BEDROCK GEOLOGY OF THE PAWLET QUADRANGLE, VERMONT

PART I
CENTRAL AND WESTERN PORTIONS

By
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DEPARTMENT OF WATER RESOURCES
Montpelier, Vermont

Bulletin No. 30 1967

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ABSTRACT

Taconic rocks of the Pawlet quadrangle are tightly folded, faulted and metamorphosed Cambrian and Ordovician argillite, impure argillite, and graywacke sediments. The rocks were deposited, either in a trough on the Champlain shelf, or seaward of the shelf, on the margin of the Magog eugeosyncline. Absence of facies change, in either the marginal Taconic sediments or the Champlain carbonates, is suggestive of a depositional site between the shelf and the eugeosyncline.

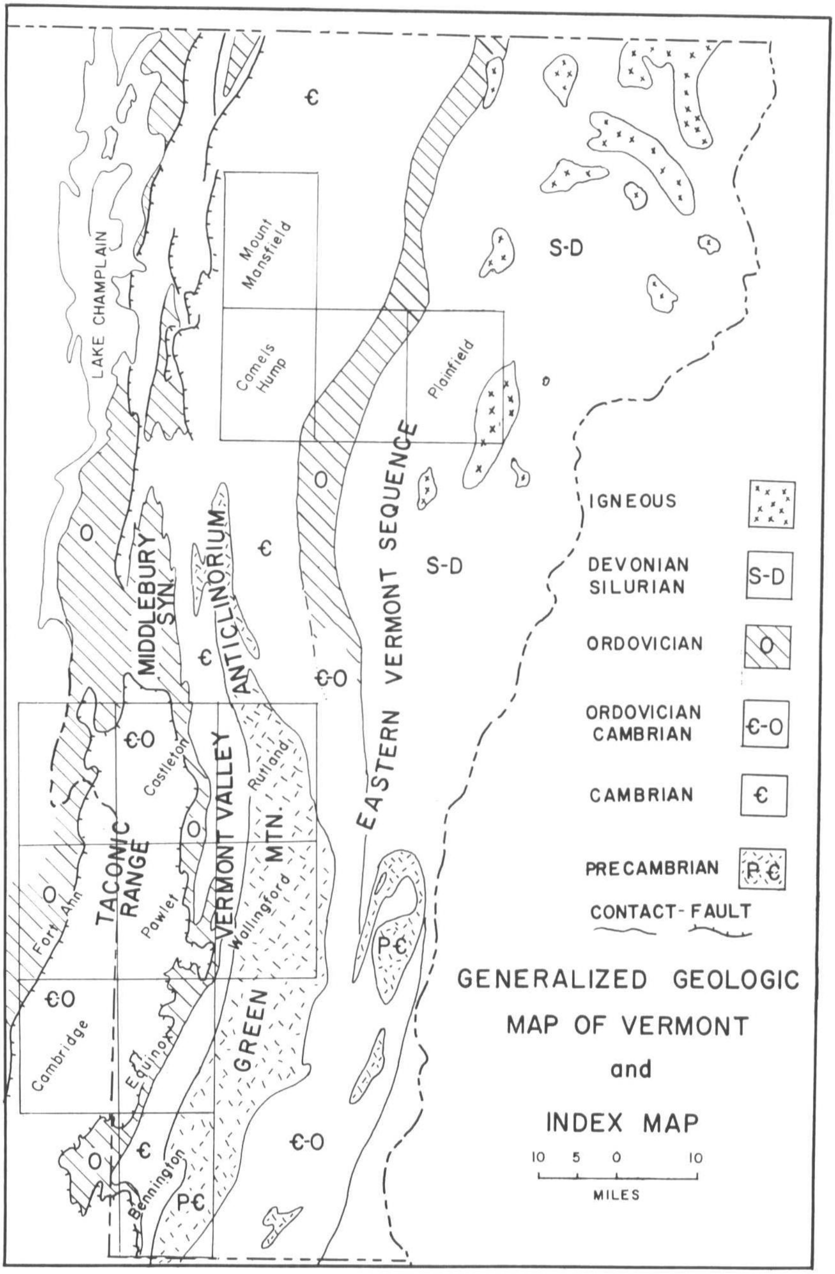
Early major folding of the sediments formed a large recumbent "Taconic fold" overturned toward the west. This "Taconic fold" probably developed during the Middle Ordovician, and was thrust westward from the site of deposition as a detached sheet during the Trenton, onto sediments of the Champlain miogeosyncline. The final stage of the emplacement was accompanied by shearing of the "Taconic fold" into a series of imbricate thrust plates.

Abundant folds, faults and cleavage of the last, post-Taconic deformation were superimposed on, and often conceal, earlier structural features. Generally, axial plane fracture cleavage dips steeply to the east. On the eastern edge of the quadrangle, near the Green Mountain anticlinorium, the folds in Shelburne and younger sediments are nearly recumbent. The time of the last folding and the regional metamorphism was post-Trenton, but may have been as late as Devonian.

INTRODUCTION

The Pawlet quadrangle covers approximately 225 square miles between 43° 15′ and 43° 30′ north latitude and 73° and 73° 15′ west longitude. Parts of Bennington and Rutland counties and the towns of Dorset, Rupert, Pawlet, Tinmouth, Middletown Springs, Danby, Ira, Clarendon,

* Now with Humble Oil and Refining Company, Midland, Texas.



LAKE CHAMPLAIN
Mount Mansfield
Camels Hump
Plainfield
MIDDLEBURY SYN.
ANTICLINORIUM
EASTERN VERMONT SEQUENCE
TACONIC RANGE
VERMONT VALLEY
GREEN MTN.
Fort Ann
Pawlet
Castleton
Rutland
Wallingford
Cambridge
Equinox
Bennington
IGNEOUS
DEVONIAN SILURIAN
ORDOVICIAN
ORDOVICIAN CAMBRIAN
CAMBRIAN
PRECAMBRIAN
CONTACT- FAULT
GENERALIZED GEOLOGIC MAP OF VERMONT and INDEX MAP
10 5 0 10
MILES

Figure 1.

and Wallingford are included within its boundaries. This region is well known for its colored slate and white marble.

Physiography

The Pawlet quadrangle lies wholly within the Taconic section of the New England Physiographic Province (Fennerman, 1938). The northern portion of the Taconic range (Fig. 1) is bounded on the north and west by the broad, open Champlain Valley. The Vermont Valley (Fig. 1), a narrow southern extension of the Champlain Valley, separates the eastern high Taconic mountains from the heavily forested Green Mountains.

Within the Taconic range only two peaks, those of Dorset and Equinox mountains, are over 3800 feet. Dorset Mountain forms the southern end of Tinmouth Valley and nearly blocks the Valley of Vermont. Near North Dorset, where the restriction is greatest, the summit is over 3100 feet above the narrow Vermont Valley floor. Two series of hills trend north-south across the quadrangle. The easternmost mountains vary in elevation from 2000 to 3800 feet, whereas the western series are approximately 1500 to 2000 feet. A central valley of low rolling hills separates the eastern and western mountain ridges and a slate belt lowland borders the quadrangle on the west. About half of the roads shown on the 1897 topographic map are passable by modern car, so that even the higher and more remote eastern sections of the eastern Taconics are readily accessible.

In general the topography of the region reflects the lithology of the underlying bedrock. Uplands are either the result of more resistant formations, or they are remnants of thrust sheets (Fig. 2). The valleys reflect belts of non-resistant marble and fine-grained slate, or erosional breaching of the thrust slices (Fig. 3 and Pl. 1). West-facing escarpments, which are prevalent throughout the region, are caused by easterly dipping subparallel cleavage and bedding. The shear bold surfaces of these cliffs are a reflection of the north-south vertical joints.

Although Pond (1927, 1928) states that nine erosional bedrock terraces all developed within the northern Taconics, it was found that most terraces and shoulders are a reflection of local resistant bedrock layers or of glacial deposits.

The topography is generally mature and the Mettawee River flows on a flood plain. Derangement of the preglacial pattern has caused limited reaches of this river to be characteristically youthful. Furthermore, glacial deposition and erosion have formed several lakes. The largest of these, Lake St. Catharine, is a product of scour and overdeepening by of Pleistocene glaciers.

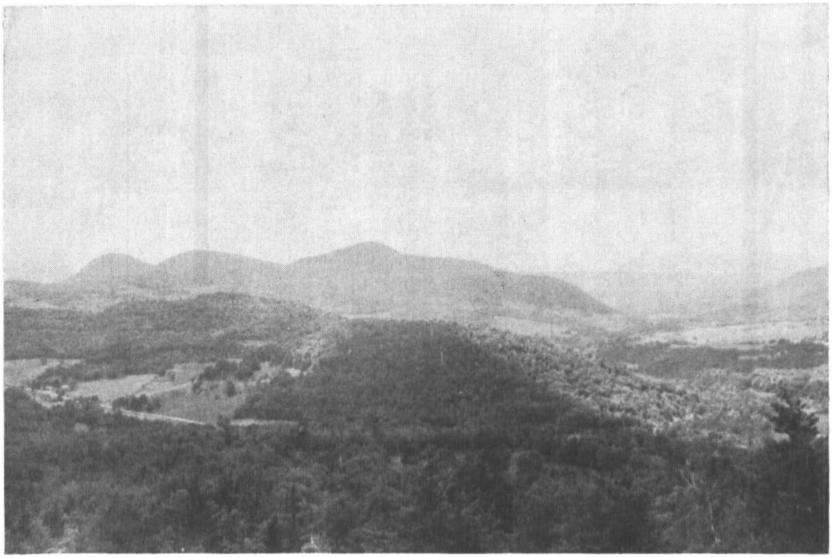


Figure 2. View west from the Purchase. Western Bird Mountain Slice in middle background. Slate Belt lowland is seen beyond Wells Brook valley. Adirondack Mountains on horizon.

