THE GEOLOGY OF THE BENNINGTON AREA, VERMONT

By

JOHN A. MACFADYEN, JR.

VERMONT GEOLOGICAL SURVEY
CHARLES G. DOLL, *State Geologist*

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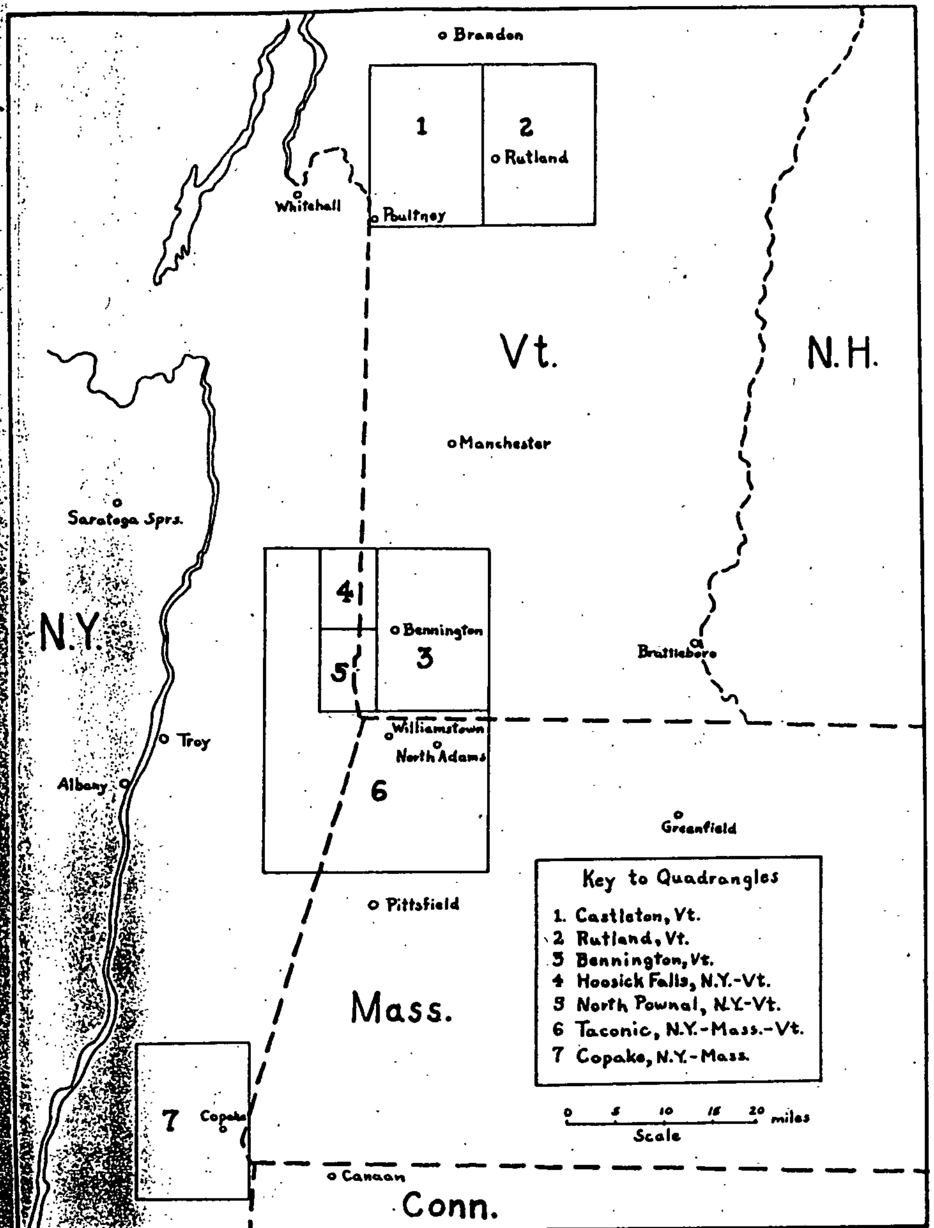

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THE GEOLOGY OF THE BENNINGTON AREA, VERMONT

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ABSTRACT

The bed rock geology of the Bennington area consists of an unknown thickness of pre-Cambrian gneisses and schists overlain by approximately 10,000 feet of younger sediments. These range in age from Lower Cambrian to Middle or Upper Ordovician. The basal Cambrian is composed of arkose, quartzite and phyllite, some of which was locally derived from the underlying gneisses. Overlying this basal section is a thick sequence of Lower and Upper Cambrian dolomites and interbedded quartzites. The non-carbonate portion of these rocks appears to have been derived from a westerly source, probably the pre-Cambrian craton. A marked change in lithology occurs along the Cambrian-Ordovician contact. The Lower Ordovician sediments are dominantly calcitic limestones and marbles with interbedded calcitic dolomites. Overlying these limestones with a profound regional unconformity are Middle and Upper (?) Ordovician phyllites and schists containing lens-like masses of limestone in their lower part. No rock younger than the Ordovician has been recognized in the area.

Within the region there have been at least two periods of metamorphic activity. The pre-Cambrian rocks give evidence of having been raised to a medium- to high-grade metamorphism prior to the deposition of the Lower Cambrian strata. The Cambro-Ordovician sediments have undergone a low-grade metamorphism as is shown by the presence of chlorite, chloritoid, and biotite in the Ordovician phyllites and schists. At the same time retrogressive metamorphism is indicated in the pre-Cambrian rocks. This second period of metamorphic activity is probably Taconic in age. Metasomatism was also important as there was a widespread tourmalinization of the Paleozoic rocks at this time.

The major structures of the region are the southern end of the Green Mountain anticlinorium and the eastern flank of the Taconic synclinorium. Superimposed on these two major folded structures are a

number of orders of smaller folds, ranging in magnitude from a "wave length" of several miles down to minor plications measuring an inch or less from crest to crest. This folding was accompanied by extensive thrust faulting.

The thrust faulting is believed to be of two major types, 1) deep-seated thrusts extending into the pre-Cambrian basement, and 2) surficial, flat-lying overthrusts affecting only the Paleozoic formations. The former type are found along the west flank of the Green Mountain anticlinorium near the contact with the pre-Cambrian, while the latter occur farther west in the synclinorium. It is suggested that the major thrusts found farther west and north in Vermont and New York are of the second type. At least one high-angle normal fault is known in the area. This structure is probably later than the folding and thrusting, and may indicate crustal relaxation or upwarping following the cessation of orogenic activity. The age of this fault is not known but may be as late as Triassic.

The evidence for the presence of the Taconic thrust in the Bennington area and surrounding regions is examined and an alternative hypothesis is advanced. It is suggested that this new hypothesis is more consistent with the structural and stratigraphic information now available than is the thrust idea. The present hypothesis depends: 1) upon a reinterpretation of the stratigraphic history within the Taconic synclinorium during Cambro-Ordovician times; and 2) upon the belief that the rise of the Green Mountain anticlinorium was the controlling factor in the development of the Lower Paleozoic structures found farther west. On the basis of this hypothesis the phyllites and schists overlying the Ordovician limestones are now considered to be in their normal stratigraphic position, and are not a part of a "Taconic klippe" as previously supposed.

INTRODUCTION

Location

The area under discussion in this report covers about 210 square miles in the southwestern corner of Vermont, including the Bennington quadrangle (latitude 43°45', longitude 73°00' to 73°15'), and those parts of the North Pownal and Hoosick Falls quadrangles (1:31680) lying within that state. In general, it extends eastward from the New York–Vermont boundary to the headwaters of the West Branch of the

Deerfield River in the heart of the Green Mountains, and northward from Pownal to Shaftsbury Center.

Two main routes traverse this region: U.S. north-south Route #7; and U.S. east-west Route #9. The western half of the area and its southeastern corner are readily accessible by automobile over all-weather black top or dirt roads.

Physiography and Glaciation

The Bennington area is made up of three topographic belts trending approximately north-south. From east to west they are: 1) the Green Mountains; 2) the Bennington Valley, and 3) the Taconic Range.

The eastern belt, that of the Green Mountains, forms a major divide separating the drainage eastward to the Deerfield and Connecticut rivers from that draining westward into the Hudson. The two major streams flowing through the region are the Walloomsac and Hoosic rivers. The Hoosic, rising in the valleys to the west of the Hoosac Range in Massachusetts and Vermont, flows northwesterly across the southwestern corner of the map area and on into the Hudson. The Wallomsac rises in the mountains east of the town of Bennington and following a sinuous course generally north of west joins the Hoosic at Hoosick Falls, New York.

The most conspicuous topographic feature in the region is the abrupt front forming the western margin of the Green Mountains. It rises precipitously some 1800 feet from the valley to the west, separating the high, rugged Green Mountain belt from the Bennington Valley. In this Green Mountain belt the highest points in the region are found. Glastenbury Mountain in the northeast rises to an elevation of 3764 feet; to the south seven other hills reach 3000 or more feet. Along the bold front, Bald Mountain, directly east of North Bennington, and The Dome, southeast of Pownal Center, are conspicuous high points. In general, the ridges here trend somewhat east of north. To the southwest the upland area of the Green Mountains is broken and separated from the Hoosac Range by the valley of the North Branch of the Hoosic River.

Directly west of the Green Mountains and paralleling them is the relatively low Bennington Valley, varying in elevation from approximately 600 to 1000 feet. This valley extends northward varying in width until it joins the Champlain Valley north of Rutland. Southward, at Pownal Center, it is terminated by the Pownal Upland, a series of low hills which connect the Taconic Range to the west with the Green Mountains to the east and separate the Hoosic Valley from the Bennington Valley. Within the valley itself are a number of small but conspicuous hills. These hills are most prevalent north of Bennington.

Like the Green Mountains, the Taconic Range, which comprises the western margin of the Bennington area, forms a marked topographic high. It is, however, broken by the broad valley of the Walloomsac into a southern and a northern portion. The northern division, rising west of Shaftsbury Center, constitutes the ridge of West Mountain, which extends northward with a fairly constant elevation toward Arlington and Manchester. The southern division rises abruptly just west of Bennington to form Mount Anthony, with an elevation of 2345 feet. The Mount Anthony ridge extends southward, decreasing in elevation until at Pownal Center it swings sharply to the east forming the Pownal Upland.

In general, the topography of the region reflects the structure and lithology of the bedrock. The low valleys are underlain by carbonate rocks, while the minor hills within the Bennington Valley are supported by sandstone beds within the carbonate sequence. The major uplands or ridges, such as the Green Mountains and the Taconic Range, are underlain by more resistant rocks; the former by the pre-Cambrian crystalline complex and the latter by younger phyllites and schists.

While it is not the purpose of this paper to deal in detail with the evidence and effects of the Pleistocene glaciation, it is impossible to ignore them completely. Evidence for this glaciation is widespread throughout the whole region. The upland areas are either scraped bare, occasionally showing striated rock outcrops, or covered by a thin veneer of till. Much of the till is quite coarse, showing many large erratic blocks.

It is in the valleys, however, that a picture of the late Pleistocene can best be reconstructed. The fact that drainage is generally toward the northwest apparently affected the area profoundly following withdrawal of the ice from the Bennington and Hoosic Valleys. Retreating slowly, the ice obstructed this northwest drainage system, and seems to have caused the formation of a large glacial lake. Into the lake large amounts of sediment, much of it fairly coarse, were poured by the heavily laden streams. These sediments can be seen today in the many sand pits that have been opened in the valley. Fresh pit exposures commonly show well-sorted sands interbedded with gravel. Occasionally there is evidence of small deltas.

With the further retreat of the ice toward the north, outlets at successively lower elevations became available, and the lake was gradually drained, leaving behind a series of terraces. These terraces were then further modified by stream erosion due to the lowering of the local base level, which in turn resulted from a lowering of the lake level.

Purpose of Study

The primary purpose of this study was an investigation of the structure of the lower Paleozoic rocks in southwestern Vermont. The analysis of the structure of this region requires three steps:

1) The resolution of the structure of the southern end of the Green Mountain anticlinorium.

2) The recognition of the effect of the rise of this anticlinorium on the Paleozoic sediments to the west.

3) The explanation of the stratigraphy and structure of the rocks believed to be a part of the Taconic thrust sheet.

Method of Study

The area was mapped over a period of about nine months during the summers of 1951, 1952, and 1953. Various combinations of quadrangle topographic maps (Scales 1:62,500 and 1:31,680) and aerial photographs (Scale approximately 1:22,000) were used to plot field observations. Final formational boundaries and other pertinent data were plotted on the appropriate quadrangle maps in true horizontal position.

Extremely thick glacial cover and the consequent scarcity of exposures in many parts of the region account for at least a few of the inconsistencies on the map and resultant errors in interpretation. Due to poor exposures it was not possible to determine in any detail the structure or stratigraphy of the pre-Cambrian complex. As a result the treatment of these rocks is somewhat superficial.

Regional Geologic Setting

Southern Vermont can be divided from east to west into three geologic provinces, each characterized by distinct rock types. The eastern half of this region is underlain by metamorphosed argillaceous sediments ranging in age from Lower Cambrian to Middle or Upper Ordovician. The predominant lithology is a low- to medium-grade schist. Amphibolites, quartzites, and marbles occur in subordinate amounts. In general the structure of this area is that of an eastward-dipping homocline with the youngest rocks being found along the Connecticut River. This province extends north into Canada and south into Massachusetts and Connecticut, lying between the Triassic lowland to the east and the pre-Cambrian crystalline highlands to the west.

In Vermont the argillaceous lower Paleozoic rocks of the eastern sequence are separated from those of the west by the Green Mountain anticlinorium. This structure, forming the core of the present Green Mountains, brings to the surface the crystalline metamorphosed sediments and igneous rocks of the pre-Cambrian in a north-south trending belt. They plunge south beneath the Paleozoic sediments at North Adams, Massachusetts. Still farther south, similar pre-Cambrian gneisses reappear, outcropping in a more or less continuous belt in western New England and southeastern New York. To the north the pre-Cambrian core of the anticlinorium disappears under later sediments near Rochester, Vermont, and reappears again at Sutton Mountain in southern Quebec.

West of the Green Mountains lies a broad synclinal belt of Lower Paleozoic sediments. Although this western sequence is of the same age as the eastern there is a marked difference in lithology. Quartzite, limestone, dolomite, and phyllite predominate; rocks of igneous origin are extremely rare. These rocks in the north extend westward to the Adirondack Mountains. Farther south they underlie the Hudson Valley and appear to continue southwestward in a band into New Jersey and eastern Pennsylvania.

Across these three provinces there is a general increase in the grade of metamorphism from west to east. Unmetamorphosed sediments occur in the Hudson and Champlain Valleys while eastward along the west flank of the Green Mountains the chlorite and biotite zones are encountered. The pre-Cambrian shows evidence of an earlier medium- to high-grade metamorphism upon which a later low grade has been impressed. Eastward the metamorphic grade increases showing rocks of the garnet and staurolite zones. Along the Connecticut River the metamorphic intensity decreases and only rocks of the chlorite zone are seen.

The Paleozoic rocks of this region have probably been affected by three periods of deformation. The most important was the Taconic orogeny which began in mid-Ordovician time and reached its peak during the latter part of that period. In Vermont the relative effects of

the Acadian and Appalachian orogenies cannot be defined as rocks younger than Ordovician have not been preserved.

Within the Bennington area parts of all three of the above geologic provinces are present. The lower Paleozoic sediments of the western sequence underlie the western half of the map area. On the other hand most of the eastern half is underlain by the pre-Cambrian core of the Green Mountain anticlinorium. In the southeastern corner a small area of the eastern sequence is exposed.

Previous Work

During the middle and latter part of the nineteenth century western New England was the scene of much of the early systematic geologic mapping. Ebenezer Emmons, on the basis of studies in this area and eastern New York State, proposed the term "Taconic System" to include the rocks of the Champlain Valley and the Taconic Range (Emmons, 1842, 1844). This new subdivision of the Paleozoic Era was based on the assumption that these rocks were all older than the fossiliferous Potsdam Sandstone of the eastern Adirondack border. Thus Emmons believed they constituted a new system. The validity of his proposal was hotly debated for many years. However, by the latter part of the century it became apparent that it was not justified. Logan's (1863) discovery of the Champlain Fault helped resolve some of the difficulties in correlation from western Vermont into eastern New York, and work by Dana (1877, 1887) and Walcott (1888) brought about the final rejection of the Taconic System. Since then the term Taconic has had no accepted stratigraphic significance, although Schuchert (1919) attempted without success to revive the term as the name for the Lower Cambrian epoch.

