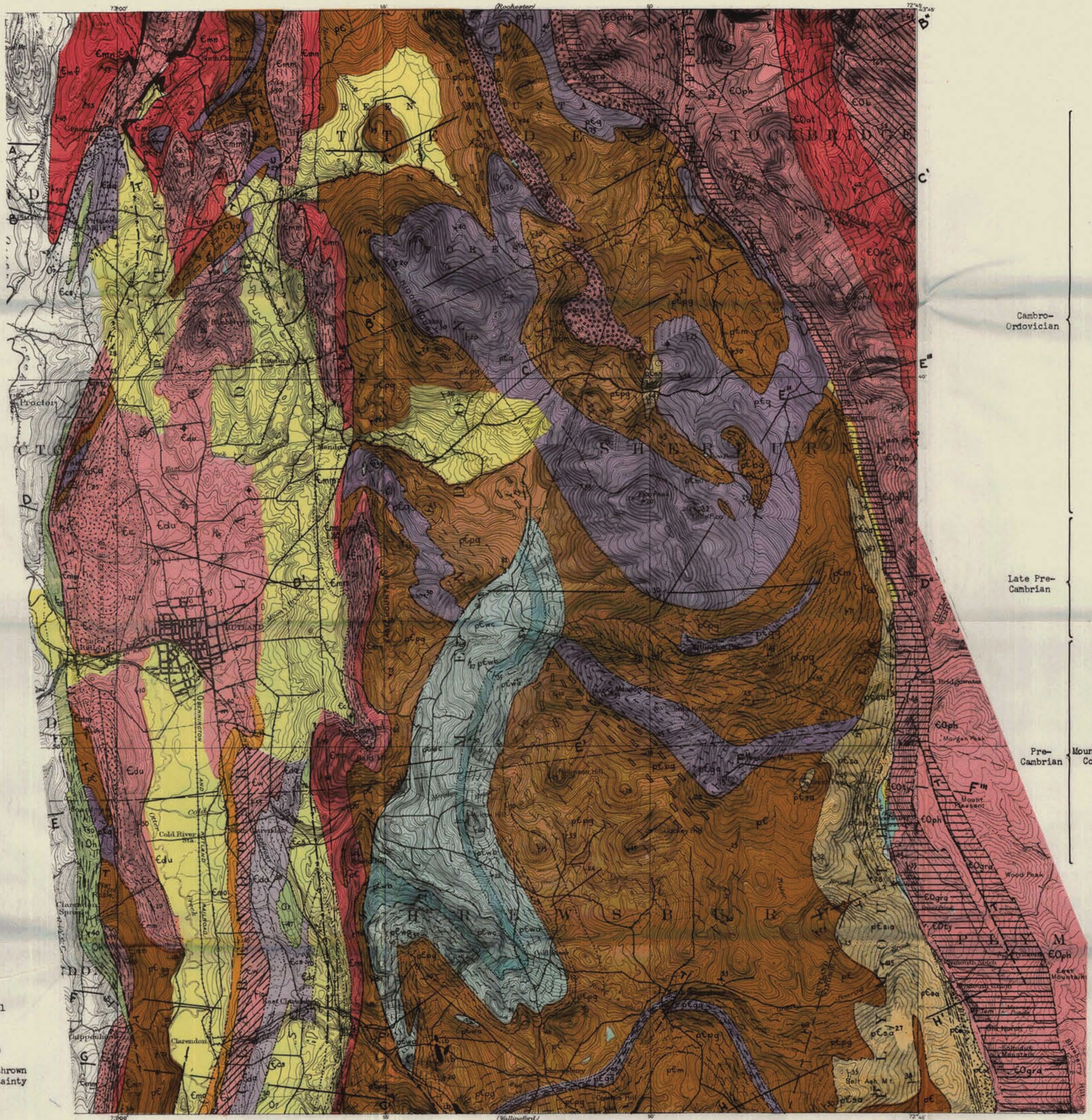
Compositional banding, where horizontal

Compositional banding, where vertical

Foliation

Foliation, where vertical

Cross Section, Plate 2



EXPLANATION

METAMORPHIC ROCKS

Cambro-Ordovician — ƐOb: "Bethel" Formation. Green phyllite, sandy finely banded green sandstone

ƐOot: Ottauquechee Formation. Black phyllite, gray, black quartzite

ƐOph: Pinney Hollow Formation. Green, white phyllite, gray, green sandstone, albitic schist

b: Black phyllite

g: Epidote-chlorite-calcite greenstone

ƐOgra: Grahamville Formation. Albite grit, dark sandstone, finely banded quartzose black phyllite, thin quartzite and dolomite near top