Figure 3. View south over Little Pond. Western fault-line scarp of Bird Mountain Slice.

Regional Geology

The rocks of the Pawlet quadrangle consist of tightly folded and faulted metamorphosed sediments of Cambrian and Ordovician age. Those of the western three-fourths of the quadrangle are dominantly pelitic sediments which have been metamorphosed to green-schist facies slates or phyllites. These Taconic sediments form a broad belt from Brandon, Vermont (Fig. 4), approximately 150 miles southward to the vicinity of Poughkeepsie, New York.

Vermont Valley marbles and well-sorted metaquartzites predominate in the eastern quarter of the quadrangle. These units are comparable in age and in metamorphic grade to the adjacent pelitic equivalents in the Taconic terrain, although the entire Vermont Valley Sequence appears to underlie the Taconics. The contact between the two sequences is considered to be a fault, but physical evidence for this conclusion is obscured by regional metamorphism. This "contact" can be traced northward to Brandon, some 20 miles, thence around the north end and southward along the western edge of the Taconic range (Fig. 4), where at least part of it is known to be a low-angle thrust fault (Walcott, 1888, p. 317). Here again Lower Cambrian pelitic sediments overlie Ordovician carbonates and orthoquartzites. Extending northward into the Champlain Valley, carbonates and orthoquartzites enclose the northern end of the Taconic Mountains (Fig. 1). Near Middlebury, the rocks are folded into a south-plunging synclinorium. This major structure causes the younger carbonates to plunge under the thrust sheet of older pelitic rocks.

Flanking the carbonates on the east is the structural complement of the synclinorium, the Green Mountain anticlinorium (Fig. 1). In southern Vermont the axis of the anticlinorium is marked by outcrops of Precambrian schists and gneisses. Northward in central Vermont, near Lincoln, these rocks plunge under Cambrian (?) garnetiferous pelitic metasediments, similar to those within the Taconic terrain.

The eastern limb of the Green Mountain anticlinorium is composed of a thick sequence of metagraywacke, phyllites, marbles, and metaigneous rocks. Although the formations are generally within the biotite and garnet metamorphic zones, post-kinematic acid intrusions have raised them locally to the staurolite facies. This series of rocks is for convenience named the Eastern Vermont Sequence and is considered to be of Ordovician, Silurian, and possibly Lower Devonian age.

Previous Geologic Work

The first mapping of the Taconic rocks within the Pawlet quadrangle was compiled by Walcott and published in 1888. Cambrian Georgia Ter-

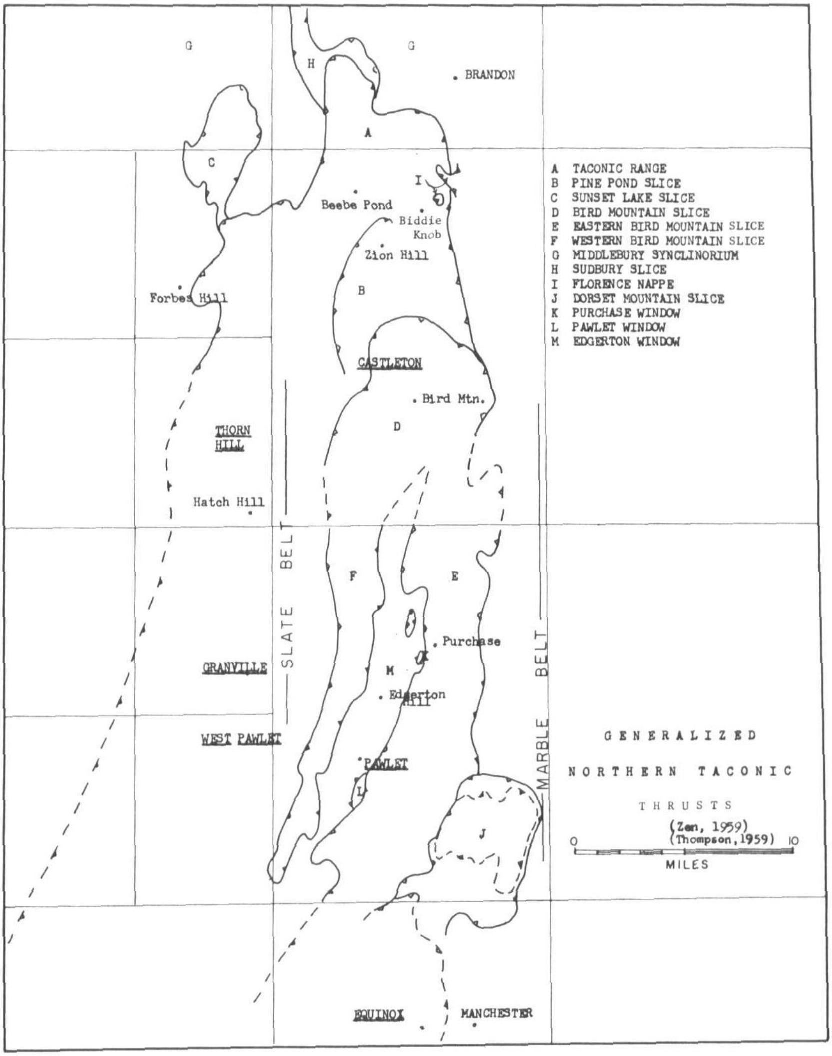

G
G
H
BRANDON
A
C
I
Beebe Pond
Biddie Knob
Zion Hill
B
Forbes Hill
CASTLETON
Bird Mtn.
D
THORN HILL
Hatch Hill
F
E
Purchase
M
GRANVILLE
Edgerton Hill
WEST PAWLET
PAWLET
L
J
SLATE BELT
MARBLE BELT
EQUINOX
MANCHESTER
A TACONIC RANGE
B PINE POND SLICE
C SUNSET LAKE SLICE
D BIRD MOUNTAIN SLICE
E EASTERN BIRD MOUNTAIN SLICE
F WESTERN BIRD MOUNTAIN SLICE
G MIDDLEBURY SYNCLINORIUM
H SUDBURY SLICE
I FLORENCE NAPPE
J DORSET MOUNTAIN SLICE
K PURCHASE WINDOW
L PAWLET WINDOW
M EDGERTON WINDOW
GENERALIZED
NORTHERN TACONIC
THRUSTS
(Zen, 1959)
(Thompson, 1959)
0
10
MILES

Figure 4.

rain and Ordovician Hudson Terrain were differentiated, and Walcott gave the locations of two Cambrian fossil localities in the Pawlet quadrangle.

In 1899, T. W. Dale published his classic report on the Slate Belt of Eastern New York and Western Vermont. His detailed stratigraphic subdivisions and datings of the Taconic rocks are still used, with only minor modifications (Fig. 5). Dale was aware of the age equivalence of the Vermont Valley and Taconic sequences, for he explained their relationship as normal, time equivalent, sedimentary facies.

> ". . . while the sediments east of the Taconic range were largely calcareous in Cambrian and Trenton time, west of it they were argillaceous and arenaceous."
>
> "The conclusion is that there were thereabouts [between the two sequences] changes from calcareous to argillaceous, and arenaceous sedimentation during Cambrian and again in Ordovician time." (Dale, 1899, p. 295).

No detailed explanation was given for a lack of exposed gradation or interfingering of the two phases, but it is clear that he thought the Berkshire schist showed this facies change.

> "On Rupert Mountain, about 3 miles north-northwest of Rupert village, near the Berkshire schist boundary, the grits [his Hudson River] show a transition to schist, becoming more sericitic, while schist shows sedimentary quartz grains." (Dale, 1899, p. 187).

Dale mapped (1899, pl. XIII) the Berkshire schist as a younger formation unconformably overlying the Cambrian units, and therefore obscuring any evidence of facies change in that part of the section. This interpretation has been accepted by Bain (1938) and MacFadyen (1956) in the Bennington quadrangle (Fig. 1).

On the other hand, the anomalous position of the Taconic and Carbonate sequences prompted Ruedemann (1909) to suggest a thrust fault between the two sequences, at the north end of the Taconics. However, it was Keith (1911, 1912, 1932) who clearly stated the evidence for faulting and the conclusion that the entire northern Taconics were thrust in from "east of the Green Mountains" (Keith, 1932, p. 364). Since then, the eastern phyllites of the northern Taconic Mountains were presumed to be allochthonous Lower Cambrian phyllites similar to those of the slate belt.

Subsequent work by Larrabee (1939–40), Kaiser (1945), Fowler (1950), and Billings, Rogers and Thompson (1952) at the northern Taconics has supported the thrust hypothesis.

