STRATIGRAPHY AND STRUCTURE OF A PORTION OF THE CASTLETON QUADRANGLE, VERMONT

By

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CONTENTS

Illustrations

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ABSTRACT

Bedrock of the southern half and northeastern portion of the Castleton quadrangle in western Vermont, exclusive of the Coxe Mountain area, comprises two distinct sequences: the Taconic sequence, and the synclinorium sequence.

The Taconic sequence is predominantly argillite, but includes minor greywacke, conglomerate, quartzite, and limestone. Major units in this sequence are, from oldest to youngest, the Biddie Knob Formation, the Bull Formation, the West Castleton Formation, the Hatch Hill Formation, the Poultney Slate, the Indian River Slate, and the Pawlet Formation. These units range in age from Early Cambrian through Late Cambrian, Early Ordovician, into the Middle Ordovician; field mapping reveals a major unconformity within this sequence, below the Pawlet Formation. A second possible unconformity is between the West Castleton Formation and the Hatch Hill Formation.

Rocks of the Taconic sequence have been thrust in from outside, presumably from the site of the present Green Mountain anticlinorium. Thrusting may have occurred as submarine gravity slides and was a mid-Trenton event. As a result of this diastrophism, a number of imbricate tectonic slices were produced; within the map area at least two slices are identified. These are the Giddings Brook (lower) slice and the Bird Mountain (upper) slice; the Spaulding Hill-Middletown Springs area in the south-central portion of the Castleton quadrangle is the north end of a window of the Giddings Brook slice, framed by the Bird Mountain slice. Within each slice, the rocks are recumbently and isoclinally folded as well as locally thrust-faulted; these early folds are tight and are believed to be concomitant with thrusting. They have been refolded during the main phase of the development of the Middlebury synclinorium.

The synclinorium sequence consists of rocks of the same age range as the Taconic sequence. By contrast, however, it is largely of clean

carbonate and orthoquartzite. Apart from the basal Lower Cambrian Mendon Series, the only major exceptional rock type is the youngest unit, the mid-Ordovician Ira Formation of black phyllite with its basal limestone, the Whipple Marble. The Ira Formation marks a permanent break in the early Paleozoic sedimentary pattern of this region and it no doubt records orogenic activities in the source area to the east; these activities may be directly associated in time and space with Taconic thrusting.

Despite local complexities, the structure of the synclinorium sequence is essentially simple and within the map area constitutes the west-facing east limb of the south-plunging Middlebury synclinorium. The "anticline" of marble at West Rutland is reinterpreted as a homoclinal sequence of pre-Middle Ordovician rocks. The true nature of the structure has been obscured by the masking Ira Formation and by axial culminations and depressions within the pre-Ira structures.

Within the synclinorium sequence, there are a number of north-south trending high angle normal and reverse faults, of which the Pine Hill reverse fault is one. The inception of these faults cannot be accurately dated but may have been in mid-Trenton time, an hypothesis which fits all data and explains certain peculiar rock types of that age in the area. If proven, then in mid-Ordovician time the area of the east limb of the Middlebury synclinorium may have been one of graben and horst, in addition to one where the major structures of the synclinorium began to develop, and into which slices of the Taconic thrust moved.

The Taconic sequence now occurs in the center of the Middlebury synclinorium and rests above the right-side up Ira Formation, with major inversion of the sequence near the contact. The delineation of the contact between the two sequences is not easy, partly because of the similarity of rocks on the two sides of the dislocation and partly perhaps because no sharp surface of contact ever did exist, due to the particular manner of emplacement of the allochthone. The contact, however, probably lies within the black phyllite rather than at either of its boundaries. Local outcrops of Ordovician marbles near this contact, in the town of Ira, are interpreted as slivers dragged along the sole of the thrust slices, rather than as indigenous anticlines.

INTRODUCTION

This report describes the bedrock geology of an area covering most of the southern portion of the 15-minute Castleton quadrangle in western Vermont, as well as a narrow strip extending northward from West Rutland to Florence, in the northeastern part of the Castleton quadrangle or the eastern half of the 7 1/2-minute Proctor quadrangle. The area is approximately bounded, to the north, by the low, east-west ridge north of the Castleton River; to the east, by Pine Hill and its southward continuation, called by T. N. Dale (1894) the "ridge between the Taconic and Green Mountain Ranges"; to the west, by the 7 1/2-minute Thorn Hill quadrangle (Vermont-New York; southeastern quarter of the 15-minute Whitehall quadrangle); and to the south, by the 15-minute Pawlet quadrangle (Fig. 1).

Geologically, the area is underlain by two distinct and contrasting stratigraphic and structural units. These are the "Taconic sequence" and the "Valley sequence" of authors (see, for example, Kaiser, 1945; Fowler, 1950; Billings, Rodgers, and Thompson, 1952). There are strong topographic expressions of this geologic junction. A line of valleys and swamps follows Whipple Hollow south from Florence to West Rutland. This line is continued by the valley of Clarendon River southward to Chippenhook, and leaves the Castleton quadrangle at the north end of Tinmouth Channel, an extensive swamp. To the west of this line rise the steep and often rugged hills of the Taconic Range, which is the eastern limit of the Taconic sequence. East of the line of swamps is relatively open country, underlain by rocks of the "Valley sequence."

The name, "Valley sequence," applies to a section of Lower Paleozoic sedimentary rocks, ranging from Early Cambrian to Middle Ordovician in age, and consisting largely of orthoquartzite and carbonate. The only significant argillaceous and quartzofeldspathic rocks in this sequence are the basal Cambrian (?) Mendon Series (Brace, 1953; Fowler, 1950; Osberg, 1959) and the Middle Ordovician Ira Formation (Keith, 1932; Zen, 1961). The sequence overlies the Precambrian gneisses of the Mount Holly Series with profound unconformity (Thompson, 1959; Brace, 1953; Osberg, 1952). Despite a number of important faults in this area (Brace, 1953; Fowler, 1950; Thompson, 1959; Zen, 1959a) and at least one regional unconformity at the base of the Ira Formation, the sequence is remarkably simple in structure and presents an overall west-facing section. The "Valley sequence," in fact, is the east limb of the south-plunging Middlebury synclinorium (Cady, 1945). The west limb of this structure lies in the vicinity of Lake Champlain, about 15 miles to the west. The stratigraphy of the two limbs is directly comparable. For this and also other reasons, the writer has urged the use of the term "synclinorium sequence" in place of "Valley sequence" (Zen, 1961).

Rocks of the Taconic sequence are of very different nature, consisting

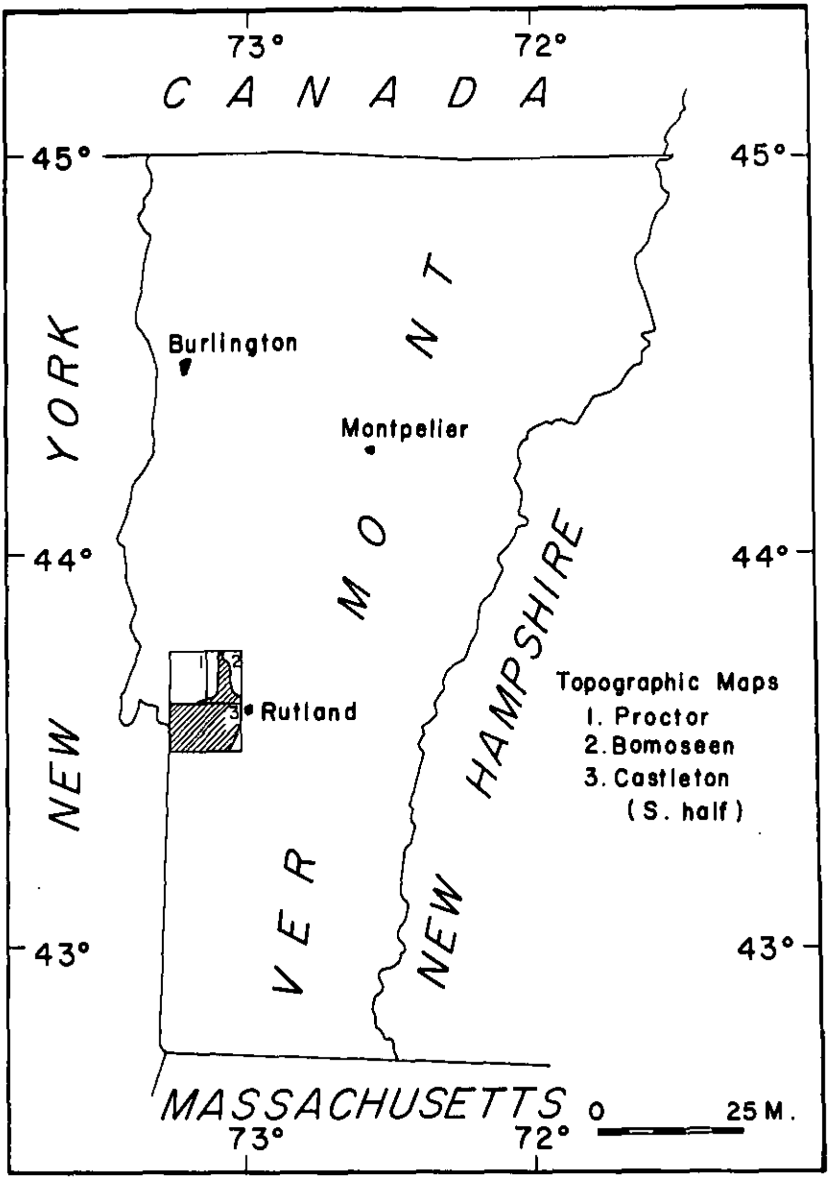

Figure 1. Location of the map area in Vermont.

largely of slate and phyllite, and with only minor carbonate and orthoquartzite beds. The rocks also range in age from Early Cambrian to Middle Ordovician. Structurally, these rocks occur nested within the Middlebury synclinorium, yet everywhere they rest upon the Middle Ordovician Ira Formation of the synclinorium sequence with pseudoconformity and even gradational contact. The position of the Taconic sequence is thus a distinct anomaly, a problem whose magnitude is underscored by the fact that rocks of this sequence continue all the way to Dutchess County, New York, a distance of some 150 miles.

The writer, mapping in the adjoining area to the north of the present one, found evidence for primary sedimentary tops in the Taconic sequence. The evidence indicates that the Lower Cambrian rocks of the Taconic sequence lie upon the Ira Formation in inverted position. The Taconic sequence is thus allochthonous. In fact, a number of distinct thrust sheets exist in the area; within each sheet the rocks are recumbently folded during at least two cycles of deformation. The writer has proposed a stratigraphic section in the Taconic sequence of this area; the section differs from that of Dale (1899), Larrabee (1939), Kaiser (1945), and Fowler (1950). For this reason, it became desirable to remap the adjoining area to the south. The result of this study is the substance of the present report.

The area is easily reached. Major highways include U. S. Route 4, Vermont Routes 30, 133, and 140, and New York Route 22A. Many secondary roads feed into these arteries. With the exception of the main range of the Taconic Mountains, the area is generally open pastureland. These facts have made it possible to take advantage of the generally good outcrops, and to map the geology in some detail.

The field data are plotted on aerial photographs, scale about 1:20,000, taken by the Soil Conservation Service. Dips and strikes are taken to the nearest 5° with a Brunton compass, and altitude recorded by a 50′ Taylor aneroid altimeter. Unfortunately, the topographic base map for the Castleton quadrangle, edition of 1891, is grossly inaccurate at many places, and the culture is much out of date (for example, many roads shown no longer exist). These facts often make it all but impossible to locate outcrops precisely on the map. At places they also make it impossible to show the detailed geology accurately. It is is hoped that better base maps may soon become available so that the geology may be replotted and published correctly. The report was brought up to date as of December, 1961; later references have not been incorporated.

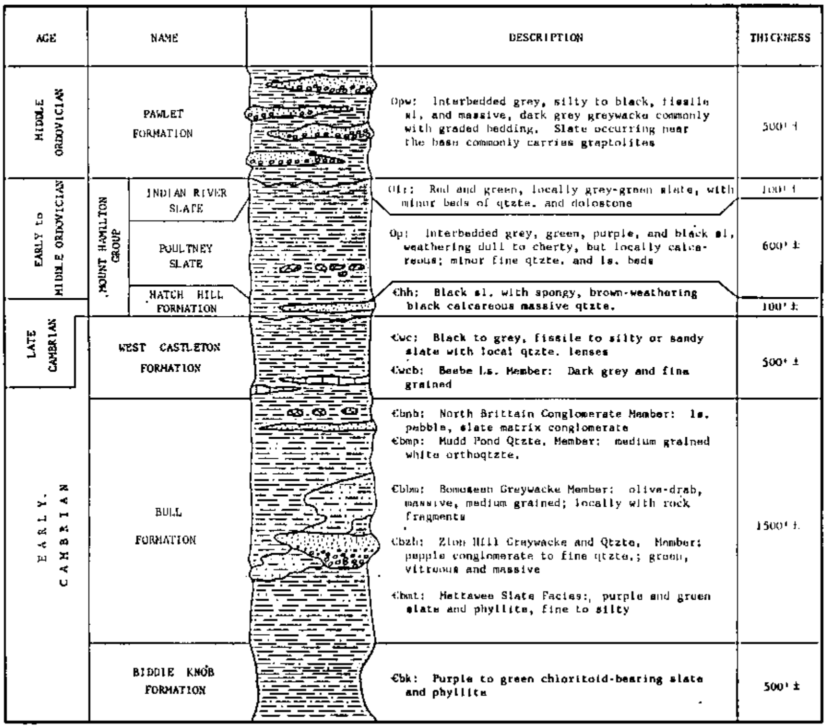

Figure 2. Columnar section of the Taconic sequence.

Acknowledgments

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Frequent field conferences with P. H. Osberg, L. B. Platt, D. B. Potter, R. C. Shumaker, G. Theokritoff, and J. B. Thompson, Jr., have helped me to maintain the proper regional perspective, and made the field season particularly enjoyable. To all of these people, my thanks. Others who have generously contributed ideas on regional relations are W. B. N. Berry, M. P. Billings, W. F. Brace, W. M. Cady, P. C. Hewitt, and J. Rodgers; their interest is appreciated.

W. M. Cady, and J. Rodgers kindly reviewed the manuscript and made many important suggestions for improvement, in style, in contents, and in rigour of expression. Although in no way responsible for my mistakes, these people share any credit for this report.

Former Governor Joseph B. Johnson of Vermont, while in office, spent an entire day with me in the field to study the geologic relations. His enthusiasm and support will be long remembered with appreciation.

To the people of the area, I wish to extend my thanks once more for permission, tacit or overt, to trespass on their land, climb over their fences, and smash their ledges.

STRATIGRAPHY

Introduction

Two major stratigraphic sequences exist in the map area: the Taconic sequence and the synclinorium sequence. In age, both sequences range from Early Cambrian to Middle Ordovician; in the area of the synclinorium sequence, however, small areas of Precambrian rocks are included for convenience. Despite their like age, the two sequences are strongly dissimilar in rock type: the Taconic sequence consists largely of argillite, greywacke, quartzite, with subsidiary limestone; whereas the synclinorium sequence is predominantly limestone and dolostone, with lesser amounts of orthoquartzite. In the synclinorium sequence, units of argillite and greywacke are confined to the basal Cambrian (?) or Precambrian (?) Mendon Series and to the Middle Ordovician Ira Formation, which is the youngest unit of the synclinorium sequence (Figs. 2 and 3).

With the exception of possibly parts of the Precambrian gneisses and of a number of cross-cutting, post-tectonic and post-metamorphic mafic dykes, there are no igneous rocks in the area. The dykes are of no interest in this study and will not be further discussed.

Taconic Sequence: Lower Cambrian Units

Biddie Knob Formation

Name and Lithology: The name Biddie Knob was proposed by Zen (1961, p. 299) for a rock unit which, north of the Castleton River, appears to be the oldest formation of the Taconic sequence. It was named after the excellent exposures on the west and south slopes of Biddie Knob in the Taconic Range. The name is used in this report without modification.

The Biddie Knob Formation is principally a purple and green, rarely grey, chloritoid-bearing phyllite and slate. By definition, the presence of chloritoid distinguishes this formation from the slate of the overlying Bull Formation. In the present map area, the Biddie Knob Formation

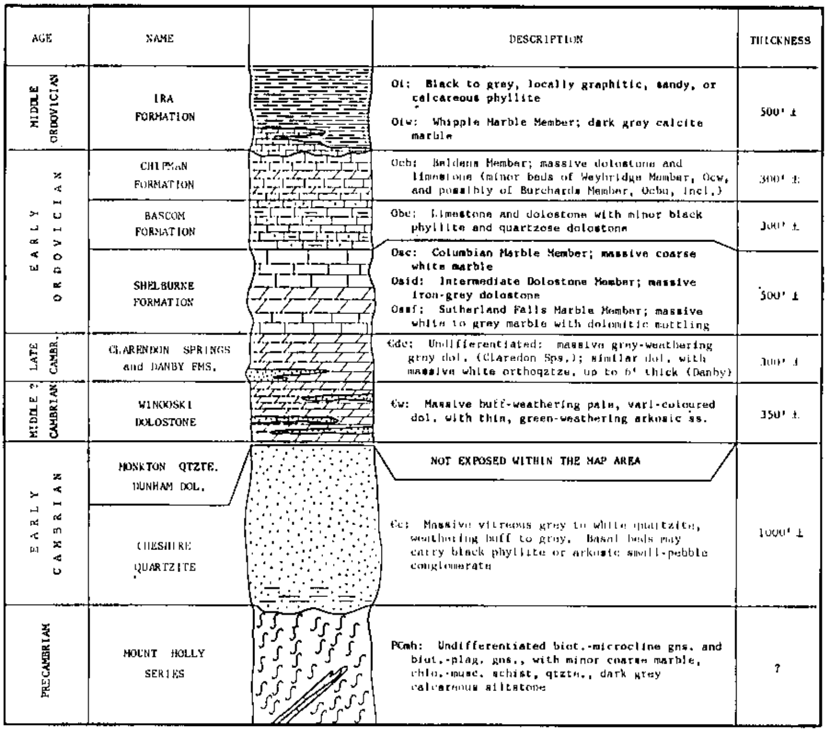

Figure 3. Columnar section of the synclinorium sequence.

also contains minor quartzite beds as well as the Bird Mountain Grit, correlated with the Zion Hill Quartzite and Greywacke Member, to be discussed under the Bull Formation. The detailed description of the Biddie Knob Formation has been given elsewhere (Zen, 1961, p. 299).

Distribution: The Biddie Knob Formation occupies large areas between Herrick Mountain and Hampshire Hollow, in the core of a large recumbent fold. At the south end of this area, the formation disappears because the fold axis plunges. Because of the manner in which the formation is defined, its stratigraphic significance relative to the overlying Bull Formation might be questioned, as the presence of chloritoid might conceivably be due to higher metamorphic grade or to original irregularities in the chemical composition of the sediments (Zen, 1960a). The presence of chloritoid-free green phyllite east and on the higher metamorphic grade side of the Biddie Knob Formation, between Spruce Knob and

West Rutland, shows that metamorphism alone cannot explain the distribution pattern of the Biddie Knob Formation. Furthermore, the contact between the Biddie Knob and Bull formations, as mapped, concurs with other contacts mapped on the basis of key beds (Zen, 1961). The Bull-Biddie Knob demarcation therefore is stratigraphically valid.

Northwest of West Rutland, a small area of the Biddie Knob Formation is shown on the map north of a major thrust fault called (Zen, 1961, p. 321) the Bird Mountain fault, thus occurring in the underlying tectonic unit. This area of the Biddie Knob Formation is part of the Giddings Brook fold complex (Zen, 1961).

Age: No direct evidence exists for the age of the Biddie Knob Formation. This formation grades upward into the Bull Formation, however, and the distinctive Zion Hill quartzite and greywacke occurs either near the base of the Bull Formation or near the top of the Biddie Knob Formation. The higher strata of the Bull Formation are dated as Early Cambrian. The Biddie Knob Formation therefore is assigned to the same age.