ƐOty: Tyson Formation. Green, gray grit, thin conglomerate, buff-weathering pink dolomite

unconformity

Late Pre-Cambrian — pƐsc, pƐsb, pƐsa: Saltash formation

c: Vitreous gray quartzite

b: Black graphitic phyllite, thin dolomite and limestone

a: Massive white grit, sericite-quartzite, thin conglomerate

unconformity

Pre-Cambrian — Mount Holly Complex

pƐq, pƐgq: Quartzite, schist. Garnetiferous quartzite

pƐm: Marble, lime-silicate rocks

pƐbg, pƐpg: Biotite-microcline gneiss. Plagioclase-garnet gneiss

pƐ: Grits; undifferentiated

Drift. Glacial and recent

Topography by U.S.G.S.
Surveyed: 1881-1925

APPROXIMATE MEAN DECLINATION 1944

1 ½ 0 1 2 3 4 Miles

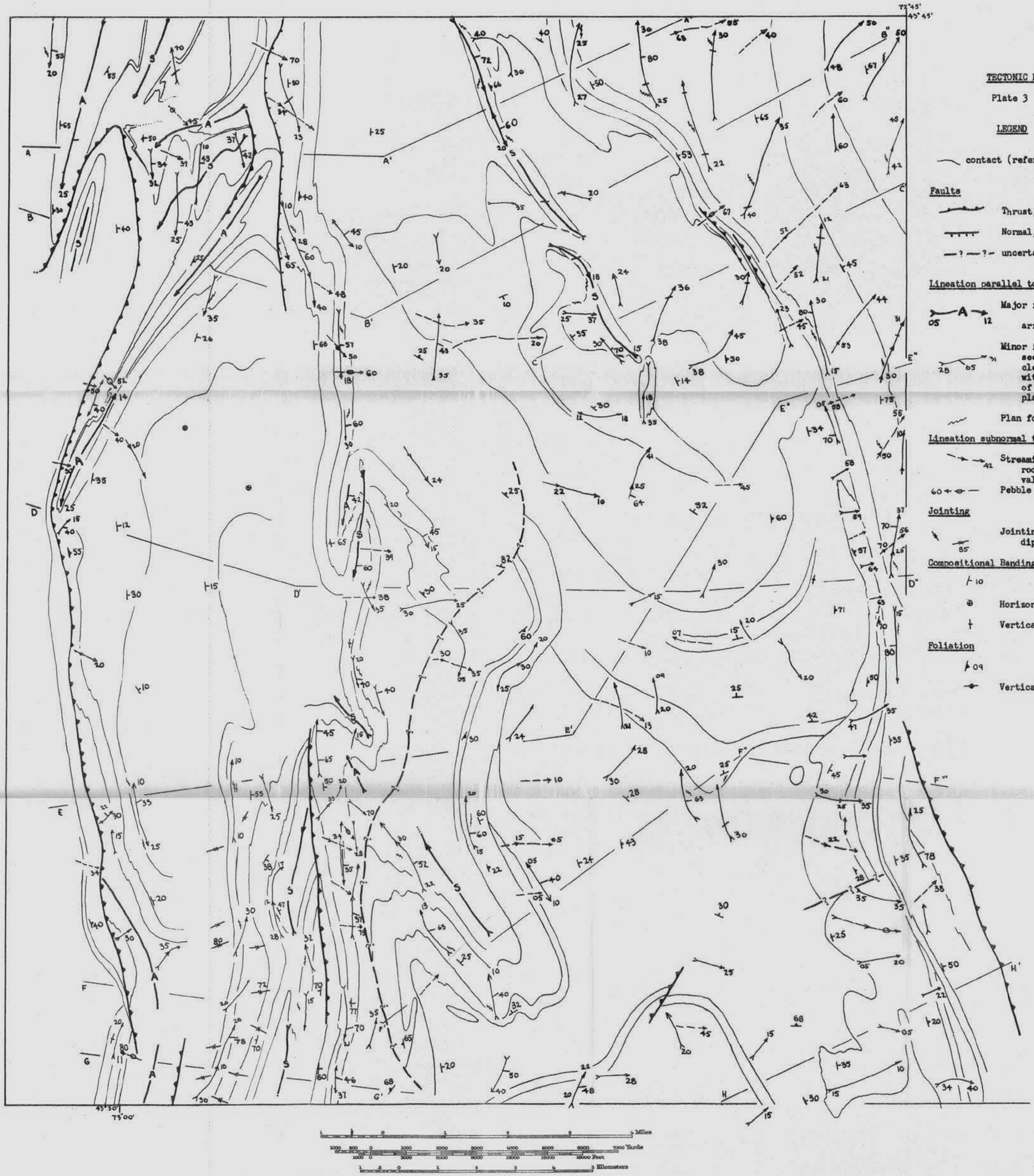
Geology by W. F. Brace
Surveyed: 1950-1952
Area East of Plymouth by Chang (1950)

AREAL GEOLOGY OF THE RUTLAND AREA, VT.

Plate I.