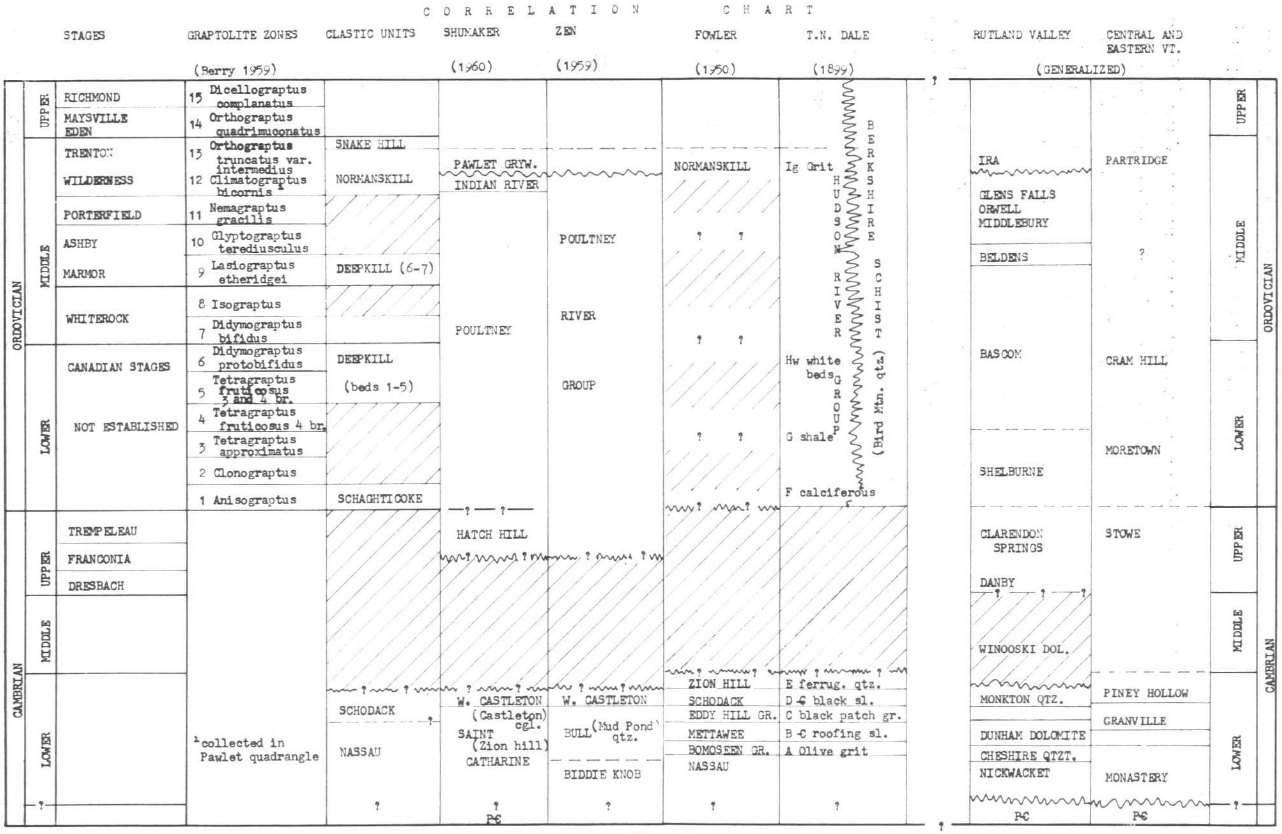

CORRELATION CHART
STAGES
GRAPTOLITE ZONES (Berry 1959)
CLASTIC UNITS
SHUMAKER (1960)
ZEN (1959)
FOWLER (1950)
T.N. DALE (1899)
RUTLAND VALLEY
CENTRAL AND EASTERN VT.
(GENERALIZED)
ORDOVICIAN
CAMBRIAN
UPPER
MIDDLE
LOWER
RICHMOND
MAYSVILLE
EDEN
TRENTON
WILDERNESS
PORTERFIELD
ASHBY
MARMOR
WHITEROCK
CANADIAN STAGES
NOT ESTABLISHED
TREMPELEAU
FRANCONIA
DRESBACH
15 Dicellograptus complanatus
14 Orthograptus quadrimucronatus
13 Orthograptus truncatus var. intermedius
12 Climatograptus bicornis[1]
11 Nemagraptus gracilis
10 Glyptograptus teretiusculus
9 Lasiograptus etheridgei
8 Isograptus
7 Didymograptus bifidus
6 Didymograptus protobifidus
5 Tetragraptus fruticosus 3 and 4 br.
4 Tetragraptus fruticosus 4 br.
3 Tetragraptus approximatus
2 Clonograptus
1 Anisograptus
[1]collected in Pawlet quadrangle
SNAKE HILL
NORMANSKILL
DEEPKILL (6-7)
DEEPKILL (beds 1-5)
SCHAGHTICOKE
SCHODACK
NASSAU
PAWLET GRYW.
INDIAN RIVER
POULTNEY
HATCH HILL
W. CASTLETON (Castleton) cgl.
SAINT (Zion hill) CATHARINE
P€
POULTNEY RIVER GROUP
W. CASTLETON
BULL (Mud Pond qtz.)
BIDDIE KNOB
NORMANSKILL
ZION HILL
SCHODACK
EDDY HILL GR.
METTAWEE
BOMOSEEN GR.
NASSAU
Ig Grit
Hw white beds
G shale
F calciferous
E ferrug. qtz.
D € black sl.
C black patch gr.
B € roofing sl.
A Olive grit
HUDSON RIVER GROUP
BERKSHIRE SCHIST (Bird Mtn. qtz.)
IRA
GLENS FALLS
ORWELL
MIDDLEBURY
BELDENS
BASCOM
SHELBURNE
CLARENDON SPRINGS
DANBY
WINOOSKI DOL.
MONKTON QTZ.
DUNHAM DOLOMITE
CHESHIRE QTZT.
NICKWACKET
P€
PARTRIDGE
CRAM HILL
MORETOWN
STOWE
PINEY HOLLOW
GRANVILLE
MONASTERY
P€

Figure 5

Field Work

A total of ten months during the summers of 1957, 1958, and 1959 was spent mapping the western two-thirds of the Pawlet quadrangle. The report was written on that portion of the quadrangle during the winter of 1959–60.

Dr. J. B. Thompson mapped all the geology east of 73° 05′ west longitude. At Mill Brook the line separating the areas mapped swings westward, passing between Woodlawn and Dorset mountains, and then southward to the town of Dorset (Plate 1).

During the investigation it was discovered that there are numerous inaccuracies on the topographic map. Displacement of minor topographic features and massive generalization of the topography are quite common. The air photos supplied by the Vermont Geological Survey were invaluable for mapping, planning traverses, and locating stations to be plotted on the topographic map. Glacial and vegetative cover restricted the reflection of underlying lithologies on aerial photos. Geologic data were located as closely as possible on the topographic map from barometric altitude readings and resection, wherever necessary.

Acknowledgments

The author is indebted to Dr. Charles Doll, Vermont State Geologist, who suggested the problem and arranged for financial assistance from the Vermont Geological Survey. He was privileged to work and write the report under the direction of Dr. Charles M. Nevin. This study was improved by helpful suggestions and criticisms of Drs. R. A. Christman and J. W. Wells.

J. B. Rather III assisted in the field during the summer of 1958, and H. L. Thung during the summer of 1959. The author is deeply indebted to Drs. J. B. Thompson and E. Zen who gave freely of their time and knowledge of Taconic geology. Frequent discussions with W. B. N. Berry, G. Theokritoff, and P. C. Hewitt gave the author an understanding of the geology of adjoining areas. W. B. N. Berry, G. Theokritoff, and H. B. Whittington kindly identified the graptolites collected during this study.

The author and his family take this opportunity to express their thanks to the people of Wells, Vermont, and in particular the R. N. Williams, Miss T. Callihan and the C. Buckleys for their kind hospitality.

STRATIGRAPHY

The Taconic metasediments of the Pawlet quadrangle include rocks from Precambrian or Lower Cambrian to Upper Ordovician age. The entire section is approximately 3000 feet thick, and the units are generally fine-grained meta-argillites. Less than one percent of the total thickness is limestone,[1] and no more than 20 percent are interbedded fine-grained quartzites and graywackes.[2]

The basic stratigraphy of the western or slate portion of the quadrangle has been established since Dale's 1899 report (Fig. 5). His Cambro-Ordovician stratigraphy has been extended into the eastern portion of the quadrangle, which he mapped (1899, 1912) as Berkshire schist. Some new units have also been proposed for local use (Fig. 5.) Most of the area mapped by Dale as Berkshire schist is now interpreted as a thrust slice which has been overturned and faulted, so that the stratigraphic sequence is inverted. The correlation of these eastern units with fossiliferous Taconic units on the west is made solely by their lithologic similarity.