Pumpelly, Wolff, and Dale (1894) mapped the northwestern corner of Massachusetts in considerable detail and determined at that time the general structure and stratigraphy of the area. Over a period of years Dale (1892, 1894, 1899, 1904, 1912) published a series of excellent papers on the structure and stratigraphy of western Vermont. While the information contained in the above reports has direct application to the geology of the Bennington area, relatively little work was actually done in that region. Beginning in 1915 Gordon (1915, 1921, 1925) brought out a series of papers dealing with the geology of the Bennington area and other parts of western Vermont. These papers, however, tend to be merely a description of outcrops. The author seems to have had little idea of the general structure or stratigraphy of the region.

The announcement by Keith (1912, 1913) of the overthrust nature of the northern end of the Taconic Range stimulated further detailed work in western New England. In 1932 Prindle and Knopf published their famous paper on the Taconic quadrangle, claiming that much of the argillaceous rock in the region was actually part of a thrust sheet and not a part of the normal stratigraphic succession as had been previously supposed. Since this time the area covered by the supposed Taconic allochthone has been extended to include much of western New England and eastern New York (Kay, 1935; Knopf, 1935). According to Kay (1941) and Stose (1946), a small portion of the Taconic allochthone is present in eastern Pennsylvania.

While recent work in the northern end of the Taconic region (Kaiser, 1945; Cady, 1945; Fowler, 1950) has tended to support the idea of the Taconic thrust, Balk (1953) working in eastern New York and southwestern Vermont has rejected it on the basis of structural observations.

Work by Keith (1932) and Cady (1945) in western and northwestern Vermont established the stratigraphy of the carbonate sequence west of the Green Mountains. This stratigraphic succession has been traced south of the type areas and is followed as far as possible in this report.

Acknowledgments

The author wishes to acknowledge his indebtedness to Professor Walter H. Bucher of Columbia University who first suggested this problem and has guided the study.

Financial aid for field and laboratory work was provided by the Vermont Geological Survey and the Lamont Geological Fund of Columbia University. Professor Charles G. Doll, Vermont State Geologist, spent several days in the field with the author and has provided much constructive criticism.

Professors Marshall Kay and Arie Poldervaart of Columbia University have visited the area and discussed with the author various aspects of the study.

Many geologists working in Vermont or eastern New York State have visited the area and have been most helpful in discussion of the many field problems. Among these are Professor James B. Thompson of Harvard University, Professor John Rodgers of Yale University, Dr.

Wallace M. Cady of the United States Geological Survey, Dr. John D. Weaver of Mount Holyoke College, Dr. William F. Brace, and the author's colleagues, Elwyn Perry, Freeman Foote and Robert Ramsdell of Williams College.

To the geologists who visited the area during the 1952 New England Intercollegiate Geological Field Conference, the author extends his appreciation for many suggestions and criticisms.

The author's wife, Nancy MacFadyen, has given continual aid during the course of the study and typed the final manuscript. Mrs. H. T. Gerrish has read the manuscript and made some suggestions as to its literary style.

STRATIGRAPHY

General Statement

The pre-Cambrian basement complex in the Bennington area consists for the most part of a thick section of gneisses and schists into which have been intruded both acid and basic igneous rocks.

The Paleozoic rocks, unconformably overlying the pre-Cambrian, can be divided into a western and an eastern sequence. The former, exposed west of the Green Mountain anticlinorium, consists of about 10,000 feet of Cambrian and Ordovician sediments. The base of the Cambrian section is composed of quartzite, arkosic quartzite, and phyllite. Above this lie sandy dolomites and dolomitic sandstones. The Ordovician sequence is made up of interbedded dolomites and calcite marbles overlain by black, gray, and green phyllites.

The eastern sequence, lying to the east of the pre-Cambrian, is essentially the same as the western with the exception that the higher Cambrian dolomites and the Ordovician marbles and dolomites are missing. They are present, however, farther south in Massachusetts. This area will not be treated in detail as the author was able to make only a reconnaissance survey of these rocks. The contacts plotted on the map (see Plate 1) are those of Prindle and Knopf (1932). Detailed mapping in northern Massachusetts around the end of the pre-Cambrian will be necessary in order to correlate the eastern and western sequences.

The thicknesses given for the various formations are based on the width of outcrop. These figures probably represent a maximum as overturned folding and thickening and thinning of the beds are common. For the Cambrian carbonate sequence the thicknesses are those deter-

TABLE 1
TABLE OF FORMATIONS

Correlation	Formation	Thickness (in feet)
Trentonian or younger?	Mount Anthony formation	0–1200
	—Unconformity—	
Trentonian	Walloomsac slate	900?
	—Unconformity—	
Canadian	Canadian limestone	0–2000
Upper Cambrian	Clarendon Springs dolomite	0–1300
	Danby formation	0–300
	Disconformity?	
Lower Cambrian	Winooski dolomite	0–700
	Monkton quartzite	0–1000
	Dunham dolomite	700–1500
	Cheshire quartzite	800
	Mendon formation	
	Moosalamoo member	300–400
	Nickwacket member	500–800
	—Unconformity—	
pre-Cambrian	Stamford granite gneiss	
	Mount Holly formation	

mined in the northern part of the area where a fairly complete section is exposed.

The various formations have been determined on the basis of lithologic similarity with those of western Vermont. Because of the lack of paleontological control there is no assurance that they are of precisely the same age as in the type areas. They do, however, represent recognizable rock units which can be used in field mapping.

Pre-Cambrian Sequence

Mount Holly Gneiss (*Whittle, 1894*): The Mount Holly outcrops throughout much of the eastern half of the Bennington section and is the dominant rock type in the pre-Cambrian basement complex. It has been mapped in the core of the Green Mountain anticlinorium as far north as the Rutland and Rochester areas in Vermont (Brace, 1953; Osberg, 1952) and south into the Cheshire area of Massachusetts (Prindle and Knopf, 1932).

It consists of an unknown thickness of metamorphosed sedimentary and igneous rocks. The sediments are represented by mica schist, pink or gray biotite-microcline gneiss containing abundant blue quartz, and massive vitreous quartzites. Occasionally the gneisses show thin layers of granitic material, giving the rock the appearance of a migmatite. Although lenses and beds of altered limestone have been reported (Prindle and Knopf, 1932; Brace, 1953) none were found in the gneiss of the Bennington area.

The Mount Holly has been intruded by both acid and basic igneous rocks. A common type of possible intrusive origin is a coarse-grained white gneiss containing quartz, microcline, albite, and biotite. Thin bands of fine-grained amphibolite are common throughout the formation, and probably represent altered basic dikes and sills. Several small areas of massive amphibolite were observed. Blue quartz-bearing pegmatites have invaded many of the above rocks.

Stamford Granite Gneiss (*Hitchcock, 1861*): The Stamford granite gneiss is exposed in the southern end of the Green Mountain anticlinorium along the Vermont-Massachusetts boundary, and also east of Adams, Massachusetts, on the western slope of Hoosac Mountain.

The typical Stamford is a gray, coarse-grained, porphyritic granite gneiss with large, rectangular or rounded, perthitic microcline phenocrysts. The granitic groundmass contains blue quartz, albite, microcline, biotite, and epidote. It also shows a weak foliation, parallel to that of the Mount Holly. Near the contact with the Mount Holly, the Stamford becomes finer grained and the porphyritic nature of the rock is less apparent. While the contact was never actually observed, it is possible that a transition zone occurs between the two rock types.

Lower Cambrian Sequence

Mendon Formation (*Whittle, 1894*): The Mendon formation is exposed along the west flank of the Green Mountain anticlinorium in western Vermont. Although locally it may appear to be conformable, it actually overlies the pre-Cambrian with a profound unconformity.

The Mendon is probably the stratigraphic equivalent of the Dalton formation of western Massachusetts described by Emerson (1892). They both overlie the Mount Holly gneiss or other pre-Cambrian gneisses and are in turn overlain by the Cheshire quartzite. If they are equivalent the name Dalton would take precedence.

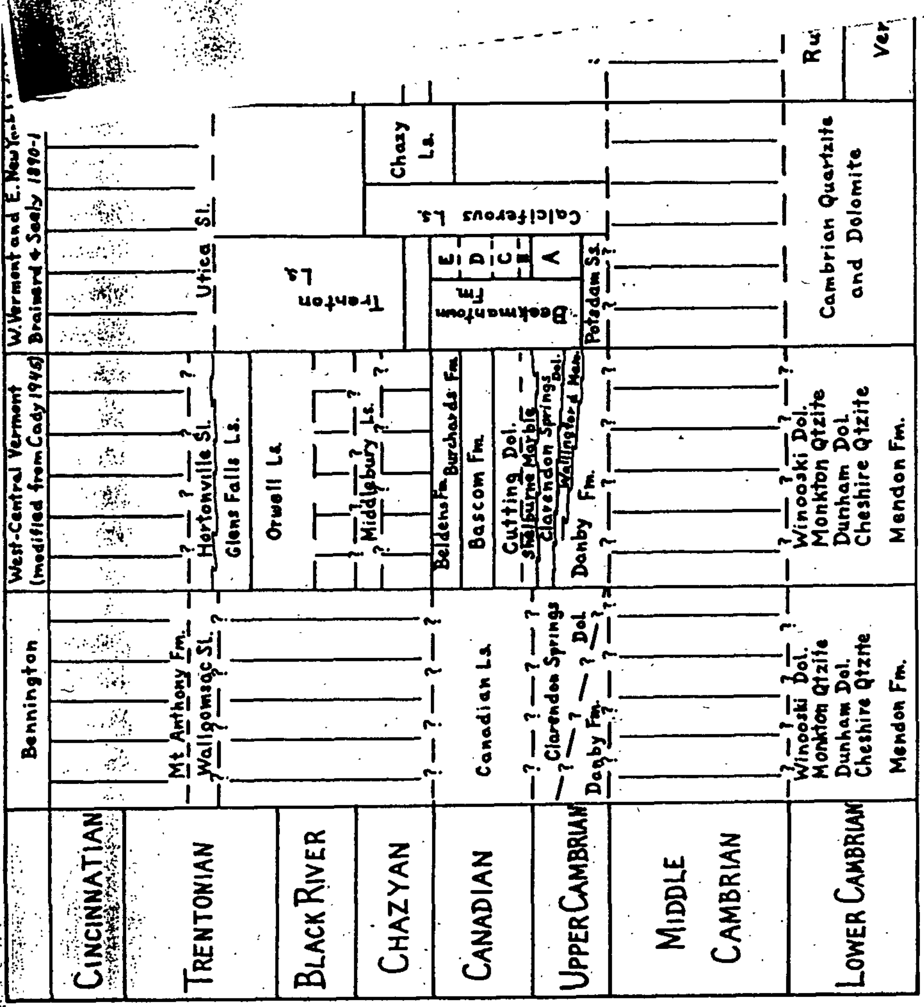

Figure 2. Correlation Chart

… …estdale, … member was not found in the … According to Keith (1932) its maximum thickness occurs in the vicinity of Brandon and decreases to the southward. Brace (1953) shows the Forestdale as intermittently present in the Rutland area, apparently disappearing near the southern end of the quadrangle.

The Nickwacket member (Keith, 1932) in the Bennington section is largely thin- to medium-bedded quartzite, arkosic quartzite, and sericite-tourmaline schist with minor amounts of graywacke. The schist and graywacke are most prevalent near the bottom of the unit. Thin bands of conglomerate are common, lying just above the contact with the pre-Cambrian. These consist of rounded blue quartz pebbles in a matrix of graywacke or arkosic quartzite. Also appearing near the contact are crystals of hematite imbedded in the schistose facies of the rock. These crystals are excellent indicators of the proximity of the pre-Cambrian.

The actual contact with the basement complex is not often evident even where adequate exposures exist. The lower part of the Nickwacket was apparently locally derived from the erosion of the pre-Cambrian. Consequently this material, following induration and metamorphism, has taken on the appearance of its parent rock.

Complete exposures of the Nickwacket along the Green Mountain front are rare. In the bed of City Stream east of Bennington this member measures about 800 feet, while southward east of The Dome approximately 500 feet appear to be present. In the Rutland area Brace (1953, p. 31) reports a variation in thickness from 25 to 800 feet, while in the Castleton quadrangle Fowler (1950, p. 16) indicates the thickness may locally be considerably greater.

The Moosalamoo member is a black, sandy, sericite-biotite phyllite with thin interbeds of white and grayish quartzite. The sand in the phyllitic facies occurs in thin laminae giving the rock a finely-banded appearance. Pyrite is occasionally present and is often drawn out into thin streaks because of deformation.

Throughout the area the Moosalamoo appears to vary in thickness from 300 to 400 feet, somewhat less than the 500 to 800 feet found by Brace (1953, p. 31).

While several previous workers in the western Vermont area (Whittle, 1894; Keith, 1932; Fowler, 1950) have assigned the Mendon to the late pre-Cambrian, the author considers it to be of Lower Cambrian age. Walcott (1888) found fragments of *Olenellus* about 100 feet above the pre-Cambrian contact near North Adams, Massachusetts. The rock here is a quartzitic graywacke and is stratigraphically beneath a band of black phyllite, probably the equivalent of the Moosalamoo. In this area no evidence was found of the reported unconformity between the Mendon and the overlying Cheshire quartzite. An exposure of this unconformity above the Falls of Lana described by Foye (1919) and Keith (1932) was re-examined by Osberg (1952, p. 37). He concluded that "this unconformity is of minor extent, and careful mapping shows no angular relations."

The Mendon is present throughout the length of Vermont and its probable equivalents extend northward into southern Quebec. Here the Pinnacle graywacke, the White Brook dolomite and the West Sutton slate represent the Nickwacket, Forestdale, and Moosalamoo members of the Mendon.

In the Bennington region Prindle and Knopf (1932) included the Mendon in the Cheshire quartzite, but recognized the typical Nickwacket and Moosalamoo lithologies. The author has followed their terminology in the southeastern part of the map area. The quartzites here are more feldspathic than the normal Cheshire, but do not have the characteristic greenish, schistose beds of the Nickwacket. In thickness they measure less than 500 feet, indicating either that there has been a considerable thinning of the units eastward or that one or more of them are missing. This question can be resolved only by further mapping.

Cheshire Quartzite (Emerson, 1892): The Cheshire is a relatively pure, completely recrystallized, medium- to thick-bedded quartzite with occasional thin beds of black phyllite. Minor amounts of sericite, altered feldspar and pyrite are present in the rock. It is the weathered pyrite that gives the characteristic brown-stained color to the rock in outcrop. On a fresh surface the color is generally white or buff.

Primary structural features are not often apparent. In the absence of minor lithologic variations the several joint sets commonly present can easily be mistaken for bedding. Although never found in place, examples of cross-bedding have been observed in float varying in amplitude from six inches to two feet.

The contact with the underlying Moosalamoo may be gradational. Thin beds of white quartzite similar to the Cheshire occurs in the upper

part of the phyllite. According to Cady (1945, p. 528), a thin zone of interbedded quartzite and sandy dolomite marks the boundary between the Cheshire and the overlying Dunham dolomite. This zone was never found in the Bennington area. If present, it is generally covered by drift.

The average thickness of the Cheshire was found to be between 750 and 800 feet, somewhat less than reported farther north (Cady, 1945; Brace, 1953).

The lower Cambrian age of the Cheshire has been well established. In the Bennington area Walcott (1888, p. 285) found *Hyolithes, Nothosoe* and *Olenellus*. Near Lake Dunmore Walcott with Seely (1910, p. 307) discovered *Hyolithes*, and a "*Lingula*" was reported by Hitchcock (1861), p. 356) near Monkton. The author has found "*Scolithus*" in the Cheshire in a road cut on U. S. Route #7, two miles north of Bennington.

Dunham Dolomite (Clark, 1934): In the Bennington area the Dunham dolomite is very poorly exposed. Here the term is used to include all the dolomites and sandy dolomites that lie between the Cheshire and recognizable Monkton quartzite.

According to Cady (1945, p. 528), the Dunham is "mainly a siliceous buff-weathering dolomite containing well-rounded sand grains irregularly distributed."

The lower part of the formation consists of massively-bedded dolomite which may be cream-pink or gray in color. The top of the zone is generally gray, and upon weathering has a sandpaper-like surface due to the abundant sand grains. The fresh rock is often mottled, and thin partings of greenish or grayish phyllite are not uncommon.

The upper part of the formation is made up of thinner bedded, cream or gray, sandy dolomite, with thin, siliceous partings. In this zone the dolomite occasionally shows cross-bedding several inches in height. The upper part of the Dunham grades into the overlying Monkton. The quartzite beds become thicker and more prevalent as the contact is approached. In this respect the upper Dunham is in many ways similar to the Winooski dolomite.

In the Bennington region the rocks mapped as Dunham would correspond to the lower part of the Rutland dolomite, described in this area by Prindle and Knopf (1932).

While in the Bennington section no fossils were found in this formation, lower Cambrian fossils have been found farther north in the vicinity of Rutland (Wolff, 1891; Foerste, 1893).