Bull Formation

Description and Name: The Biddie Knob Formation is succeeded upwards by the Bull Formation, named after Bull Hill north of the Castleton River (Swinnerton, Harvard University thesis, 1922, p. 69; Zen, 1961, p. 300). Within the map area, the Bull Formation consists preponderantly of a slate, argillite, or phyllite facies, called the Mettawee Facies. A number of distinct and mappable rock types exist in addition; these are referred to as members in the Bull Formation. The Mettawee Facies constitutes the "matrix" of these members.

1) The Mettawee Facies: west of the Taconic Range, the Mettawee Facies is dominantly a purple, green, or variegated purple and green, fine-grained slate. Chloritoid is absent in this rock. In a number of areas, particularly in the higher hills immediately north of the Castleton River and east of North Brittain Brook, the rock is a monotonous mass of siltstone, with poor development of slaty cleavage but strong development of "cleavage banding" (Dale, 1895, p. 561). The purple, green, and grey colours are here subdued, giving the rock a bleached appearance. These different rock types appear to grade laterally into one another.

East of the Taconic Range, rock of the Mettawee Facies is metamorphosed to a phyllite; the mica flakes are coarser grained, and vein quartz becomes common. The rock is crinkled on a fine scale and commonly acquires a glossy luster. Purple colour disappears or becomes exceedingly rare in this area, and the rocks are generally green and more rarely grey.[1] Even within the areas of soft roofing slates, however, there is occasional interbedded phyllite; the difference is attributed to initial compositional differences such that the more micaceous and less quartzose layers, at a given metamorphic grade, are the more phyllitic. Such textural data therefore may be used to delineate metamorphic grade only with great caution.

The Mettawee Facies also includes several minor rock types:

a). Quartzite, green or white, a few inches to several feet thick. These quartzite beds do not form traceable units and occur randomly in the rock.

b.) Limy slate which, but for the higher carbonate content, is similar to the bulk of the Mettawee. Iron in the carbonate, upon slight weathering, gives the rock a light brown colour, and rotten appearance. These strata are up to several feet thick, but the writer has not been able to assign these beds to particular stratigraphic horizons.

c.) A grey or grey-green phyllite with prominent porphyroblasts of albite up to 1/2 mm across. This rock unit occurs only east of the Taconic Range, near the contact of the Mettawee phyllite and the spatially underlying black phyllite. Over large areas southeast of Clark Hill, this unit is unknown and it seems to pinch out on strike. For more detailed description of the various rock types, see Zen, 1961.

2) The Bomoseen greywacke member: The Bomoseen greywacke is Dale's Cambrian olive grit (1899, p. 179). Its type locality is west of the outlet of Lake Bomoseen (Dale, 1899, p. 180; Ruedemann, 1914, p. 69). The rock has been described before (Dale, 1899, p. 179; Zen, 1961, p. 301). Briefly, it is a massive, olive to green-grey, fine-grained greywacke with characteristic mica spangling on fresh surfaces. It weathers to a white or brick-red cast, and bedding is rare in the rock.

Within the Bird Mountain thrust slice, the Bomoseen Greywacke is identified only in two isolated patches near the south end of Fennell Hollow. West of the Bird Mountain fault, however, the Bomoseen Greywacke is widespread and occurs at the core of a complex anticline. It is the lowest stratigraphic unit exposed here; on the west flank of this same structure (Zen, 1961, Pl. 1), the Zion Hill Member of the Bull Formation

[1] This colour change is due to the conversion of hematite to magnetite whereupon the purple colour, due to the addition of red from hematite and green from chlorite, gives way to the green of chlorite alone (Zen, Harvard University thesis, 1955, p. 105; 1960a, p. 167).

overlies the Bomoseen. However, in the Fennell Hollow area, the map pattern suggests that the Zion Hill is older than the Bomoseen. Such contradictory age relations between these two units have also been found in the area to the north, and Zen (1961, p. 302) has concluded that the two lithofacies must cross each other.

A related conclusion is that the Bomoseen Greywacke, instead of being the lowest stratigraphic unit of the Taconic sequence, as mapped by Dale (1899, plate facing p. 178), Kaiser (1945), and Fowler (1950), is actually a lithofacies within the Bull Formation (Zen, 1961, p. 302).

3) The Zion Hill Quartzite and Greywacke Member: The type locality for this member, Zion Hill, is in the town of Hubbardton north of the Castleton River. The name was proposed by Ruedemann (1914, p. 70) for the bulk of Dale's "ferruginous quartzite" (1899, p. 183). The rock is a medium- to coarse-grained, green quartzite or greywacke, with slate, quartz, feldspar, and mica pebbles up to 6 inches across, although most of the grains in the greywacke are less than 1/4 inch across. The beds commonly are graded through changes in maximum grain sizes. The rock has been described many times before (Dale, 1899, 1900; Zen, 1961, p. 302).

West of the trace of the Bird Mountain thrust fault and within the map area, the Zion Hill Member occurs sporadically within the purple and green Mettawee Slate. These outcrops cannot be traced, however, and their designation as Zion Hill is based on lithologic comparison. East of the fault, on the other hand, extensive areas of this rock exist; these outcrops have been previously mapped as the Bird Mountain Grit.

The petrography of the Bird Mountain Grit has been given by Dale (1900). The rock is identical in appearance to the Zion Hill Member, and has been so correlated by Swinnerton (Harvard University thesis, 1922, p. 196), Keith (1932, p. 401), and Kaiser (1945, p. 1090). Fowler (1950, p. 41), however, mapped the rocks surrounding Bird Mountain as "Nassau Formation," which he regarded as possibly Precambrian in age by a questionable correlation with the dubiously dated Nassau Formation in New York state; because of this correlation, Fowler rejected the correlation of the Bird Mountain Grit with the Zion Hill quartzite (1950, p. 54).

According to Fowler, the main argument against correlating the slate and phyllite surrounding the Bird Mountain Grit with the "Mettawee Slate" (Biddie Knob Formation and Bull Formation of the present study), mapped by Kaiser north of Castleton River, is a need to thicken the Mettawee east of the area around Poultney at least ten fold in order to make the sections comparable (1950, p. 43–44). This need to thicken the strata eastward, based on the incorrect map of Kaiser, is illusory, as the thickness of the Mettawee Facies of the Bull Formation in the Taconic Range is in fact comparable with that in the slate belt (Zen, 1961). Moreover, the Bird Mountain Grit occurs in a sequence that proceeds from the Biddie Knob Formation through the Bull Formation (including key units like the Mudd Pond Member and the North Brittain Member; see below) into the West Castleton Formation. Thus, even though fossils are wanting in the Bird Mountain slice, lithologic as well as stratigraphic evidence strongly indicates correlation of the Bird Mountain grit with the Zion Hill Member of the Bull Formation.

Zen (1961, p. 302) notes that, north of the Castleton River, the Zion Hill Member locally occurs within, though near the top of, the Biddie Knob Formation rather than in the Bull Formation. The geologic map, Pl. 1, shows that the same is true south of the Castleton River. In addition to minor undulations in stratigraphic position, the Zion Hill Member also varies rapidly in thickness (Zen, 1961, p. 302). North of the Castleton River, the thickness is generally 100 feet or less (Swinnerton, Harvard University thesis, 1922, p. 82, 85, 90, 92, 99, 104, 105, 115). South of the river and within the Bird Mountain slice, however, thicknesses may be greater. Thus, at the south end of Bird Mountain, it may be as much as 500 feet: this is the difference in relief between the top of the south peak and the base of the rock unit, and the continuous outcrop on the cliffs show the beds to be flat-lying and apparently without repetition. All the known bodies of the Zion Hill Member, finally, seem to be elongate in shape and peter out rapidly on strike. These points, taken together, suggest that the Zion Hill Member represents channel-fills, perhaps due to density currents; different areas of this unit therefore may not be precise time-equivalents. Nonetheless, in general the bodies occur in the same stratigraphic zone, at the upper part of the Biddie Knob Formation or the lower part of the Bull Formation. Correct recognition of this unit is therefore useful.

4) The Mudd Pond Quartzite Member: the name, Mudd Pond, was proposed (Zen, 1961, p. 303) for an orthoquartzite found near the top of the Bull Formation, stratigraphically below the North Brittain conglomerate, to be next described. The Mudd Pond is medium grained, vitreous, and white to pale green. Locally, it carries limy pods or lenses which may have been concretions. A lithologic variant is the Eddy Hill Grit (Zen, 1961, p. 303; Larrabee, 1939, p. 51), which is a dark grey arkosic grit with variable amounts of brown-weathering limy and argillaceous

cement. Fowler (1950) mapped patches of "Eddy Hill Grit" in the present map area. These patches occur in the overlying black slate, however, and the writer assigned them to the overlying West Castleton Formation (1961, p. 305). As now defined, the Eddy Hill Grit is a definite stratigraphic as well as lithologic unit, unrelated to the "Eddy Hill lithofacies" of Lochman (1956, p. 1336).

West of the trace of the Bird Mountain thrust fault, outcrops of the Mudd Pond Member are common, but with few exceptions these outcrops could not be traced into one another. East of the fault, the Mudd Pond Member is rare; only three outcrops with this lithology are known. The distinctive lithology, combined with their stratigraphic position, justify the correlation of these quartzite bodies with the Mudd Pond.

5) The North Brittain Conglomerate Member: This name was proposed by the writer (Zen, 1961, p. 303) for an intraformational, polymikt limestone-pebble, slate-matrix conglomerate found at or near the top of the Bull Formation. The matrix is generally green or purple slate, free of carbonate; the pebbles are either cream-colored, fine-grained limestone weathering buff, or pale to deep grey, fine- to coarse-grained limestone weathering darker grey. The pebbles are commonly deformed. Thicknesses of the conglomerate range from a few inches to several feet, the true thickness however is generally difficult to ascertain owing to the common lack of bedding, and to the ease with which the rock deforms and weathers. The intraformational nature of the conglomerate is demonstrable especially in a section in the Poultney River northwest of Fair Haven (Zen, 1959a, p. 8; 1961, p. 304).

Within the area of the Giddings Brook fold complex included in the present map, the North Brittain Conglomerate Member occurs at the proper stratigraphic position near the top of the Bull Formation and a short thickness above the Mudd Pond Member. East of the trace of the Bird Mountain thrust fault, however, the North Brittain Conglomerate Member commonly occurs in areas of monotonous purple and green slate, without any of the next younger West Castleton Formation in the immediate vicinity. This is so generally true that an explanation is needed. Three possibilities exist: the conglomerate may occupy here a lower stratigraphic level, within the bulk of the Mettawee Facies; the West Castleton Formation may have been eroded off; or the West Castleton Formation may merge lithologically into the Bull Formation within the confines of the Bird Mountain slice.

Black slate, like the West Castleton and so mapped here, does occur within the Bird Mountain slice south of Gorhamtown. The last of the explanations is thus inadequate by itself. The second explanation is improbable, inasmuch as the rocks are intensely folded, and the present erosion surface is not likely coincident with the Bull-West Castleton contact over large areas. Wherever the relations can be determined, moreover, the conglomerate occurs at its correct relative position, between the Zion Hill Member and black slate. The writer therefore interprets the relation as due to a thickened section of purple and green Mettawee Slate between the conglomerate and the top of the Bull Formation, in the upper, or normal limb of the large recumbent fold that constitutes the Bird Mountain slice. In the inverted limb of this fold, southeast of Spruce Knob, however, the North Brittain Conglomerate Member occurs in its more common position, near the contact of the Bull Formation and the black slate (Fig. 4).

Age of the Bull Formation: The limestone "pebbles" of the North Brittain Conglomerate Member have yielded abundant fossils of the *Elliptocephala asaphoides* fauna, of Early Cambrian age (Lochman, 1956; Theokritoff, 1959a, p. 54, 1959b; Schuchert, 1937, p. 1039). Since the North Brittain conglomerate is demonstrably intraformational, the "pebbles" representing broken up beds, the same age would apply to the enclosing slate. The uninterrupted sedimentary sequence below the North Brittain conglomerate, extending into the Biddie Knob Formation, suggests then that these unfossiliferous strata are also of Early Cambrian age.

West Castleton Formation

Name and Lithology: The name West Castleton Formation was proposed by Zen (1961, p. 304) for the black and grey slate unit that overlies the Bull Formation, and is of Early Cambrian age. The type locality is at the hamlet of West Castleton. The name supplants the name Hooker of Swinnerton (Harvard University thesis, 1922) and Keith (1932) and used previously by the writer (1956), as well as the name Schodack of Ruedemann (1914), Larrabee (1939), Kaiser (1945), and Fowler (1950). The inappropriateness of these names has been discussed by Theokritoff (1957, p. 1804; 1959a, p. 55) and by Zen (1961, p. 304).

The West Castleton Formation, in the present map area, is largely a grey to black, fine-grained slate; locally it is graphitic and may carry pyrite. Thin (½ mm or less) white silty laminae occasionally occur in the fine-grained slate. As a whole however the rock rarely shows bedding. Outcrops of the rock weather easily into low, rounded hills; weathered rock surfaces may be dull grey, rusty, or reddish-coloured.

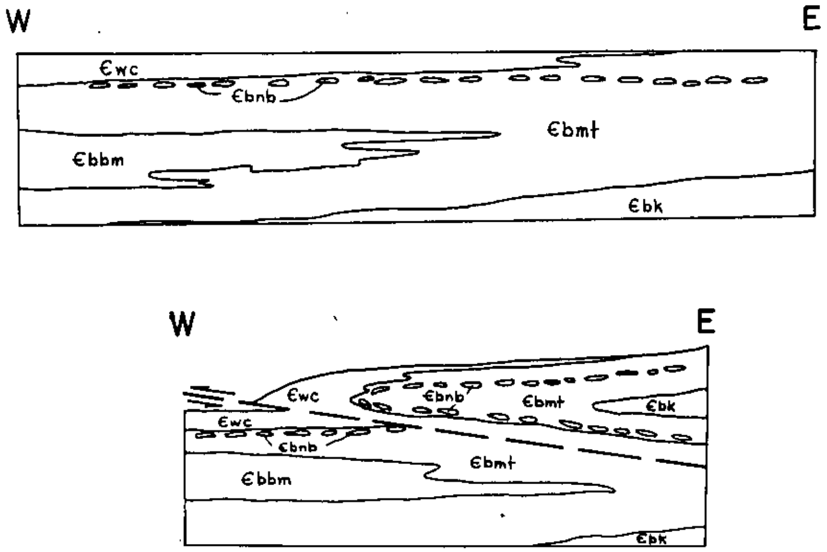

Figure 4. An interpretation of the sedimentary-facies relationship between the upper Bull Formation and the West Castleton Formation in the map area. A thickening of the Mettawee Slate Facies of the Bull Formation eastward, above the horizon of the North Brittain Conglomerate Member and at the expense of the West Castleton Formation, could explain the field relations.

Besides slate, the West Castleton Formation also includes other rock types:

1) A massive, dark bluish grey limestone, weathering light blue-grey, a few inches to over 10 feet thick. The rock is locally fossiliferous; it has been called the Beebe Limestone by Keith (1932, p. 402). Zen (1961, p. 304) designated this limestone a member in the West Castleton Formation. In general, it occurs near the base of the formation.

2) A black conglomerate consisting largely of fine pebbles of quartz. The subrounded to rounded pebbles are up to 2 mm across; and the matrix may be finer grained quartz, or it may be argillaceous. The rock appears to be local lenses within the West Castleton Formation and has been mapped incorrectly as Eddy Hill Grit by Fowler (1950, p. 50).

Distribution: West of the Bird Mountain thrust fault, the West Castleton Formation occurs in a continuous belt from Hydeville to Poultney, which is apparently the southern continuation of the Cedar Mountain syncline (Zen, 1961, p. 317). A second important area of similar slate, between Castleton and Lewis Brook just west of Gorhamtown, is so far unfossiliferous, but it is correlated across the alluvium-filled Castleton River with the fossiliferous West Castleton Slate west of the Pine Pond thrust (Zen, 1961, p. 314), in the Giddings Brook fold complex. Other scattered patches of the West Castleton Formation are mapped primarily on the basis of lithologic similarity, and in particular on the occurrence therein of limestone like the Beebe.

East of the Bird Mountain thrust fault, the West Castleton Formation is very scarce. The only continuous belt is immediately above and east of the fault. Other isolated areas of black slate are mapped as the West Castleton strictly by lithologic correspondence and their assignment is not proven.

Age: The Beebe Limestone of the West Castleton Formation carries the Early Cambrian *Elliptocephala asaphoides* fauna (Zen, 1961, p. 305; Swinnerton, Harvard University thesis, 1922, p. 77, 78, 79; Schuchert, 1937, p. 1039; Fowler, 1950). At least one of the localities reported by Fowler, 1¼ mile southeast of Blissville (1950, p. 52), is in the present map area, and has yielded these forms: *Linnarssonia taconica* Walcott 1887, *Lingulella granvillensis* Walcott 1887; *Pagetides connexa* (Walcott) 1887; *Bonnia clavata ?* (Walcott) 1887; *Elliptocephala asaphoides* Emmons 1844, etc. The age of the Beebe Limestone, and therefore presumably the West Castleton Formation as a whole, must be Early Cambrian. Theokritoff (1959b, p. 1686) finds evidence of a second, younger *Paedeumias* fauna in the West Castleton Formation in the Thorn Hill quadrangle west of the present area. These ages all concur with the age of the underlying North Brittain Conglomerate Member of the Bull Formation.

Taconic Sequence: Post-Lower Cambrian Units

General Remarks

At the north end of the slate belt, rocks of the Taconic sequence which are younger than the Early Cambrian are now known to range from Late Cambrian to Middle Ordovician (Theokritoff, 1959a; Berry, 1959, 1961). These rocks were mapped by Dale (1899) under the following names and division labels: Calciferous (F), Hudson shale (G), Hudson white beds (Hw), Hudson thin quartzite (Hg), Hudson grit (Ig), and Hudson red and green slate (Irs). Larrabee (1939), Kaiser (1945), and Fowler (1950) mapped a Normanskill Formation for the equivalent units. In the vicinity of Poultney, Keith (1932) recognized three formations: the

Poultney Slate, the Indian River Slate, and a black slate, unnamed, stratigraphically above the Indian River.

Theokritoff (1959a, p. 55) recognized the Poultney and Indian River formations, but proposed the name Hatch Hill for a formation of black slate and interbedded calcareous quartzite. This formation, as defined, is between the Poultney and the West Castleton formations, and is of Late Cambrian age. Shumaker (1959) recognized these same units in the western part of the Pawlet quadrangle; in addition he mapped a black slate—greywacke unit above the Indian River Slate. This is presumably the same as Keith's unnamed black slate.

Zen (1961, p. 306) mapped six rock types under the name of "Mount Hamilton Group," as well as an overlying black slate—greywacke unit called the Pawlet Formation. Earlier, the name Poultney River Group (Zen, 1959a, p. 2; 1960a, p. 131) had been used; this name however is now abandoned because of possible confusion with the name Poultney Slate. Some of the units included in the Mount Hamilton Group were tentatively correlated with Theokritoff's formations (Zen, 1959a, p. 2). However, from Zen's table (1961, p. 307), it appears that the Hatch Hill—Poultney—Indian River succession of Theokritoff does not hold precisely in Zen's area to the north. New data bearing on these problems have accumulated; the problems will be discussed after a description of the rock units.

Hatch Hill Formation

Name and Lithology: Theokritoff (1959a, p. 55) proposed the name Hatch Hill for a formation of "sooty black pyritic rusty-weathering shales interbedded with rotten-weathering bluish dolomitic sandstones, locally cross-bedded, and characteristically traversed by numerous quartz veins." The type locality, west flank of Hatch Hill, is in the Thorn Hill quadrangle just west of the present area. The Hatch hill is the lithic equivalent of units 2 and 5 of Zen (1961, p. 306).