Cambrian System

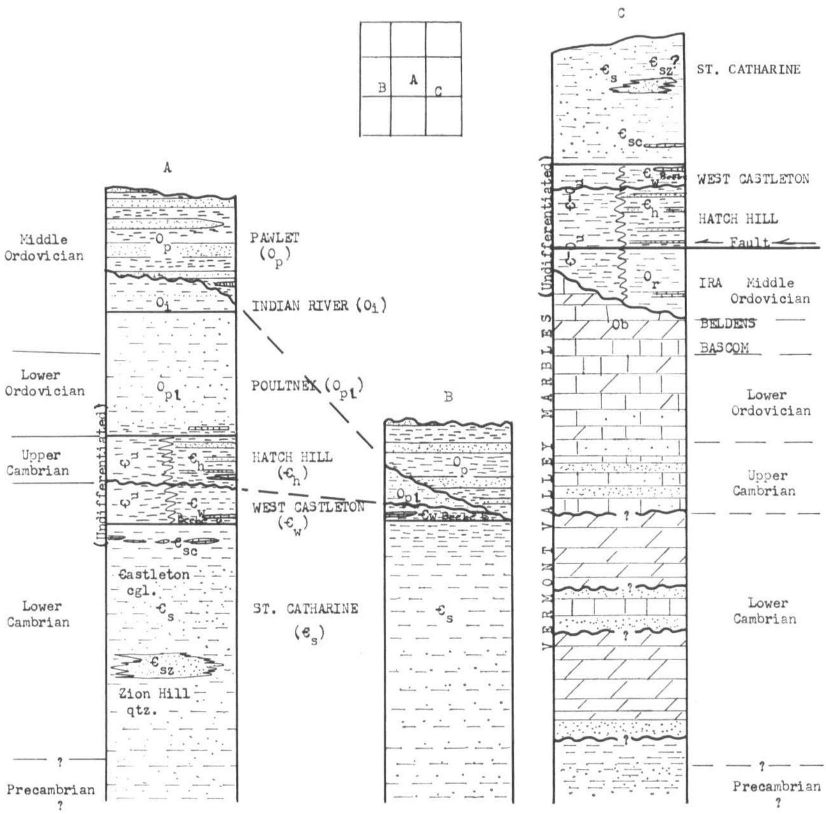
St. Catharine Formation €s (new name)

The St. Catharine Formation crops out in the western chain of hills which include such mountains as Moosehorn, Haystack, and the Pattern; and it is exposed in anticlines near Pawlet, Rupert, and Lake St. Catharine (Plate 1.) The St. Catharine Formation is found in the higher elevations of the eastern Taconic Mountains. Originally, the St. Catharine Formation was marine clay deposits with intercalated thin silty beds. The purple slates of this formation occur only in the western one-third of the quadrangle. Eastward, metamorphism has raised the grade of the rocks to green chlorite-muscovite phyllites and fine-grained schists. The Castleton Conglomerate (€sc) and Zion Hill Quartzite (€sz) (Fig. 6) are distinctive but thin members of the formation.

Lithology: The St. Catharine Formation has two dominant lithologies: a lower section of green phyllites with thin stringers of fine-grained, white-weathering quartzite (Fig. 7); and an upper section of purple and green slates, phyllites and fine-grained chlorite schists (Figs. 8 and 9). The thin fine-grained quartzite stringers occur about every 4–5 millimeters in the phyllite, and generally are one or two millimeters thick, although some are 5 to 6 millimeters thick. Pyrite cubes and pseudo-

[1] Finely recrystallized carbonate rock. In Vermont the term "marble" is generally used when referring to commercial carbonate rock which will take a polish.

[2] A clastic sedimentary rock of high clay-sericite content which has grains of silt size or larger, and which is characterized by vertical grading of sedimentary grains, poor sorting, and a general lack of bedding.



GENERALIZED COLUMNAR SECTIONS
(no vertical or horizontal scale)
Figure 6

morphs of limonite after pyrite are common in the phyllite and, occasionally, large crystals, as much as 15 mm., are found in the schist. The phyllite is composed of sericite-muscovite, chlorite, quartz, chloritoid, and albite, with minor amounts of carbonate, magnetite, pyrite, and hematite (Fig. 10). Rosenfeld, Thompson, and Zen (1958, p. 1637) reported coexistent muscovite and paragonite from Dorset Mountain. These were identified by the basal spacing of 9.981A. and 8.623A. respectively, as determined by X-ray. Zen (1959, p. 1) considers the chloritoid-bearing strata to be restricted to the lower portions of the formation, and on this basis he has differentiated a basal Lower Cambrian chloritoid-bearing unit called the Biddie Knob Formation (Fig. 5).

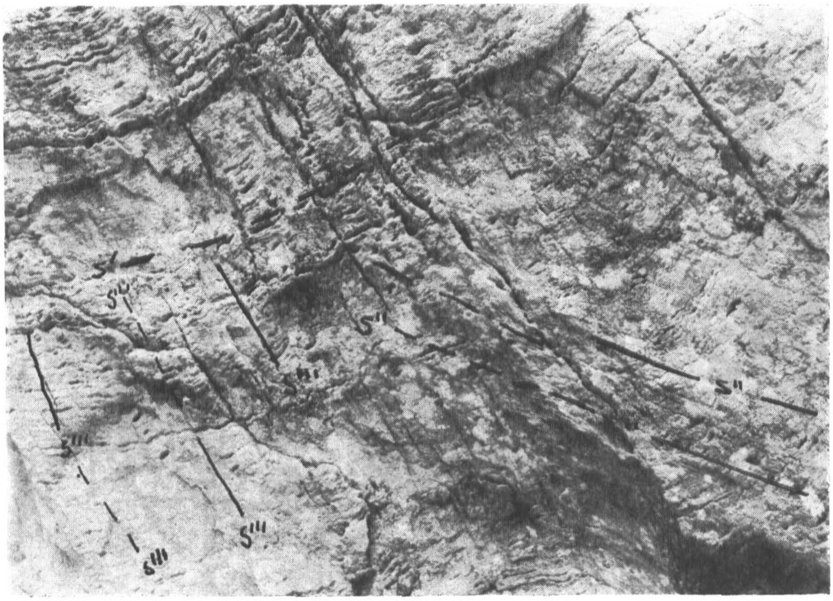


Figure 7. *Impure phyllite of the St. Catharine Formation near East Wells, Vermont.* Early shear surfaces (s'') dip eastward at about 20 degrees. Later regional axial plane fracture cleavage (s''') dips eastward at approximately 55 degrees. Bedding (s') dips gently to the west.

The top 200 feet of the St. Catharine Formation is composed of green and purple slate or phyllite. Axial plane flow cleavage is the most prominent planar surface in the slate (Fig. 8). Development of flow cleavage in this portion of the St. Catharine reflects the high clay content of the original sediment. Where apparent, the bedding is seen as faint brown weathering, limy horizons or uncleaved quartzite beds. Generally, the purple and green colors do not now reflect bedding, but indicate the oxidation state of the contained iron. Dale's report of 1899 includes chemical analyses of various slates, and Larrabee (1939–40) describes the quarrying processes. The mineral assemblage of the slate is sericite, chlorite, albite, quartz, and carbonate, with minor rutile, tourmaline, apatite, magnetite, pyrite, marcasite, and hematite.

Coarse Zion Hill Quartzite (€sz) was mapped at three localities (Plate 2). Dale first described the quartzite at Zion Hill as:

> "... dark grass-green, chloritic calcareous grit ... pebbles measuring up to 2 ⅜ inches in diameter, mostly quartz but a few are quartzite or altered grit consisting of quartz and feldspar." (1899, p. 184).

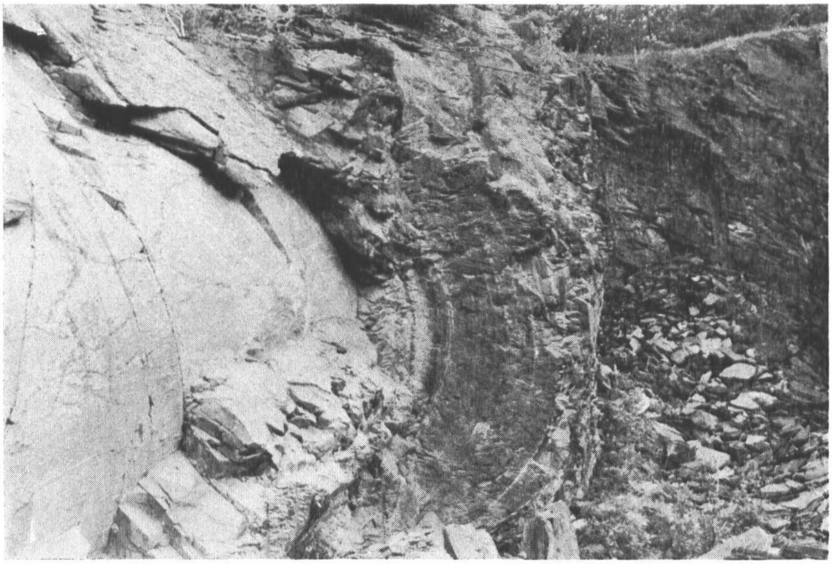


Figure 8. Cedar Mountain Quarry near Lake Bomoseen in the Castlston Quadrangle. The St. Catharine Formation is nearly recumbent and overturned toward the west. Note the lack of apparent bedding on quarry wall in the background.