In general, the Dunham is very poorly exposed, with most of its outcrop area under a thick mantle of glacial debris. Nowhere is a complete section to be seen. Any picture of the formation must be pieced together from scattered outcrops throughout the area. The lower part of the formation is best seen on the western flank of The Dome. The upper part is exposed along the Rutland Railroad northeast of South Shaftsbury and on the west side of Trumbull Mountain.

Throughout the northern part of the area the thickness of the Dunham seems to be on the order of 1500 feet. This compares favorably with the 1700 feet in the Rutland area (Brace, 1953, p. 29), and the 1700 to 2000 feet in west-central Vermont (Cady, 1945, p. 525). Southward on the west side of The Dome where the upper Dunham has been cut out by the pre-Walloomsac unconformity, approximately 700 feet are present.

Monkton Quartzite (*Keith, 1923*): The Monkton quartzite in this area consists of white, gray and buff, fine-grained, sandy, thin-bedded dolomite, with inter-bedded dolomitic sandstone. The dolomite habitually weathers to a reddish-buff color, and commonly shows well-developed cross-bedding several inches high. The interbedded sandstones, while gray on a fresh surface, weather deeply to a soft, friable, rusty, sandy mass.

The age of the Monkton in the Champlain Valley has been determined as lower Cambrian. Kindle and Tasch (1948) have reported finding four species of *Olenellus* as well as a number of other lower-Cambrian forms. In this area no undoubted fossils were found. In several localities, however, the author found structures in the sandstones that might possibly be of organic origin.

Both Cady (1945, p. 531) and Fowler (1950, p. 20) report that the Monkton dies out southward in a series of dolomites. It is, however, present in the Wallingford area; and the rocks mapped as Monkton around Bennington have the same stratigraphic position and lithologic characteristics as the formation farther north. It is evident, though, that the amount of sand decreases toward the south. In general, the sandstones are thinner and more widely spaced than is usual in the western Vermont section. This thinning-out probably represents an increased distance from the source.

In this area all varied-colored dolomites containing dolomitic or deeply weathered sandstone beds more than several inches in thickness were mapped as Monkton. In the vicinity of Trumbull Mountain approximately 1000 feet of the formation are present. This is somewhat greater than that reported farther north. Southeastward it probably pinches out beneath the transgressing Walloomsac slate. It is quite possible that the Monkton as mapped here includes some Dunham and Winooski.

Winooski Dolomite (*Cady, 1945*): The Winooski is a thin-bedded, somewhat sandy dolomite that is characterized by thin siliceous partings along the bedding planes. In general, in the lower two-thirds of the formation, the dolomite is cream- to pink-colored on a fresh surface and is almost identical with the upper part of the Dunham. In the upper third the rock is a gray calcitic dolomite.

The contact between the Monkton and the Winooski is gradational, and merely represents a decrease upward in detrital material. For the purpose of mapping, the last quartzite bed more than two inches in thickness was taken as the top of the Monkton. Similarly the gray dolomites of the upper Winooski grade into the typical Danby with its accompanying white, vitreous, quartzite beds.

To date, no fossils have been found in this formation. The gradational relationship with the Lower Cambrian Monkton and Upper Cambrian Danby presents a certain age problem. It is of course possible that all or a part of the Winooski is of Middle Cambrian age. If, as suggested by Cady (1945, p. 531), the Monkton grades northward into the Parker Slate, the Winooski might be considered equivalent to the upper part of the latter. At South Wallingford, Stanley (1954) found evidence of a minor unconformity, or disconformity, between the cream and gray phases of the Winooski. He also found a significant change in heavy mineral content across this boundary. It is possible that the lower, cream-to-pink phase of the Winooski represents a continuation of Monkton sedimentation with a decreased amount of detrital material, and is, therefore, of the same age as the Monkton. If this is true, the upper, gray phase would represent the beginning of the Upper Cambrian, which during Danby times contained an appreciable amount of detrital material. Whether or not the pink- or cream-colored phase of the Winooski extends into the Middle Cambrian cannot be determined without paleontological evidence.

On the east side of Trumbull Mountain the formation is about 700 feet thick. Southward no estimates are possible because of poor exposures and faulting. It is probable that originally the Winooski, like the

Monkton, was cut out by the unconformity at the base of the Walloomsac.

Upper Cambrian Sequence

Danby Formation (Keith, 1932; Cady, 1945): In the Bennington region the Danby formation was recognized only in the northern portion east of Trumbull Mountain. It is a gray calcitic dolomite with interbedded white or gray vitreous quartzite. Southward the quartzite beds disappear, probably due to non-deposition, and the formation becomes indistinguishable from the lower part of the Clarendon Springs dolomite.

Although in Vermont no identifiable fossils have been found in the Danby, lithologic comparisons and the tracing of the beds into other areas indicate that the upper Danby, or Wallingford member, is of Franconian age, while the lower Danby is Dresbachian (Rodgers, 1937, p. 1575; Cady, 1945, p. 536).

The apparent thickness of the Danby where exposed is less than 300 feet as compared with the 1000 or more feet found by Brace (1950, p. 29) in the Rutland quadrangle.

Clarendon Springs Dolomite (Keith, 1932): This formation lies between the uppermost Danby quartzite and the first marble bed of the Canadian limestone. It is a massive, gray, somewhat calcitic dolomite. Probably the most distinguishing characteristic is the prevalence of quartz-filled geodes and knots distributed irregularly through the upper part of the formation. West of Maple Hill, just below the Canadian limestone, a sandstone bed was found in the upper Clarendon Springs. Also in the upper part, near the contact with the overlying marbles, it is not unusual to find beds containing large masses of gray-to-black chert. These beds in weathered outcrop or float present a highly scoriaceous appearance due to the removal of the dolomite in solution. This chert member present over most of the area, is an excellent marker for the top of the formation.

The age of the Clarendon Springs is probably Upper Cambrian, and lithologically is similar to the A zone of the Beekmantown (Brainard and Seely, 1890; Rodgers, 1937). It is very possible that much of the lower part of the Clarendon Springs in the Bennington region is equivalent to the upper Danby farther north, but is not recognized because of the absence of the quartzites and sandstones. According to Rodgers (1937), the lower four-fifths of the Clarendon Springs correlates with the Little Falls formation of the Mohawk Valley in New York. The upper part containing the massive chert is lithologically similar to the White

Hall formation and may be its correlative. Thus in this area, the Clarendon Springs may range in age from Franconian through the Trempealeauan.

The thickness of this formation is in the neighborhood of 1300 feet. This is considerably more than reported in the Rutland area (Brace, 1953) and in west-central Vermont (Cady, 1945). It is probable that much of the rock of Danby age in the Bennington region has been included in the Clarendon Springs. This would account for the abnormal thickness of the latter. East of Mann Hill it is covered by the Walloomsac and probably pinches out in this section.

Lower Ordovician Sequence

Canadian limestone: This unit lies stratigraphically between the Clarendon Springs dolomite and the Walloomsac slate. It would be equivalent to the calciferous limestone as mapped by Prindle and Knopf (1932). The term calciferous is not used in this report since as originally used in the early New York State reports, it included all rocks lying between the Potsdam sandstone and the Chazy limestone, thus including part of the Upper Cambrian. The best section is exposed on the east and northeast flanks of Carpenter Hill, located three miles south of the town of Bennington.

The lower part of this unit consists of interbedded, sandy marble and calcitic dolomite. The beds range from about one foot to three or four feet in thickness. In the type area the marble bands of this lower part are never more than ten feet in thickness. They are generally white to light gray, and are usually deeply weathered, often leaving a residue of clean white quartz sand. In a pasture just east of South Shaftsbury, the lowest part of this zone is well exposed. Here about 80 feet of the Canadian limestone outcrops above the Clarendon Springs. The characteristic dolomite zones as seen in the Carpenter Hill area are missing; and the section as exposed is almost all a white, gray or buff marble, containing thin chains of dolomite crystals.

Above the lower zone of interbedded marbles and dolomites there is a succession of dirty gray and banded gray marbles, alternating with bands of gray, calcitic dolomite. In the upper third of the unit, interbedded calcitic sandstones and quartzites are not uncommon. These are exposed in the northwestern corner of the area. At two horizons just south of West Mountain and on the eastern flank of Mount Anthony, beds of black phyllite were observed in the upper part of the limestones. As these beds could not be traced any great distance, they probably represent shaly lenses in the dominantly carbonate sequence.

Due to the lack of sufficient paleontological control, exact correlation of the Canadian limestone with the formations north and west of the Bennington area is not possible at this time. It seems probable that it is roughly equivalent to divisions B, C, D, and E of the Beekmantown (Brainard and Seeley, 1890) as Prindle and Knopf (1932, p. 273) reported finding fragments of trilobites of these ages in the vicinity of Bennington. The author has found several poorly-preserved, flat, coiled gastropods that are probably *Lecanospira*. Walcott (1888) discovered gastropods in limestone in the neighborhood of Pownal which he identified as *Hormotoma* (*Murchisoma*) *bellicincta* and *Loxoplocus* (*Murchisoma*) *milleri*, thus indicating a Trentonian age for the rock. However, from Walcott's paper and accompanying map, it is impossible to determine whether or not these fossils came from the Canadian limestone or from one of the large limestone lenses in the overlying Walloomsac slate. As several of these lenses are exposed in this area, the author believes that the latter is the case.

On the basis of this limited evidence, it is believed that these limestones are for the most part Canadian, although their upper part may include some Chazyan. They would be the stratigraphic equivalent of the Canadian carbonate formations of west central Vermont (Fig. 2).

The contact between the Canadian limestone and the underlying Clarendon Springs appears to be conformable throughout the area, and no evidence for a disconformity was found. There was apparently no break in sedimentation during the transition from Cambrian to Ordovician times.

The thickness of this unit varies considerably because of the unconformity between it and the overlying Walloomsac slate. Around the sides of West Mountain it may be as much as 2000 feet thick; in the Carpenter Hill area approximately 750 feet. Farther south, east of Mann Hill, it has been pinched out entirely by the Walloomsac. The above maximum thickness is of the same order of magnitude as that found by Cady (1945, p. 524) for the Canadian rocks in west-central Vermont. In the Castleton area Fowler (1950, p. 13) estimated a maximum of 1450 feet, somewhat thinner than the maximum in the Bennington region.

Middle Ordovician Sequence

Walloomsac Slate (*Prindle and Knopf, 1932*): The typical Walloomsac is a black, graphitic, sericite phyllite. While the rock occasionally contains thin, sandy laminae that serve to indicate the attitude of the bedding, the original bedding has generally been obliterated by later cleavage or foliation. Lenses of vein quartz are very prevalent, usually lying concordant with the foliation. These lenses vary in thickness from several inches to fifteen feet. Similarly, lense-like masses of blue, sandy, micaceous limestone are quite common. The larger of these usually contain beds of gray calcitic dolomite. These lenses vary in thickness from several feet to over 250 feet as in the case of the large mass exposed at North Pownal. These limestone bodies are quite similar in appearance to the underlying Canadian limestone, except that they contain many thin, micaceous seams that are particularly evident on a weathered surface.

The only paleontological evidence for the age of the Walloomsac is the presence of *Diplograptus foliaceous* found by Emmons (1844) in a slate quarry, near Hoosick, New York. This would indicate a Normanskill age for the formation. A few poorly-preserved crinoid stems were found by the author in the limestone lens at North Pownal, indicating a probable Ordovician age. The Walloomsac is considered to be equivalent to the Hortonville slate in west-central Vermont which is in turn correlated with the Canajoharie shale of the Mohawk Valley (Cady, 1945, p. 558). The included limestone would then correlate with the Whipple marble described by Fowler (1950, p. 32).

While the contact between the Canadian limestone and the overlying Walloomsac appears to be locally conformable, over a larger area it marks a major unconformity. Along the eastern flank of the Taconic Range the Walloomsac is found lying in apparent normal stratigraphic succession on the limestone section. Southeastward, just north of Mason Hill, it overlies the Upper Cambrian Clarendon Springs; on the western flank of The Dome, a small remnant of the phyllite with one of the characteristic limestone lenses lies on the Lower Cambrian Dunham dolomite. In the Hoosick Falls, New York, area Prindle and Knopf (1932, p. 274) reported the Walloomsac overlying limestones of Chazyan age. Thus the base of the Wallomsac lies on the truncated edges of progressively older formations going eastward.

This unconformity at the base of the Walloomsac or its equivalents has been recognized by many geologists working in western Vermont or neighboring regions. In the Castleton area Fowler (1950, p. 35) found the Hortonville slate resting on rocks ranging in age from Lower Cambrian to Trentonian. To the north Cady (1945, p. 559) recognized a similar break at the base of the Hortonville, while in other areas to the west and north the stratigraphic break beneath the Canajoharie has been well described (Ruedemann, 1901, p. 546, 1930, p. 104; Kay, 1937, p. 276; Clark and McGerrigle, 1936, p. 672). According to Currier and Jahns (1941, p. 1510), this unconformity may be present east of the Green Mountains. Kay (1937, p. 291) has suggested that a land mass, *Vermontia*, formed in the vicinity of the present Connecticut Valley during middle Trentonian times. It was possibly a source of the argillaceous sediment now found overlying the Ordovician and older carbonate sequences in western Vermont, southern Quebec and eastern New York. The fact that the pre-Trentonian uplift was apparently greatest toward the east and that here the Trentonian phyllites rest on the older carbonate formations is considered to be evidence in favor of Kay's hypothesis.

Mount Anthony Formation (new name): The term Mount Anthony is proposed here to include the argillaceous-schistose rocks overlying the Walloomsac slate in the Bennington area. While the name Taconic slate (Emmons, 1842) has been used in the past for at least some of these rocks, a new formational name is considered advisable because of the age and structural connotations inherent in the term, Taconic. The formation is best exposed on Mount Anthony and its southern extension.

The Mount Anthony is dominantly an altered argillite and sandy argillite. Occasional limestone and dolomite lenses are present, but are not as common nor as large as in the Walloomsac. In color, the rock varies from gray to green, with the green facies most extensive in the upper part of the formation. On the western margin of the area, it is a fine-grained sericite and chlorite schist often containing chloritoid. The chloritoid is generally found in rosettes with the crystals up to several millimeters in length. Going eastward, the formation changes into a chlorite-biotite schist containing many thin sandy laminae. Locally albite is well developed in the gray facies. Like the Walloomsac, pod or lens-like masses of vein quartz are common. In general, the rock has a much more granular texture than the Walloomsac, and is considerably harder due to the many thin, sandy layers.

Within limited field exposures, the contact between the Walloomsac and the Mount Anthony appears to be gradational. Nowhere was the writer able to draw an exact line of demarkation. However, over a larger area it appears that the Mount Anthony lies unconformably on the Walloomsac. Along portions of the eastern flank of Mount Anthony, the Walloomsac is absent, and the higher formation overlies the Canadian limestone with occasionally a very slight angular discordance.

The age of the Mount Anthony is unknown as no fossils have ever been found in it. The writer believes, however, that it is probably middle to late Ordovician, as it appears to overlie the Walloomsac in normal stratigraphic succession. The term, Mount Anthony, as used in this report includes much of what has been mapped as Berkshire schist (Dale, 1893) in western Massachusetts, and as Taconic slate (Emmons, 1842) in eastern New York and southwestern Vermont. In the Taconic quadrangle report, Prindle and Knopf (1932, p. 268, Fig. 2) mapped parts of the Mount Anthony formation as both Rowe schist and Normanskill shale.

The total thickness of the formation prior to erosion is unknown. On West Mountain and on Mount Anthony the maximum thickness preserved is in the order of 1250 feet.

Rocks of Uncertain Age

Hoosac Schist (Emerson, 1892): The Hoosac schist as exposed in the Hoosac Range of northwestern Massachusetts and southwestern Vermont is typically a black or gray, siliceous biotite schist containing abundant porphyroblasts of albite. The quartz is concentrated in thin beds giving the rock a finely laminated appearance.

According to Prindle and Knopf (1932, p. 285), near Heartwellville, Vermont, the Hoosac overlies a massive conglomerate containing "large pebbles and slabs of the underlying Stamford granite gneiss and Mount Holly gneiss, pebbles of blue quartz, microcline, and vein quartz, all in a cement mostly of biotite, quartz, and albite." This conglomerate apparently grades upward into the overlying schist. Farther south the normal Hoosac rests, unconformably or in fault contact, on the various formations of the Cambro-Ordovician carbonate sequence. In the vicinity of Spruce Hill, east of North Adams, Massachusetts, the Hoosac contains a zone of carbonaceous, garnetiferous albite schist at its base. The relationship of this rock to the rest of the formation and the underlying rock is not clear (Prindle and Knopf, 1932, p. 293).