Age and Stratigraphic Position: Theokritoff (1959a, p. 55) and Berry (1959, p. 61) report *Callograptus*, *Dendrograptus*, and rarely *Dictyonema* from the Hatch Hill Formation. Berry considers these forms to be of Late Cambrian, pre-Schaghticoke age (1959, p. 61; table G-1). Stratigraphically, the formation is between the West Castleton Formation and the Poultney Slate; the relative ages therefore agree with the paleontological evidence so far presented.

Zen (1961, p. 307) finds, however, that the sandstone, typical of the

Hatch Hill, may occur also higher up in his Mount Hamilton Group. For instance, in Poultney River, at the core of the Mount Hamilton syncline slate and quartzite typical of the Hatch Hill occur above a typical Poultney slate; graded bedding and channel filling features prove a normal sequence here. In the field, isolated outcrops of the Hatch Hill lithology could and should be mapped only as rock units without time-stratigraphic connotations.

Distribution: In the map area, the only Hatch Hill-like rocks are on the south and east slopes of the 2405′ peak, west of Chippenhook, in Clarendon. Here the characteristic quartzite, 6 feet or more thick, occurs topographically below a green, Mettawee-like phyllite, and above a grey-green soft phyllite. This latter greyish green phyllite is labelled Op? — Ꞓb? on the geologic map because of its uncertain status. If the correlation of the upper green phyllite with the Mettawee is correct, then the black phyllite with its calcareous quartzite could correspond to the Hatch Hill Formation and the section here would be inverted, concurring with the pattern deduced by Shumaker (1959, p. 59) in the Pawlet quadrangle, in areas apparently structurally continuous with the area of the 2405′ peak.

Correlation: Dale (1899, p. 183) includes in his Cambrian ferruginous quartzite two different rock types. One, the more important type, has been named the Zion Hill Quartzite by Ruedemann (1914, p. 70). A second rock type of Dale (1899, p. 184) however is clearly the Hatch Hill; confusion of the two rock units has led to Larrabee's assignment of the Zion Hill to the Ordovician (1939, p. 49). Theokritoff (1959a, p. 63) has corrected this error.

The Eagle Bridge Quartzite (Prindle and Knopf, 1932, p. 277) at Eagle Bridge, New York, is very similar to the Hatch Hill and likewise is interbedded in sooty black shale. Lochman (1956, p. 1336) points out that this rock may be Ordovician in age. Potter (1959, p. 1658) apparently also assigns the Eagle Bridge Quartzite to the Ordovician. In the company of Potter and Theokritoff, the writer (1957) has found cross-bedding in this quartzite at the type locality, thus confirming the stratigraphic sense here. The rock is in sequence below argillites typical of the Poultney. A correlation of the Eagle Bridge with the Hatch Hill seems fully justified, as has also been pointed out by Theokritoff (1959a, p. 55). Similar rocks appear in the Cossayuna quadrangle to the north (Platt, Yale University thesis, 1960, p. 34) in the same stratigraphic position.

Poultney Slate

Name and Lithology: Keith (1932, p. 403) applied the name Poultney Slate to a sequence of Lower Ordovician grey slates found in the town of Poultney. Larrabee (1939) and Fowler (1950) mapped these units as part of the "Normanskill Formation." Theokritoff (1959b) revived the name Poultney Slate; because of the predominance of argillite, this designation will be continued in this report. The rock is characteristically a fine-grained argillite, grey, green, or motley-coloured, weathering dull white. Buff-weathering, slightly calcareous, fine-grained quartzite beds ½ inch to 6 inches thick are common and interlaminate with slate at intervals of 6 inches or less. These quartzite beds weather out in relief and may give the rock a pin-stripe appearance. For further description, see Theokritoff, 1959b; Zen, 1961, p. 306 ("unit 1") and Plate 2, fig. 4.

Another rock type, included in unit 2 of the Mount Hamilton Group (Zen, 1961, p. 306), is a jet black, pyritiferous and chert-bearing, siliceous, fine-grained argillite, with a glazed appearance where fresh and with a dull white cast where weathered. The lower part of this unit carries quartzose limestone seams up to 6 inches thick. The limestone is dark grey, very compact and fine grained, and weathers pale grey, locally showing cross-bedding (Zen, 1961, Pl. 2, fig. 3). The upper part of the argillite merges into dark grey, typical Poultney Slate with interbedded quartzite layers. This rock is lithologically like the type Schaghticoke beds of New York (Theokritoff, 1959b). Although not found in the present map area, this rock unit is mentioned here to complete the stratigraphic sequence. Following Theokritoff (1959a, p. 56), this unit is interpreted as local, basal beds of the Poultney Slate; this agrees with Berry's correlation based on graptolite zonation (1961, p. 226). Where the Hatch Hill is absent, this unit rests directly on the Lower Cambrian beds (Zen, 1959a, p. 8).

Distribution: In the map area, only the belt of slate just west of the village of Poultney definitely belongs to the Poultney Slate. Green and grey slates also occur however (1) southwest of Gorhamtown south of Lewis Brook, (2) around East Poultney, (3) about ½ mile east of East Poultney and south of the Poultney River, and (4) in the area of Vail Brook and North Brook south of Spruce Knob. These areas of slate have been mapped as Poultney Slate because of the hard, cherty, and glazed appearance, white weathering, and predominantly green colour; however, part of the overlying Indian River Slate is also green and, lacking guide fossils or key beds, the classification is tentative.

The upper part of the Poultney Slate is locally a soft grey-green slate, without the thin quartzite beds or the characteristic white cast on weathered surfaces. This slate grades upward into the Indian River Slate. In isolated outcrops, these beds are difficult to distinguish from the Mettawee slate of the Lower Cambrian Bull Formation. The patch of "Poultney Slate" east of East Poultney, for instance, may in fact be Mettawee, at least in part. On the other hand, the elongate belts of green slate west of North Brook southeast of Spaulding Hill, and on the east flank of the 2405′ knob west of Chippenhook in Clarendon, may be at least in part the Poultney Slate. On the map, these areas are labelled Op? and Op? — Ꞓb? to indicate this uncertainty.

Age: Graptolites from beds mapped as Poultney Slate have been fully reported by Berry (1959; 1961) and by Shumaker (1959; Cornell University thesis, 1960). According to Berry (1961, p. 225, 226), the Poultney Slate ranges in age from basal Early Ordovician ("Schaghticoke") to Normanskill (*Nemagraptus gracilis* or *Climacograptus bicornis* zone). Despite this wide age range, nonetheless, it is a distinct rock unit, and is comparable with, though not equivalent to, the Deepkill beds of New York (Ruedemann, 1914, p. 140).

Indian River Slate

Name, Lithology, and Distribution: Keith (1932, p. 403) applied the name Indian River Slate to the red slate between the Poultney and the overlying black unnamed slate. The type locality is on the Indian River, "a few miles" south of Granville, New York. Theokritoff (1959b) extends the name to include bluish green slates which are stratigraphical equivalents of the red, commercial slate. Theokritoff calls the unit "Indian River Formation," but the name Indian River Slate should be retained.

In the present area, soft red slates associated with green slates occur at three places:

(1) South of the Poultney River, ½ mile east of East Poultney. Here soft red slate, merging into maroon, and carrying thin green quartzite seams, occur in an area of green and grey slate tentatively mapped as Poultney. The red slate is much like "unit 3" of Zen (1961, p. 306), except for its softness. It may be a transition phase between the Poultney and the Indian River.

(2) Below the bridge across Poultney River at East Poultney, and in the pasture immediately southeast (see Larrabee, 1939, p. 52), soft, red slate, associated with green quartzite, occurs in two bands with a band of soft green slate in between. The green slate is interbedded with

2- to 6-inch beds of buff-weathering, cream-coloured dolostone. This green rock is not found elsewhere, and is here assigned to the Indian River above the red slate in the core of a syncline. The two bands of red slate are flanked on the outside by carbonate-free greenish-grey slates which merge outward into grey siliceous slates. These latter green and grey slates are assigned to the Poultney Slate.

(3) South of Lewis Brook and southwest of Gorhamtown, along the same structural belt as (2) above, another area of soft red slate occurs with green Poultney Slate. The red slate has thin green quartzites but otherwise is identical with red Indian River slates quarried south of the present area (Shumaker, 1959, p. 67); in fact several prospect pits for slate exist here.

Age: The Indian River Slate near Granville and near Hampton, New York, has yielded graptolites characteristic of the Normanskill *Climacograptus bicornis* zone (Berry, 1959, p. 61). Presumably, the same age applies to the red slates here mapped as Indian River.

Pawlet Formation

Name: A grey to black slate and greywacke unit overlies the Indian River Slate. This unit is presumably the "black slate" of Keith (1932, p. 403), and the "Hudson grit" of Dale (1899, p. 187). Shumaker (1959 p. 59) discusses this unit without giving it a name. Field consultation leads Shumaker and Zen to adopt the name Pawlet, called a formation by Zen (1961, p. 307). Shumaker (Cornell University thesis, 1960, p. 36) calls it the Pawlet greywacke. The name Pawlet Formation will be retained here.

Lithology: The Pawlet Formation has been described in detail by Dale (1899, p. 187). It consists of interbedded slate and greywacke. The slate may be silty and grey, or it may be black, fissile, soft, and locally graphitic and pyritiferous. The two types may interbed at intervals of a few inches or more; they also laterally grade into each other. The base of the unit is marked by a black slate which is commonly fossiliferous (Shumaker, 1959, p. 59).

The greywacke, which in the map area constitutes at least half of the bulk of the formation and stands out in relief in outcrops, is dark grey when fresh and weathers to dull grey or light greyish-brown. It contains black and green, angular chips of slate up to half-inch across. The bulk of the material is subangular quartz and more rarely feldspar grains, in a dark-grey argillaceous, slightly calcareous matrix. The rock thus closely resembles the Eddy Hill Grit (see Dale 1899, p. 181). The beds of greywacke are generally between 6 inches and 3 feet thick, and are separated by slate layers. Graded bedding, due to differences in maximum grain size, and yielding sedimentary top sense, is common in the greywacke; otherwise the rock is massive.

Distribution: Three belts of the Pawlet Formation occur in the area, as shown on the geologic map. The delineation of slates of the Pawlet Formation from those of the West Castleton Formation is not always easy especially where diagnostic beds like the Beebe Limestone or the greywacke are absent. Locally, therefore, the boundary of the Pawlet Formation against the West Castleton is subject to revision. The Pawlet is not recognized in the Bird Mountain thrust slice.

Age: The only fossils from the Pawlet Formation in the present area, discovered by Berry, are from an outcrop about 500 feet south of the bridge at East Poultney, in a cut on the east side of the paved road. The rock is a black slate unit basal to the Pawlet. The fossils include *Nemagraptus gracilis* but belong to the *Climacograptus bicornis* zone (Berry, 1959, letter; 1961, letter). South of the present area, this same basal unit is very fossiliferous (Shumaker, 1959, p. 67) and the forms belong to the *Climacograptus bicornis* zone (Berry, 1959, p. 62). The Pawlet Formation is therefore of Late Normanskill age or younger.[1] Potter (1959), in the Eagle Bridge and Hoosick Falls quadrangle, New

[1] An important point on the biostratigraphic dating of the Indian River Slate and the Pawlet Formation requires clarification. Berry (1959, p. 61) considers the Indian River Slate as belonging to the *Nemagraptus gracilis* zone or the *Climacograptus bicornis* zone (zones 11 and 12, resp., of his standard section of the Marathon region, Texas; Berry, 1960). Berry also states (1959, p. 62) that the Pawlet Formation belongs to the *C. bicornis* zone. These data agree well with the field relations, which includes evidence for an unconformity below the Pawlet, generally accepted by workers in the northern Taconic region (Berry, 1959, p. 62; Zen, 1961, p. 308; Shumaker, Cornell University thesis, 1960, fig. 6). The graptolites from south of East Poultney, collected from the basal beds of the Pawlet, however, include *N. gracilis* (Berry, 1959, letter). Shumaker (Cornell University thesis, 1960, p. 40), in a foot-by-foot collection of graptolites from a single outcrop within the basal beds of the Pawlet Formation, reports the finding of both *N. gracilis* and *C. bicornis* in the same bed, according to the identifications of Berry. These two index forms for zones 11 and 12 of Berry's Marathon succession thus are non-diagnostic in the Taconic region (Berry, letter, November 1961), although in the Marathon section *N. gracilis* is diagnostic of, and coextensive with, zone 11 in the Woods Hollow shale, distinct from the overlying *C. bicornis* zone (Berry, 1960, p. 26). Because of the limited area in which these discrepant data are obtained, vertical transgression of the pre-Pawlet unconformity through Berry's zones 11 and 12 cannot be the correct explanation. I am indebted to Berry and to John Rodgers for discussions of the implications of this problem.

York, finds the same formation grading laterally eastward into the Walloomsac Slate.

Correlation: By lithology and also by fossils, the Pawlet Formation is correlated with the Austin Glen Member of the Normanskill (Ruedemann, 1942, p. 102). This correlation, however, must remain tentative until more definitive study of the stratigraphy and paleontology of the type Normanskill is completed by Berry.

Taconic Sequence: Discussion

Stratigraphic Relations Within the Post-Lower Cambrian Units

In an earlier section, the significance of the Hatch Hill Formation as a time-stratigraphic unit has been discussed. Although in many areas the Hatch Hill may represent a unit below the Poultney Slate, similar rocks also occur within or above the Poultney Slate. At least some of the Hatch Hill-looking rocks, therefore, represent a lithofacies without definite time value. In addition, the Hatch Hill is by no means always present at the base of the Mount Hamilton Group; the Poultney Slate may rest directly on Lower Cambrian units.

The contact between the Poultney Slate and the Indian River Slate is difficult to map in the field, due in part to the lithologic variability of the Poultney, such that it at places closely approaches the Indian River. At Scotch Hill south of West Castleton (Zen, 1961, p. 307), for example, deep red to maroon slate with thin green quartzite layers, resembling the Indian River, is seen to be both overlain and underlain by typical Poultney Slate; in fact the two rock types grade into each other laterally within a distance of some 50 feet. The green slate of the Indian River is also often difficult to tell apart from similar slates of the Poultney.

The Pawlet Formation, on the other hand, forms a distinct unit clearly separately mappable from the Mount Hamilton Group, although its slates are not always easily separated from Lower Cambrian units or some of the surrounding autochthonous units. In the map area, the Pawlet Formation locally rests directly upon the Bull Formation or the West Castleton Formation of the Lower Cambrian succession. It also rests, without visible discordance, upon green and grey slates which are probably part of the Poultney but which may be lower beds of the Indian River. Just south of Poultney village, the Pawlet is separated from type Poultney Slate by a narrow belt of alluvium, hiding the contact completely.

The conclusion that the Pawlet Formation is younger than the Indian River Slate is based primarily on field relations. The Pawlet *never* occurs between the Indian River Slate and the Poultney Slate in an unbroken succession. In areas to the south of the Castleton quadrangle (Shumaker, 1959, p. 67; Berry, 1959, oral communication; also the writer's own observations), the Pawlet seems to be demonstrably younger than the Indian River.

The fact that the Pawlet Formation cross cuts structures of the Lower Cambrian Taconic sequence (see geologic map), and also of rock units of the Mount Hamilton Group, suggests that a regional unconformity exists at the base of the Pawlet. The fact that graptolites of the same zone are found both above and below this unconformity at one outcrop (Berry, 1959, p. 62), however, shows that the time lapse may not have been great. The hypothesis that the base of the Pawlet Formation represents a fault is improbable, because everywhere younger rocks rest upon older, and only the older structures are truncated. A possible explanation is that the Pawlet Formation was deposited during the time of emplacement of the Taconic allochthone. This would explain the abundance of greywacke, the similarity of fossils in this unit and in the upper strata of the Mount Hamilton Group, the nature of the contact between it and the underlying beds, and the apparent gradation of this unit into the autochthonous Walloomsac Slate (Potter, 1959). The Pawlet Formation would thus be a hybrid rock, bridging the gap between the allochthonous and autochthonous units, not only in lithology, but in its tectonic-sedimentological framework[1].

The absence of the Upper Cambrian Hatch Hill Formation at the base of the Mount Hamilton Group at many places, with the Lower Ordovician Poultney Slate resting directly upon older rocks, suggests that a second unconformity might exist here (see also Platt, Yale University thesis, 1960, p. 37, 90). However, the field relations might also be interpreted as due to facies overlap; the Hatch Hill Formation, in the time-stratigraphic sense of Theokritoff, may have been replaced laterally by the facies of the basal Poultney Slate.

Stratigraphic Relations Between the Lower Cambrian Succession and Later Rock Units

The Pawlet Formation rests on the older units of the Taconic se-

[1] After this section was written, my attention was called to the term "neoautochthonous," used by the Italian geologists to describe a similar concept applied to certain rocks in the northern Apennines. See Maxwell, 1959.

quence with an angular unconformity. Examples of this rock overlying a beveled older sequence are shown on the geologic map, for example, just south of Lewis Brook and east of the Delaware-Hudson Railroad track, where the Pawlet cuts across the West Castleton-Bull contact. Again, half a mile southeast of Poultney River along Route 30 (the road between Poultney and Wells), an outcrop 200 feet south of the road shows the Mettawee Slate within 2 feet of the Pawlet Formation; the Pawlet is to the east and overlies the Mettawee. This is the closest that the unconformity has been located in the present area.

The unconformity, if any, between the Mount Hamilton Group and the Lower Cambrian Taconic sequence, however, cannot be demonstrated satisfactorily. The arguments for and against such an unconformity may be briefly summarized.

In the vicinity of Hampton, New York, along the Poultney River, the complex anticlinal structure defined by the contact of the Bomoseen Greywacke Member and the Mettawee Slate strikes into, and apparently disappears under the Poultney Slate that enters the Castleton quadrangle from a northwesterly direction. This relation may be interpreted as an unconformity; but the contact is nowhere found.

The quartzite of the Hatch Hill Formation is locally missing, and the Poultney Slate may rest directly above the Lower Cambrian units. This may be due to an unconformity between the Poultney and the Hatch Hill (and by implication an unconformity against the Lower Cambrian); however, the Poultney may be in part a lateral facies of the Hatch Hill. The Hatch Hill-type quartzite is not persistent, and in its absence it is locally very difficult to draw any contact, let alone an unconformity, between this formation and the underlying West Castleton Formation.

Recent mapping by Potter (1959) in the Eagle Bridge quadrangle, New York, and by Platt (Yale University thesis, 1960) in the Cossayuna quadrangle, New York, shows that the Hatch Hill (= Eagle Bridge) Formation overlies different units of the Lower Cambrian sequence, which includes equivalents of the Bull and West Castleton Formations, as well as a monotonous grey-green slate unit (Potter, 1962, written communication). The latter grey slate is apparently the same unit reported by Platt (Yale University thesis, 1960, p. 31) at the extreme south end of the Cossayuna quadrangle. This unit wedges out farther north and is not known at the north end of the Taconic sequence. The field relations may reasonably be interpreted as beveling of units at the base of an unconformity below the Hatch Hill; but other explanations are readily conceivable and so the evidence is not compelling.

The fossiliferous Beebe Limestone Member occurs near the base of the West Castleton Formation. The fossils from the Hatch Hill Formation came from beds near the top of that formation (Theokritoff, 1959a, map; also personal communication). Therefore, the absence of Middle Cambrian fossils does not prove an unconformity, for the Middle Cambrian might be within the barren intervening beds, although the evidence can be interpreted either way.

The difficulties in interpreting the field data are compounded by the fact that the Mount Hamilton Group is known to show rapid lateral facies changes, even within the confines of the present map area. On the other hand, the West Castleton Formation, as a lithostratigraphic unit, is not persistent; its occurrence is restricted in the Pawlet quadrangle (Shumaker, 1959, p. 59), and probably even more so farther south. In the absence of fossils, it is clearly very difficult to prove an unconformity between two units both of which show lateral variation within the same general range of lithic types.

The writer tentatively suggested an unconformity between the base of the Mount Hamilton Group and the underlying units (Zen, 1961, p. 297); this however is chiefly a working hypothesis.