Within the Pawlet quadrangle, this member is a massive, coarse-grained, vitreous green quartzite. Eighty to ninety percent of the grains are quartz, and the remainder is predominantly feldspar with an average grain size of one to five millimeters. From more extensive exposures in the Castleton quadrangle, Zen made the following observations regarding the Zion Hill Quartzite:

> "This is a massive greywacke or subgreywacke . . . occurs as discontinuous beds. . . . The base of the rock may be a pebble conglomerate, with load casting features . . ., whereas the top may be a mudstone. Graded bedding is common. The Zion Hill is thus a crucial unit in working out the sense of the stratigraphic succession." (1959, p. 1).

Minor amounts of chlorite, sericite, sphene, zircon, magnetite, pyrite, limonite, and biotite were noted in thin section. The coarse phase of this member grades into fine-grained, green quartzites. Frequency of the latter, and resemblance to quartzites which occur higher in the section, prompted differentiation of only the coarser phase on the geologic map (Plate 1). The pebble conglomerate has no counterpart within dated Ordovician units in the northern Taconic rocks. Its occurrence, then, is indicative of the St. Catharine Formation.

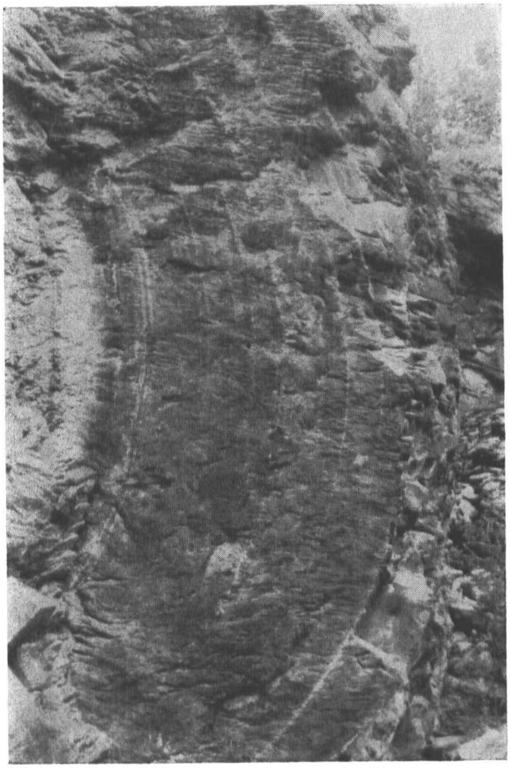


Figure 9. Close up of Fig. 8. Axial plane flow cleavage in slate. An exceptional instance where color banding and bedding are parallel.

The Castleton Conglomerate (Ꞓsc) member is found near the top of the St. Catharine Formation (Fig. 6). It, like the Zion Hill, is limited in distribution and sufficiently distinctive to be considered characteristic of the St. Catharine Formation. This member is a lenticular limestone and intraformational conglomerate. Individual lenses may be as much as 6 inches thick and two or three feet long, and the blue-gray to yellowish-white pebbles are one or two inches thick, and three or four inches long. These limestone pebbles and beds are usually scattered through a five-foot zone of green to brownish weathering, limy slate, but occasionally the slate zone may be 10 to 15 feet thick. In thin section, the carbonate is recrystallized with minor sericite, chlorite and quartz. Occasionally micro- and mega-fossils are present (Fig. 11, Plate 3).

A few exposures of the Castleton Conglomerate occur in the eastern

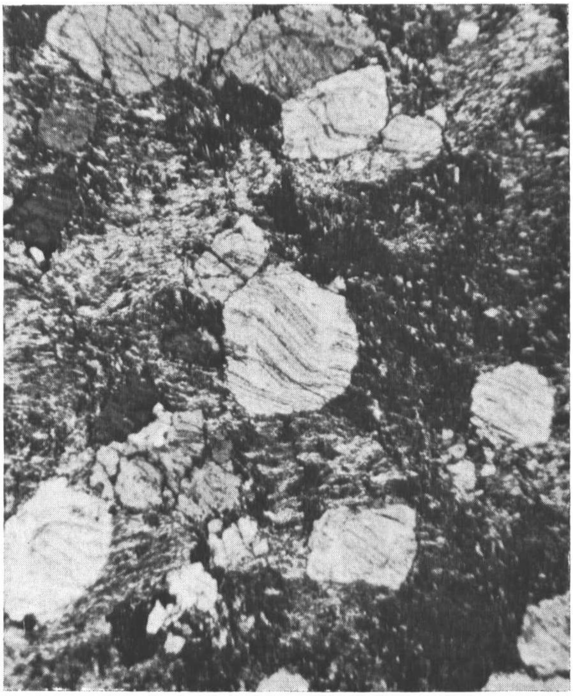


Figure 10. Photomicrograph of the St. Catharine Formation. Albite porphyroblasts in sericite-chlorite phyllite. A few porphyroblasts show slight rotation (×55).

area and in the northern and central portions of the quadrangle, but the best exposures are in the vicinity of Rupert (Plates 1 and 2). The distinctive appearance, fossils and consistent stratigraphic position make this member an important key horizon for determining the structural relations of the entire sequence.

Thickness and contacts: The lower stratigraphic contact of the St. Catharine formation is not exposed in the northern Taconics, for everywhere it is believed to be faulted. The upper contact is conformable with the overlying West Castleton (Ꞓw) Formation. Total thickness is not known, but a minimum of 1500 feet is estimated from exposures within the quadrangle. A similar estimate was made by Fowler (1950, p. 45) for the equivalent formation in the Castleton quadrangle.

Type section, age and correlation: The St. Catharine Formation is typically exposed on the east and west sides of Lake St. Catharine. Excellent exposures occur along Route 30 at the base of St. Catharine Mountain.

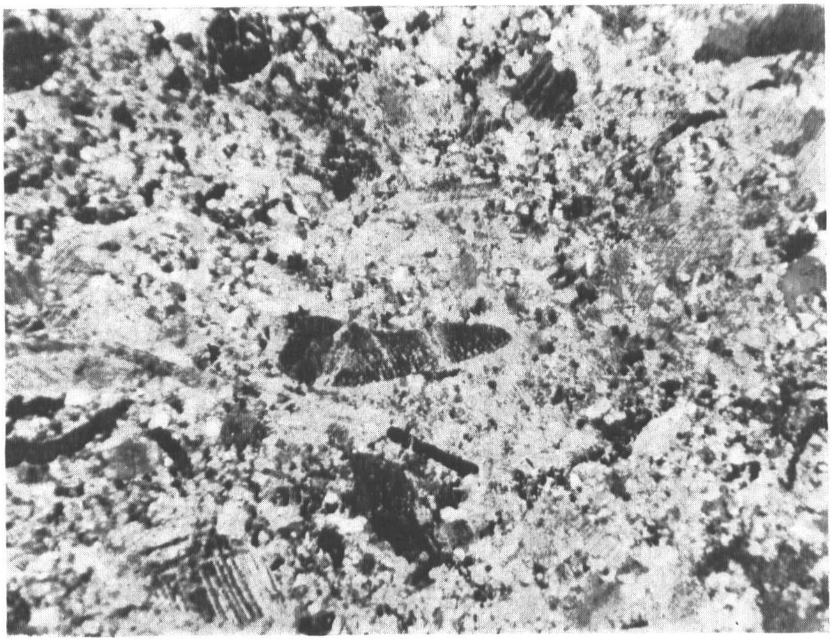


Figure 11. Photomicrograph of Castleton conglomerate limestone. Note the single ray of a sponge spicule and tubular shaped fossil fragments. (×55). Locality 7.

The Zion Hill Member outcrops approximately one-half mile northeast of Lilly Pond at the north end of the lake (Plates 1 and 2). Lower Cambrian fossils were found by Walcott (1888, Pl. 3) south of the lake near the gorge of the Mettawee River.

The unit is tentatively assigned to Lower Cambrian even though some of the lowest part may be Precambrian. A correlation chart of formations equivalent to the St. Catharine Formation is given in Figure 5. The description of typical Nassau Formation (Ruedemann, 1914, p. 70) is identical to the St. Catharine. Although rocks which have been mapped as Nassau (Fowler, 1950) are equivalent to the St. Catharine, the age and stratigraphic position of typical Nassau, some 60 miles to the south, is not known precisely (Price, 1956). Therefore, any correlation with typical Nassau is guesswork. The purple and green slates of the upper part of the St. Catharine Formation are equivalent to the Mettawee Slate Member of the Bull Formation (Zen, 1959; Theokritoff, 1959). The Bomoseen and Eddy Hill grits, common to the west, were not found within the Pawlet quadrangle. Minor amounts of a dark quartzite were found near the upper contact within the St. Catharine Formation on the west side of the Cobble, and in a stream bed at 940 feet on the northwest side of the Pattern. Correlation of these quartzites is uncertain, but they may belong to the Eddy Hill Formation (Ruedemann, 1914, p. 69) or Mudd Pond Quartzite (Zen, 1959, p. 1). Massive white, coarse-grained quartzites were found near the top of the St. Catharine Formation on the west side of Woodlawn and Spoon mountains, on Sargent Hill, and near Spruce Top. It is probable that these quartzites belong to the Zion Hill Member (Fig. 6), but they may be Mudd Pond (Zen, 1959, p. 1). Inasmuch as the quartzites are not coarse, vitreous and green, they were not mapped as Zion Hill.