The age of the Hoosac is open to question. The earlier workers in the region (Emerson, 1892; Pumpelly, Wolff and Dale, 1894) considered it to be Cambro-Ordovician. Emerson (1917, p. 40) on the basis of lithologic similarity correlated it with the Berkshire schist of western Massachusetts and eastern New York, which at that time was believed to be Ordovician, as it apparently overlies the Stockbridge limestone in normal stratigraphic succession. In 1932 Prindle and Knopf changed the age of the Hoosac to Lower Cambrian because of their structural and stratigraphic interpretations in the Taconic quadrangle. Northward in Vermont along the eastern flank of the Green Mountain anticlinorium, the Hoosac becomes the Grahamville formation which is considered by workers in that area to be Cambro-Ordovician in age (Billings and Thompson, 1952, p. 40; Brace, 1953, p. 43). It is unfortunate that the age of the Hoosac will continue to be in doubt unless identifiable fossils are found in it or it can be traced westward into formations of unquestioned age. On the basis of lithologic similarities, the writer feels that at least part of the Hoosac is equivalent to the Walloomsac slate and the lower Mount Anthony formation of the Bennington area.

METAMORPHISM

General Statement

Any discussion of the metamorphic history of the Bennington region must, of course, include both the pre-Cambrian and the Paleozoic rocks. The lower Paleozoic formations have here undergone a low-grade regional metamorphism. The pre-Cambrian strata give evidence of an earlier, more intense metamorphism, upon which that of the Paleozoic has been impressed.

A detailed study of progressive metamorphism in the Cambro-Ordovician rocks is not practicable here due to the outcrop pattern and the nature of the rocks involved. As can be seen by the map (Plate I), the various formations outcrop in bands running approximately north-south. The argillaceous rock of Middle Ordovician age is exposed along the western margin of the area. Unfortunately, with the exception of the small section of Hoosac shist in the southeastern corner of the map, these rocks do not outcrop far enough eastward to show any major change in metamorphic grade. Since the correlation between the Hoosac, the Walloomsac, and the Mount Anthony is still somewhat doubtful,

any tracing of progressive metamorphism in the Paleozoics across the pre-Cambrian core of the Green Mountain anticlinorium becomes of questionable value. Between the argillaceous rock on the west and the pre-Cambrian lie the dominantly carbonate and quartzitic sediments of the Cambrian and Lower Ordovician. Unfortunately, these lithologies are relatively poor indicators of metamorphic grade.

Metamorphism of Pre-Cambrian Rocks

The pre-Cambrian rocks exposed in the core of the Green Mountain anticlinorium have undergone at least two periods of metamorphism. During pre-Cambrian time the vast thickness of the Mount Holly series underwent a profound orogeny in which the rocks were converted to gneiss and schist of medium to high metamorphic grade. During late pre-Cambrian time the area was deeply eroded; and on the ancient surface the Cambrian and Ordovician sediments were deposited. During the latter part of the Ordovician, this region again underwent orogeny and metamorphism. This time the grade was less intense. The lower Paleozoic rocks show a low grade of metamorphism, while the previously altered pre-Cambrian strata give evidence of a retrograde metamorphism.

One of the most common rock types in the Mount Holly series is a quartz, biotite, microcline gneiss. The gneiss commonly contains graphite, and almandine garnet is present in thin bands. An extensive development of blue quartz is often associated with the garnet. Included in the gneiss are biotite layers also containing blue quartz. Much of the gneiss contains many thin bands of granitic material, which give the rock the appearance of a migmatite.

In many sections the Mount Holly contains interbedded amphibolites, showing actinolitic hornblende. Farther north, Brace (1953) reports the development of tremolite in the metamorphosed limestone beds and lenses. Completely recrystallized vitreous quartzites occur throughout the series.

The presence of almandine garnet and actinolitic hornblende would indicate a medium grade of metamorphism. The rocks have been elevated at least to the garnet zone or the amphibolite facies of Eskola (Turner, 1948).

The effect of the Stamford granite gneiss on the adjacent Mount Holly is not clear at this time. As stated above, the Stamford seems to be a concordant structure and probably represents a syntectonic intrusion.

It is very possible that there is a genetic relationship between the Stamford and the granitic layers or bands in the Mount Holly. No evidence was found in the latter indicating a higher grade of metamorphism along the contact with the Stamford. On the other hand, the decreasing grain size in the Stamford as the contact with the Mount Holly is approached, might well indicate a gradational or transitional contact.

During the latter part of the Ordovician, the area again suffered a period of metamorphism. This time the intensity was less than before, with the result that the medium-grade, metamorphic minerals in the pre-Cambrian rocks have been partly or wholly altered to those indicative of a lower grade. In general, the trends have been for a change from the mineralogy of the amphibolite facies to that of the green schist facies. The almandine garnet of the gneisses and the actinolitic hornblende of the amphibolites are generally rimmed and partly replaced by chlorite. Albite is common in association with epidote, zoisite, and calcite, possibly indicating the earlier presence of a more calcic plagioclase. The evidence for retrogressive metamorphism is not as clear in the Stamford granite gneiss. It is almost impossible to secure fresh samples of this rock, due to its extreme susceptibility to weathering. The freshest samples obtained showed the development of a fine, sericitic-like mica. How much of this was due to deep weathering, and how much to retrogressive metamorphism cannot now be ascertained. According to Turner (1948, p. 95), both biotite and microcline are stable in the green-schist metamorphic facies.

In conclusion, the pre-Cambrian rocks show evidence of two periods of metamorphism. The earlier period has elevated the rock to the garnet grade or the amphibolite facies. Later, during the late Ordovician, the region underwent a low-grade, chlorite zone of metamorphism with the result that the minerals formed during the earlier period are partly altered. The effect on the pre-Cambrian has been to approach a mineralogical equilibrium with the younger Cambro-Ordovician rocks of the region. During this later period, water and other volatiles have been active and available. Most of the reactions involved necessitated the addition of water for the formation of the new minerals.

Metamorphism of the Cambro-Ordovician Rocks

The Cambro-Ordovician rocks of the region have undergone a low grade of metamorphism. The mineralogy of the argillaceous rocks indi-

cates that they should be assigned to the green schist facies. On the western margin of the area, the Mount Anthony is composed of lustrous, green and gray, sericite-chlorite and sericite-chloritoid-chlorite phyllites or slates. The chloritoid generally occurs as porphyroblasts in the otherwise fine-grained rock, and often shows a radiating or rosette pattern. Following the formation eastward across the southern flank of Mount Anthony into the Pownal Upland, so named by Balk (1953), the rocks are observed to become coarser and more quartzitic. The greater amount of quartz in the eastern portion is probably due to a higher percentage of quartz in the original sediments. Along with the enlargement in grain size, biotite begins to make its appearance on the eastern flank of Mount Anthony. The fine-grained phyllites or slates of the west have become quartzitic biotite, chlorite schists. In the southern portion of the Pownal Upland, albite appears in the form of small porphyroblasts. While there is a definite change in the mineralogy and appearance of these rocks from west to east, it is probable that they are the result of compositional variations in the original sediments. A gradual increase in iron and decrease in aluminium going eastward would tend to suppress the sericite and promote the formation of biotite. The enlargement in grain size may indicate a slightly more intense metamorphism toward the east.

In contrast to the Mount Anthony, the Walloomsac shows very little change throughout the area. It reveals itself as a fine-grained black, carbonaceous or graphitic slate or phyllite, appearing to be of a lower grade of metamorphism than that of the overlying Mount Anthony. According to Eskola (1932) and others, graphite or other finely-divided, carbonaceous matter may inhibit the formation or growth of new minerals during metamorphism.

The carbonate rocks of the region have undergone varying amounts of recrystallization. The limestones are generally recrystallized. The dolomites and calcitic dolomites show little or no evidence thereof, and even where dolomite beds have been broken up and scattered through a highly deformed and recrystallized marble, little recrystallization has taken place.

Aside from the recrystallization of the limestone, a minor development of sericite is the only other evidence of metamorphism in these rocks.

The Cheshire quartzites and those in the Mendon are completely recrystallized. Wherever these rocks are composed largely of quartz, as in the case of the Cheshire, the resultant rock shows no macroscopic evidence of the individual grains and appears to be a homogeneous mass of quartz. Thin sections show a mosaic of interlocking grains. This complete recrystallization is present even in areas where the rocks have undergone relatively little deformation.

The presence of tourmaline in the Paleozoic rocks presents a problem as to its origin. Black tourmaline in large, euhedral crystals often occurs in the foliation and bedding planes of the Nickwacket. Throughout the Cambro-Ordovician sequence, tourmaline is the most common heavy mineral. Original detrital grains are present, as well as secondary euhedral crystals. Often the secondary tourmaline crystals have recrystallzed around one or more detrital grains. Small, euhedral, colorless tourmalines are prevalent throughout the Mount Anthony, and in certain zones make up three to four percent of the rock. By calculation, the amount of boron necessary to form this much tourmaline is many times that of a normal argillaceous rock (Rankama and Sahama, 1950, p. 491).

There are three possible explanations to account for the abnormal amount of tourmaline in the rocks of this region.

The first of these is that the sediments were derived from a source that was rich in tourmaline. This would probably mean a granitic area containing unusually large numbers of tourmaline-bearing pegmatites. If many of these sediments had their source toward the west, as is indicated by the westward increase in the sand content of the carbonate rocks, then such tourmaline-bearing igneous rocks should be exposed in the Adirondacks. However, although there are tourmaline-bearing pegmatites present in the Adirondacks, they constitute only a small proportion of the rocks exposed (Buddington, 1939). Moreover, it is interesting to note that while the most common Adirondack tourmaline is schorlite, the secondary tourmaline most prevalent in the Bennington region is the lithium-bearing elbaite. Detrital grains of schorlite are present in the insoluble residues of the carbonates. On the basis of our existing knowledge regarding the source areas for these sediments, the writer feels that the above explanation is insufficient to account for the amount of tourmaline present.

A second possible source of boron might be the sea water in the geosyncline during Cambrian and Ordovician times. If, as is believed by Kay (1951) and others, there was a zone of volcanic activity in the east-

ern part of the geosyncline at this time, it is possible that this volcanism was the source of appreciable quantities of boron. Normal sea water contains approximately .0015% B_2O_3. It seems feasible that during periods of intense marine volcanism the concentration might increase significantly over limited areas. According to Rankama and Sahama (1950), the boron, once in the sea water, could be precipitated as a calcium or magnesium borate, or it might undergo an exchange reaction in which the boron is picked up by fine detrital or precipitated material entering the area of deposition. Later, during the metamorphism of these sediments, tourmaline would be formed.

A third possibility is that the presence of tourmaline indicates widespread introduction of boron from an unknown source. According to Turner (1948) and Hutton (1939), extensive tourmalinization of schists other than of pelitic origin is the result of pneumatolytic activity from deep-seated or "distant granitic intrusions." Hutton (1939) reports the tourmalinization of rocks in the Otago region of New Zealand as much as 120 miles from the nearest known outcrop of granite. While there are no known Paleozoic granitic intrusions in the Bennington area, there are several dome-like structures near Wilmington, Vermont, (Skehan, 1953) with monzonitic or dioritic cores that may be of igneous origin. A similar structure is found around Shelburne Falls, Massachusetts (Balk, 1946). In west central Massachusetts, the Williamsburg granite is quite extensively exposed. Also a line of small mineral deposits extending down through eastern Vermont and west central Massachusetts gives indication of early or middle Paleozoic hydrothermal activity. According to this hypothesis, the rather selective distribution of tourmaline, a maximum in the Mount Anthony and a minimum in the Cheshire and Canadian limestone, could be explained on the basis of variation of lithologies and the favorability of certain rock types as a host.

The writer feels that, in view of available evidence, the third hypothesis is the most probable. The possibility of boron enrichment of sea water due to volcanic activity and its subsequent inclusion in the local sediments should, however, not be overlooked.

On a much smaller scale, hydrothermal activity is common in this area as a result of the free mobility of H_2O and CO_2 during metamorphism. The abundant quartz veins are undoubtedly a result of the leaching of silica from the surrounding rock due to metamorphism, and subsequent deposition as quartz in areas of least stress. In the Canadian limestone and the carbonate facies of the Walloomsac, boudinage blocks of calcitic dolomite in marble wherever deformation has been intense are generally surrounded by haloes of coarse-grained calcite and quartz with some microcline and sericite. The calcite and quartz have been derived by solution from the surrounding rock, and the microcline has formed due to the alteration of the sericite.

STRUCTURAL GEOLOGY

General Statement

The major structural features in the Bennington quadrangle are the Green Mountain anticlinorium and the eastern flank of the Taconic synclinorium. These comprise the major structures of western Vermont. In the former, the pre-Cambrian core is exposed along its axis for at least 100 miles, and has a maximum outcrop width of approximately 13 miles. The latter extends southward from the Champlain Valley for an unknown distance and, at its northern end, occupies the area between the Green Mountains and the Adirondacks.

Superimposed upon these larger folded structures are smaller folds, the largest of which are indicated on the tectonic map (Pl. 3). Smaller folds are superimposed on these down to five or six orders of decreasing magnitude. The scale varies from the major structures noted above down to minor plications measuring a half-inch or less from crest to crest.

The above folded structures have been in turn broken by high-angle reverse faults, overthrusts, and steeply-dipping normal faults.

While the many minor structures such as cleavage, minor folding, lineation, and jointing have been recorded and plotted, the emphasis in the delineation of the various structural units has been on stratigraphic relations. It is felt that in areas of rather limited exposure these smaller structures may be misleading unless supported by stratigraphic information. This is particularly true when the mapping is done on a relatively small scale. In general, the characteristics of the minor structures mirror and support those of the major structural features.

MAJOR STRUCTURES

The Green Mountain Anticlinorium

The Green Mountain anticlinorium is the major structural feature of the eastern half of the map area. It was first described by Adams (1846,

p. 167), and since then has been studied in varying degrees of detail by Keith (1932, p. 404), Cady (1945, p. 564), and Osberg (1952, p. 75), and by Brace (1953, p. 65), who in the Rutland area was able to determine in some detail the complex structure and stratigraphy of the pre-Cambrian rocks exposed in the core.

The proof of the anticlinal nature of the Green Mountains lies in metamorphic, structural and stratigraphic considerations. The pre-Cambrian rocks in the anticlinorium have undergone several periods of metamorphism (see above), and show evidence of having been raised originally to at least a medium grade. In contrast, the younger formations are of the chlorite-biotite grade, and have undergone only one period of intense deformation. The structural trends within the pre-Cambrian basement differ in part from those of the overlying Paleozoics. At some distance normal to the contact with the overlying quartzites these trends are predominantly toward the west and northwest, as opposed to the northeasterly trend of the Paleozoic rocks. Their significance will be discussed presently.

The quartzites overlying the gneisses have yielded Lower Cambrian fossils which clearly date the underlying rocks as pre-Cambrian. These quartzites wrap around the gneiss core of the structure just south of the map area in the vicinity of Williamstown and North Adams, Massachusetts, where the gneisses plunge beneath the younger Paleozoic formations.

In section, the anticlinorium tends to be asymmetrical with the steepest dips on the western flank. This is consistent with the smaller Paleozoic folds in the region. The eastern flank appears to slope off with moderate dips beneath the lower Paleozoic sediments. The western flank, on the other hand, is somewhat more complicated. Overturning of the pre-Cambrian-Cambrian contact and associated thrusting is not uncommon. At the northern boundary of the area the contact dips steeply westward. East of Bennington along City Stream, the Lower Cambrian sediments are overturned and the contact dips eastward. South of Bennington along the front of the Green Mountains there is no evidence of overturning until the vicinity of The Dome is reached.

The great quartzitic mass of The Dome is an overturned anticline of Nickwacket with the pre-Cambrian exposed in its core. This anticline has been thrust westward over the Lower Cambrian sediments in the valley (Plates 1 and 2). The southwestern end of the thrust was later cut off by the Reservoir Brook fault, and thus its continuation in this direction is not known. The thrust plane is nowhere visible, but outcrops of the over-ridden Walloomsac slate are found within a few yards of the Nickwacket. The thrust mass seems to have ridden on a phyllitic band in the Nickwacket. The approximate thrust contact can be followed northeastward until it disappears into the pre-Cambrian crystalline complex. Balk (1952, 1953) has mapped this thrust along the lower contact between the pre-Cambrian and the Nickwacket. However, as the overturned basal Nickwacket is present immediately beneath the pre-Cambrian contact, and above the Dunham and Walloomsac to the northeast, the plane of the thrust was located as is shown on the map (Plate 1).

To the east of The Dome there is an overturned syncline with the Cheshire exposed in its core. On the eastern flank of this fold the pre-Cambrian contact appears to dip toward the eastward.