The relation between the Pawlet Formation and the Mount Hamilton Group cannot be demonstrated beyond doubt. Although the Pawlet does rest at places on rocks as old as the Mettawee, no truncation has been found between the Pawlet and beds of the Mount Hamilton Group. The difficulty, again, is in part interpretative, due to the lateral variability within the Poultney Slate (and perhaps also the Indian River Slate). Shumaker (1960; Cornell University thesis; also Doll et al., 1961) shows the Pawlet as truncating the contact between the Poultney and the Lower Cambrian Taconic units. This may be due to transgressive overlap, but the weight of evidence seems to favor an unconformity between the Pawlet and all the underlying units.

The above discussion serves to emphasize the difficulty in differentiating the black slate units in the field, previously discussed by Zen (1961, p. 334). This is particularly vexing when one tries to draw an upper contact for the West Castleton Formation. Lacking fossils, recognition of key beds like the Beebe Limestone, quartzite in the Hatch Hill, or greywacke in the Pawlet, is indispensable for the separation of these units.

The writer has tended, in doubtful cases, to map in favor of the West Castleton Formation, so that its areal extent of the geologic map may be taken as a maximal one. It is not believed, however, that errors due to these uncertainties would radically affect the regional map pattern.

Synclinorium Sequence

PRECAMBRIAN: MOUNT HOLLY SERIES

Although the Precambrian rocks are not part of the Synclinorium sequence, the small areas of these rocks shown on the geologic map are conveniently discussed here.

Precambrian rocks are found only on the upthrown and east side of a major fault, called the Pine Hill thrust (Wolff, 1891, p. 337; Dale, 1892, p. 516; Fowler, 1950; Brace, 1953). Fowler (1950, Pl. II) mapped these rocks as part of the basal Cambrian (?) Mendon Series; Brace (1953, p. 78) correctly recognized the Precambrian gneisses here. The name, Mount Holly Series, is used by correlation with similar-looking rocks in the Green Mountain anticlinorium (Osberg, 1952; Brace, 1953). In the present map area, the rocks consist of biotite-chlorite-microcline gneiss, biotite-chlorite schist, calcareous and feldspathic, locally graphitic siltstone, biotite gneiss, and coarse-grained calcite marble. Most of the rocks are clearly of sedimentary origin. A green, chlorite schist with pebbles of blue quartz up to 5 mm across is found on the north slope of Boardman Hill associated with a black, calcareous, silty schist; this rock may belong to the overlying Mendon Series (Brace, 1953; Osberg, 1959). However, outcrops of this rock are extremely limited and are here grouped with the undifferentiated Precambrian units.

The lithology of the Mount Holly Series was described by Brace (1953) and will not be repeated.

LOWER CAMBRIAN: CHESHIRE QUARTZITE

Within the map area, the Lower Cambrian Cheshire Quartzite (Fowler, 1950) was found only east of the Mount Holly Series of the Pine Hill thrust block. A major unconformity exists between these rock units. The Cheshire Quartzite is generally a massive, homogeneous, white to buff vitreous quartzite. The summit of Boardman Hill is underlain by thin-bedded quartzite with interbedded gneissic layers; immediately east of the summit, the quartzite is dark grey though otherwise similar to the typical Cheshire; this area of quartzite was traced to typical Cheshire outcrops on the north slope of Boardman Hill. Brace (1953) shows this quartzite as Precambrian, but does not consider the evidence compelling (1959, oral communication). The assignment of the quartzite to the Cheshire is in agreement with the map pattern immediately south, in the Flat Rock area (Thompson, 1959, map).

The Early Cambrian age of the Cheshire Quartzite is amply documented (see, for instance, Brace, 1953, p. 35) and needs no repetition.

Although in the Chippenhook-Clarendon Springs area no fossil is found in the Cheshire Quartzite, the Dunham Dolostone, which overlies it with gradational contact (Thompson, 1959, p. 79) *west* of the Pine Hill fault, has yielded several species of *Hyolithes* and *Salterella pulchella* Billings (Dale, 1894, p. 535).

LOWER CAMBRIAN: WINOOSKI DOLOSTONE

The Winooski Dolostone has been described by Cady (1945, p. 532), Bain (1938, p. 8; 1959, p. 36) and Fowler (1950). Although no fossil has been found in this rock unit, its conformable relation with the underlying and older fossiliferous Dunham and Monkton suggests an Early Cambrian age; on the other hand the gradational contact with the younger, Danby and Clarendon Springs formations, correlated with Upper Cambrian rocks of the Champlain valley sequence (Cady, 1945; Fowler, 1950) suggests that the Winooski may be at least in part Middle Cambrian (Thompson, *in* Billings *et al.* 1952, p. 39).

In the present area, the Winooski consists of massive beds of dolostone, grey, pink, or buff in colour and weathering grey to buff. Thin beds of arkosic sandstone, up to three feet thick but more commonly less than 6 inches, weathering to a pale greenish cast, characterize this formation. Rarely, beds of black silty slate also occur in the Winooski, for example in the railroad cut just west of Gorham Bridge across Otter Creek near Florence. For further description of this rock unit, see Fowler (1950, p. 21) and Cady (1945, p. 532).

In the present area, the Winooski Dolostone occurs in two tracts, both on the downthrown side of the Pine Hill fault. One area is west of Boardman Hill, in the core of a north-plunging anticline. Because of the difficulty in determining the upper contact of this formation (see below), however, the areal extent of the Winooski here may be subject to revision. Fowler (1950) showed the same area as underlain by the younger Danby Formation; the presence of characteristically Winooski rock types here, such as the arkosic sandstone, indicates the present classification.

The second tract of Winooski, west of Pine Hill, is in the core of a south-plunging anticline whose west limb is the homoclinal sequence on the east limb of the Middlebury synclinorium to the north. The Winooski here is typical in appearance even though its upper contact is again somewhat arbitrarily drawn.

Upper Cambrian: Undifferentiated Danby—Clarendon Springs Formations

Name and Description: These two formations have been described in the literature (Keith, 1932, p. 396; Cady, 1945; Fowler, 1950; Thompson, *In* Billings *et al.* 1952, 1959; Bain, 1938; Brace, 1953) and correlated (Cady, 1945, p. 535; Rodgers, *In* Billings *et al.* 1952, p. 35–36) with the Potsdam-Little Falls and lower Whitehall formations (Rodgers, 1937, p. 1575–1576; Division A of Brainerd and Seely, 1890) of the Lake Champlain area. These formations, therefore, are supposedly of Late Cambrian age.

The Danby Formation consists of interbedded vitreous quartzite and grey massive dolostone. The overlying Clarendon Springs Dolostone, into which the Danby grades, is a homogeneous sequence of massive grey dolostone. The rocks are well described by Fowler (1950, p. 22–23) for the present area. The writer, however, finds that the quartzite beds diagnostic of the Danby are commonly poorly developed or exposed, so that separation of these two formations is not generally practicable. They are therefore mapped together.

Many beds of the Clarendon Springs Dolostone contain irregular aggregates of milky quartz that weather out in relief. Such aggregates break up easily into a friable mass, due apparently to the loss of interstitial carbonate cement. Mapping of the area near West Haven, Vermont, shows that the correlative of the Clarendon Springs Dolostone (unpublished data; see also Rodgers, 1937, p. 1576; *In* Billings *et al.* 1952, p. 35) commonly contains fine-grained, grey chert nodules of similar shape and abundance. The quartz bodies in the Clarendon Springs are probably recrystallized chert and are thus useful in stratigraphic correlation.

The dolostones of both the Danby and Clarendon Springs formations are commonly cross-bedded. The cross-beds occur in zones up to a foot thick. Some of the apparent cross beds, when traced in detail, are festoon shaped in cross-section; they are channel fills. While of little use for current direction determinations, they remain valuable in determining the sedimentary top-direction.

The quartzite of the Danby Formation differs from that in the Winooski, in that it is a clean orthoquartzite, medium grained and vitreous, weathering to a grey massive, waxy surface. Beds of the quartzite occur sporadically in the map area, though confined to the same general stratigraphic zone; individual beds are commonly about one foot thick though they may range to as much as 4 feet.

Between unquestioned Winooski Dolostone and Danby-Clarendon Springs formations, there is, especially between Florence and Proctor, a zone of rocks of hybrid character. This zone is up to 100 feet thick. Thick quartzites in this zone are commonly the Danby type, but thinner beds are impure and resemble those in the Winooski. Thin, greenish, silty to phyllitic partings in dolostone, like those in the Winooski, also occur. The dolostone beds themselves are variable, ranging from the buff-to-brown weathering, vari-coloured type of the Winooski to the grey-weathering grey dolostone of the Danby. They may be cross-bedded as in the Danby. This transition zone may be seen, for example, on the knolls southwest of the 364′ road corner about one mile south of Florence, and also south of the 372′ road corner just to the south.

Lower Ordovician: Shelburne Formation

Name: Keith (1932, p. 397) first proposed the name Shelburne Marble for a calcite marble unit between his "Clarendon Springs Dolomite" and his "Williston Limestone." Bain (1938, p. 10) mapped, in the West Rutland-Proctor belt, three distinct lithologic units, which he called the Sutherland Falls Marble, the Intermediate Dolomite, and the Columbian Marble. Cady (1945, p. 540) includes these units of Bain in the Shelburne Marble, which underlies the Cutting Dolomite (1945, p. 541); both are assigned by Cady to the Lower Beekmantown. Fowler (1950, p. 24) includes the Cutting and the Shelburne in his Boardman Formation; the Cutting being regarded as a lateral facies of the Columbian Marble Member of the Boardman (1950, p. 26). Thompson (*In* Billings *et al.*, 1952, p. 38) follows Fowler but later (1959, p. 75) uses the name Shelburne *Formation*. South of the Brandon area, the Cutting loses its distinctive lithology and usefulness as a mapping unit; it merges into the Bascom Formation, not the Columbian marble. Cady and Zen (1960, fig. 4) use the name Shelburne Formation specifically excluding the Cutting Dolostone; this usage is followed here. The Shelburne Formation includes the Sutherland Falls Marble, the Intermediate Dolostone, and the Columbian Marble as its three members, named from oldest to youngest.

Lithology: The nature of the three members of the Shelburne Formation has been described in the literature (Fowler, 1950, p. 25; Cady, 1945, p. 540; Bain, 1933, p. 77; 1938, p. 10; 1959, p. 37). Bain (1938, p. 10) states that the Sutherland Falls Member "is a 90 feet, cream colored band with contorted chains of dolomite grains across any surface. It has one siliceous band (the Hen Hawk layer) through the middle." The Intermediate Dolomite "is a thick bedded, grey dolomite containing some siderite. . . . The sand content of this formation rarely exceeds 15 per cent"; and the Columbian Marble Member "is 500 to 600 feet thick. The lower 50 feet is lithologically similar to the Sutherland Falls Marble but contorted forms are not conspicuous and the markings usually have a linear pattern. The entire formation weathers white but the color varies from white to dark gray. The upper part is usually light with numerous green silicate bands. Thick, buff weathering, argillaceous, gray dolomite bands mark the upper limit of the formation."

The "contorted chains of dolomite" do not always occur in the Sutherland Falls and so it is not always possible to distinguish the Sutherland Falls from the Columbian Members, or indeed either from local white marble bands in the overlying Bascom Formation. The criterion of contorted versus linear dolomitic chains clearly must be used with caution. The dolomite chains are probably recrystallized "dolomite curdling" (Cady, 1945, p. 543, 548) or "dolomite mottling" (Cady and Zen, 1960 p. 728; Zen, 1959b, p. 668) such as is found throughout the carbonate strata of Late Cambrian through Early Ordovician age in west-central Vermont and eastern New York (Cady, 1945; Brainerd and Seely, 1890; Rodgers, 1937; van Tuyl, 1914, p. 350; Fisher, 1954, p. 82). The stratigraphic value of such features is dubious.

The problem of distinguishing the Sutherland Falls from the Columbian Member is compounded by the lithologic similarity between the Intermediate Dolostone and the Clarendon Springs Dolostone. Thus, if a stratigraphic section is incomplete, or if the Sutherland Falls Member is locally missing, the identity of the Columbian Member cannot always be established. In general, however, the Columbian marble is the thickest and purest white marble unit in the area.

Distribution: In the map area, the Sutherland Falls Member occurs sporadically, pinching and swelling rapidly along strike, in agreement with Bain's observations (1959, p. 37). As the marble has been plastically deformed and shows features of flowage, such as boudinage, relative to the interbedded dolostone, it is not feasible to determine how much of this variation in thickness should be ascribed to original sedimentation.

The top of the Columbian marble and the Shelburne Formation as a whole is clear-cut where the Cutting Dolostone is present. In the present area, however, where the Cutting is absent, the Columbian marble may merge into the gray-streaked marble of the overlying Bascom Formation. Good examples of this gradation may be seen west of the Clarendon and Pittsford railroad tracks just north of the Proctor town (not village) line. The occurrence of brown-orange weathering dolostone, if present, then becomes the best means of establishing the presence of the Bascom Formation.

The Columbian marble contains solution caverns of considerable dimensions, one of which occurs in the present map area, near Chippenhook. The location and access to these caverns have been compiled by Scott (1959).

Lower Ordovician: Bascom Formation

The Bascom Formation (Cady, 1945, p. 542) was defined to overlie the Cutting Dolostone (1945, p. 541). Fowler (1950, p. 26) includes the Cutting and perhaps part of the Bascom in his Boardman Formation. For reasons already stated, in this report the Bascom Formation includes the lateral correlatives of the Cutting and is distinguished from the underlying Shelburne Formation by the highest stratum of the Columbian marble.

The Bascom Formation is a heterogeneous unit. In the present area, it consists largely of brown to orange weathering quartzose and silty, or pure and massive, dolostone a few inches to 4 feet thick, interbedded with grey-streak, blue-grey, sandy weathering marble. White massive marble locally occurs near the base of the formation. Black phyllite and white siliceous partings and rare orange-weathering, grey, calcareous siltstone beds are diagnostic of this unit, which may otherwise resemble the overlying Chipman Formation. The stratigraphic-structural problem involving the Bascom Formation at the north end of Whipple Hollow, near Florence, has been considered earlier (Zen, 1961, p. 308, 322).

The Bascom Formation occurs in three belts. The smallest area is about 1.5 miles west of Chippenhook. This area of the Bascom Formation is separated from the Shelburne Formation to the east on the basis of the characteristic dolostone beds. This outcrop, though areally limited, demonstrates the unconformity at the base of the black phyllite, which overlies it as well as the Shelburne Formation.

A second area of Bascom Formation is on the east side of the West Rutland marble belt, mapped by Fowler (1950) in part as Whipple Marble and in part as Beldens Marble. In this report, the structural interpretation of this belt is changed, and the sandy weathering, grey-streak marble with its siliceous partings is assigned to the Bascom, the oldest unit in a homoclinal section. The rock appears identical with the

Bascom formation on the east limb of the Middlebury synclinorium, near the Proctor–Brandon quadrangle boundary (Zen, 1961, p. 308).

The third and areally most extensive belt of Bascom Formation occurs north of Proctor and extends to the abandoned quarries northwest of Florence, considered elsewhere (Zen, 1961). This belt contains all the lithologic varieties of this formation, and on the west side it is truncated by black phyllites of the Taconic Range.

Lower Ordovician: Chipman Formation

The Chipman Formation was re-defined by Cady and Zen (1960) to include as its members four previously recognized lithologic units: the Bridport Dolostone (Cady, 1945, p. 545), the Beldens Formation (Cady, 1945, p. 550), the Weybridge Member of the Beldens (Cady, 1945, p. 550), and the Burchards Formation (Cady, 1945, p. 548; Kay and Cady, 1947; Kay, 1950). The Chipman is the highest unit of the Lower Ordovician rocks in the synclinorium sequence.

The Chipman Formation occurs in the West Rutland belt. Here it includes the Beldens, the Burchards, and the Weybridge Members. The Beldens Member, in which the principal quarries at West Rutland are located (Fowler, 1950, p. 79; Bain, 1938, p. 7; 1959, p. 36) is a snow white to pinkish white marble, with green streaks due to aggregates of chlorite and actinolite. Pale gray massive dolostone beds up to 3 feet thick are common; these beds weather typically buff and craggy. Local bands of snow white marble with white dolomitic mottling are probably the Burchards Member. With the higher metamorphic grade here compared to the Brandon–Sudbury area (Zen, 1959b, 1960a), however, the Burchards Member, easily recognized in the latter area (Cady and Zen, 1960), becomes increasingly difficult to distinguish from the Beldens, as the dolomite in the mottling patches reacts with other minerals in the rock to form magnesian and calcium-aluminum silicates including chlorite, actinolite, and zoisite (Zen, 1960a, p. 155).

The Weybridge Member occurs east of the quarry zone of Beldens, in the West Rutland belt, and stratigraphically underlies the Beldens and Burchards members. It is easily distinguished from the other members by the alternating 1- to 2-inch beds of pale grey marble and brown-weathering sandy, slightly dolomitic layers.

Middle Ordovician: Ira Formation

Name and Definition: Keith (1932, p. 398) suggested the name Ira Slate for black phyllites and slates occurring in the town of Ira and adjoining areas. By correlation with the black slates at Hortonville (Cady, 1945, p. 559), the name Hortonville has been applied to these same black slate and phyllite units east of the Taconic Range in recent years (Fowler, 1950; Thompson, *In* Billings *et al.*, 1952; MacFadyen, 1956).

The age of the type Hortonville slate, at the village of Hortonville, is itself unproven (Zen, 1961, p. 311). Moreover, the black slate and phyllite east of the Taconic Range cannot be traced unambiguously into the Hortonville, and any correlation between these areas of black slates must be based on the comparison of the meager shelly fauna in the black slates of the Orwell area (Kay, 1959, p. 18)[1] with the equally sparse graptolite fauna in the black phyllite near Hoosick Falls, New York, which is an extension of the Ira belt (Prindle and Knopf, 1932, p. 274; MacFadyen, 1956, p. 27; Potter, 1959, also personal communication), and perhaps with the fauna in basal limestones associated with the black phyllite (see below). These faunas have no element in common. The writer therefore suggested to revive Keith's name Ira for the black, mid-Ordovician slate and phyllite east of the Taconic Range (1959a, p. 3; Thompson, 1959, p. 75).

Fowler (1950, p. 35) correctly recognizes the essential stratigraphic unity between the black "Hortonville" slate and the limestone and marble unit which he calls the Whipple (1950, p. 32). The Whipple, as Fowler defines it (1950, p. 33) is basal to the Hortonville, but it is not always present at the base; "the Whipple is in part a facies of the Hortonville Slate" (1950, p. 33). The writer agrees with this interpretation of the relationships. The name Ira has therefore been emended (Zen, 1961, p. 310) to include both the Whipple marble and the black slate and phyllite which is partly interbedded with and partly overlies the marble (Ira Slate of previous workers); it is accorded the rank of a formation.

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[1] Kay reports *Dinorthis* sp. from two localities at the southeast corner of the town of Orwell, about 1/2 mile northeast of Johnson Pond (1959, Pl. B-1), as well as *Reuschella* sp. from an interbedded limestone bed nearby. The writer, in 1959, collected from the southernmost of Kay's starred localities (Kay, 1959, Pl. B-2; in pasture at el. 410′ 0.3 mile east-southeast of 745′ knob of Felton Hill and 0.03 mile west of road running east of Felton Hill, Orwell town, Sudbury quadrangle). These were kindly identified by G. A. Cooper, of the U. S. National Museum. Cooper (memorandum, January 5, 1960) reports as follows: "The specimens you sent over represent two species of brachiopods. The commoner one is the strong-ribbed orthoid, which I identify as *Dinorthis* and relate it to *D. sweeneyi*, which is fairly common fossil in the Decorah (Ion member). The other species I am unable to identify satisfactorily. It is a rather fine-ribbed dalmanellid, but I can go no further than that with it." The specimens are in deposit with the National Museum.

Whipple Marble Member: Fowler (1950, p. 32) defines the Whipple Marble to include the "dark blue-grey marble that is best exposed in a band about 2.5 miles long extending northward along the east side of the West Rutland valley from a point about 2 miles north of West Rutland." It includes Bain's True Blue Marble of the quarry belt (Bain, 1938, p. 10; 1959, p. 35). Fowler tentatively correlated the Whipple Marble with the Shoreham Member of the Glens Falls Limestone, and thus it is supposed to be of mid-Trenton, Sherman Fall age (1950, p. 34; Twenhofel *et al.*, 1954).