Dale (1899) may have confused typical Zion Hill Quartzite with a quartzite which occurs between Lower Cambrian black shales of horizon D and Ordovician black shales of horizons F and G (Fig. 5). He named these quartzites the Ferrugenous quartzite (Fig. 5). In 1914 Ruedemann applied the name Zion Hill Quartzite to Dale's Ferrugenous quartzite. Larrabee (1930–40) and Fowler (1950) assigned different ages to the Zion Hill, but neither corrected the cross correlation. Recent work by Zen (1959) and by Theokritoff (1959) indicates that the two quartzites are different. Zion Hill is within the Lower Cambrian St. Catharine Formation, and the other quartzite, the Hatch Hill Quartzite, is Upper Cambrian (Fig. 6). Furthermore, quartzites which Dale (1899) mapped as Bird Mountain, within the Berkshire schist, are identical with Zion Hill (Zen, personal communication and observation of the authors).

The Castleton Conglomerate contains fossils characteristic of the Lower Cambrian *Elliptocephala asaphoides* fauna (Theokritoff, 1959, p. 54). This fauna is found exclusively in Lower Cambrian limestones of the Taconic rocks. The trilobite assemblage is intermediate between the Atlantic and Pacific faunal province assemblage (Lochman and Wilson, 1958, p. 320). Similar fossils are found in limestones within green slates at the type Schodack Formation (Theokritoff, 1957, p. 1804–5). The latter is therefore considered to be equivalent to the Castleton Conglomerate. The name *Schodack* is rejected because previous workers considered the Schodack Formation to be black slates, which here are designated West Castleton (Fig. 5). Walcott's map shows the location of two Cambrian fossil localities within the Pawlet quadrangle. A collection was made from one of these localities near Rupert (Locality 7 on Plate 2) during the summer of 1958. The fossils found were *Acrotreta*, fragments of a *Hyolithid*, *Calodiscus*, *Pelagiella* cf. Lochman (1956, pl. 1), *Hyolithellus* cf. Lochman (1956, pl. 2), numerous sponge spicules, miscellaneous perforate and imperforate phosphatic plates of unknown affinities (Plate 3, and Fig. 11). These phosphatic and silicified fossil remains were etched from the limestone by using a dilute solution of acetic acid.

Two new fossil localities were found during the field work. Well-preserved fragmentary remains of phosphatic tubes of *Hyolithellus* cf. Lochman (1956, p. 12) and hexad sponge spicules were collected from lenticular limestones exposed on Town Hill (Locality 8, Pl. 2). Poorly preserved tubular phosphatic fragments, and a single rayed sponge spicule were found near Woodlawn Mountain (Locality 9, Plate 1) in rocks considered to be Castleton Conglomerate.

Fossils from these two localities are not sufficiently diagnostic to prove a Lower Cambrian age for the conglomerate. However, the fauna is similar to part of the one found within the established Lower Cambrian Castleton Conglomerate. The limestones at these two localities were mapped as Castleton Conglomerate prior to identification of the fossil remains; thus the micro-fossil fragments tended to confirm this mapping. Further study of the phyllites of the eastern, high Taconics range should include etching of lenticular blue-gray limestone found in the stratigraphic position of the Castleton Conglomerate. If well preserved and diagnostic fossil remains can be found clearly within the Taconic sequence, a positive age assignment may be possible, and the opposing viewpoints (Thrust and Non-Thrust) may be finally resolved.

Undifferentiated Black Slates and Phyllites (Є-Ou) and (Єu)

The Cambrian (Єu) undifferentiated black slates include rocks of the Lower Cambrian West Castleton Formation (Єw) and Upper Cambrian Hatch Hill Formation (Єh) (Fig. 6). The Cambro-Ordovician (Є-Ou) undifferentiated black slates include both the West Castleton and Hatch Hill formations and, along the eastern margin of the Taconics, some autochthonous Middle Ordovician black slates of the Ira Formation (Or) (Keith, 1932, p. 398). Individual formations are differentiated wherever possible, and assignment to an undifferentiated unit means that the particular rocks are homogeneous, graphitic black slates or phyllites with minor amounts of gray slate or phyllite (Fig. 6).

Some Cambrian black slates outcrop within the Edgerton window, and slate belt regions (Fig. 4). These occur in sequence between the St. Catharine and Poultney formations, but unfortunately, sufficiently distinctive lithologies are not present to differentiate West Castleton or Hatch Hill.

The Cambro-Ordovician black pelites outcrop extensively in the eastern portion of the Bird Mountain slice (Plate 2). Inasmuch as the Taconic sequence of the Bird Mountain slice is interpreted as overturned, Lower and Upper Cambrian black phyllites of the Taconic sequence may be in thrust contact with the younger, underlying black phyllites, the Ira Formation, of the Vermont Valley sequence (Fig. 6). The Ira Formation unconformably overlies truncated marbles of the Vermont Valley sequence. Middle Ordovician fossils have been found in limestones of the Ira Formation (Zen, 1959, Thompson, 1959). The black phyllites and slates of the West Castleton, Hatch Hill and Ira formations are so similar that it is impossible to locate the fault plane exactly, or to separate the Taconic and Valley sequences with any degree of certainty. Other criteria could not be established to map the thrust. For these reasons the sole of the thrust is somewhat arbitrarily placed at the nearest mappable horizon, the top of the Valley marbles. This contact is interpreted to be a possible lower limit for the Taconic thrust zone; however, some rocks above this thrust certainly include part of the autochthonous Ira Formation.

West Castleton Formation (€w) (Zen, 1959, p. 2)

Rocks of the West Castleton Formation crop out on Town Hill, The Cobble, and on Pond Mountain, and are best exposed near the town of Rupert. Lithologies similar to those of the West Castleton Formation were mapped near North Rupert, within the Bird Mountain slice (Fig. 4, Plate 2).

Lithology: Black slates and phyllites predominate in the West Castleton Formation, but some limestone and minor amounts of rusty-weathering quartzite and pyrite are associated with the slates of this formation. Zen (1959, p. 2) reports the occurrence of thin dolomite beds and pebble conglomerate within the West Castleton Formation in the Castleton quadrangle. The black lenticular Beebe Limestone Member (€wb) (Keith, 1932, p. 402) distinguishes black slates of the West Castleton Formation from similar black slates of other units. The importance of this limestone was emphasized by Keith (1932, p. 402) in his original description:

> "The limestone, which is only 5 to 20 feet thick, . . . is such an exceptional change from the usual character of the sediments and it is so fossiliferous that it is the most important formation in the entire Taconic sequence. It is named for its exposures near Beebe Pond in Hubbardton, Vermont."

> ". . . but it is everywhere present at the proper horizon so far as known."
>
> ". . . although the formation is small, it commonly makes an impress upon the topography either as a line of knolls or a shelf along the sides of hills made by another formation." (Keith, 1932, p. 402).

Thickness and contacts: The estimated maximum thickness of the West Castleton Formation is approximately 30 feet. Where exposed, the lower contact is conformable with the underlying St. Catharine Formation. However, in some places the West Castleton black slates are absent. The individual beds of the Beebe Limestone Member range in thickness from two inches to perhaps two feet, but several beds of limestone interbedded with black slate may be found in a zone up to 8 or 10 feet thick.

Age and correlation: Fossils similar to those found within the St. Catharine Formation occur in the Beebe Limestone of the West Castleton Formation. Theokritoff reports the following from the Granville and Thorn Hill quadrangles.

> ". . . Several beds of fine bluish-gray limestone-conglomerate yield fossils of the *Elliptocephala asaphoides* fauna, and at one locality, fossils of an overlying *Paedeumias* fauna . . . both of these faunas indicate a position at the top of the Lower Cambrian." (Theokritoff, 1959, p. 55).

The West Castleton Formation is equivalent to Fowler's (1950) and Ruedemann's (1914) Schodack Formation, Keith's (1932) Beebe and Hooker formations, and Dale's (1899) Lower Cambrian black slates (Fig. 5).

HATCH HILL FORMATION (Ꞓh) (Theokritoff, 1959, p. 55)

The Hatch Hill Formation is best exposed in the Pawlet quadrangle along the Rupert-Pawlet road one and one-fourth miles north of Spruce Top, and on Town Hill in the central part of the quadrangle. Lithologies similar to, and tentatively assigned to the Hatch Hill, form low shoulders and terraces on the hillsides of the mountains in the eastern Bird Mountain slice (Plates 1 and 2).

Lithology: The formation is composed of black slates and phyllites with characteristic massive, rusty-weathering dolomitic and limy quartzite, sandy dolomite, and minor amounts of sandy limestone. Percentages of carbonate and subangular quartz vary considerably, but generally they comprise 80 or 90 percent of the rock, with 5 to 15 percent of detrital feldspar, mostly microcline and plagioclase. One of the distinctive characteristics of the quartzite is the presence of carbonate, either disseminated in the rock or as white calcite-quartz vein mosaic which is etched into relief by differential weathering. Fresh surfaces of the metaquartzite are gray to dark blue, fine grained, and vitreous. The quartzite is frequently brecciated, with the angular fragments ranging in size from a fraction of an inch to several inches. These are seen as darker masses in a gray-blue to white matrix of recrystallized carbonate and quartz. The fragmental quartzite occurs within the eastern Bird Mountain slice and is best exposed on the southwestern end of Spoon Mountain and on the southeastern side of Sargent Hill. It is possible that breccia-like texture was caused by Taconic thrusting, as the sole of the thrust is believed to be just below this unit.