Further evidence for the anticlinal nature of the structure has been supplied by gravity investigations in central Vermont (Bean, 1953). Bouguer anomaly profiles across the region show strong maxima over the outcrop of the pre-Cambrian along the crest of the fold. The asymmetry of the anticlinorium is suggested by the configuration of the above profiles (Bean, 1953, Pl. I). To the west of the gravity maxima the values drop off rapidly to a minimum in the Champlain Valley. Eastward the decrease is more gradual and reaches a minimum in the vicinity of the Connecticut River Valley. According to Bean (1953, p. 523), the observed anomaly is one that would be produced by a major warping of the earth's crust. He assumes that both the granitic and intermediate layers of the crust are involved.

While the original direction of the pre-Cambrian trends was west of north, along the contact with the overlying Cambrian strata a pseudo-conformable relationship is observed between the pre-Cambrian and Paleozoic rocks. This phenomenon was recognized by Balk (1936, p. 734) in Dutchess County, New York, and has been discussed in some detail by Brace (1953, p. 75) in the Rutland area. Brace recognized a marginal zone in the pre-Cambrian in which the Paleozoic trends are discernible. Within this margin there is a "central zone which seems structurally intact and may represent part of an ancient fold belt." The exposed Mount Holly in the Bennington area lies for the most part in the marginal zone. As the anticlinorium plunges beneath the over-

lying Paleozoics just south of the map area, most of these rocks lie within a few hundreds of feet from the original pre-Cambrian-Cambrian contact. In the vicinity of Glastenbury Mountain, however, the basement complex has been more deeply eroded and the central zone showing the earlier trend is exposed. The parallel alignment between the pre-Cambrian and Paleozoic structures near their contact has been explained by Balk (1936, p. 734) as being due to shearing within the pre-Cambrian rocks. Broedel (1937, p. 162) described the apparent structural parallelism along the contact between the Setters quartzite and the older Baltimore gneiss. According to his interpretation, the original folds in the gneiss have been stretched out along the flanks of the dome until the attenuated limbs of the folds became parallel to the original unconformity. This explanation is similar to that advanced by Eskola (1949).

The proximity of most of the Mount Holly to the overlying Cambrian is indicated by the patch of Paleozoic rock preserved near Woodford. Structurally it is a doubly-plunging syncline with the basal Cambrian sediments in the core. In section it is asymmetrical with the axial plane dipping steeply toward the eastward. The dip of this plane is somewhat steeper than that of the next syncline to the west, and may indicate a fanning of the axial planes across the anticlinorium.

The question arises as to what type of stress was responsible for this large structure. To the east, Skehan (1953), in studying the gneiss-cored domes of the Wilmington area, postulated that they were the result of vertical movements. The evidence for this was that the gneiss contacts on the margins of the domes and the surrounding sediments showed reverse dragfolding (Fig. 2). This seemed to indicate that there had been a movement of material up into the cores of these structures relative to the surrounding rock. The possibility of a similar origin for the Green Mountain anticlinorium was considered. However, in spite of the occasional small reverse drags observed along the western margin, the large folds of the structure are of the normal drag type, and would indicate that the dominant force was lateral compression. It may be that the small reverse drags are a reflection of a vertical component in the lateral compression, and that the net direction of stress was westward and up.

To summarize, the Green Mountain anticlinorium is a more or less typical anticlinorial structure, dipping moderately toward the eastward

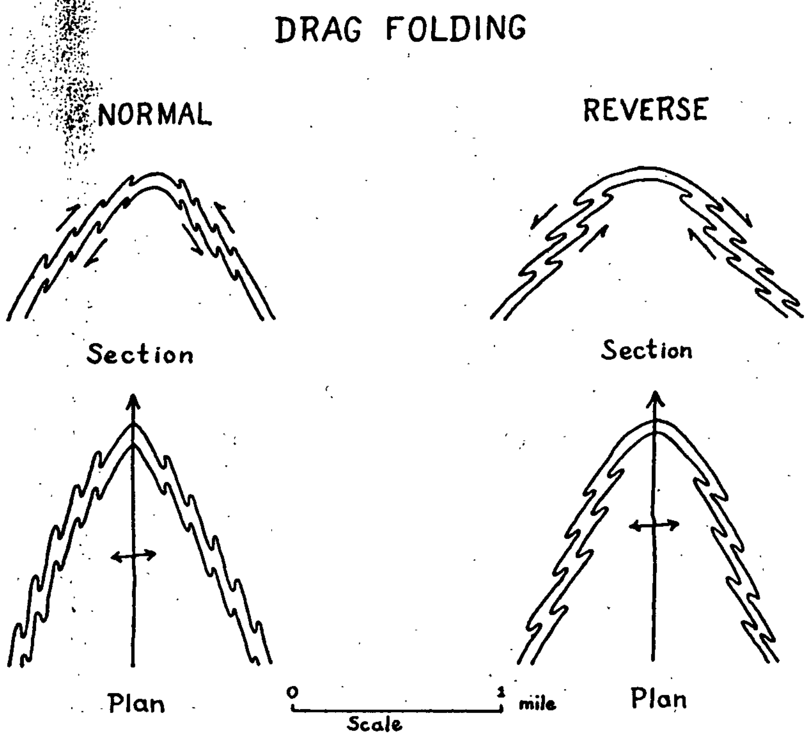

Figure 3

along that flank, and showing local overturning and associated thrusting on the western flank. There is evidence that the Paleozoic deformation has brought about a parallelism between the pre-Cambrian and Paleozoic structures in the marginal zone, while the central zone has been relatively undisturbed by later deformation. The structure as a whole plunges gently southwestward, and was probably formed by lateral compression.

The Taconic Synclinorium

West of the Green Mountain anticlinorium lies a broad complex synclinal structure that will be referred to as the Taconic synclinorium. It is without much doubt the southern continuation of the Middlebury synclinorium described by Cady (1945, p. 562). Only the eastern flank of the structure is present in the Bennington area. The western flank is

found along the complexly-faulted eastern margin of the Adirondack massif.

Within the map area its eastern flank includes the following structural units: the North Bennington anticline; the West Mountain and the Mount Anthony synclines; the Maple Hill thrust and syncline; the North Pownal overthrust; the structural and stratigraphic complex of the Pownal Upland; and the Reservoir Brook fault.

North Bennington Anticline: In the northwestern quarter of the map area a major anticline extends northeastward from the Walloomsac River through the village of South Shaftsbury to the northern boundary of the area. The author has named this structure the *North Bennington anticline*. The northern extremity is not known, but probably lies somewhere in the Equinox quadrangle. To the south, the structure disappears under the thick glacial debris in the valley of the Walloomsac River and farther southwest is obscured by poor outcrop and glacial drift. According to Prindle and Knopf (1932), it is terminated in this direction by complex faulting in eastern New York.

The best exposures are found north of South Shaftsbury along the eastern flank of the anticline. Here a fairly complete section can be seen, from the Dunham in the core of the fold up to the lower Canadian limestone. Going eastward, the structure is abruptly cut off by the Maple Hill thrust. This rupture brings the Lower Cambrian carbonate and quartzite rocks into juxtaposition with the Lower Ordovician limestones and dolomites.

The western flank of the fold is poorly exposed because of the thick cover of glacial sediment. It is not until the steeper portion of the eastern flank of West Mountain is reached that bed rock is again found. Here the section begins in the Canadian limestone and extends upward into the Mount Anthony formation which underlies the upper part of the mountain. The Cambrian formations of the western flank are presumably buried beneath the Pleistocene sediments.

The fold strikes about N30°E, and plunges gently toward the southwest. It is probably somewhat asymmetrical with the axial plane dipping steeply toward the east. The rocks on the eastern flank show much drag folding of several orders. The "wave lengths" of these folds vary from several hundred feet down to three or four inches. The amount of drag folding is not everywhere constant as many large dip slope exposures show no sign of it. Generally speaking, wherever the drag folding is intense, axial plane cleavage is well developed. The axes of the folds plunge either northeastward or southwestward, and locally are uniform in direction. The amount of plunge varies from horizontal to 25 degrees. This variation in the direction of plunge probably indicates a rather gently undulating axis of the major anticline.

The West Mountain Syncline: The West Mountain syncline is best displayed around the southern end of West Mountain in the northwestern corner of the map area. The axial plane appears to be vertical, and the fold plunges gently southwestward in accordance with the other larger structures of the region. While the fold as a whole is approximately symmetrical and shows little overall evidence of severe deformation, the strata involved have suffered a considerable degree of disturbance. In the limestones and dolomites overturned folds are common, and cleavage is often seen particularly in the thin dolomites (Plate V, fig. 2). Above the carbonate rocks, the phyllites in the core of the structure have undergone similar deformation. Whereas the "wave lengths" of the folds in the carbonates are in the order of several tens of feet to several hundreds of feet, those in the phyllites can be measured in inches. The contrast between the apparently simple structure of the syncline and the complex structures shown by the beds within it points to intense surficial deformation. This may be the result of a former northwest extension of the Maple Hill thrust with the beds beneath the overriding thrust plane undergoing intense deformation.

South of West Mountain the structure is not well exposed, but the phyllites occurring in this direction probably lie along its axis.

The Mount Anthony Syncline: The synclinal nature of Mount Anthony is well exposed around its northern end. On both flanks the various formations dip gently under the upper part of the mountain. The fold plunges toward the south, following the crest of the mountain and parallel to it. At the southern end the syncline has been obscured by the North Pownal overthrust. In this area a number of smaller asymmetrical folds have been formed, possibly as a result of the thrusting.

As in the case of the West Mountain syncline, both the carbonate and phyllitic rocks show considerable evidence of deformation. This deformation is, of course, most intense in the neighborhood of the thrust.

The Maple Hill Thrust and Syncline: The Maple Hill thrust follows the eastern flank of the North Bennington anticline until it is lost beneath the alluvium and glacial debris of the Walloomsac Valley. It appears to

be a high-angle reverse fault, and is named for the hill in the northern part of the region where it is best exposed.

The western flank of Maple Hill is underlain by the lower-most Ordovician limestones and dolomites. Just over the crest and on the summit, the Monkton formation is well exposed. While the thrust plane is not visible, the distance between the last outcrop of the Canadian limestone and the first of the Monkton is only a few yards. Both above and below the fault the rocks are tightly folded with some fracture cleavage, but no fault gouge or mylonite is exposed.

The thrust extends southwesterly from the above location, and is next picked up between South Shaftsbury and Bennington. Here, the Cheshire quartzite outcrops, apparently overlying first the Canadian limestone and then, slightly farther south, the Walloomsac slate (Plate I). Between this section and Maple Hill it is effectively concealed by drift, but probably passes just east of Harrington and Bucks cobbles.

The net slip of the fault is not known, but in the Maple Hill area the dip slip is probably in the order of 4500 feet. To the south this increases to about 6000 feet.

The eastern or hanging-wall side of the thrust has brought the western limb of the Maple Hill syncline up and over the eastern limb of the North Bennington anticline. This syncline appears in the vicinity of Maple Hill as a minor flexure in the Dunham between the thrust and the Green Mountain front. Southward, as the trace of the thrust swings westward away from the Front, the syncline broadens and rocks as young as the Monkton appear in its core.

Prindle and Knopf (1932) show the fault continuing southwestward into New York State, where it becomes involved with several other thrusts. The author could find no stratigraphic evidence for any continuation west of Mount Anthony, although it must be admitted that exposures in this region are poor. If the fault does not continue in this direction there is a distinct possibility that it swings southeastward under the Walloomsac River (Plate III), passing into a strike-slip fault or flaw. This solution is attractive as it would explain the sudden termination of both the Maple Hill syncline and the large anticline that plunges south from Bald Mountain (Plates II and III). These structures apparently disappear in the outskirts of Bennington. South of here, the stratigraphy shows that the rocks are going into a large syncline to the west under Mount Anthony.

The northward extent of the thrust is not known; and its delineation will have to wait until it can be traced in the Equinox quadrangle.

The North Pownal Overthrust: There is a small overthrust sheet well exposed along the northern side of the Hoosic Valley in the southwestern corner of the map area. This is the only flat-lying overthrust that is recognized in this region. The thrust plane can be seen along the foot of the cliffs overlooking the village of North Pownal, from which the structure takes its name. This plane is a gently-undulating surface, rising from beneath the valley fill east of the village, and continuing westward until it disappears into the air west of the town. Above the fault plane lies a large limestone lens of the Walloomsac formation (Plate IV, fig. 1), which has overridden the Mount Anthony. The latter is well exposed in a cut on the Boston and Maine Railroad just east of the North Pownal Station.

The present structural interpretation of the North Pownal region depends on the fact that the limestones exposed here are a part of an isolated carbonate lens within the Walloomsac slate.

Previous investigators (Prindle and Knopf, 1932; Balk, 1953) have assumed that these limestones belonged stratigraphically below the phyllites. Thus it was necessary for them to invoke various structural complications in order to get them into their present position. Prindle and Knopf (1932, p. 288) postulated a recumbent anticline, with the limestone in its core. This necessitated the recumbent folding of the "Taconic thrust" plane in order to explain the presence of the limestone overlying what they mapped as the Hoosac or Rowe schists. Balk (1953, Pl. 1) shows two thrusts associated with the limestone mass, one to the west and another on the east.

Evidence that the limestone at North Pownal is actually a lens within the black Walloomsac slates is abundant along its eastern and southern margins. The lower contact between the limestone and the black slate or phyllite is well exposed in a road cut on State route #346, 0.7 miles southeast of North Pownal. The contact dips westward toward the rising plane of the overthrust. At this location there is no evidence that this is a fault contact. In the abandoned quarry just above the road cut, and on the cliff face farther west, black phyllite layers are interbedded in the limestones. Northward along the eastern edge of the carbonates, this interbedding of the limestones and phyllites is again observed in stream exposures. In most cases the dip of the bedding is toward the west.

Figure 1. Limestone hill on North Pownal Thrust sheet. Photograph taken from the east.

Figure 2. Dolomite bed in Canadian limestone on the east flank of Carpenter Hill.

PLATE IV

Similar evidence is found along the northern margin. In all probability the eastern and northern limits of the limestones lie along the change in facies between carbonate and argillite deposition. Eastward only phyllites are found.

Limestone lenses of this size do not appear to be uncommon in the Ordovician argillaceous rocks. A similar lens in the Walloomsac is well exposed around the northern and western flanks of Mount Anthony, and is lithologically the same as that of North Pownal. It is very possible that these two lenses were once continuous, and have since been separated by the thrust.

T. N. Dale (Pumpelly, Wolff and Dale, 1894) proposed a similar explanation for the Bellow Pipe limestone on Mount Greylock in northwestern Massachusetts. His map and sections show it lying in normal succession between the Berkshire and Greylock schists. Like the limestones of North Pownal and Mount Anthony, the Bellows Pipe is somewhat micaceous and is apparently conformable with the surrounding schists. Prindle and Knopf (1932) rejected this explanation, and thus mapped it as the core of a recumbent anticline.

Toward the southeast the thrust disappears under the alluvium of the Hoosac Valley, and is not seen again. No evidence for it was found on the south side of the valley in the phyllite sequence; nor were any limestones noted in this section. The Walloomsac slate is exposed north of the limestone area on the thrust plate, the edge of which trends along the western flank of the mountain, finally dying out somewhere south of the summit of Mount Anthony.

Aside from stratigraphic evidence, there is ample structural evidence for thrusting. The limestones are severely deformed above the thrust plane at North Pownal. Extensive and complicated flow-folding is common. In many exposures the limestone layers look as if they had been kneaded like dough. The interbedded dolomites are often drawn out and broken in complex boudinage structures. On the northern side of the abandoned quarry mentioned above, a black phyllite layer has been so deformed by flowage that it looks like a pendant hanging on the limestone mass (Fig. 3). Along the North Pownal cliffs, the general structural trends in the limestone are somewhat west of north. Bedding and cleavage planes often show a lineation plunging eastward or east-southeastward, which may correspond to Balk's thrust structures (see below).

Intense deformation is readily apparent in the phyllites below the

thrust. Minor, steeply-dipping thrusts have buckled the main thrust plane in at least two localities. Balk (1953, p. 846) has made a detailed study of the smaller structural features in the previously-mentioned railroad cut at North Pownal. The phyllites exposed here are perhaps 60 to 100 feet below the thrust plane. He recognized (p. 847) here what he has called "regional and thrust structures." The regional structures are characterized by small overturned folds generally striking northeastward and plunging at low angles in either direction, and by occasional axial plane cleavage, dipping steeply toward the east or east-southeast. The thrust structures are revealed as shear zones dipping 30°–50°, with an "intense schistosity" that is later than the other structural elements. A strong lineation plunging in the dip direction is developed on this schistosity.