Fowler includes (1950, p. 32) in the Whipple grey marble and brown-weathering dolostones found (1) in the West Rutland belt, (2) at the north end of Whipple Hollow near Florence, and (3) along the marble belt west of Boardman Hill, as far south as Chippenhook. However, Zen (1961, p. 308), on structural and lithologic grounds, re-assigned the marble mapped as Whipple near Florence as part of the Bascom Formation.

On the other hand, Fowler (1950, p. 30) mapped a narrow band of Chazyan Middlebury Limestone on the west side of the West Rutland belt, which he states to grade downward into the upper Beldens (1950, p. 30), and shows as unconformably underlying the Whipple Marble. The Chazyan age of these marble outcrops is based on the presence of *Raphistoma* (*Pleurostomaria*) *staminium* (Hall) and a plate of *Paleocystites tenuiradiatus* (Hall), identified by Billings (1872; Fowler, 1950, p. 31), the latter fossil supposedly a guide to the "Chazy." Accordingly, Fowler assigned the Middlebury to the late Valcour age (1950, p. 31).

Fowler's age assignment may be questioned. The Valcour age is based on comparison with the type Chazy and with the Chazy of the Highgate Springs sequence (Fowler, 1950, p. 31, reference to Cady, 1945, p. 553). The comparison is not justified until the Chazyan age of the Middlebury Limestone at West Rutland is itself established. Billings himself was guarded in his identification of the fossils (1872):

> (The fossils) consist of numerous obscurely preserved forms like *Pleurostomaria staminia*, small encrinal joints, and a single plate of *Paleocystites tenuiradiatus*. I think this collection is Chazy. . . . The plate of the Cystidean, *P. tenuiradiatus* Hall, is a never-failing guide to the Chazy; at least it is so on the west side of Lake Champlain.

It is not clear what Billings had in mind when he used the name Chazy. In view of the field relations, namely the lithologic identity between the "Middlebury" and typical Whipple Marble in Whipple Hollow, their great faunal similarity, the sharp break between the "Middlebury" and the underlying Chipman Formation contrary to Fowler's statement, and finally the continuity of "Middlebury" and Whipple outcrops along the strike, it is suggested that they are here one and the same rock unit.

The writer, in the summers of 1958 and 1959, has found numerous fossil localities in the blue grey marble of the West Rutland belt and farther south in the town of Ira. To list all of these would take too much space, as most of the outcrops, when closely examined, show fossil fragments; only a few significant ones are given below.

1. At the 1100-foot level in a gully just east of the gravel road, ½ mile west of the 1408′ knob, in the town of Ira and about 2 miles northwest of Clarendon Springs. This is a limestone lens in black phyllite, and the fossils are largely cup corals and crinoid stems. This locality is here designated the type locality of the Ira Formation.

2. In a brook about ¼ mile north of the gravel road leading north of Clark Hill, just south of the southernmost quarries (now abandoned) in the Beldens Marble. Elevation 520 feet. Abundant echinoderm (cystid?) plates and gastropods, as well as crinoid stems.

3. On the west side of Whipple Hollow about 1 mile north of the paved east-west connecting road and ½ mile south of the 553′ triangulation station; elevation 530′, by junction of path leading east across Whipple Hollow. The limestone is east of the road, underlying black phyllite which occurs along the road. Fossils include cup corals, cystid? plates, colonial bryozoa, gastropods, and orthoconic cephalopods, as well as crinoid stems.

4. On the 550′ knob in Whipple Hollow, ¼ mile southeast of the 545′ benchmark directly east of Grandpa Knob. Black limestone interbedded with three-inch beds of black phyllite has yielded cystid? stems. About ⅓ mile farther south-southwest, and east of the 569′ road corner, a low knoll in the flats of Whipple Hollow is underlain by similar limestones associated with black phyllite; the limestone carries abundant cystid? stems.

5. At the 620′ level in the mountain brook about ½ mile north of the 950′ knob, and about 1 mile east of Clark Hill, a slump block of black limestone in an area of limestone outcrop, evidently little-transported, yielded a single cup coral.

In addition, there is the fossil locality in the "Middlebury" limestone, west of the athletic field of West Rutland High School (Cady, 1945, p. 553) which presumably yielded the fossils identified as Chazy by Billings.

Between here and locality 2 above, a distance of half a mile, all similar limestones, which are easily traced into one another, carry similar fossils. Bain (1938, p. 10) also reports "crinoid stems, orthocerid, gonioceras, turritiforme and other gastropods and colonial corals" in the "blue marble" which unconformably overlies the Beldens in the West Rutland quarries (1938, p. 10, 11; 1959, p. 78).

Harry B. Whittington of Harvard University made collections from several of these localities. The coralline components of the faunae were studied by Helen Duncan of the United States Geological Survey, who reports as follows (Duncan, written communication, Feb. 25, 1960):

> Lot 2329 (locality 1 above): "horn coral, indeterminate (chip from side of corallum). . . . The fragment . . . obviously is not identifiable as to genus."
>
> Lot 2462 (locality 3 above): "*Streptelasma*? sp. indet."

Duncan comments that "the beds that yielded [these] collections of solitary rugose corals presumably are not older than Black River and might be considerably younger." "So far as I can tell, these specimens have no features that would not be present in Ordovician streptelasmatids, [and] if the specimens are Ordovician Streptelasmas, they exhibit a stage of complexity I would expect to find in early Trenton forms." Duncan (oral communication, 1960) considers these corals as "indicative of post-Chazyan rocks; they could be late Lowville or Chaumont in age but might be as young as Rockland or Kirkfield. So far as is known, solitary corals have not been found below the zone of *Nemagraptus gracilis*."

The age assignment is in complete accord with the age of limestone and quartzite at Carver Falls and near Shaw Mountain, in West Haven in the Champlain valley belt (Zen, unpublished data). These rocks also rest unconformably above the Late Canadian rocks (Bridport Member of the Chipman Formation), and contain corals which Duncan considers to be probably Late Black River or Early Trenton in age (Duncan, written communication, Feb. 25, 1960).

Because of the continuity of "Middlebury" and Whipple outcrops, the two are clearly one and the same unit in the West Rutland area. Because of the fossil evidence, the correlation of the Whipple with the upper Shoreham Member of the Glens Falls Limestone is not established. The mid-Ordovician unconformity, commonly assigned to the Trentonian (Fowler, 1950, p. 77; Cady, 1945, p. 560; Zen, 1959a, p. i) may therefore be older in the West Rutland area, of Black River or even possibly Chazyan age, a contingency speculated on by Thompson (*In*

Billings *et al.*, 1952, p. 16). The Whipple Marble may range as young as the Glens Falls; however, the age for lot 2329, at the base of the Ira Formation suggests that the base of the slate sequence of the Ira may be slightly older than the black slates in the Hortonville Slate (Cady, 1945, p. 558) on the west side of the Middlebury synclinorium, where it overlies the Shoreham Member of the Glens Falls Limestone of Sherman Fall age (Twenhofel *et al.*, 1954, Chart 2). It would be desirable to re-examine Billings' identification of *Paleocystites tenuiradiatus* and also its age significance.

The assignment of the "Middlebury" limestone near West Rutland to a younger age, and its combination with the supposedly younger Whipple Marble, does not imply that the principal tracts of the Middlebury Limestone in the Middlebury synclinorium are necessarily post-Chazyan in age, although it does deprive this formation of the bulk of its published paleontological age evidence (Cady, 1945, p. 553).

Finally, two features about the distribution of the Whipple Marble should be mentioned.

First, there are the isolated patches of limestone mapped by Fowler as Whipple and Orwell? (1950, p. 31; Pl. II) in the town of Ira. The more southerly of the two areas carries abundant gastropods that may be *Maclurites logani* (Fowler, 1950, p. 32). Both patches are located at the contact between green and black phyllites; at the northern patch an outcrop shows green phyllite resting directly upon the marble and the entire marble outcrop is surrounded by a belt of green phyllite which may be as narrow as a few feet. The green phyllites are part of the allochthonous Taconic sequence, here regarded as Lower Cambrian. The limestone patches are thus interpreted as slivers dragged along the sole of the Taconic thrust, rather than as local anticlines (Fowler, 1950, p. 32); structurally they are part of the allochthone.

Second, a belt of Whipple Marble exists on the east side of the West Rutland marble belt, resting discordantly upon grey marble here mapped as the Bascom Formation. A similar band of Whipple rests upon Shelburne Marble in a quarry just south of Florence. Yet another band underlies the *inverted* sequence of Bascom marble east of Biddie Knob, between the Bascom above and a window of black phyllite below (Zen, 1961, Pl. 1). These relationships furnish clues on the evolution of the structures in the West Rutland-Whipple Hollow area; this problem will be examined later.

Slate and Phyllite of the Ira Formation: The bulk of the Ira Formation consists of slate and phyllite, which Keith called Ira Slate (1932, p. 398). The rocks are black to dark grey, micaceous to silty, and weather dull to rusty due to pyrite. Bedding is commonly absent in this rock; where visible it is due to (1) interbedded quartzite a few inches thick; (2) rusty, porous-weathering limy bands, or (3) thin (½ mm or less) white sandy laminae. North of the Clarendon River, about half a mile east of West Rutland, there is an area of grey vitreous quartzite more than 20 feet thick. This was mapped by Fowler (1950, Pl. II) as a klippe of the Cheshire Quartzite in front of the Pine Hill thrust. However, this quartzite is actually interbedded with black phyllite identical with the bulk of the Ira. It is here interpreted as a local quartzite lens in the Ira Formation.

Locally, orange-weathering dark dolomitic beds, up to 6 feet thick, occur in the phyllites of the Ira Formation. Another important rock type is a dark grey to jet black phyllite with albite porphyroblasts less than ½ mm across, interbedded with black or grey non-albitic phyllite (Zen, 1961, p. 310). This rock occurs chiefly along the contact with green phyllites that belong to the Taconic sequence. An extensive belt however occurs within the black phyllite terrane, on the west slope of the sharp ridge between Otter Creek and Whipple Hollow, in the vicinity of Butler Pond. This is undoubtedly the rock which Fowler mapped as a klippe of the Mettawee Slate (1950, p. 66; Pl. II). However, the rock is not of the Mettawee type, and occurs as a narrow but traceable band within the black slates (Zen, 1961, p. 323). It is now mapped with the Ira Formation.

At a number of outcrops, the phyllite of the Ira Formation is seen to rest directly on older units, without the intervention of the Whipple Marble. The contact between these rocks, where actually exposed, appears conformable;[1] however the map pattern clearly shows that the Ira Formation, within the map area, rests variously on the Chipman, the Bascom, the Shelburne, and the Danby-Clarendon Springs formations. Immediately south of the present area, Dale (1892, 1894) and Thompson (1959, Pl. H-1) show the same unit to rest on all older units including the Precambrian. Thus, a regional unconformity clearly exists.

A persistent band of black phyllite, extending from the west side of Pine Hill through Center Rutland and to the west side of Boardman Hill, was mapped by Fowler in part as "Hortonville" and in part as his "type 3" of the Mendon Series (1950, p. 14), underlying the Cheshire Quartzite.

[1] One of the few actual exposures of the contact is between the phyllite of the Ira Formation and the Columbian Marble Member of the Shelburne Formation; this is on the 530-foot subsidiary knob directly under the telephone line west of a local swamp, 1¼ miles south of Florence. These features are shown on the Proctor topographic sheet. Another good exposure, involving these units, is on the quarry wall immediately west of the West Rutland-Chippenhook road and just south of it crossing with the Clarendon River.

Similarly, Brace mapped this band as partly "Hortonville" and partly the Moosalamoo Member of the Mendon Formation (1953, p. 32; Pl. I).

The writer interprets all the lustrous black and grey phyllites in this belt as part of the Ira. Black phyllite interbedded with the Cheshire is generally silty and contains pebbles of blue quartz, rather than a lustrous micaceous black phyllite. This interpretation accords with the mapping of the black phyllite as Ira (Thompson, 1959, p. 79) in the Chippenhook area which is continuous with that west of Boardman Hill (Brace, 1953, Pl. I; Thompson, 1959, oral communication; also the writer's reconnaissance observations). The mid-Ordovician age of this black phyllite is established by fossils in the associated Whipple Marble Member (Dale, 1892, p. 517), found by Foerste, in the area immediately southeast of the present map and reported on currently by Thompson (1959).

The age of large areas of black phyllite adjacent to the Taconic sequence is unknown. Part is indubitably Ira, but part may be in the Taconic sequence. On the geologic map, the contacts are drawn rather arbitrarily; however, a contact must exist somewhere within the black slate terrane. Keith (1913) places all the black phyllite in the allochthonous Taconic sequence and regards the phyllite - marble contact as the sole of his Taconic thrust. Cady first suggests (1945, p. 560), and Fowler demonstrates that the black phyllite is at least in part mid-Ordovician in age, resting unconformably upon older units (1950, p. 36). Fowler therefore places the sole of the Taconic thrust at the contact of green and black phyllites (1950, p. 65).

The writer (1961, p. 313) shows that the sole of the Taconic thrust probably lies *within* the black phyllite rather than at either of its contacts. The exact location of this dislocation however is conjectural, and it may be intrinsically impossible to place precisely because of tectonic commingling of the allochthonous and autochthonous units (Zen, 1959c; 1961, p. 313). This uncertainty must be borne in mind when the map is perused. The ambiguity, luckily, should not affect the main argument of regional structural interpretation.

STRUCTURAL GEOLOGY

Taconic Sequence

Introduction

Within the region underlain by rocks of the Taconic sequence, four areas may be separately discussed, because of their structural as well as stratigraphic distinctness. These are (1) the area west of a line approximately from Gorhamtown to East Poultney; (2) bulk of the area east of this line; (3) the small area in the vicinity of Vail Brook and North Brook just north of the village of Middletown Springs (in the Pawlet quadrangle); and (4) the area around the 2405′ peak west of Chippenhook (Pl. 1).

Structures West of the Gorhamtown-East Poultney Line

The structure west of the Gorhamtown-East Poultney line (shown on the geologic map as the trace of the Bird Mountain thrust fault), and within the present map area, has been discussed briefly in an earlier report (Zen, 1961, p. 320) after preliminary work there. Minor revisions have been made with additional work; what follows is supplementary to the previous study.

At the west side of this area, the Lower Cambrian Bomoseen Greywacke occurs extensively and appears to be the oldest unit here exposed. It is interpreted as occupying the center of a complex anticline.[1] This interpretation agrees with the fact that, immediately to the east there occur several belts of the younger, and locally fossiliferous, West Castleton Formation. These belts of the West Castleton merge to the south, due to the south plunge of the structures. The synclinal structure of the westernmost of these belts is outlined in the field by a continual series of quarry holes in the uppermost part of the underlying Bull Formation. A double-pronged belt of Bomoseen Greywacke plunges under the structure just north of Blissville; this belt of the Bomoseen can be traced into the anticline just mentioned (Zen, 1961, Pl. 1).

In a large quarry of the Vermont Structural Slate Company, just south of the road leading west from Route 30 at its crossing with the Delaware and Hudson Railroad tracks south of Blissville,[2] operations in 1957 revealed a band of black slate with interbedded black limestone, like the Beebe Member of the West Castleton Formation, between the main pit of green slate and the higher, eastern pit of variegated purple and green slate; the slates in the two pits are rather dissimilar. Because

[1] The terms, anticline and syncline, will be used in this report strictly in the geometric sense; thus an anticline is a fold whose limbs merge in the air and a syncline one whose limbs join underground. For folds containing stratigraphically older or younger beds in the core, the terms bottoming or topping folds, respectively, have been proposed (Zen, 1961, p. 313).

[2] The road on the topographic map no longer exists; its site is now the quarry. The road has been relocated a few hundred feet to the north.

the black slate and limestone can be traced north into the main belt of the West Castleton Formation in a well-defined syncline, interpretation of the exposure in the quarry as a local facies within the slates of the Bull Formation is ruled out. This outcrop had previously been explained as the center of a recumbent syncline (Zen, 1961, p. 320); the contrasting nature of the slates on either side was considered as due to lateral facies variation. Yet another explanation, which I now favor, is that at the south end of the syncline of the West Castleton Formation a local thrust fault has sheared off its reverse limb and brought the purple and green slates of the Bull Formation on top of the West Castleton. This is the same sort of explanation given (Zen, 1961, Pl. 1) for the southward termination of a syncline of the West Castleton Formation northeast of Fair Haven.

Two extensive belts of the Pawlet Formation occupy the area east of Route 30 (main road between Castleton Corners and Poultney). These belts are open synclines; the rocks in them are the youngest in the Taconic sequence and are right-side up as shown by graded bedding in the greywacke beds. The Pawlet is unconformable upon older rocks, truncating the structure of the underlying units. Near the village of Poultney, the Pawlet Formation comes close to the Poultney Slate; however, the contact is not exposed. It is possible that a belt of Lower Cambrian rocks actually intervenes between these Ordovician strata.

The eastern belt of the Pawlet Formation directly underlies the Bird Mountain fault to its east over a considerable distance. Near East Poultney, and again west of Gorhamtown, however, areas of green and grey Poultney? or Indian River? Slate, and red Indian River Slate, occur between the fault and the Pawlet Formation. The following observations pertain to the structural interpretation of these outcrops.

1. At East Poultney, two bands of red slate occur. On the outer sides of both bands, the red slate passes into rather typical Poultney grey and green slate; between the two bands is a peculiar green argillite which strongly resembles the green Indian River Slate. The relations indicate a topping fold here.

2. To the east of both areas of the Indian River Slate, a sharp topographical break exists, with higher hills to the east underlain by an inverted section of the Bull and the West Castleton formations. This line of scarp is a continuous expression of the trace of the Bird Mountain fault.

3. The relation between the Indian River Slate and the Pawlet Greywacke in the slate belt as a whole (Berry, 1959, p. 62, and oral com-

munications; also the writer's own observations) indicate that the Indian River is older than the Pawlet. In the present area, data are consistent with this conclusion, although they do not demonstrate it. Significantly, however, the only graptolite collection found in the present area, south of the East Poultney bridge across the Poultney River, is from an outcrop occurring between the red Indian River Slate and greywacke beds of the Pawlet, in a rock like the basal Pawlet elsewhere, which so far has proven to be the most fossiliferous stratum of the formation (Shumaker, 1959, p. 59).

The writer therefore interprets the two areas of Indian River Slate (and Poultney Slate?) just west of the trace of the Bird Mountain fault as simply overturned synclines, which are truncated to the west by the base of the Pawlet Greywacke and to the east by the Bird Mountain fault. While it is conceivable that these areas are part of the Bird Mountain slice and lie east of the Bird Mountain fault, occupying centers of anticlines in the inverted sequence, this seems an unnecessarily complex explanation.

The alterative explanations are diagrammatically represented in Figure 5.

Structure of the Bird Mountain Slice

Evidence for a Fault: A major thrust fault has been proposed (Zen, 1961, p. 321) to separate, as a distinct slice, part of the Taconic sequence south of the Castleton River. This area was called the Bird Mountain slice. The slice is bounded to the north, approximately by Belgo Road and North Brittain Brook (both north of the Castleton River), and to the west by a line running roughly from Gorhamtown to East Poultney. No breccia or mylonite marks this fault, and the arguments for its existence are indirect. These are now briefly summarized.