Limestone, with beds 6 inches to 4 feet thick, is associated with the quartzite and black slate at Town Hill and near Spruce Top. The limestone is well exposed in a stream three-fourths of a mile north of Spruce Top. On a fresh surface the rock is light gray to dark blue. Thin dark seams containing organic material may be apparent on a weathered surface. Resistant beds of sand up to one-half inch thick give outcrops of this rock a ribbed appearance. This limestone occurs above the Beebe Limestone and below quartzites of the Hatch Hill. Inasmuch as it has a high sand content, and outcrops of the limestone are limited to association with the Hatch Hill, it is tentatively assigned a basal position within the Hatch Hill Formation.

Thickness and contacts: The Hatch Hill is overlain, apparently conformably, by the Poultney Formation (Op) (Fig. 5). The contact is drawn between massive limy quartzites of the Hatch Hill and basal argillites, thin limy quartzite or ribbon limestones (Schaghticoke lithology) of the Poultney Formation. Thin limy quartzites are exposed directly above the unconformity in several places. These quartzites may represent incompletely developed Hatch Hill, but their similarity to lithologies found in the Poultney necessitates their being mapped as Poultney Formation. Exposures within the eastern Bird Mountain slice suggest a maximum thickness of 300 feet for the formation. Elsewhere, the maximum thickness is 100 feet, and generally the unit is missing (Plate 2).

Type locality, age and correlation: The type locality is in the southeastern part of the Thorn Hill quadrangle (Fig. 4) at Hatch Hill. The unit has been described by Theokritoff from exposures in the Granville quadrangle as:

> ". . . sooty black pyritic rusty-weathering shales interbedded with rotten-weathering bluish dolomitic sandstone, locally crossbedded,

and characteristically traversed by numerous quartz veins. The type locality is Hatch Hill in the Thorn Hill quadrangle where the formation can be seen outcropping on its western flank." (Theokritoff, 1959, p. 55).

Graptolites found within the black slates of the Hatch Hill Formation of the Granville quadrangle are Upper Cambrian in age (Berry, 1959, p. 61). The Hatch Hill Formation, as found within the slate belt region, is identical with the unit which Fowler (1950), Larrabee (1939–40), and Ruedemann (1914) identified as the Zion Hill Formation. It is, therefore, the same as the upper unit of Dale's (1899) Ferrugenous quartzite. The Hatch Hill Formation has the same stratigraphic position and lithology as the Eagle Bridge Quartzite (Potter, 1959, p. 1658) of the Eagle Bridge quadrangle, and as exposed within the Bird Mountain slice, it is similar to Dale's (1894) description of the Bellowspipe Formation.

Ordovician System

Poultney Slate (Opl) (Keith, 1932, p. 403)

In the Pawlet quadrangle the Poultney Slate outcrops in the Edgerton window and in the slate belt region. This unit has not been recognized in the Bird Mountain or Dorset Mountain slices (Plate 1, Fig. 4).

Lithology: Several different lithologies occur within the formation. Black argillite and ribbon limestone (Schaghticoke) have been reported to occur at the base of the unit within adjacent quadrangles (Zen, 1959; Theokritoff, 1959). In the Pawlet quadrangle, brown-weathering, limy slates with occasional limestone conglomerate (Deepkill?) occur near the base of the Poultney Slate. Limestone conglomerate outcrops two miles northwest of the Town of Rupert and in the valley west of Spoon Mountain. Angular fragments in the conglomerate are predominantly light- or dark-gray limestone, and occasionally tan-colored, sandy limestone and small nodules of black phosphate (Fig. 12). The lithologic variability, smaller size, and more angular shape of the rock fragments distinguish this conglomerate from the Castleton Conglomerate of the St. Catharine Formation.

Keith's (1932, p. 403) original description characterizes the dominant lithology of the unit as:

> ". . . mainly gray slate which becomes lighter or even white on exposure. The most prominent feature of the formation is white or

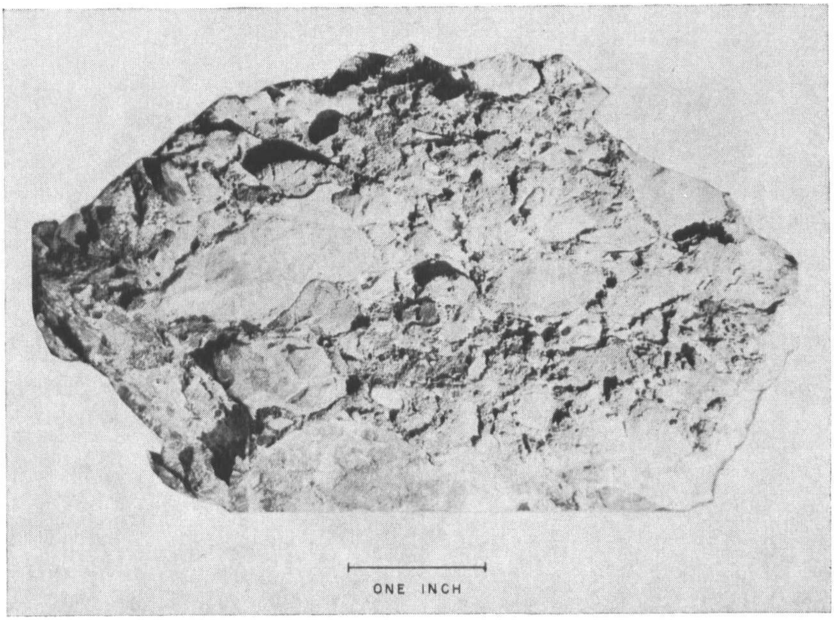


Figure 12. Limestone conglomerate (Deepkill?) from the lower part of the Poultney Slate. Note the small size, rounded shape, and greater relief of the phosphate pebbles in comparison to the limestone fragments. The matrix is mostly limestone.

light gray chert which appears in very thin seams or in massive beds a foot or so thick."

Quartzite stringers give outcrops of this formation a characteristic pinstriped appearance (Fig. 13), and they serve to distinguish the Poultney Formation from other units. Usually the quartzites are white, but they may be light green, yellow or, with increasing carbonate content, a light brown. Within the Castleton quadrangle, the quartzites show abundant graded bedding and channel filling (Zen, 1959, p. 28).

Generally, the argillite matrix of the Poultney Slate is gray-blue, but the color may grade rapidly to green or black. A single specimen may have irregular patches of various shades of green and gray. Green and red argillites occur generally at the top of the formation.

Differentiation of the Poultney Slate from the St. Catharine Formation is difficult wherever the characteristic dark-gray color and pinstripe are absent. In fact, some of the commercial slates which Dale (1899) mapped as Lower Cambrian are considered to be Poultney Slate. The most ex-

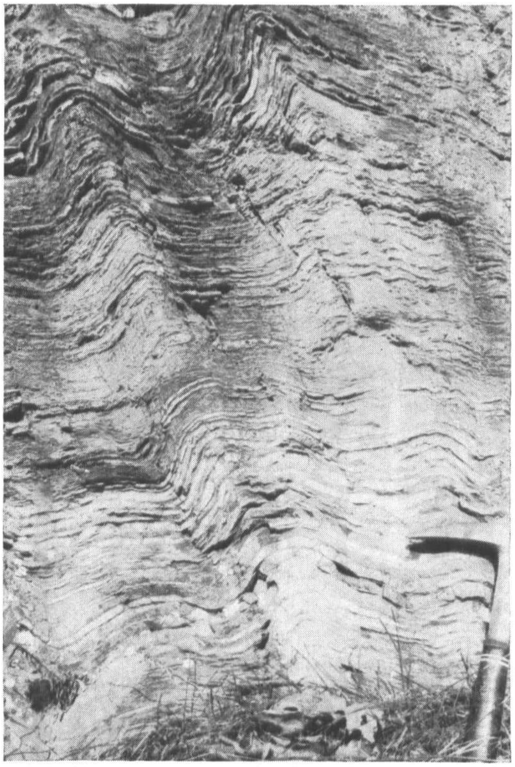


Figure 13. Outcrop of Poultney Slate on the south side of Indian Hill. Pinstriped red argillite near the top of the Poultney which shows minor folds. Note the fracture cleavage in the argillite beds, and the spacing of cleavage planes in relation to the abundance of quartzite stringers.

tensive changes were made in assigning the roofing slate at Bullfrog Hollow to the Poultney Formation for the following reasons:

1. A thin black slate (West Castleton?) separates the slates in question from purple and green pelite of the St. Catharine Formation.
2. The upper contact of the black slate (West Castleton?) is marked by a pyrite zone which gives the surface the appearance of an unconformity.
3. Irregular dark markings of an organic (?) origin are common both to these slates and to known Poultney Slate farther to the west. These markings, however, occur to a lesser degree in slates of the St. Catharine Formation.

4. The lowest 10 to 15 feet of the slate in question contain quartzites very similar to those of the Hatch Hill and Poultney formations.
5. Very thin seams of brown-weathering, limy sands gives the unit a subdued pinstripe appearance.
6. The upper contact of the slate is apparently gradational with red slates of the overlying Indian River Formation.