The writer is in agreement with Balk in that these shear zones and associated lineation are genetically related to the thrust. A similar lineation and shearing in limestone just below the Maple Hill thrust is seen on Harrington Cobble and at South Shaftsbury (Plate VII, fig. 2). He does not feel, however, that these structures are sufficient evidence for thrusting unless accompanied by corroborative stratigraphic information.

The Pownal Upland: Balk (1953) has given the name *Pownal Upland* to "a tract of high level ground that extends southward for seven miles from Mount Anthony, to connect with the Green Mountain front." The northern end is located on the Mount Anthony syncline; the southern end extends eastward onto the western flank of the Green Mountain anticlinorium and is terminated by the Reservoir Brook fault. Both the Walloomsac and the Mount Anthony formations in this region show a considerable degree of deformation. Intense crinkling of the thin quartzitic beds in the Mount Anthony is common, and slip cleavage is often observed. Limestone or dolomitic lenses in these rocks are complexly folded or are sheared with the development of a strong lineation in the dip direction. All the evidence points toward a higher degree of deformation than is seen on the western flank of Mount Anthony or on West Mountain. It should be kept in mind that the area is bounded both on the east and on the west by thrusts. It is not surprising that the relatively incompetent phyllites reflect the stresses that caused the thrusting.

The larger structures of the area are a series of flat southwestward-plunging anticlines and synclines as is indicated by the distribution of the two phyllitic formations (Plate I). However, it should be emphasized that the contact between the two appears to be gradational in the field, and that the map pattern is therefore only a rough approximation of the true condition.

Reservoir Brook Fault: A major break separates the Pownal Upland from the folded and thrust quartzites of The Dome. This structure can be traced southward into Massachusetts, and is again well exposed along U. S. Route 7, just south of Williamstown. Between Mason Hill and the next hill to the east, the fault follows Reservoir Brook, from which it takes its name. On the western bank of the brook and over Mason Hill, highly fractured Walloomsac slate is found in scattered outcrops. Above the eastern bank near the southern border of the area, the massive Cheshire quartzite forms a bold escarpment. Farther north along the same bank, outcrops of the Moosalamoo phyllite are encountered. Just north of these quartzites and phyllites, occasional blocks of highly slickensided or brecciated dolomites are found. The fault is lost under the glacial debris still farther north in the vicinity of Barber Pond.

The fault appears to dip steeply to the west with the rocks on that side being down-dropped. Exposed "joint" planes in the massive quartzites have a steep westerly dip, and strike parallel to the fault. These often show a high degree of polish. Nowhere else in the Bennington section has this phenomenon been observed in the Cheshire. These planes occasionally show striations or a step-like surface, indicating that the western block moved down and slightly toward the north.

At several locations smaller escarpments of the Cheshire, just east and above the main one, may indicate the presence of subsidiary faults in a zone paralleling the principal break.

The displacement on this fault is not accurately known, but is probably between 2500 and 3000 feet in the dip direction. The amount of strike slip is unknown.

This fault has been indicated as a normal fault on the map on the basis of surficial evidence cited above. However it may represent a secondarily steepened reverse fault along which, locally, the fault plane has been rotated from an easterly to a steep westerly dip. Evidence for this latter interpretation lies in the fact that its strike is parallel to most of the major regional structures and that it lies where the plunging nose of the Green Mountain anticlinorium juts out into the Paleozoic sediments.

This evidence suggests a genetic connection with the major fold structures.

MINOR STRUCTURES

Bedding: In most of the rocks studied the original bedding is indicated by readily visible lithologic variations. This is particularly true in the case of weathered outcrops. Here, for example, sandstone bands in the dolomites and phyllites stand out in bold relief. The Walloomsac and Cheshire formations, on the other hand, show relatively little variation over considerable thicknesses of section; therefore the attitude of the bedding may be in doubt. In the Cheshire the highly developed joints may easily be mistaken for bedding, while in the Walloomsac often the only visible structure is the easterly dipping cleavage. Occasionally planes separating beds of contrasting competency are crumpled or otherwise disturbed due to the different behavior of the two rock types upon folding.

Bedding in the pre-Cambrian may be indicated by the distinct banded characteristic of the gneisses and schists. It is quite possible that the different lithologies observed are due in part to metamorphic differentiation (Turner, 1948, p. 137).

Cleavage: Both slip cleavage and fracture cleavage have been recognized in the Bennington region. The former appears in the argillaceous rock; the latter is generally confined to the more competent dolomites and dolomitic quartzites (Plate V, fig. 2).

Billings (1947, p. 217) has defined slip or shear cleavage as a cleavage in which there is differential movement along the fractures, and the "platy minerals may be dragged into parallelism with the cleavage." According to Mead (1940, p. 1011), it is generally oriented about 45° to the axis of the "compressive resultant force," and the individual shear surfaces can be considered as minute thrust faults. White (1949, p. 593), in a detailed study of slip cleavage and schistosity in the argillaceous rocks of eastern Vermont, came to the conclusion that under varying conditions such cleavage "may be either normal or oblique to the axis of maximum total shortening in any given unit of rock." He demonstrated that in the development of slip cleavage two components of deformation must be considered: 1) the displacement along the slip planes which is a shear phenomenon, and 2) the shortening at right angles to the cleavage planes which is the result of flowage. He has shown

(p. 591) that there is a transition between slip cleavage and schistosity, with the latter representing the ultimate stage in the development of the former.

In southwestern Vermont and western Massachusetts, slip cleavage in the phyllitic rocks was first recognized and described by Dale (1894, p. 139; 1896, p. 563). It appears as a minor parting, generally parallel or subparallel to the axial planes of the small folds. The slippage and dragging of the micas in the more quartzitic facies of the Mount Anthony formation is readily seen under the microscope. In the Walloomsac and the highly micaceous western facies of the Mount Anthony, this phenomenon is not so evident due to the fine-grained nature of the rock and the absence of thin quartzite bands. The mica in these rocks appears to be oriented parallel to the cleavage surfaces, and may represent a type of schistosity.

In both the Walloomsac and Mount Anthony there is a considerable variation in the attitude of the cleavage planes. On an average, they strike east of north, and dip approximately 25° toward the east. Neither formation differs significantly from the other in this respect.

Fracture cleavage as defined by Leith (1905, p. 119) was seen only in the carbonate sequence. It is most commonly developed on or near the crests of smaller folds in the dolomites and dolomitic quartzites, where it is invaluable for the determination of tops and bottoms of beds. In the Canadian limestone it is occasionally developed in limestone layers between beds of dolomite.

Joints: Joints are a conspicuous feature in the Paleozoic rocks of the area. No generalizations can be made regarding the jointing in the pre-Cambrian due to the limited number of observations.

The Cheshire quartzite is by far the most highly-jointed rock studied. There are always at least two directions of jointing present; these two generally strike at right angles, one parallel to the fold axis, and the other normal to it (Plate V, fig. 1). They usually show quite steep dips. The cross joints, those normal to the fold axes, probably have formed as a result of tension or stretching parallel to these axes. The longitudinal joints, parallel to the fold axes, are due either to stretching during folding or to release of load. Shear and cross joints are common on the crests of tightly folded anticlines. Often both shear directions are visible. The angle between the shears in the probable direction of stress varied from 75° to 99°. While these angles are somewhat larger than the theoretical

Figure 1. Bedding plane exposure of Cheshire quartzite showing two directions of jointing.

Figure 2. Fracture cleavage and recumbent folding in the Canadian limestone in the core of the west mountain syncline.

PLATE V

60° it is possible that the joint planes have undergone rotation as a result of continued stress.

According to Bucher (1920, p. 718), the normal to the line of intersection of these shears is the attitude in space of the effective direction of stress at that locality. It is interesting to note that in the two cases where this was tested the apparent stress "plunged" in the same direction as the dip of the bedding. In one case, on the western flank of an anticline, the indicated stress plunged down toward the west. In the other, on the eastern flank of a fold, the plunge was up from the east. This suggests the apparent rotational effect of stress during this type of folding.

In the higher formations the same joint directions are present, but not so well developed. This is possibly due to the fact that these rocks, being less competent than the massive quartzites of the Cheshire, tended to yield to stress more by flowage than by fracture. In general, the cross joints are the most obvious.

Folds: The folds of the region are the dominant and most conspicuous of the structural features present. They vary in size as has been stated above, from those of the first order, i.e., the Taconic synclinorium and the Green Mountain anticlinorium, down to those of the fifth or sixth order showing wave lengths or crestal distances of less than an inch. The latter is most characteristic of the higher phyllitic formations. In general the geometry is that of westward overturned to recumbent structures, with a gentle to moderate plunge toward the southwest. Where quartzites are interbedded with dolomites as in the Monkton formation, these folds may approximate the textbook illustrations of chevron folding. While the anticlinal and synclinal axes are not perfectly angular, they closely approach this condition. Isoclinal folding is rare in the Paleozoic rocks, and was recognized only in the immediate vicinity of the thrust faults. It may, however, be present in the pre-Cambrian.

In any section showing moderate to high degree of deformation it is difficult to classify the folds as to type. In general those of the carbonate and quartzite sections would correspond most closely to flexure folds (Billings, 1947, p. 87). The more competent rocks like the Cheshire and Nickwacket quartzites give little evidence of flowage. On the other hand, the Cambrian dolomites and Ordovician limestones or marbles have undergone some flowage with a resulting thickening on the crests of the smaller folds, and a corresponding thinning on the limbs (Plate VI, fig. 2). In formations having considerable variation in lithology,

Figure 1. Weathered siliceous zone in the upper part of the Clarendon Springs dolomite.

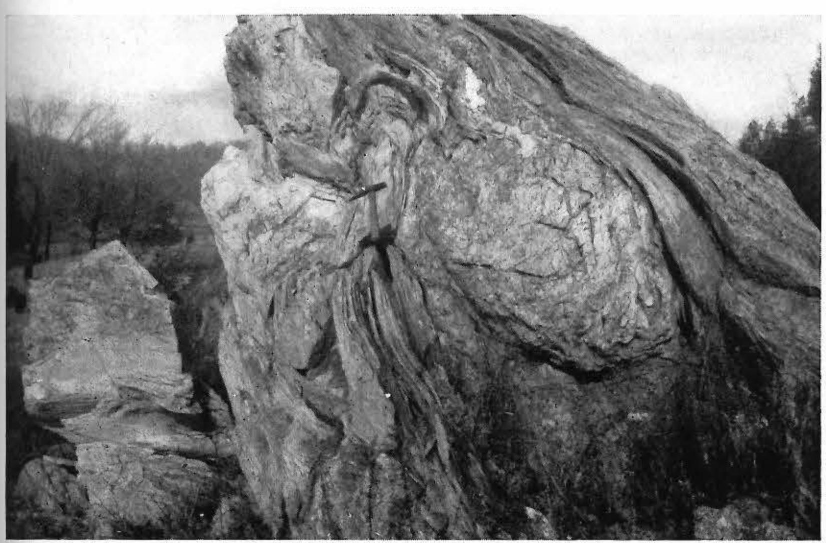

Figure 2. Complex folding in the limestone facies of the Walloomsac slate.

PLATE VI

Figure 1. Extreme boudinage in the Canadian limestone just below the Maple Hill Thrust at South Shaftsbury, Vermont, showing quartzcalcite replacement rims around the ruptured dolomite. A thin bed of dolomite has been broken up into isolated blocks as the result of flowage in the surrounding marble.

Figure 2. Fracture cleavage in sandy marble at the base of the Canadian limestone beneath the Maple Hill Thrust.

PLATE VII

most of the flowage occurs in those beds having the lowest competency. Thus in the Monkton formation the quartzites are relatively competent and show little flowage, while the dolomites have suffered thickening and thinning. In the Canadian limestones and in the carbonate facies of the Walloomsac, on the other hand, the dolomite is the competent lithology. It has been bent and sometimes broken. The limestones and marbles have adjusted by flowage. Extreme examples of this can be seen in the marble exposed 0.25 miles east of South Shaftsbury and in a recrystallized limestone lens in the Walloomsac at Pownal. In both cases the flowage of the marble has been so intense that interbedded dolomites have been broken up and distributed at random through the rock (Plate VII, fig. 1), thus making it impossible to trace out the trend of the original beds. In the South Shaftsbury example this extreme flowage is probably due to the proximity of the Maple Hill thrust, while the Pownal case is due to the effects of shearing in the phyllites.

Along Broad Brook just east of The Dome in the lower quartzitic facies of the Moosalamoo, extremely complex flow-like folding was observed. As similar structures were not seen in similar lithologies elsewhere, it is suggested that these may well represent subaqueous slumping or sliding of the sediments at the time of deposition (Hills, 1953, p. 16).

In both the Walloomsac and Mount Anthony formations the smaller folds appear to be due, at least in part, to shearing. While the displacement along any given shear or slip cleavage plane is on a microscopic scale, the total amount of movement over a larger area may be considerable. It is probable that the folds in these formations began as flexure folds, and that when stress along these planes was relieved by the development of cleavage, further folding was accomplished by shearing.

Throughout the region the evidence points toward an increasing intensity at least in the minor folds, going progressively upward in the stratigraphic column. This phenomenon is possibly due to a combination of two factors: first, that the most competent rocks lie in the lower part of the section, the least competent at the top; second, that the apparent local stress causing the deformation was due to the Green Mountain anticlinorium. During the formation of this structure, its western flank was locally overturned and not uncommonly thrust westward. Thus the least competent rocks were caught in the core of the synclinorium forming on the west, and yielded by intricate crinkling, shearing, and overthrusting.

Figure 1. Contact between the limestone and phyllite facies of the Walloomsac slate in road cut on Route 346, 1 mile east of North Pownal, Vermont.

Figure 2. Typical exposure of Mount Anthony schist showing abundance of small quartz veins.

PLATE VIII

Lineation: The fold axes are the major linear elements in the Bennington section. The general trends and plunges of the more important of these can be seen on the tectonic map (Plate III). While most of them are roughly parallel, there seems to be some convergence in the vicinity of Bennington. This is probably related to the dying out of the large Anticline underlying Bald Mountain. The strikes of the minor fold axes tend to parallel those of the larger folds although local variations are common. Reversals in direction of plunge occur within relatively short distances. All of the axes, both large and small, with the exception of minor folds in the North Pownal area, are consistent with the trend of the Green Mountain anticlinorium. In the North Pownal area the minor fold axes of the limestone along the southwestern margin of the thrust plate trend northwestward as contrasted with the generally northeasterly trend elsewhere. This local variation in trend direction is undoubtedly related to the differential movement within the overthrust. It is suggested that while the general movement of the sheet was westerly as indicated by the lineation in the underlying phyllite, there were local deviations in the apparent direction of movement along the edges.

The intersection of bedding and fracture cleavage in the carbonate sequence gives rise to a strong lineation on the bedding planes trending more or less parallel to the axes of the folds. Where the cleavage is parallel to the axial planes, this lineation is particularly useful in the determination of the attitude of these fold axes. Minor lineation in the form of striations on the bedding or cleavage planes is general. Most of it is normal or nearly normal to the linear trace of the fold axis and represents differential slippage between the various beds or cleavage plates as the result of folding. As has been discussed above, a strong easterly plunging lineation is found wherever thrust faulting has been important.

Faults: Several possible minor faults were observed in this area. Between Harrington Cobble and Hale Mountain a thin section of the Clarendon Springs dolomite seems to be in fault contact with the Monkton quartzite. It is possible, of course, that the contact represents a local unconformity between the Lower and Upper Cambrian rocks. However, as no such unconformity has been recognized elsewhere, it seemed preferable to map it as a fault with question marks. It is probably a normal fault with the nearly vertical fault plane dipping steeply to the east. Such 'normal' faults trending like the thrust faults, but indicating 'stretching,' are commonly found locally in strongly folded belts.

On the northeastern flank of Mount Anthony the Canadian limestone may be locally faulted against the Walloomsac slate as the Walloomsac at two locations appears to pass beneath the older limestones. The apparent faulting, however, may be due to complex folding. Balk (1953, Pl. 1) shows these in his cross sections as high-angle reverse faults although the attitude of the fault plane cannot be ascertained from field observation.

In the southeastern corner of the map area near Heartwellville the Hoosac schist has probably been thrust over the Stamford granite gneiss and the Cheshire quartzite. Fault breccia is present, but cannot be traced any great distance. In the Wilmington quadrangle Skehan (1953) found no evidence for faulting along this contact. Farther south toward North Adams, Massachusetts, this zone is completely hidden by glacial drift.

The Taconic Thrust Problem

The idea of the Taconic overthrust was first announced by Keith (1912). It is apparent, however, that what Keith (1913) had actually discovered was the pre-Walloomsac or Hortonville unconformity. In recent years (Fowler, 1950; Billings and Thompson, 1952) the tendency has been to place the thrust contact between the Trenton slates and the overlying phyllites of the Taconic Range.