1. Truncation of structures. North of the Castleton River, structures belonging to the Giddings Brook slice (Zen, 1961, p. 314), which are geometrically continuous with the area west of the Gorhamtown-East Poultney line (Zen, 1961, p. 320), are truncated by structures belonging to the Bird Mountain slice. The main belt of the Biddie Knob Formation of the Taconic Range disappears abruptly to the south; similarly the east-west striking belts of the Bull Formation and the West Castleton Formation, southeast of Hooker Hill, disappear to the east, under the hills which are part of the Bird Mountain slice. South of the Castleton River, both the structures of the Giddings Brook slice and the trace of the Bird Mountain fault trend nearly north-south; therefore no clear

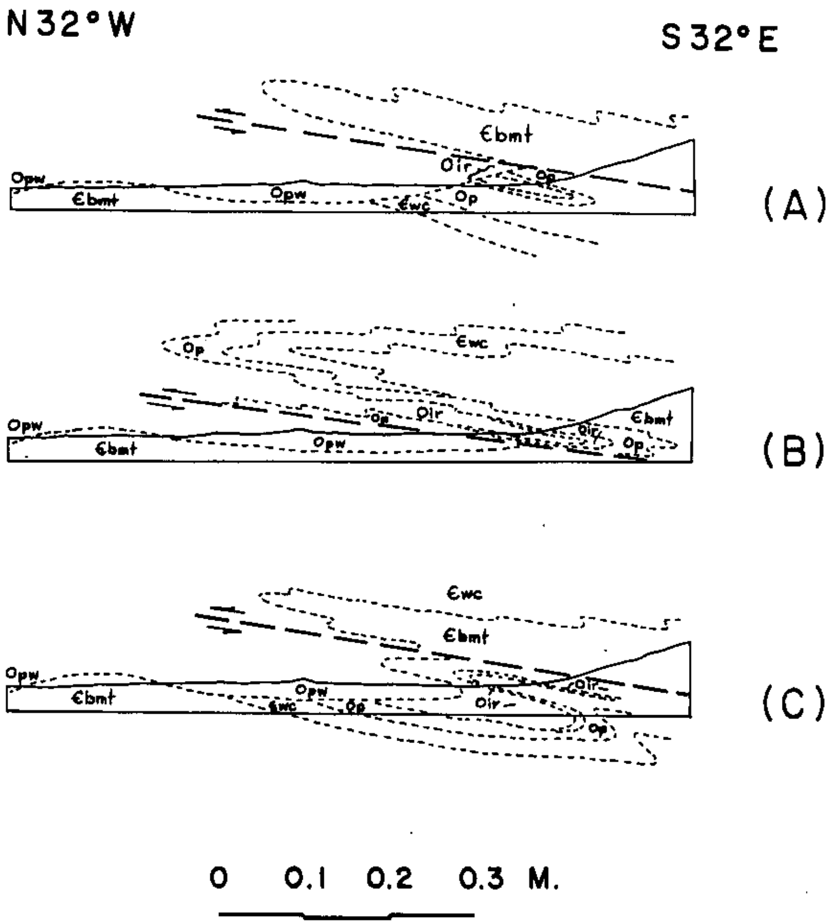

Figure 5. Three possible interpretations of the structure near East Poultney. (a) The Indian River Slate is in a syncline below the Bird Mountain thrust. This is the favoured interpretation. (b) The Indian River Slate is in a syncline under the inverted limb of the major recumbent fold of the Bird Mountain slice; the Bird Mountain thrust is below the Indian River Slate syncline. (c) The Indian River Slate is part of an anticline instead of a syncline and the fold is under the Bird Mountain thrust.

evidence of structural truncation exists. The juxtaposition of various units along this line with older units generally overlying the younger, however, is best interpreted as due to a low-angle thrust fault.

2. Detailed stratigraphic differences. Although rocks on the two sides of the proposed fault are broadly similar, in detail many differences exist.

The Pawlet Formation is totally absent and only a single area of Indian River and Poultney slates exists east of this fault, although these are important units immediately west thereof. The near-absence of the West Castleton Formation, and the apparent thickening of the slates of the Bull Formation east of the fault, discussed in an earlier section, are also noteworthy. Finally, east of the fault the slates of the Bull Formation tend to be silty and massive, instead of soft, fine grained and finely cleaved, as is generally true west of the fault. As a result, even though the metamorphic grade is not noticeably higher, east of the fault not a single slate quarry exists.

3. Differences in internal structure. As will be seen later, the structure within the Bird Mountain slice is of a different style from that of the west, and involves large-scale recumbent folds. The area to the west is marked by linear belts of relatively simple anticlines and synclines. Although these latter belts are continuous on the ground with the Giddings Brook fold complex, these two groups of recumbent folds are not geometrically continuous; in the present map area, west of the Bird Mountain fault only the upper, normal limb of the Giddings Brook recumbent fold is exposed.

4. The structure in the Vail Brook-North Brook area is anomalous, as will be considered later. The relations are simply explained, however, by assuming the validity of the Bird Mountain slice.

Structure Within the Bird Mountain Slice: The structure within the Bird Mountain slice is suggested principally by the outcrop pattern of the Biddie Knob Formation, the oldest unit of the Taconic sequence. The map shows that it occupies an extensive area in the center of the Bird Mountain slice. Although local, small-scale folds are generally abundant (as may be seen, for example, along Gully Brook with its nearly 2 miles of continuous exposure), on the whole the formation must be nearly flat-lying. Thus, near the center of the area of the Biddie Knob Formation, a broad syncline defined by the Zion Hill Quartzite (Bird Mountain Grit) occurs. The three-dimensional control afforded by the cliffs of Bird Mountain leaves no doubt of the synclinal structure here, a fact confirmed by primary sedimentary tops indications, such as graded bedding, in the quartzite exposed on the cliffs.

The relations at Bird Mountain indicate that the Biddie Knob Formation surrounding the mountain is in normal succession. This is supported by the relation of younger units to the west. Thus, the belt of the Bull Formation west of the Biddie Knob overlies the latter, and their relations suggest a slight westerly regional dip to the formation contact. The various patches of black West Castleton Formation are local synclines; the two areas of the Bomoseen Greywacke are regarded as local lenses in the Bull Formation.

This picture of superposition, however, is deceptively simple. It was already noted that along the Bird Mountain fault, rocks in the Bird Mountain slice are inverted and commonly it is the West Castleton Formation, rather than the Biddie Knob or even the Bull Formation, which is in contact with units of the lower tectonic plate. Moreover east of the extensive area of the Biddie Knob Formation, a persistent belt of green phyllite, free of chloritoid and representing the higher-metamorphic grade equivalent of the slates of the Bull Formation, intervenes between the Biddie Knob Formation to its west and the black phyllite to its east. Where the relationships can be ascertained, the black phyllite dips gently west under the green phyllite; the contact is commonly gradational (Zen, 1961, p. 313). The green phyllite locally contains quartzite beds like typical Zion Hill and the Mudd Pond (both members of the Bull Formation), for example, about 1 mile southeast of Clark Hill, where the Mudd Pond occurs to the east of, and underlies, the Zion Hill. In addition, between the 1850′ knob south of Clark Hill and Moss Hollow five miles to the southwest, green phyllite near the flat contact with the underlying black phyllite commonly includes the North Brittain Conglomerate Member. Although no fossil has been found so far, the conglomerate is so characteristic that there can be no doubt as to its identity. On the basis of stratigraphic succession, *the belt of the Bull Formation east of the Biddie Knob Formation represents an inverted sequence.* Thus, even though in the center of the Bird Mountain slice the section is unambiguously right-side up, near the margin of the Bird Mountain slice the succession seems always to be upside-down.

The overturning of sections near the margin of the Bird Mountain slice might, of course, be caused by local drags during the thrust movement. However, if this is the sole explanation, a radially outward movement sense would be necessary; the Bird Mountain slice would in fact be a large mushroom fold. This contradicts the regional structural relations of the Taconic sequence as a whole; specifically, arguments have been advanced against such an interpretation for the underlying Giddings Brook nappe (Zen, 1961, p. 325), and therefore even less probable for the Bird Mountain slice. In addition, a geometric restriction to this picture is afforded by the Bird Mountain syncline which must occupy the "cap" of this mushroom. Because of the small topographic relief in this area as a whole, the reader can see the improbability of this interpretation by constructing cross-sections.

Therefore, even though one could explain the inversion at the west margin of the Bird Mountain slice (assuming an east-over-west sense of movement) as due to local drags along the fault, the similar inversion between Clark Hill and Moss Hollow remains inexplicable. On the other hand, the conflicting evidence can be reconciled by supposing that the Bird Mountain slice represents a large recumbent, isoclinal, bottoming fold. The larger areas of the Zion Hill Quartzite, at Bird Mountain and at the 1962′ knob to the southwest, represent broad synclines on the upper, or right-side up limb of this nappe. The areas of the Zion Hill Quartzite at the 2089′ knob, northeast of Herrick Mountain, may belong to the same structure, although in this last area the outcrops of quartzite are too scattered and broken up for unequivocal interpretation. The intersection of the axial plane of the recumbent fold with topography forms a closed loop; it separates the normal sequence within the loop from the inverted sequence outside of it. The course of this loop is shown on the geologic map. On the east side of the Bird Mountain slice, the trace of this loop lies within the Biddie Knob Formation. It runs generally southward, between Clark Hill and Hampshire Hollow Road, where it climbs into higher strata. The precise course of this trace from here on becomes uncertain; it is however probably coincidental with, or slightly offset to the east of, the trace of the west margin of the Bird Mountain fault as far north as Gorhamtown. Here it must swing northeast to re-enter the Biddie Knob Formation at Pond Hill, continue northeast to the Castleton River, and finally bend to the east to close the loop,

Under this interpretation, the anticlinal axis in the Biddie Knob Formation, east of Pond Hill, represents a secondary structure on the upper limb of the recumbent fold; it is the structural complement to the Bird Mountain syncline. The narrow, east-west trending belt of the Biddie Knob Formation north of the Castleton River in the Bird Mountain slice may represent a digitation on the main bottoming fold. Although the alluvium of the Castleton River intervenes, the continuation of this belt of the Biddie Knob Formation with the main tract is indicated by the outcrops of the Biddie Knob Formation at around the 1000′ level south of the 2135′ knob, north of the Castleton River. This interpretation requires that the Zion Hill Quartzite south of the 1405′ knob to the south should be right-side up, which is verified by observed graded bedding.

The discontinuous but up-side down belt of the West Castleton Formation along the western margin of the Bird Mountain slice is the carapace of the lower limb of the bottoming fold. The horizon of the black slate projects underground into the black phyllite immediately southeast of the high ridges of the Taconic Range between Herrick Mountain and

Spruce Peak, where the section is also upside-down, although the physical continuity of the two areas of black slates is not required.

The isolated area of green phyllite with its North Brittain Conglomerate, is interpreted in an identical manner. The green phyllite here is as an erosional outlier of the inverted Bull Formation. The large, crudely horse-shoe shaped area of green phyllite and slate between Ira Brook and North Brook, with its single outcrop of the North Brittain Conglomerate, is interpreted in an identical manner. The green phyllite here is typical of the Mettawee, and the low ground of Train Brook is underlain by black phyllite. The contact, actually exposed, is nearly horizontal. This area of green phyllite is nearly traced into the main belt to the northwest, at a point south-southeast of Herrick Mountain; at the east and west ends of a pasture near the trail due north of the 1904′ knob, black slate occurs in the lowland, but north and south of the pasture, on the wooded slopes, green slate occurs. Although these outcrops are separated by about 500 feet of alluvial cover, it is safe to infer the former continuity of the green phyllite.[1]

The structure of the Bird Mountain slice, as interpreted, strongly resembles that of the Giddings Brook fold complex to the north (Zen, 1961, p. 314), which similarly is a recumbent bottoming fold, with westerly closure. The two structures however are distinct; the Biddie Knob Formation occupying the cores of the two folds are separated north of the Castleton River by the lower limb of the bottoming fold of the Bird Mountain slice. This lower limb, here consisting of the Bull Formation, occupies the east-west ridge between Hooker Hill and the 2135′ summit northwest of West Rutland.

Assuming that the rocks were derived from the east and arrived at the present site through a westerly sense of movement (Zen, 1961, p. 330), a palinspastic development of the Bird Mountain slice would indicate an eastward thickening of the Bull Formation to the east (now part of the upper limb), and a diminished importance of the West Castleton Formation in the same direction. Such a picture would accord with the greater importance of the West Castleton Formation in the Giddings Brook slice, if we assume that slices now occurring relatively to the west and tectonically at lower levels were derived from a more westerly portion

[1] Thompson (in Doll et al., 1961) suggests that the green phyllite southeast of the black phyllite belt belongs to a separate, lower thrust slice. The geometry of this explanation however is implausible, as a cross-section across the contacts would readily show.

of the original basin of deposition. The scheme is represented in Figure 4, simplified so as to include only the Lower Cambrian units.

Structure in the Vail Brook-North Brook Area

The structure in the small area around Vail Brook and North Brook, between Spaulding Hill and Middletown Springs, is here separately discussed because of its structural distinctness.

The greater part of the area is underlain by a green or greyish green slate and phyllite devoid of diagnostic beds. The North Brittain Limestone Conglomerate, abundant in the adjacent green phyllite of the Bird Mountain slice, is lacking. The slate and phyllite generally closely resemble the Mettawee, and some purple slate exists on the south side of Spaulding Hill; however, elsewhere, as for instance east of the 1750′ knob south of Spaulding Hill, the green slate is more like the Poultney Slate.

On the slopes of the sharp 1710′ knob northeast of the 1750′ knob just mentioned, the green slate is overlain by black slate characteristic of the Poultney. Rocks on top of the knob are hard, silty, poorly-cleaving, white-weathering, "cherty" green and grey-green slate which belongs either to the Poultney or to the Indian River. A right-side up section is thus indicated, in contrast with the structure in the surrounding Bird Mountain slice.

The Spaulding Hill belt of green slate shows variable contact relations with the surrounding black slate. At the northwest side of Spaulding Hill, green slate rests above the black slate, which outcrops in stream beds. Near the 1710′ knob, green slate seems to underlie black slate to the west. On the 1610′ knob to the south, green slate both overlies and underlies the black. On the east side of this Spaulding Hill belt, the black slate is above the green. A good place to see the relation is on the tiny 1480′ knob half a mile southeast of Spaulding Hill, immediately west of the road along Vail Brook.

The right-side up section in this area, involving units as young as the Middle Ordovician, when compared with the North Brittain Conglomerate-bearing, upside-down Bull Formation of the surrounding area, shows that the two areas cannot be directly connected. The nature of the rock units in the Vail Brook-North Brook area, their succession, and their right-side up stratigraphic sense, relate these rocks to those of the Giddings Brook nappe, exposed north and west of the Bird Mountain fault. It is now suggested that the Vail Brook-North Brook area is an erosional window in the Bird Mountain slice, exposing rocks of the underlying Giddings Brook nappe.

This hypothesis explains the age and stratigraphic sense of rocks in the Vail Brook-North Brook area. The variable contact relation between the green and black slates, west of the Spaulding Hill belt, is explained as a folded fault with east-over-west sense of movement, where rocks of both tectonic units deformed together. Shumaker, mapping in the adjacent area to the south, reaches a similar conclusion on the structural relation of the Spaulding Hill belt of rocks. He calls the structure of this belt the Edgerton Window (Cornell University thesis, 1960, p. 61); the window frame however is breached at the south end by erosion.

Fowler (1950) shows the entire North Brook-Vail Brook area as underlain by "Normanskill" slate, and interprets the structure as a simple syncline. The present study partly agrees with Fowler's age assignment. The formation contact however is more complex than Fowler shows and the structural explanation is accordingly different.

Structure of the Green Phyllite Area East of Ira Brook

The 2405′ peak east of Ira Brook, one of the highest in the Castleton quadrangle, was mapped by Fowler (1950) as part of the terrane of the Hortonville Slate. In fact, however, it is underlain by an extensive area of green phyllite. As this area is isolated from the main Taconic belt to the west by an intervening area of black phyllite, the structural relationship of the green phyllite cannot be established by direct tracing.

The green phyllite capping the peak overlies black phyllite which can be traced into the valley of Ira Brook. The green phyllite ranges from soft, lustrous, and micaceous to silty with interlaminated white coarse silt layers up to half an inch thick at one inch intervals; these are rocks assignable to either the Bull Formation or the Poultney Slate. The area is interpreted as an erosional outlier of the Taconic sequence resting on black, largely autochthonous slate of the Ira Formation.

It has been impossible to ascertain whether part of the black phyllite on the west side of the 2405′ peak is allochthonous. Southeast of the peak, however, black phyllite occurs between the 1900′ and 2100′ levels and contains dolomitic quartzite typical of the Hatch Hill Formation, first discovered in a joint traverse by J. B. Thompson, R. C. Shumaker, and the writer in 1958. If the Hatch Hill classification is correct, then an inverted sequence of the allochthonous Taconic sequence must be present.

Detailed relations on the slope east and southeast of the 2405′ peak are complex. Due east of the peak, black phyllite, traceable into the outcrops carrying Hatch Hill type of quartzite, rests directly on the Columbian Marble of the Shelburne Formation. On the same slope, to the south, however, several alternating belts of green-grey, green, and black phyllite intervene between marble outcrops and the quartzite-bearing black phyllite. Depending on location, either green or black phyllite rests on the marble. The phyllites are non-diagnostic of formations.

At least three possible explanations exist for the stated relations:

(1) The green and black phyllites to the east of, and underlying the black phyllite with Hatch Hill lithology, are autochthonous and rest unconformably above the Ira Formation. The green phyllite would then be comparable to the interpreted position of the Mount Anthony Formation of MacFadyen (1956). This idea is countered by the fact that locally green, rather than black, phyllite rests directly on the marble; so that if the green phyllite represents an autochthonous unit, a rapid lateral facies change or multiple unconformity must be assumed.

(2) The belts of green and black phyllite represent part of the inverted allochthonous Taconic sequence, stratigraphically above the Hatch Hill. Although on the map these belts are shown as truncated by the Hatch Hill just southeast of the 2405′ peak, the field relations are open to alternative interpretations. The green and black phyllite then would correspond to units in the Poultney and perhaps also the Indian River Slates; lack of diagnostic features of these units would be explained by metamorphic effects. Although this hypothesis is attractive owing to its simplicity and accordance with the structural picture in areas to the south (Shumaker, 1959, p. 59), it too is without positive evidence. Moreover, the quartzite beds and banding typical of the Poultney is not observed here; these features should not be easily obscured by metamorphism as they represent strong compositional contrasts.

(3) The units between the marble and the Hatch Hill Formation may represent a zone of imbrication or a sliver of a separate tectonic unit, perhaps even a remnant of the Giddings Brook nappe. This unit is truncated above by the Hatch Hill Formation which presumably belongs, along with the Mettawee Slate, to the Bird Mountain slice. If so, it must exist as a very narrow strip here, for on the west side of the 2405′ peak there is no trace of it.

Evidence at hand is inadequate to choose between the two latter possibilities. From the fact that the typical Poultney lithology is missing in the phyllite between the "Hatch Hill" Formation and the marble, the third interpretation is here tentatively adopted. More detailed work in the area may resolve the problem.

Synclinorium Sequence

Pine Hill Reverse Fault

A major fault that extends from north of Pine Hill, west of Rutland, to the northeast corner of the Pawlet quadrangle, was early recognized and named the Pine Hill thrust fault (Wolff, 1891; Dale, 1892, 1894; Bain, 1933; Brace, 1953; Thompson, 1959). According to Brace (1953, Pl. I), the fault brings Precambrian rocks into juxtaposition with Middle Ordovician "Hortonville" Slate, now called the Ira Formation (Thompson, 1959). Although Wolff (1891, p. 337) first suggested that the fault is a thrust, the average dip of the fault surface is unknown. Brace (1953, Pl. 2) shows the dip commonly near 45°, and Thompson (1959, Pl. H-2) agrees. The only fault surface exposure known to the writer, on the summit of Boardman Hill, dips 70° ESE. Because of this uncertainty, the feature is here called the Pine Hill reverse fault.

The eastern, or hanging-wall side of the fault consists of gneisses, schists, and lime silicate rocks of the Precambrian Mount Holly Series, as well as the Lower Cambrian Cheshire Quartzite. At Pine Hill, Brace (1953) shows units of the Lower Cambrian (?) Mendon Series to occur between the Precambrian gneisses and Lower Cambrian Cheshire Quartzite. Similarly, at the 1130′ level on the north slope of Boardman Hill, at the north edge of a large clearing, limited outcrops of a green schist containing blue quartz pebbles up to 5 mm across occur between the Cheshire Quartzite and a coarse marble; this pebbly schist may correspond to the Pinnacle Grit of the Mendon Series (Nickwacket of Brace, 1953; see, however, Osberg, 1959, p. 47, footnote).