VERMONT GEOLOGICAL SURVEY

Charles G. Doll, State Geologist



TECTONIC MAP

Plate 3

LEGEND

contact (refer to plate 1)

Faults

Thrust, with arrows on thrust plate

Normal, with hachures on downthrown block

—?—?— uncertain as to existance

Lineation parallel to fold axes

05 —A→ 12 Major folds. A denotes anticline, S denotes syncline, arrow and number are plunge

28 65 31 Minor fold axes, crinkles, intersection of foliation with slip cleavage and fracture cleavage, with value and direction of dip of slip cleavage or fold axial plane

Plan fold pattern

Lineation subnormal to fold axes

→ 41 Streaming, mineral lineation and rodding showing direction and value of plunge

60 ←⊖— Pebble elongation

Jointing

85 Jointing vertical or with value of dip on dipping side

Compositional Banding

⊢10

⊕ Horizontal

Vertical

Foliation

09

Vertical

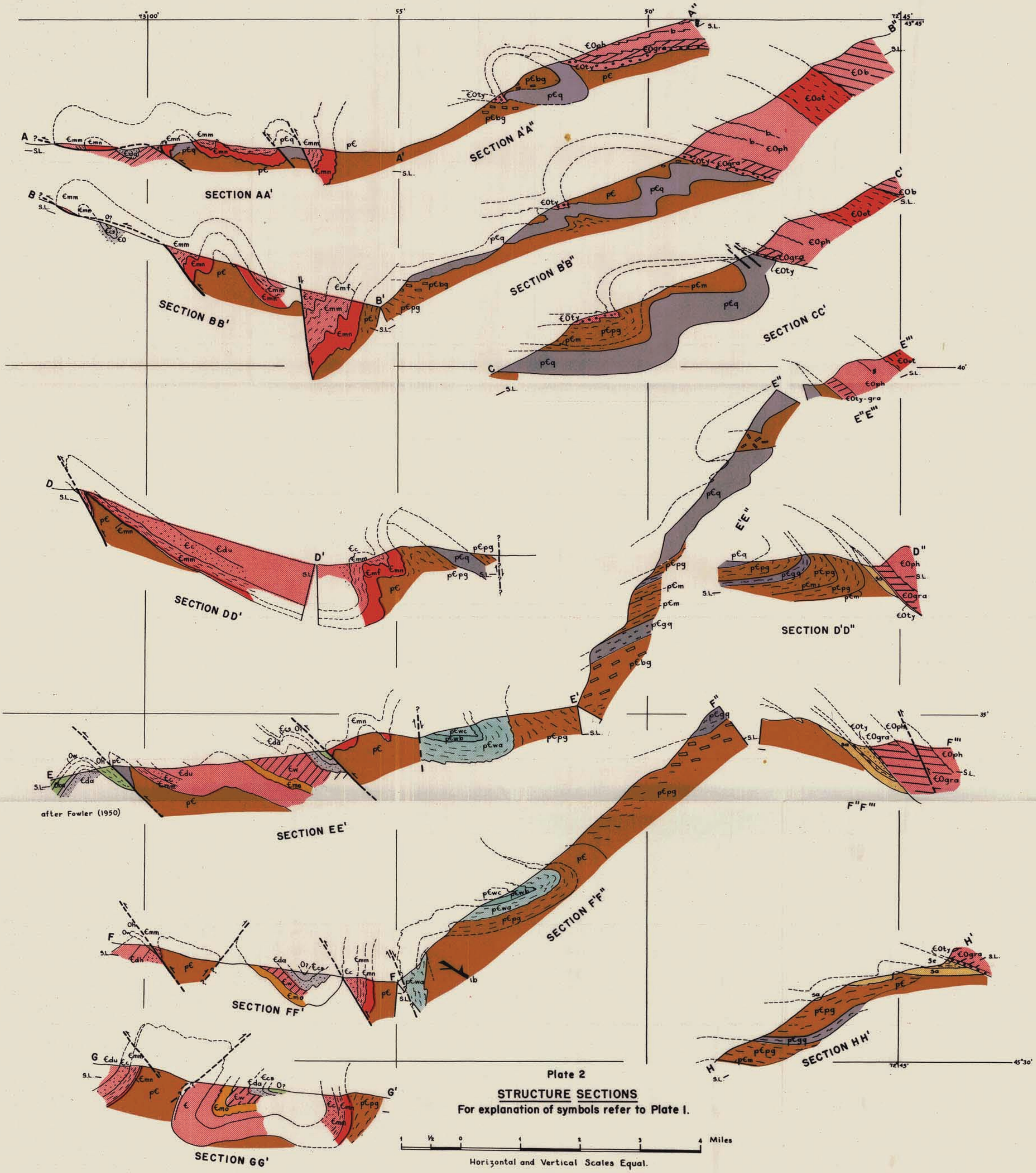


Plate 2

STRUCTURE SECTIONS

For explanation of symbols refer to Plate I.