Following Keith's lead, later workers in the Taconic area extended the proposed "thrust sheet" southward and westward until it covered much of the area along the New York–New England boundary. In 1932, Prindle and Knopf announced the presence of the Taconic thrust sheet in the Taconic quadrangle, of which the Bennington area is a part.

According to the proponents of the thrust hypothesis, much of the argillaceous rock (referred to as the Taconic sequence) exposed in the Taconic Mountains and the area to the west is an allochthonous sheet which has been overthrust from the east and is now lying unconformably on the autochthonous carbonates and shales of western Vermont and eastern New York. This sheet has apparently been separated from its roots by the folding which exposes the pre-Cambrian of the Green Mountain anticlinorium and similar anticlinal structures farther south. As now mapped (King, 1945) these rocks form a large klippe measuring more than 100 miles in a north-south direction and 40 miles from east to west.

In order to prove any such structure as indicated by the advocates of

the Taconic thrust, it must be demonstrated that: 1) at least some of the rocks of the "thrust sheet" are older than those underlying it; 2) there is a discordant relationship between the "autochthonous" and "allochthonous" rocks; and 3) there is evidence of thrusting along the contact between the two sequences of rocks. These three conditions have been demonstrated along the western margin of the supposed Taconic klippe in eastern New York, but neither the first nor the third has been unquestionably proven along its eastern margin in western New England.

Thus, it is quite possible that the thrusts which bound the area of the so-called Taconic klippe on the west are of the same type as those seen farther north, i.e., do not encircle the clastic sequence as envisaged in the klippe hypothesis. In that case, however, the presence of the "anomalous" Cambro-Ordovician series will have to be explained, fitting it into the areal sedimentary and stratigraphic picture.

Two possibilities have been indicated so far. One is that the clastic sequence contains the time equivalents of the surrounding carbonate sequence and is the result of peculiar conditions of sedimentation. While this is at present merely a speculation we must realize that it cannot be rejected on the grounds of sedimentary principles. It will be well to examine the idea, beginning with the basic fact of the areal distribution of the facies.

It has been long observed that a dominantly calcareous and quartzitic facies of Cambro-Ordovician rocks is found exposed in two more or less parallel belts. One of these is just west of the Green Mountain anticlinorium, while the other lies along the eastern margin of the Adirondacks. North of Rutland these belts merge in the Champlain Valley. Southward they are separated by an extensive area of argillaceous rock also of Cambro-Ordovician age. Many of the workers in this area have assumed that the previously mentioned carbonate quartzitic facies extends beneath the argillites, and that the two bands represent the outcroppings of a continuous sequence. If this is the case, the Cambro-Ordovician slates and shales must have reached their present position via thrusting from some other area.

To the east of the Green Mountain anticlinorium the Cambro-Ordovician rocks are dominantly phyllites and schists. Consequently this area has been assumed to be the root area for the argillites of the Taconic sequence.

If the Taconic thrust hypothesis is to be rejected, some other explanation must be advanced for the "anomalous" presence of the Taconic slates and phyllites. Weaver (1953), as a result of his work in the Copake quadrangle, New York, has suggested that the Cambrian shales of the Taconic sequence are actually autochthonous and represent a central argillaceous facies of the carbonates exposed both to the east and west. On the basis of heavy mineral analyses, he felt that he could correlate the impure sandstones of the Cambrian carbonates with similar sandstones in the shales. Weaver's hypothesis suggests that during the Cambrian and at least part of the Ordovician there was a trough between the eastern and western belts of carbonates which was receiving dominantly argillaceous sediment. This trough was probably somewhat deeper than the areas on either side, and provided a basin of deposition below effective wave base. Fine sediment from the craton to the west was carried across the relatively shallower area along the ancient coast line by current or wave action, and settled out in the deeper water farther east. During times of lowered sea level or of uplift on the craton, coarser detritus in the form of sand deposited along the coast and occasionally spread across the central trough and into the area of carbonate deposition to the east. This would give rise to the sandstone layers that are so commonly found in the Cambrian dolomites of western Vermont.

Until recently at least, geologists have been reluctant to accept the idea of rapid changes of sedimentary facies. Much of this reluctance is probably the result of the study of sediments on the stable craton where such abrupt changes are not common. On the other hand, in a subsiding and active orthogeosyncline it is not unreasonable to expect sudden and profound changes in sedimentary facies. It is possible that this phenomenon has not been recognized more in the past because of the outcrop pattern of rocks in these areas. In general continuous exposures of a given time-stratigraphic unit are parallel, not normal, to the trend of the original geosynclinal trough because of the later folding and faulting characteristic of these structures. Kay's (1951) distinction between the eu- and the miogeosynclines is based on a major change of facies across the orthogeosyncline. In the miogeosyncline carbonates, quartzites, and subgraywackes are common. Volcanic and other igneous rocks are absent or rare. In the eugeosyncline the dominant rock types are argillites and graywackes with associated volcanics. Thus there occurs along the margin between the eu- and the miogeosyncline a profound change in facies. It is suggested that similar changes on a smaller scale may be present

within the two types of geosynclines. Studies of recent marine sediments along the California coast and in Indonesia, both tectonically active areas, show a considerable variation in types of present bottom sediments. These preserved in lithofied form would constitute rather rapid changes of facies. In the Indonesian area Neeb (Kuenen and Neeb, 1943, Plate I) shows a band of terrigenous mud with globigerina ooze and calcareous mud on either side extending southeastward between Ceram and New Guinea. Here the scale and relationship between lithologies are approximately the same as in the Taconic region. Along the southern California coast a wide variety of bottom sediment is present (Revelle and Shepard in Trask, 1939, p. 248). In the shallower water, sand, sandy muds and calcareous detritus predominate, while fine-grained muds have accumulated in the local depressions on the sea floor. In this area the distribution between coarse and fine material is a function of the depth of water rather than of the distance from the shore.

According to the present interpretation none of the phyllites of known Cambrian age are exposed in the Bennington area. West of the Green Mountains rocks mapped by Prindle and Knopf (1932) as Cambrian Hoosac or Rowe schists or as Normanskill slate of the overthrust sequence are now included in the Trentonian Walloomsac or Mount Anthony formations (Prindle and Knopf, 1932, fig. 2; and this report, Plate I). According to the facies shift hypothesis, this change would occur along a north-south line just west of the Bennington area. The Cambro-Ordovician shales and quartzites outcropping west of Hoosick Falls, New York, would thus be the stratigraphic equivalents of and continuous with the carbonates found in the vicinity of Bennington.

An entirely different solution to the problem of the "Taconic clastic series" is based on the presence of two important unconformities in the Copake and Kinderhook Quadrangles, N. Y., recently mapped by Weaver and Craddock, respectively. These are said to lie at the base of the Schaghticoke-Deepkill-Normanskill sequence on the one hand and the base of a Middle Trenton shale sequence on the other, equivalent to that above the Rysedorph conglomerate in the Capitol district, New York. These unconformities are said to have brought transgressing Ordovician slates into contact with Lower Cambrian clastics exposed along major anticlinal axes, giving the impression of a "Cambro-Ordovician" clastic series corresponding to the "Cambro-Ordovician" calcareous sequence. Without the maps and texts of the two quadrangles no adequate picture of the resulting interpretation can be made at this time. The writer was surprised to learn of their alleged discovery, since he has found two unconformities in the Bennington Quadrangle, one below the Walloomsac slate, the other beneath the Mount Anthony phyllite.

In the Bennington region the Mount Anthony overlies the Wallomsac with a regional unconformity. It might be argued that the contact between the two formations is a thrust plane, and that the higher phyllites and schists are part of an allochthonous sheet that has overridden the carbonates and slates of the Bennington Valley. Fowler (1950, p. 65) in the Castleton quadrangle found a similar relationship between the underlying Hortonville and the eastern part of the Taconic sequence. On the basis of his structural and stratigraphic interpretation of that area, he mapped this contact as the trace of the Taconic thrust plane (Fowler, 1950, p. 66).

The writer on the other hand believes this contact is a simple unconformity, as no evidence of faulting was observed. As previously mentioned, the underlying Walloomsac appears in field exposures to grade up into the Mount Anthony with no discernible break or evidence of thrusting. Wherever the Mount Anthony is in contact with the Canadian limestone or the lower carbonates, there is no brecciation or mylonitization along the contact as might be expected if this contact were the sole of a major thrust. The structural phenomena observed above and below the plane of the North Pownal overthrust is completely lacking along the Mount Anthony-Canadian limestone contact. Even more important, there is no structural discordance between the two formations. The Mount Anthony has been involved in the same folding and to the same degree as the Walloomsac, and gives every evidence of having been subjected to the same stresses as have affected the latter. If the Mount Anthony is a part of a thrust sheet, it must have been emplaced prior to the major period of deformation in the area. Prindle and Knopf (1932) recognized this fact as they show the recumbent folding of their Taconic thrust plane in both the Mount Greylock and North Pownal regions. Under the present structural interpretation the movement on the North Pownal overthrust must have taken place following any possible thrusting of the Mount Anthony, as the Walloomsac is found overlying the higher formation. As a result of the above evidence, it is felt that the contact between the two phyllite formations is an unconformity, and

that to consider it as a thrust is to introduce unnecessary structural complications for which there is no local evidence.

The known thrusting along the western margin of the Taconic area (Cushing and Ruedeman, 1914; Ruedeman, 1930, 1942; Goldring, 1943; Kaiser, 1945) and the complete lack of evidence for it on the eastern margin is in accordance with the present view of the local structural and stratigraphic history. The author feels that the thrust faulting in western Vermont and eastern New York can be divided into two major categories: 1) a deep-seated thrusting involving the pre-Cambrian basement, and 2) a rather superficial overthrusting that is probably confined to the Paleozoic sediments.

In the Bennington area the first type is exemplified by the Maple Hill thrust and the thrust underlying The Dome. These ruptures extend into the basement complex, as the pre-Cambrian or the Lower Cambrian formations are exposed along the traces of the faults. Similar thrust faults are known farther north along the Green Mountain front. In the Rutland area the Pine Mountain thrust is of this type and a similar thrust extends along the western side of Lake Dunmore, near Brandon, Vermont. All of these structures are situated on the western limb of the Green Mountain anticlinorium. They are undoubtedly associated with the formation of this structure and represent the ultimate relief of stress within the basement by fracturing following the major folding of the anticlinorium.

The second group of thrust faults appear to have no connection with the underlying pre-Cambrian. They can be thought of as relatively thin sheets or slices of the Cambro-Ordovician sediments that have slid westward as a result of the westward overturning and thrusting of the Green Mountain anticlinorium. The observed faulting along the western margin of the Taconic area would fall into this category. Similarly the major ruptures such as the Orwell and Champlain thrusts along the western side of the Middlebury synclinorium may be of this type. In the Bennington region the North Pownal overthrust is a small but significant example of this kind of structure. Rodgers (1953) has considered the faulting of the Appalachian Valley and Ridge provinces from the point of view of the basement and no basement control theories of deformation. He concluded (Rodgers, 1953, p. 164) that the evidence favors the latter theory, and that these faults represent an imbricate type of shearing in the Paleozoic sediments. The shear planes presumably do not extend into the basement, but instead become parallel at depth to the pre-Cambrian contact. In his cross section of southern Pennsylvania (Fig. 2), he indicates that the westward thrusting of the pre-Cambrian of the South Mountain anticline has provided the necessary stress for the thrusting farther west. The structural setting here is very similar to that of western New England.

STRUCTURAL AND STRATIGRAPHIC SYNTHESIS

The Cambrian of southwestern Vermont began with the deposition of the Nickwacket and Moosalamoo members of the Mendon formation. The Nickwacket is composed of detritus eroded from the local pre-Cambrian rocks. Pebble bands in the lower part of the unit and the presence farther north of coarse conglomerates at its base indicate a nearby source. The period of Nickwacket deposition was followed by the laying down of the finer sediment of the Moosalamoo. The finding, north of Rutland, of the Forestdale dolomite between the two units may mean that they are separated by a minor disconformity. It is probable that during this time locally higher areas on the old, pre-Cambrian erosion surface provided most of the sediment. As the onlapping Cambrian seas became deeper and these island sources were gradually worn away, the local supply of sediment came to an end.

Following the Mendon sedimentation there may have been a brief period during which the sea withdrew to the east, and the newly deposited Moosalamoo was subjected to the forces of erosion. The sea returned, but this time the source area was the craton. The clean, well-sorted sand of the Cheshire spread across the subsiding miogeosyncline, forming the basal sand that is found everywhere beneath the later sediments. At the same time similar subsidence was taking place in the easterly eugeosyncline. The source of these sediments was possibly from local land masses or islands along the present eastern margin of the North American continent (Kay, 1951). Many of these islands may have been of volcanic origin. This detritus was dominantly argillaceous as contrasted to the arenaceous and calcareous material being deposited at the same time in the miogeosyncline. It is possible that these two geosynclinal troughs were separated by a geanticline along the present line of the Green Mountain anticlinorium. Prindle and Knopf (1932, Fig. 2) show a decided thinning of both the Cambrian and Ordovician carbonates across the axis of this structure.

Some time shortly after the deposition of the Cheshire, a minor trough developed down the middle of the miogeosyncline. While the slightly shallower areas on either side were receiving calcareous detritus and minor amounts of clean sand from the low-lying craton, muds and sand were accumulating in the central depression. This condition continued throughout the Cambrian and possibly into the Middle Ordovician. There is some evidence, however, that this pattern was interrupted during the Middle Cambrian by a general withdrawal of the sea from the miogeosyncline. A period of erosion would have followed, and thus the Lower Cambrian may be separated from the Upper by a major disconformity.

During this period of general geosynclinal subsidence, the calcareous margins of the miogeosyncline were probably rather shallow banks. Reef-like structures may have been common. The prevalence of cross-bedding and the general absence of fine continental detritus would indicate that wave and current action was continually sweeping these areas. To the east in the eugeosyncline the depth of water was considerably greater. Thick deposits of muds and impure sands accumulated, interspersed with the products of volcanism.

The relationship of the Hoosac and Rowe schists exposed east of the Green Mountain anticlinorium to the phyllites and schists of the Bennington area is not precisely known because of the uncertainty of their ages. It would appear, however, that throughout much of Vermont the present Green Mountain anticlinorium separates the sediments of the eastern eugeosyncline from the sediments of the western miogeosyncline. If this concept is valid, then the Hoosac and Rowe schists are the eastern equivalents of the western carbonate phyllite sequence.

Beginning in the Middle Ordovician the geosynclines became tectonically active. The first movement was an epeirogenic uplift of the miogeosyncline following the deposition of the Chazyan limestones. This uplift was most intense toward the east, as here the subsequent erosion cut most deeply. It is possible that this movement was activated by the rise of the geanticline that may have separated the major geosynclinal troughs. Following a period of erosion, the sea again returned to this area. It is apparent, however, that a land mass of considerable size had risen somewhere toward the east in the eugeosyncline. The fine muds and sands of the Walloomsac spread westward from this land across the miogeosyncline overlapping the carbonates of its margin and the shales in the central depression. There is no evidence that the craton was any longer a source of detritus for this area. Throughout the period of the deposition of the Walloomsac and the Mount Anthony, large areas of the eugeosyncline must have been above sea level furnishing a constant supply of fine detritus. Following the deposition of the fine, black Walloomsac shales, a second epierogenic upwarping occurred with the subsequent erosion of the shales and, locally, the underlying limestones. The deposition of the Mount Anthony followed this brief period of erosion. By this time the source areas had become higher or nearer as the resulting detritus was coarser than before. This westward movement of sediment from the rising lands to the east probably continued until the end of the Ordovician, when the whole region was profoundly affected by the Taconic orogeny.

During the Taconic orogeny the rocks of the region were highly folded and faulted, and toward the east attained a low- to medium-grade of metamorphism. Boron metasomatism was important in the formation of tourmaline in the phyllitic rocks.

From the structural point of view, the most important manifestation of the orogeny was the formation of the Green Mountain anticlinorium by a major warping of the crystalline basement complex. It formed as the result of lateral compression from the east which may have had a small vertical component. This large fold was overturned toward the west and thrusting commonly occurred on the overturned limb.

It was this upward and westward movement of the granitic crust along the western flank of the anticlinorium that has determined the major structures in the Taconic and possibly the Middlebury synclinoria. This flank, acting like a piston, formed one half of a shearing couple. The other half was provided by the relatively passive pre-Cambrian basement underlying the above synclinoria. The Cambro-Ordovician sediments were caught in between. The lower competent dolomites and quartzites were the least affected, and reacted by folding. Higher up in the section the relatively incompetent limestones and phyllites in the core of the synclinoria underwent folding, flowage and shearing. Low angle overthrusts developed and were confined to these sediments. The extensive thrusting along the western margin of the Taconic area was probably determined by the incompetent Cambrian shales, buckling and riding up and over the more competent formations farther west. In the Bennington region the North Pownal overthrust shows the same

type of structure. The extensive development of shear cleavage in the Ordovician phyllites is merely a small-scale duplication of these larger overthrusts.