A problem in mapping the Pine Hill fault has been the similarity of the Moosalamoo Phyllite, of the Mendon Series, to the Ira Formation (see Dale, 1892, p. 518; Brace, 1953, p. 78). On the present map, most of the black phyllite immediately west of the Precambrian gneiss and/or Cheshire Quartzite is called the Ira and placed on the foot-wall side of the fault. These rocks are soft and micaceous, different from the scattered outcrops of silty, hard black phyllite associated with the coarse Precambrian marble east of the fault (not separately shown on the map because of the very limited areal extent). The reference of the black phyllite immediately west of the fault to the Ira Formation is corroborated by the occurrence of fossiliferous Middle Ordovician limestone lenses

(Whipple Member) in the southern continuation of the same belt of rocks (Dale, 1892, p. 517; Wolff, 1891, p. 336).

PITTSFORD-CHIPPENHOOK ANTICLINE

Dale (1912, Pl. I) shows a major anticline that runs from Coxe Mountain, north of Pittsford, through Clarendon Springs and Chippenhook to Tinmouth and beyond. The anticline cores in "Beekmantown (?) and Lower Cambrian" dolostone (1912, p. 66) which was not differentiated on his map. Dale did, however, show an axial depression on this anticline, so that along the Clarendon River east of West Rutland, Ordovician marbles span the fold. The same general structure has been adopted by Bain (1933, 1959) and by Fowler (1950). Bain (1959, p. 38) implies that the Intermediate Dolostone (of the Shelburne Formation) spans the axial depression along U. S. Route 4 between Center Rutland and West Rutland, in agreement with Fowler's designation of these units as the "Boardman Formation." In general, except for the greater details shown, Fowler's map pattern follows Dale faithfully; like Dale, Fowler interprets Coxe Mountain as the core of the fold at its north end. Bain (1933, Pl. 13; 1959, Pl. D-1) disagrees with this interpretation; he shows Coxe Mountain as a thrust plate above the structure of the anticline of the carbonate units. Bain's interpretation of this area is inconsistent with the recent detailed work of Osberg (1959), which reveals unexpected stratigraphic as well as structural complexities.

Fowler (1950, p. 61) uses the name Pittsford anticline and Chippenhook anticline, respectively, for the folds north and south of the axial depression. The writer suggests the single term, the Pittsford-Chippenhook anticline, to emphasize the continuity of the fold. The structural picture differs from Fowler's only in detail.

Fowler shows the Whipple Marble to occur tenuously but continuously on both flanks of the Chippenhook anticline though this unit is shown as absent in the Pittsford anticline. The writer maps most of the more westerly branch of this Whipple Marble as the upper beds of the Columbian Marble; the eastern belt cannot be recognized except for isolated outcrops of a dark grey marble basal to the Ira Formation, northwest and southwest of Boardman Hill.

About half a mile northwest of Boardman Hill, the Sutherland Falls Member of the Shelburne Formation nearly bridges across the anticline, in agreement with Bain's mapping (1938). North of here, the Sutherland Falls Marble diverges again; the western branch trending northwest from this place and describing a sinistral map pattern, only to disappear under the Ira Formation to the west, about one mile southeast of West Rutland. This western belt of marble is underlain to the east by the Clarendon Springs Dolostone. These relations show that an axial depression must exist at the place where the Sutherland Falls Marble nearly spans the anticline. This structure has not been previously recognized.

The dolostone on Route 4 between Center Rutland and West Rutland, called the Intermediate Dolomite by Bain (1959, p. 38), contains quartzite beds up to 2′ thick. The writer has interpreted this outcrop as part of the Upper Cambrian Danby Formation (1959, p. 7). This interpretation accords with the location of an axial depression northwest of Boardman Hill.

Within the map area, the oldest rock unit exposed in the Pittsford-Chippenhook anticline, both north and south of the axial depression, is the Lower Cambrian (?)-Middle Cambrian (?) Winooski Dolostone. North of the axial depression, the structure is by and large simple, with the west limb of the anticline dipping west and passing northward into the east limb of the Middlebury synclinorium (Cady, 1945). Locally, however, easterly, inverted dips exist, and are demonstrable by cross-bedding in the Danby and Clarendon Springs formations. Complicated structures locally occur, for instance near Meade Bridge southwest of Pittsford, where a large drag fold is shown on the map; similarly, west of Florence, the writer has mapped a major recumbent fold, the Florence nappe (Zen, 1961, p. 322), within the carbonate sequence. Much of the Pittsford-Chippenhook anticline is masked by the Middle Ordovician Ira Formation, which rests unconformably upon rocks which, within the map area, range in age from the Upper Cambrian Danby Formation (Route 4 near West Rutland) to the Lower Ordovician Chipman Formation (West Rutland); in the Pawlet quadrangle to the south however it overlies Precambrian gneisses directly (Thompson, 1959, Pl. H-1, H-2). The development of the Pittsford-Chippenhook anticline is thus clearly at least in part pre-Ira in age. This pre-Ira deformation is also recognized by Thompson (1959, p. 72) in the east part of the Pawlet quadrangle. The relative ages of these deformations and of the emplacement of the Taconic allochthone will be considered later.

West Rutland Marble Belt

A long, narrow belt of Ordovician marble and dolostone extends from the vicinity of West Rutland northward into Whipple Hollow, for a distance of almost six miles. This marble belt, in which much of the recent quarry operation in the West Rutland area occurs, is surrounded by black phyllites of the Ira Formation. Keith (1932, p. 398–399; 1933, Pl. 8) interprets the structure of this marble belt as synclinal, the marble, which he calls the West Rutland, is considered to be Chazyan and the crinoid-bearing strata, here assigned to the Whipple, are taken as the basal beds of this formation (1932, p. 399).

Wing (Dana, 1877, p. 339), Bain (1933, Pl. 13; 1959), Dale (1912, p. 70), and Fowler (1950, p. 62), on the other hand, interpret this marble belt as an anticline, and the surrounding black slate is taken to overlie the carbonates. As Dale gives no structural or stratigraphic detail within the belt of marble, and Fowler's map pattern is very similar to Bain's (1959), these authors' ideas may be considered together.

Bain and Fowler regard the anticline as a rather simple structure, with both of its limbs preserved. The oldest bed of this anticline is the Burchards Formation according to Fowler (1950, Pl. 1, = Burchards Member of the Chipman Formation, Cady and Zen, 1960), and is the "West Blue marble" of Bain (1938, p. 7, map; also 1959, p. 36, Pl. D-1) which Fowler (1950, p. 29) correlates with the Burchards but which may in part correspond to the Bascom Formation of the present work (see Editor's note, Bain, 1959, p. 36). The extensive development of west-up drag folds, in the quarries at West Rutland, on the supposed east limb of the anticline, contradicts this structural interpretation; these folds were explained by Bain (1931, p. 515, 518; 1959, p. 38) and by Fowler (1950, p. 71) as flowage folds with a movement sense opposite that of normal drag folds.

The writer now suggests a different interpretation of the structure of the West Rutland marble belt. Briefly, the area is taken to be an overturned, essentially homoclinal sequence of Lower Ordovician Bascom and Chipman formations. This structure is masked in part by the unconformity at the base of the Ira Formation, and in part by later folding of the units as a whole, which produced an anticline in the Middle Ordovician units. The reasons for this interpretation follow.

Although, due to the swamps of Whipple Hollow, no continuous stratigraphic section exists across the West Rutland marble belt, a satisfactory composite section can be obtained. Southwest of West Rutland near the southern terminus of the belt of marble, the Beldens Member of the Chipman Formation, with its characteristic dolostone beds, occurs in direct contact with the Whipple Marble to its west. Drag folds, in massive dolostone beds inter-bedded with the incompetent limestone, indicate a west-up movement sense. The same Beldens Limestone and dolostone can be traced nearly continuously northward to the center of West Rutland village, and is there found on the north side of the athletic field behind the West Rutland High School. About 1000′ to the east, it is again found in the shaft quarry of the Green Mountain Marble Company, in the center of the village, and from here northward to the main quarries of the Vermont Marble Company. Drag folds in the dolostone beds show west-up sense persistently, but these folds now presumably become "flowage folds" of Bain and Fowler because they occur on the east limb of their anticline, even though the folds are in nowise different from those west of this supposed anticlinal axis.

Above the line of quarries of the Vermont Marble Company north of West Rutland village, the Beldens Member is overlain to the east by interbedded white-grey marble and brown-weathering sandy marble, at 1–2 inch intervals; these beds are typical of the Weybridge Member of the Chipman Formation. Farther east, on top of the hill, this rock is followed by coarse, grey-streaked, sugary-weathering marble typical of the upper beds of the Bascom Formation as found near Florence (Zen, 1961, p. 308). This rock can be followed north to the abandoned quarries about half a mile north-northeast of the *site* of the Whipple Hollow School, where it is seen to be overlain by dark grey, finer grained Whipple Marble. The drag sense in the older marble is persistently west-up.

The pre-Whipple rocks, thus, constitute a sequence from the upper Bascom beds through the Weybridge (which commonly occurs near the base of the Chipman Formation; see Cady and Zen, 1960, p. 737) into the Beldens. This agrees with the evidence of the drag folds and shows the West Rutland marble belt to be a homoclinal sequence. Whereas "flowage" folds undoubtedly occur in the Vermont Valley marble belt (see, for example, Thompson, 1959, p. 81; Brace, 1953, p. 96), they are not dominant in the West Rutland area. It is also interesting to note that, at the West Rutland quarries, the movement sense of massive, competent dolostone beds agrees consistently with that of the relatively incompetent marble beds.

It will be recalled that, northwest of Boardman Hill, the west (overturned) limb of the south-plunging anticline, defined by the outcrops of the Sutherland Falls Marble north of the axial depression, has a sinistral drag sense on the map. This limb contains the Columbian Marble as its youngest pre-Whipple rock. Projection of this structure northwestward brings the Columbian Marble east of and near the east margin of the West Rutland marble belt, whose oldest unit is the next

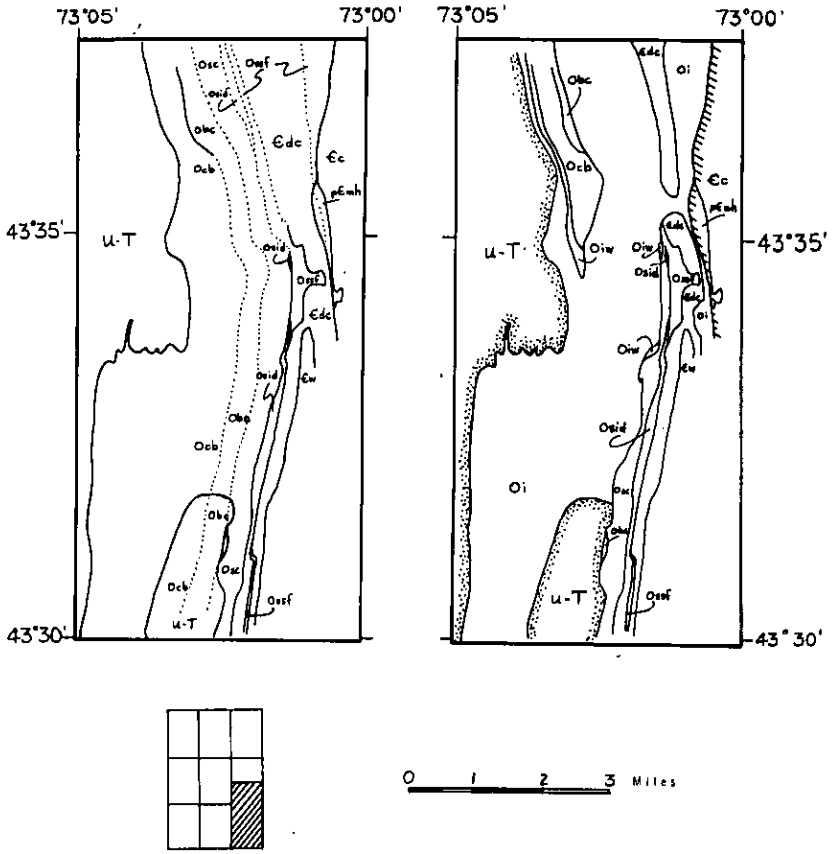

Figure 6. A re-interpretation of the West Rutland marble belt. The right-hand figure shows the field geology as observed and the left-hand figure shows the interpreted connections with the masking Ira Formation peeled off. Inset shows location of the area relative to the 15-minute Castleton quadrangle.

higher Bascom Formation. In both structures, the stratigraphic sense is tops west.

The writer now suggests that the West Rutland marble belt is the re-emerged west (and overturned) limb of the same south-plunging anticline northwest of Boardman Hill (Figure 6). The anticline was clearly formed in pre-Whipple time as the Whipple truncates the older and underlying units. The degree of overturn of the older structure however has been accentuated by post-Whipple movements. After a considerable thickness of the Ira Formation had deposited on the older structure, a second deformation, also with westward movement sense, deformed the Ira Formation and the underlying carbonate units together. Rotation of the flanking younger strata, during this deformation, resulted in an anticline in these beds, such that on the west side of this structure the Beldens-Whipple contact is now pseudoconformable, but on the east limb this same contact now dips gently east. The second deformation failed to obliterate the earlier drag folds.

As the second deformation is post-Ira at least in part, it is tempting to relate the event to the formation of the Florence nappe at the north end of Whipple Hollow (Zen, 1961, p. 322). The Florence nappe, a recumbent bottoming fold, is probably contemporaneous with or later than the emplacement of the Taconic allochthone in that area (Zen, 1959a, p. 4), and involves the same stratigraphic units as does the West Rutland anticline (Zen, 1961, p. 308). These deformations might be chronologically related to the formation of the Pine Hill fault.

Relation Among the Tectonic Units

Relation Within the Taconic Sequence

In the preceding sections, the writer tried to show that the area here called the Bird Mountain slice is structurally distinct from the areas to the north and to the west, called the Giddings Brook slice (Zen, 1961, p. 314). Structural elements of this latter unit disappear under the Bird Mountain slice. This discordance is easy to map because rocks of the two sides of the contact show differences in resistance to erosion, commonly resulting in strong topographic expression of the junction; the Bird Mountain slice occupies relatively higher ground.

This relationship, as such, could mean either a fault or an unconformity. No fossil has been found in rocks of the Bird Mountain slice. However, the various key units of the Lower Cambrian sequence, such as the North Brittain Conglomerate, the Mudd Pond Quartzite, and the Zion Hill Quartzite, have all been found in their proper stratigraphic positions in the Bird Mountain slice. This parallelism of units in two complex rock sequences indicate their equivalence; the Bird Mountain sequence must be largely Lower Cambrian. As this slice rests upon the Giddings Brook slice, which includes rocks as young as the Middle Ordovician, a fault contact is indicated. Relation of the contact with topography, shown for instance along the ravine southeast of Belgo Road, shows that the fault must dip at a low angle into the upper plate.

Rocks of the area around Vail Brook and North Brook closely resemble those of the Giddings Brook slice. The sequence is right-side up and the highest units are the Poultney Slate, in contrast to the surrounding inverted Lower Cambrian units. The right-side up rocks also occupy lowland, which continues southward as the valley of South Brook in the Pawlet quadrangle, where Shumaker (Cornell University thesis, 1960) has found, within the same sequence, rocks as young as the Pawlet Formation, so far known only from the Giddings Brook slice. This area is then a tectonic window of the Giddings Brook slice, called the Edgerton Window by Shumaker (Cornell University thesis, 1960, p. 61). The existence of a window corroborates the conclusion that the contact between the two slices is a thrust of low angle.

At the extreme northeast corner of the tectonic window, southeast of Spruce Knob, the contact between the two slices is drawn in rather arbitrarily. This is because of the similarity of black phyllites belonging to the two slices. Differentiation of black slates and phyllites of different ages and tectonic units, a difficult task in general (Zen, 1961, p. 334), is particularly uncertain as the increasing metamorphism tends to further obscure slight differences in initial lithologic characteristics. The contact, on the map, is drawn close to topographic contours, reflecting the conclusion that the sole of the Bird Mountain thrust is nearly flat-lying.

Relation Between Taconic and Synclinorium Sequences

If the internal stratigraphy of the Taconic sequence is as described, then on the east flank of the Taconic Range, within the map area, Lower Cambrian rocks of the Taconic sequence rest, in inverted order, on the Middle Ordovician Ira Formation, which is right-side up. The contact between these rocks therefore cannot be a normal one. Detailed considerations of identical geometric relations in areas to the north have led (Zen, 1961, p. 325) to the further conclusion that the observed relations are probably not due to mushroom-folding of autochthonous Taconic rocks; rather the Taconic rocks must have been thrust into their present location.

The conclusion was deduced for the Giddings Brook slice. As this slice underlies the Bird Mountain slice, with a thrust contact, this latter slice, *a fortiori*, cannot be autochthonous. The Taconic rocks probably moved in from the east, perhaps from the site of the present Green Mountain anticlinorium (Zen, 1961, p. 331).

From available evidence the writer suggested (Zen, 1961, p. 313) that the Taconic sequence, perhaps as soft, poorly-indurated material, slid westward into a Trenton sea in which black mud (Ira Formation and Hortonville Slate) was depositing. The intimate mixing of the two sequences was suggested to explain the absence, in the field, of a mappable "Taconic thrust"; rather there commonly appears to be a gradational contact between undoubted autochthonous black Trenton slate and undoubted allochthonous green Lower Cambrian units.

The reason for suggesting that the Taconic thrust is a group of submarine slides, and not a more conventional thrust fault, is largely because of the nature of the Forbes Hill Conglomerate, previously described (Zen, 1961, p. 311). This conglomerate is invariably found in the autochthonous Hortonville or Ira slate close to the contact with the allochthonous Taconic sequence. The lithologies of the boulders and pebbles can all be matched with the Taconic rocks. The matrix, locally graptoliferous, is an integral part of the autochthonous slate. The matrix does not show deformation; locally the stratification continues up to the boulders and stops abruptly. The matrix material is never found as pebbles. The pebbles, where disc-shaped, are aligned parallel to the bedding. Finally, the conglomerate, which is a discontinuous unit, is not found directly in contact with rocks of the Taconic sequence, but is everywhere some distance (perhaps 50 feet stratigraphically) below the contact; the rocks above the conglomerate zone are identical with those below and of the matrix. Taken together, the evidence indicates a submarine, sedimentary slump conglomerate, and not a simple fault breccia. The intimate lithologic, sedimentary, and geometric association of the conglomerate with the Taconic sequence then suggests the mode of emplacement of the Taconic allochthone itself.

Because of the uncertainty on the location of the Taconic thrust on the map, the dislocation is drawn somewhat arbitrarily. It is entirely possible, for instance, that at least part of the extensive area of black phyllite along Ira Brook, shown as the Ira Formation, belongs to the Taconic sequence.

The tectonic history of the Taconic sequence, previously presented (Zen, 1961, p. 324), suggests however that the Taconic sequence should have undergone at least two periods of deformation. The earlier one of these, Trentonian in age, occurred during the emplacement of the thrust slices and the style of deformation was recumbent, isoclinal folding as well as faulting. Little or no metamorphism accompanied this deformation (Zen, 1960a). The second deformation took place during the formation of the Middlebury synclinorium, and involved more open folding but more intense metamorphism. It is possible that the Ira Formation has been deformed only during the latter diastrophism. Petrofabric study of minor structures, thus, may yet reveal a criterion for differentiation of the black argillaceous units; this problem invites close study.

Summary of Chronology of Diastrophic Events

This report may be aptly closed with a summary of our present knowledge on the relative ages of the main tectonic events in the Taconic allochthone and its environs. The principal features that must be considered are these: (1) the unconformity at the base of the Ira Formation; (2) the Taconic thrust; (3) the Pittsford-Chippenhook anticline; (4) the West Rutland anticline; (5) the Florence nappe; (6) the Pine Hill reverse fault; and (7) the normal faults on the east limb of the Middlebury synclinorium. These will first be considered individually and then compared.

The pre-Ira unconformity has been dated by the corals in the Whipple as Lowville, Chaumont, or younger. It is therefore possibly as old as early Black River because the fossils are not from the base of the rock unit. The upper age limit of the Ira Formation is unknown, but may be supposed to be at least as young as the mid-Trenton Sherman Fall by comparison with similar beds near Johnson Pond in Orwell.

The age of the Taconic thrust is post-Ira on the basis of the Forbes Hill Conglomerate, discussed above, and no older than the post-Glens Falls Snake Hill (Hortonville). The thrusting is a mid- to late Trenton event.

The Pittsford-Chippenhook anticline is part of the main structure of the Middlebury synclinorium. Near Brandon and Sudbury (Cady and Zen, 1960; Zen, 1961, Pl. 1), the Ira unconformably overlies older rocks, and, as stated earlier, within the present map area and within the east limb of the Middlebury synclinorium, it progressively overlaps onto rocks as old as the Early Ordovician Shelburne Formation. The inception of diastrophic events at the site of the Middlebury synclinorium, presumably the precurser of the formation of the synclinorium itself, is thus pre-late Black River. However, the Hortonville Slate and the overlying Taconic sequence have also been deformed in the Middlebury synclinorium as a unit and thus the formation of the Middlebury synclinorium must be an extended event ranging possibly into the Acadian.

As argued in an earlier section, the West Rutland marble belt is structurally part of the Pittsford-Chippenhook anticline and may be similarly dated. Here also both the pre-Ira and post-Ira foldings are recorded.

The Florence nappe is tentatively taken as a locally accentuated phase of the Pittsford-Chippenhook anticline. Geometric relations however indicate that its formation is no older than the main Taconic thrust. It is possible that much of the deformation in this structure was in the nature of drag folds below the Taconic thrust as the latter moved in from the east. However, the metamorphism of the Taconic rocks is later than thrusting and recumbent folding (Zen, 1960a, p. 132); the grade and trend of metamorphism accord with the disposition of metamorphic grade in northern New England as a whole. In eastern Vermont, the metamorphic event is post-Siluro-Devonian (Thompson, *In* Billings *et al.*, 1952) and is accompanied by intense deformation. Presumably, then, part of the deformation of the Middlebury synclinorium is as late as the Acadian and this may be the age of the formation of the Florence nappe.

The Pine Hill fault cannot be precisely dated. At Boardman Hill, Precambrian rocks abut the Ira Formation, so the last stage of movement on this fault must be post-Black River, but the beginning of this deformation remains undated. It seems likely however that the fault is approximately contemporaneous with the Pittsford-Chippenhook anticline and the West Rutland anticline.

Immediately to the southeast of the present map area, in the Pawlet and Wallingford quadrangles, Thompson (1959, p. 83, Pl. H-1, H-2) shows a belt of north-south trending normal faults. One of these faults, north of Dorset Mountain, is shown to bring Precambrian gneisses against the Shelburne Formation; the Ira Formation however overlies the fault apparently without displacement. Following Thompson, the normal fault is pre-Ira in age. As the Bascom and Chipman Formations appear to be in normal sequence with the Shelburne, the fault may be related to the pre-Ira, mid-Ordovician movements earlier considered. Thompson (1959, Pl. H-1) however also shows a normal fault to truncate the structure on the upthrown side of the Pine Hill fault. These "normal faults" may then have been active in post-Ira time as well. Thompson shows no displacement of the Ira Formation by this fault, which may suggest a chronological discrepancy; however, faults of this sort are very difficult to detect in a mass of black slates. Recently Platt (thesis, Yale University, 1960, p. 94) concluded that normal faults were important on the west side of the Taconic allochthone at approximately the time of Taconic thrusting.

Whatever our uncertainties, it is nonetheless clear that the Middle Ordovician was a time of intense unrest in this area. The Ira Formation not only marks a major unconformity but also marks a fundamental shift in sedimentation pattern, from carbonate-orthoquartzite to shale-greywacke deposition, in western Vermont (Zen, 1960b). The Taconic thrust took place at this time and may in fact be responsible for the shift in sedimentary regime through changing the source material. Normal and reverse faulting, as well as close and open folding, also were concomitant in the autochthone. Submarine graben and horst topography may have prevailed with appreciable local relief and scarps; this could possibly account for some rapid local changes in facies and thicknesses of the Ira Formation.[1] This was a time of folding in the autochthonous carbonate-orthoquartzite terrane also; as new beds were deposited in the Trenton sea, or allochthonous elements brought in by thrusting, they became folded and faulted with persistent westward movement sense. While deformation continued into the post-Ordovician time, very possibly no new major structural features were initiated in the later period. The contrast in intensity of deformation, of the mid-Ordovician and later periods in the Taconic belt of rocks, is perhaps best illustrated by the relation at Becraft Mountain in eastern New York (Schuchert and Longwell, 1932), where the Siluro-Devonian rocks, in relatively open folds, rest unconformably upon isoclinally folded and faulted Taconic rocks of Cambro-Ordovician age.

[1] This concept explains the origin of the quartzite near Center Rutland, immediately west (downthrown side) of the trace of the Pine Hill fault, discussed previously. Fowler interpreted this as a klippe of the Cheshire Quartzite. The rock does resemble the Cheshire but is interbedded with the Ira and is here considered a lens in the latter. The proximity of the rock to the Pine Hill fault suggests that it is locally derived from the Cheshire Quartzite which may have formed a local scarp resulting in a talus-like deposit. Alternatively, the residual sand of an emergent shore underlain by the Cheshire may have slid into the Ira sea in a local turbulent flow.

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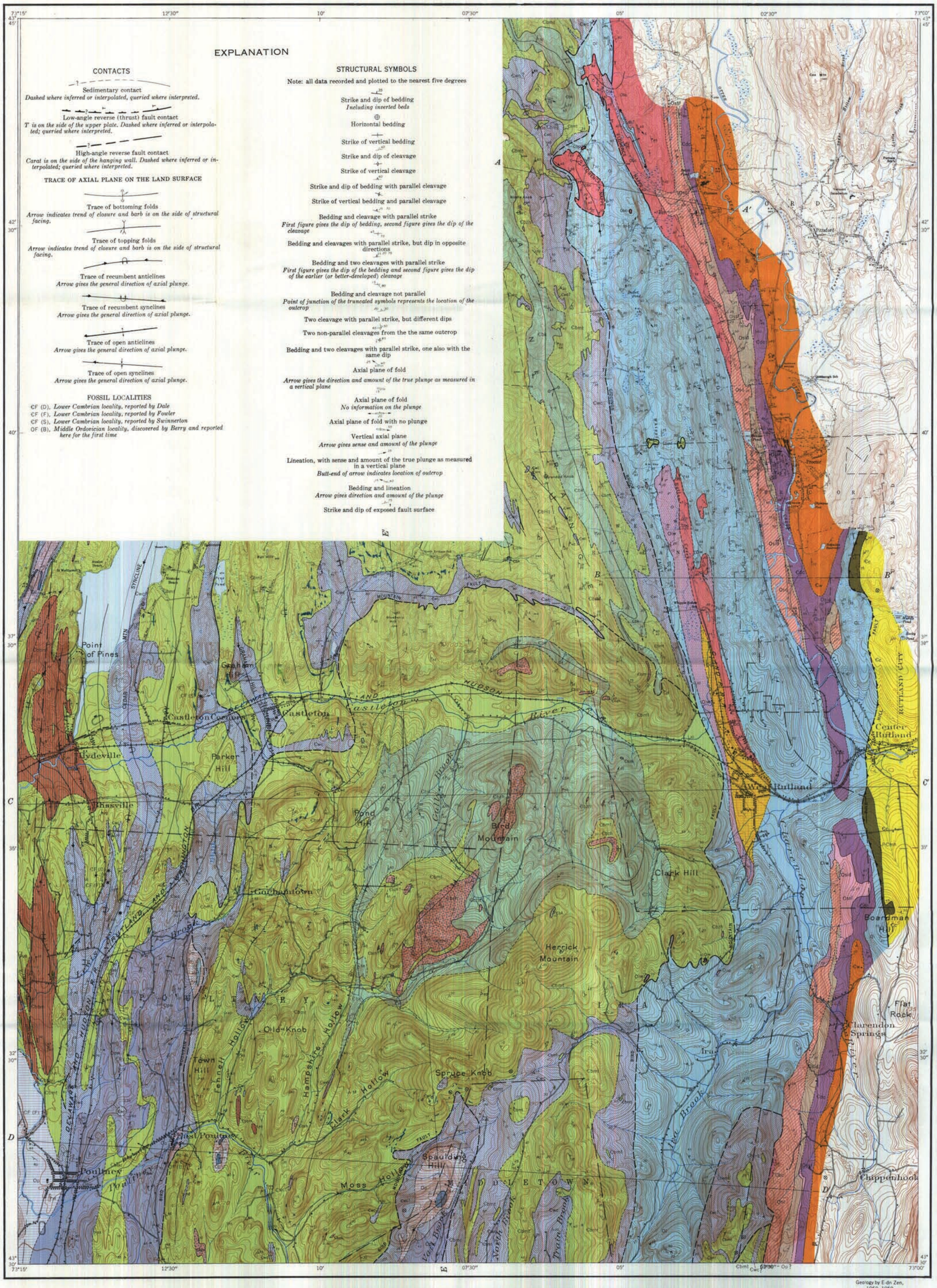
PLATE 1



BEDROCK GEOLOGY OF THE CASTLETON QUAD VERMONT

Scale 1:48 000

1 0 1 2 3 4 Miles

EXPLANATIONS

SYNCLINORIUM SEQUENCE

MIDDLE ORDOVICIAN

Oi / Oiw

Ira Formation

Oi; *black to grey, siliceous to graphitic and pyritiferous phyllite, locally sandy, limy, or with beds rich in albite porphyroblasts.*

Oiw; *Whipple Marble Member: dark bluish-grey weathering, thin-bedded marble, generally dark grey on a fresh surface and finer grained than the Lower Ordovician marbles. Basal to the Ira Formation, but locally interbedded with the black phyllite or occurs as lenses therein.*

LOWER ORDOVICIAN

Ocb

Chipman Formation

Ocb; *Beldens Member: uniform grey to green-streaked white massive marble, interbedded with massive, buff and craggy weathering tan dolostone. Beds with dolomitic mottling occur locally; these correspond to the Burchards Member in lithology but are not differentiated on the map.*

Ocw; *Weybridge Member: grey fine marble interbedded with brown-weathering, slightly dolomitic layers at about 3 inch intervals. Not separately shown on the map, but occurs locally immediately east of the quarries at West Rutland, stratigraphically overlying the Bascom Formation.*

Obc

Bascom Formation

Obc; *grey, streaky marble; grey marble with dolomitic mottles; massive white marble; interbedded massive grey dolostone and grey marble; interbedded thin grey-white marble and brown-weathering siliceous dolostone and limestone; massive spongy-weathering dark grey dolomitic quartzite; grey sugary marble with pale siliceous or micaceous partings.*

Osc / Osid / Ossf

Shelburne Formation

Osc; *Columbian Marble Member: massive, coarse marble, rarely grey-streaked or with dolomitic mottling.*

Osid; *Intermediate Dolostone Member: massive, iron-grey, uniform, grey-weathering dolostone.*

Ossf; *Sutherland Falls Marble Member: massive white to grey marble with yrep streaks and dolomitic mottling and rare interbedded dolostone. This unit is locally absent in the section.*

UPPER CAMBRIAN

Єdc

Undifferentiated Clarendon Springs and Danby Formations

Єdc; *massive iron grey, sugary, grey weathering dolostone with local channel fills and cross-bedding marked by laminae of quartz sands, in beds as thick as 1 foot; fist-sized clusters of milky quartz are abundant (Clarendon Springs lithology). This rock grades downward into similar dolostone with local interbedded vitreous quartzite as much as 6' thick (Danby lithology). North of Proctor, the basal beds of the undifferentiated unit are locally buff coloured and the quartzite is arkosic weathering greenish; the rock grades downward into the Winooski Dolostone.*

LOWER TO MIDDLE CAMBRIAN

Єw

Winooski Dolostone

Єw; *massive, grey to buff-weathering, pale dolostone in shades of orange, yellow, or grey. Thin arkosic quartzite beds weathering greenish grey, and grey silty slate beds, occur at irregular intervals.*

LOWER CAMBRIAN

Єc

Cheshire Quartzite

Єc; *massive vitreous grey to white quartzite, weathering buff to grey and smooth. Basal beds may carry black phyllite, arkosic small-pebble conglomerate, or feldspathic layers each bed as much as 1 foot thick.*

PRE-CAMBRIAN

PЄmh

Mount Holly Series

PЄmh; *undifferentiated biotite-microcline gneiss and biotite-plagioclase gneiss, with minor beds of coarse, yellowish marble, green chlorite-muscovite schist, quartzite, and dark grey calcareous siltstone.*

TACONIC SEQUENCE

MIDDLE ORDOVICIAN

Opw

Pawlet Formation

Opw; *interbedded grey, silty to black, fissile slate and massive, dark grey greywacke beds up to 6' thick that commonly show graded bedding due to a up-section decrease in maximum grain diameter. Quartz and feldspar dominate in the greywacke beds, but slate chips are common, and the matrix may be slightly calcareous. Base of the formation is a black slate which commonly carries graptolites.*

Oir

Indian River Slate

Oir; *red and green, locally grey-green slate, with minor beds of quartzite and dolostone.*

LOWER ORDOVICIAN

Op

Poultney Slate

Op; *interbedded grey, green, purple, and black slate weathering dull to cherty, but locally calcareous. Thin quartzite beds and, especially near the base of the unit, thin grey, fine-grained limestone seams, are locally abundant.*

LOWER CAMBRIAN

Єwc

West Castleton Formation

Єwc; *grey siliceous to black, graphitic, pyritiferous slate and phyllite, locally with interbedded thin dark grey dolostone and grey quartzite and arkosic layers. Thin, white sandy laminae commonly found in the graphitic beds. The Beebe Limestone Member; a massive, lenticular, black limestone weathering blue-grey, and locally fossiliferous, is not separately shown on the map.*

Єbmt / Єbbm / Єbzh

Bull Formation

Єbmt; *Mettawee Slate Facies: bulk of the formation. Purple green-grey, and variegated slate, siltstone, and phyllite, non-chloritoid-bearing; minor beds of white to green quartzite and creamy limestone near the top of the formation. Bright blue; North Brittain Conglomerate Member: limestone-pebble conglomerate in a non-calcareous green to purple slate matrix. Thickness extremely variable. The conglomerate is intraformational and may pass visibly into boudinaged beds. Locally fossiliferous beyond the confines of the map area. Bright green; Mudd Pond Quartzite Member: white to grey, vitreous, medium-grained orthoquartzite with local dolomitic pods (concretions?). Locally dark grey quartz-grit in an argillaceous and slightly calcareous matrix (Eddy Hill lithology). Rock generally weathers white and smooth with a waxy luster.*

Єbzh; *Zion Hill Quartzite and Greywacke Member: green, vitreous chloritic quartzite or greywacke with common limonite spots. Base of the unit is commonly a pebble conglomerate and top may be a siltstone. Unit as a whole is massive and lenses in and out rapidly. Weathers white and tends to form cliffs. Stratigraphic position relative to the Biddie Knob Formation variable in the northern Taconic area as a whole; in the map area, much of this member ("Bird Mountain grit") is actually within the Biddie Knob Formation.*

Єbbm; *Bomoseen Greywacke Member: massive and uniform, green to olive arkose and greywacke, carrying visible flakes of mica and rock fragments. Weathers pale red to dull white. Stratigraphic relation to the Zion Hill Member is variable, within the map area it is below the Zion Hill.*

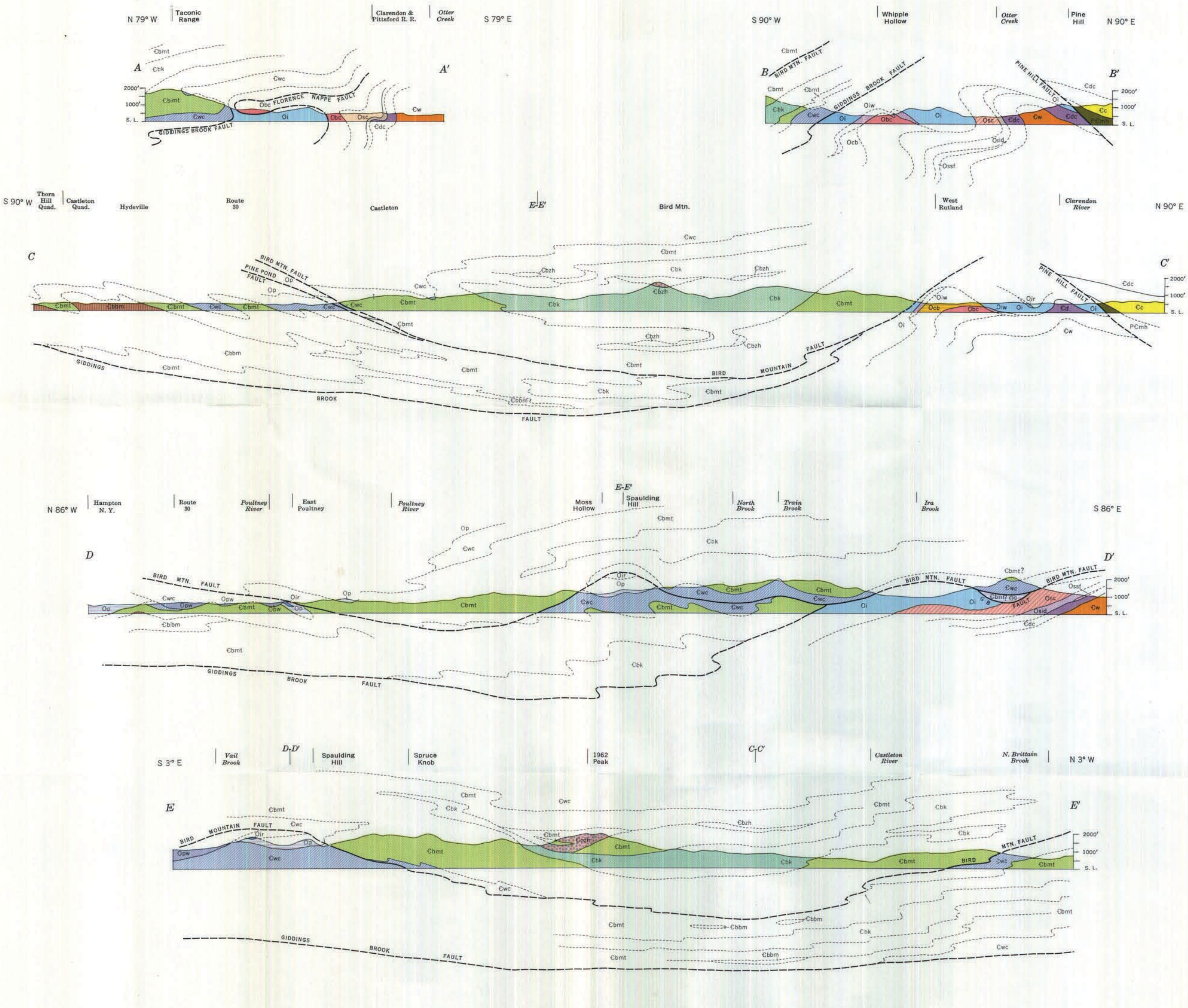
Єbk

Biddie Knob Formation

Єbk; *purple and green chloritoid-bearing slate and phyllite. Minor beds of white to green quartzite and rarely limestone seams occur at irregular intervals.*

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 25)

PLATE 2



GEOLOGIC SECTIONS ACROSS THE CASTLETON QUAD VERMONT

Scale 1:48 000

1 0 1 2 3 4 Miles

VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 25)