It has been previously shown that many orders of folding are present in this region. The wave lengths of these folds vary from a matter of inches to several miles. Similarly it appears that there may be an equivalent number of orders of shearing. Shears on a microscopic scale are seen in the slip cleavage of the phyllites. Equivalent shears in the form of major overthrusts are common. There is no reason to believe that planes of shearing with displacement intermediate between these two extremes may not be common throughout the region.

All of the above structures have been related to the Taconic orogeny. It is possible, however, that they have been modified to an unknown degree by the later Devonian and late Paleozoic periods of orogenesis. The only structure in the area that may possibly be post-Ordovician in age is the Reservoir Brook Fault. This high-angle "normal" fault could be as young as late Triassic. While it strikes almost due north-south, parallel to the Triassic trends in New England, so does the west front of the Green Mountain axis north of this point. In spite of the fact that the fault locally cuts across the Taconic folds it is in general parallel to the major Taconic structural trends of western Vermont. If it is Triassic it might reflect a general crustal relaxation or isostatic uplift following the cessation of orogeny in this region. However, one such fault is insufficient evidence for so great an event.

ECONOMIC GEOLOGY

The mineral deposits of the Bennington area have been exploited since colonial times. The mining of iron was an important local industry during the early part of the nineteenth century. The ore was found as residual concentrations of limonite in small pockets in the limestone of the Bennington Valley. These mines and their adjoining furnaces were adequate to supply the local demand during this time. However, the discovery of large high-grade iron deposits in northern Michigan and Minnesota made possible in the late nineteenth century, the rise of an iron and steel industry on a national rather than a local scale. With the influx of large quantities of cheap iron, local mines and furnaces such as were found in the Bennington area, were gradually forced out of business. At the present state of iron and steel technology it is highly improbable that the Bennington deposits will ever again be of commercial importance.

No other evidence of metallic mineral deposit was found in the area with the exception of some copper-bearing vein quartz. It was in the bed of a small stream on the south flank of Mount Anthony. The material was not in place and there is no evidence as to where it originally came from.

In the past numerous small quarries have been opened in the local limestones and dolomite for dimension stone and road metal. None of the dimension stone quarries have operated in recent years but some road metal has been quarried recently. In general the variation in amount of dolomite in the limestones precludes the use of this stone where rigid chemical control is necessary.

At the present time, the sand and gravel deposits of glacial origin are the most important mineral products of the region. Many small pits provide washed gravel for road metal and sand and gravel for concrete aggregate. While the actual amount of this material present is difficult of access, there is without question more than enough to supply local needs for the future.

In the past the Bennington clay deposits were an important source for pottery and brick clay. These deposits have not been worked since the last war and the pits have caved so that no detailed examination could be made. For a complete discussion of the location and origin of these clay deposits see Burt, 1929.

The oil and gas possibilities of the Bennington region are practically nil. The degree of metamorphism found is more than sufficient to have destroyed any petroleum products that may have once been present.

In this area the ground water resources have probably not been fully exploited. Adequate supplies are found in drilled wells in the limestone and dolomites in the valleys as well as in shallow wells in the glacial sediments. It is very possible that large supplies might be found in the buried, pre-glacial valleys to the north, south and west of Bennington beneath the Pleistocene lake deposits. On the mountains, water from wells drilled in the Walloomsac or Mount Anthony formations is not abundant and seems to depend largely on local fracturing of the bedrock. In the Pownal Center region the contact between the Walloomsac and the Mount Anthony is water bearing. Wells drilled to this contact find adequate amounts of water for domestic consumption and where it comes to the surface springs are common.

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EXPLANATION

PALEOZOIC ROCKS

Middle to Upper? Ordovician

Omal / Oma

Mount Anthony Formation

Green, gray and black phyllite

--------- unconformity ---------

Middle Ordovician

Owl / Ow

Walloomsac Slate

Black carbonaceous phyllite, blue calcite marble

--------- unconformity ---------

Lower Ordovician

Ocl

Canadian Limestone

Gray, white calcite marble; gray calcitic dolomite; black phyllite

Upper Cambrian

Ɛcs

Clarendon Springs Dolomite

Dark gray calcitic dolomite

Ɛd

Danby Quartzite

Thin bedded gray calcitic dolomite; gray, white vitreous quartzite

--------? disconformity ?--------

Lower Cambrian

Ɛw

Winooski Dolomite

White, buff, gray thin bedded quartzitic dolomite

Ɛm

Monkton Quartzite

White, buff, gray sandy dolomite; gray dolomitic sandstone; black, green phyllite

Ɛdh

Dunham Dolomite

Thick bedded gray, buff sandy dolomite

Ɛc

Cheshire Quartzite

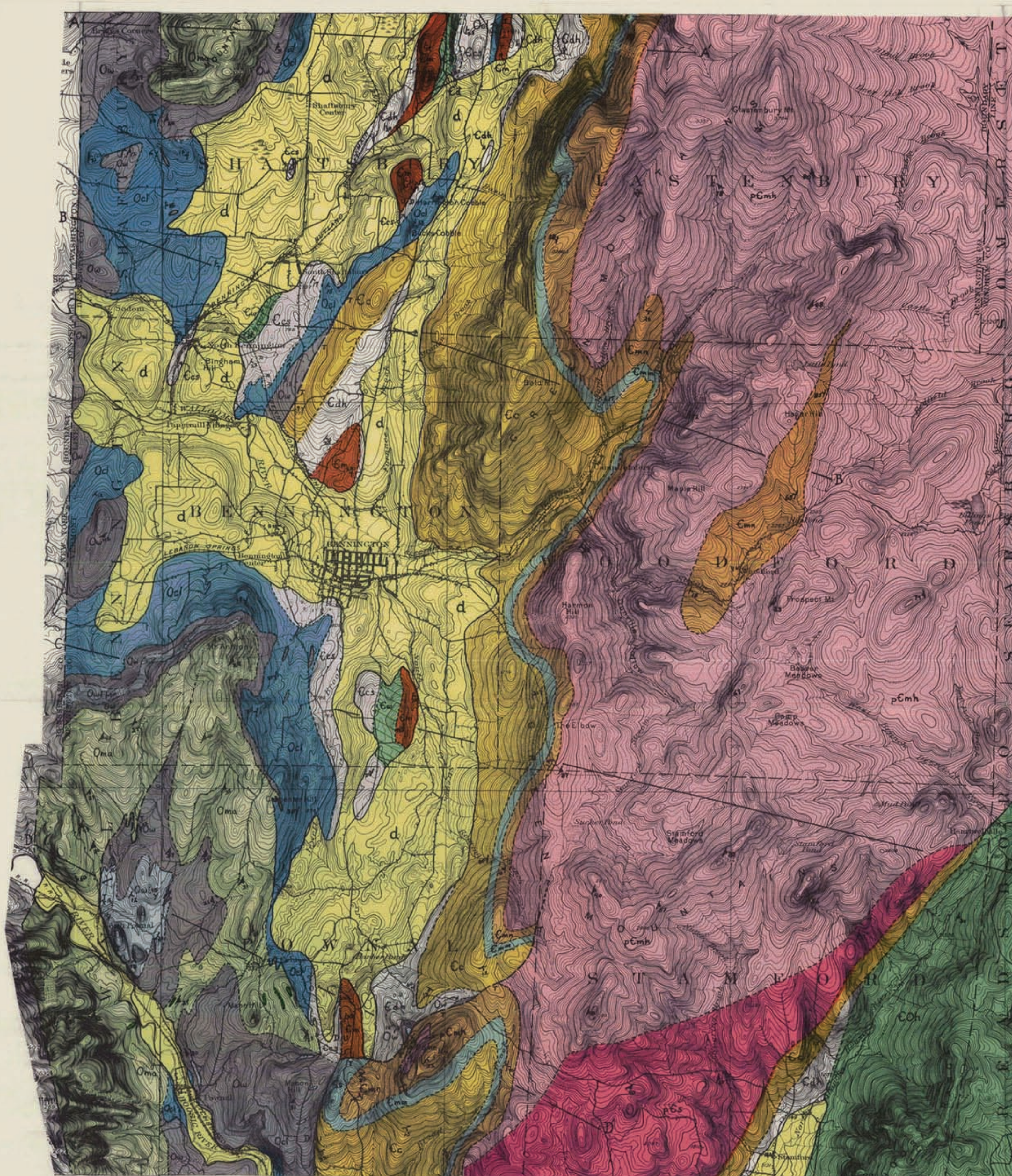
Massive white, buff vitreous quartzite

Ɛmm / Ɛmn

Mendon Formation

Moosalamoo member: black fine banded phyllite; dark quartzite
Nickwacket member: gray quartzite, arkosic quartzite, green quartzitic phyllite, thin bedded conglomerate

--------- unconformity ---------



EXPLANATION

PALEOZOIC ROCKS

Cambro-Ordovician ?

ƐOh

Hoosac Schist

Gray, black albitic schist

PRE CAMBRIAN ROCKS

Sedimentary and igneous

pƐmh

Mount Holly Gneiss

White gneiss, banded gneiss, schist, quartzite

Intrusive

pƐs

Stamford Granite Gneiss

Porphyritic biotite-microcline gneiss

d

Drift

Glacial and recent

STRUCTURAL SYMBOLS

Contact

Concealed or gradational contact

Thrust or reverse fault, T on upper plate, dashed where approximate

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

Bedding

Foliation

A— —A′ Line of cross section

Topography by U.S.G.S. 1897-1898

Geology by J. A. MacFadyen, Jr. 1951 - 1953
SE corner from Prindle and Knopf 1932

True North / Magnetic North 15°

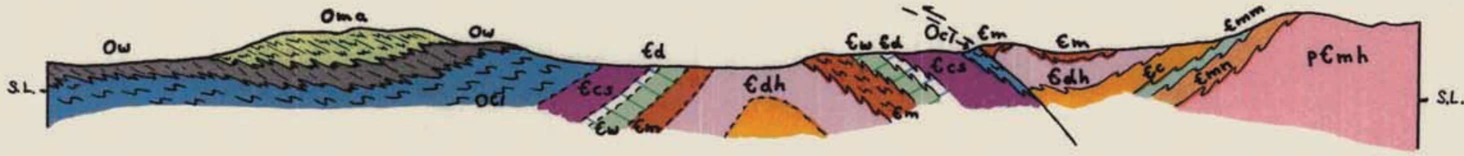
Plate I

Geology of the Bennington Area, Vermont

Vermont Geological Survey

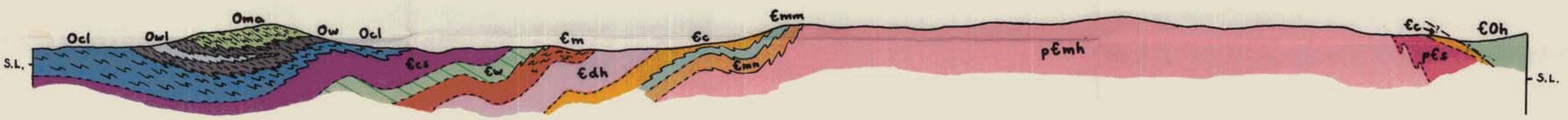
Charles G. Doll, State Geologist

Scale 0 1 2 3 4 Miles

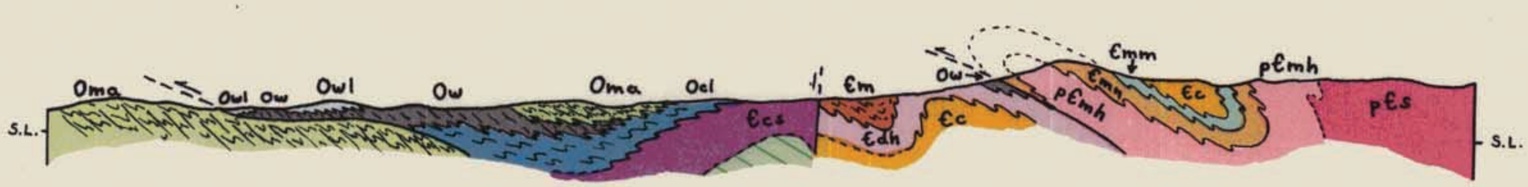


A - A'

B - B'



C - C'



D - D'

Plate II
Structure Sections

Scale

0 1 2 3 4 miles

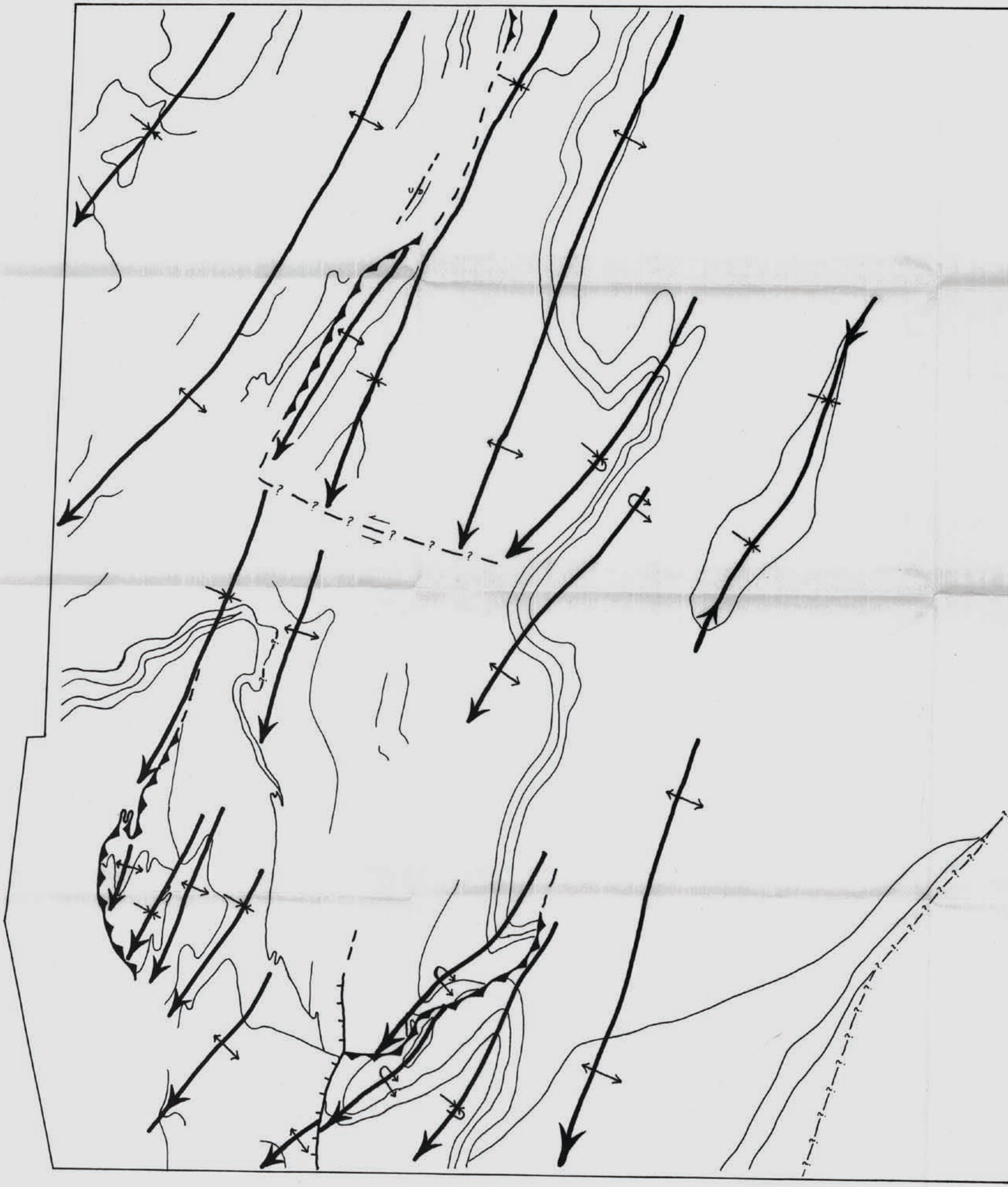


Plate III
Tectonic Map

EXPLANATION

- Trace of crest of plunging anticline (short arrow on side of steeply dipping flank)
- Trace of crest of overturned plunging anticline (arrows convex in direction of overturning)
- Trace of trough of plunging syncline (short arrow on side of steeply dipping flank)
- Trace of trough of overturned plunging syncline (arrows concave in direction of overturning)
- Thrust or reverse fault; dashed where inferred
- Normal fault (hashures on downthrown side)
- U/D Fault (U, upthrown side; D, downthrown side)
- —?—?— Possible fault
- Formational contact