Thickness and contacts: The Poultney Slate is conformable with both the underlying Hatch Hill and overlying Indian River units. Quite frequently the Hatch Hill Quartzite is missing, and in such instances, Poultney Slate rests upon Lower Cambrian units (Fig. 6). An irregular pattern of deposition and non-deposition rather than localized unconformities best explains the sporadic occurrence of the Hatch Hill Quartzite. That such a period of limited deposition did occur is suggested by the occurrence of small phosphatic pellets in the angular limestone conglomerate (Deepkill?) near the base of the Poultney Formation (Fig. 12). An angular unconformity between the overlying Indian River Slate and the Pawlet Graywacke (Op) truncates the metasediments, so that the thickness of the Poultney Formation varies from zero to approximately 700 feet.

Type locality, age, and correlation: The Poultney Slate was named by Keith (1932, p. 403) for: "... good exposures in the Town of Poultney at the boundary of New York, 7 miles southeast of Castleton."

The basal Ordovician age has recently been established by graptolite discoveries from Zones 2, 3, and 4 within the adjoining Thorn Hill and Granville quadrangles (Berry, 1959, p. 61). However, graptolites collected from the highest beds of the Poultney are characteristic of Zones 11 and 12 (Berry, 1959, p. 61), and therefore the top of the formation is Middle Ordovician.

The Poultney Slate includes rocks which Dale (1899) named Hudson River white beds, thin quartzites, and shales. Ruedemann's (1914) Schaghticoke Formation, Deepkill Formation and perhaps the lowest part of the Normanskill Formation are probably equivalent to this unit. The Shelburne, Bascom and Beldens marbles of the Vermont Valley sequence are of similar age, and the Moretown and Cram Hill formations of the Eastern sequence are comparable both in age and lithology to the Poultney Slate (Fig. 5).

Bascom-Beldens Formations (Ob)

Marbles of the Vermont Valley sequence are described because they occur in Taconic terrain within the Pawlet and Purchase windows

(Plate 2). Marbles exposed in the two windows are identical with rocks which Thompson (1959, p. 75) describes as follows:

> "Bascom formation: Blue-gray, locally white marble characterized by dolomitic mottling and minor beds of gray- or orange-weathering dolomite. Minor rock types include thin beds of black phyllite and cross-bedded calcareous sandstone." (Thompson, 1959, p. 75).
>
> "Beldens formation: Blue gray and white marble alternating with beds of gray- or orange-weathering dolomite that may be several feet thick." (Thompson, 1959, p. 75).

Age, correlation and thickness: In the eastern portion of the Pawlet quadrangle the Bascom and Beldens formations vary in thickness from 350 to 450 feet (Thompson, 1959, p. 75). A thickness of 200 is a minimum figure for the marbles in the two windows, and the lower contact of the Bascom Formation is not exposed.

Fowler (1950, p. 28–30) suggests that the Bascom Formation is equivalent to part of the Beekmantown (Canadian), and that the Beldens Formation is Chazyan (Fig. 5). In so far as it is possible to determine from limited paleontologic evidence, both marbles are equivalent to part of the Poultney Slate. Significantly, neither the Poultney nor the Bascom and Beldens formations appear to change lithology as they converge in the central part of the quadrangle (Plate 2).

Paleontology: Crinoid columnals occur within the marbles outcropping in the Pawlet window (Thompson, personal communication).

INDIAN RIVER SLATE (O_i) (Keith, 1932, p. 403)

Indian River Slate outcrops in the synclines of the western slate belt region, and they are exposed on Town Hill and near Pawlet. The Indian River Formation has not been identified in the Bird Mountain or Dorset Mountain slices.

Lithology: The formation includes commercial red slates, deep blue-green slates, and minor amounts of interbedded light-colored quartzite and black slate. Flow cleavage is well developed within the slate lithology, and in most places it is the most prominent planar surface. Bedding can usually be seen either as thin black seams of slate or as thin uncleaved quartzite beds.

Thickness: The thickness varies from zero to 80 or 100 feet. The underlying Poultney Slate grades upward into the fine-grained, red and green slates of the Indian River. An angular unconformity separates the Indian River from the overlying Pawlet Graywacke.

Type locality, age, and correlation: Indian River Slate was named by Keith (1932, p. 403) for the type locality at: "... the Indian River, a few miles south of Granville, New York, where several red slate quarries are located on the banks of the stream." Keith's correlation chart (*op. cit.*, p. 360) shows the Indian River to be Chazyan, whereas fossils recently discovered within the adjoining quadrangles (Berry, 1959, p. 61) are characteristic of the twelfth graptolite zone of the Ordovician (Fig. 5). Worm trails (?), as subrounded sinuous ridges up to 4 mm. in diameter, were found at two localities, on the surface of thin quartzites. These markings are unrelated to cleavage direction or crinkle axes and are probably of organic origin. They seem to be restricted to bedding surfaces of the quartzites.

The Indian River red slate correlates with the lower Normanskill Formation, the Ira Formation and Walloomsac black slates (MacFadyen, 1956) of the Vermont Valley sequence, and perhaps the Cram Hill Formation of Eastern Vermont (Fig. 5). Potter (1959, p. 1658) reports the occurrence of a Normanskill red slate in the Eagle Bridge quadrangle some 30 miles to the southwest.

Angular Unconformity

In the Pawlet quadrangle an angular unconformity lies below the Pawlet Graywacke (Op) within Middle Ordovician rocks (Fig. 6, 14). The unconformity can be traced northward into the Castleton quadrangle (Zen and personal observation), but its presence to the south and west has not been recognized. Fowler (1950) and Thompson (1959) report a similar angular unconformity within the Middle Ordovician units below the Ira Slate of the Vermont Valley sequence (Fig. 6).

The erosion interval does not span much time, as rocks on either side of the surface contain graptolites from the same zone (Berry, 1959, p. 62). However, localized erosion during this time interval must have been severe, because in places Middle Ordovician Pawlet Graywacke rests upon Lower Cambrian St. Catharine slates (Fig. 6).

Pawlet Graywacke (Op) (new name)

The Pawlet Graywacke outcrops along the western border of the quadrangle, east and southeast of Lake St. Catharine, and in synclines within the Edgerton Window (Fig. 4, Pl. 1).

Lithology: Approximately 70 percent of the Pawlet Formation is massive, tan-weathering, fine-grained metagraywacke. Beds of graywacke, up to 6 feet thick, alternate with ½ inch to 2 feet of gray-black slates. The

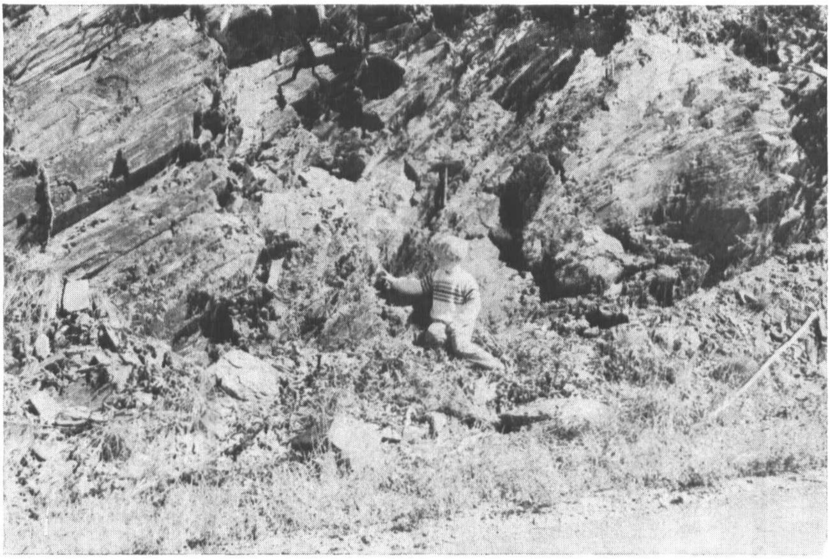


Figure 14. Angular unconformity between black slates of Pawlet (right) and Indian River (left), exposed near bridge over Poultney River at East Poultney (Castleton quadrangle). Graptolites from both slates are of zone 12. Hand of the child and hammer are on the unconformity.

base of the formation is marked by a 2- to 12-foot fossiliferous black slate. To the east, near Tadmer Hill, the basal beds are limy and at a few places outcrop as an impure limestone. Sand-size particles of white feldspar, glistening dark quartz, and rock fragments stand out against the dull, blue-gray groundmass on a fresh surface. Composition of the graywacke is quite variable, but generally there is approximately 25 to 35 percent subangular quartz, 20 percent feldspar, and 40 to 50 percent detritus and clay-sericite matrix and cement. Cleavage in fragmental slate detritus may not be oriented with cleavage of the graywacke bed. Carbonate content varies, but frequently the graywacke effervesces slightly in dilute hydrochloric acid.

In the massive graywacke, fracture cleavage may be the only planar structure. In such instances, differential weathering of feldspar, clay spots, or quartz grains may indicate the trend of bedding. Occasionally, graded bedding is well developed, and coarse, sand-size particles in graywacke beds grade upward into black slate. Generally, the graywacke beds are fine grained throughout, showing little change in grain size or gradation into the overlying slates. In some of the fine-grained graywacke,

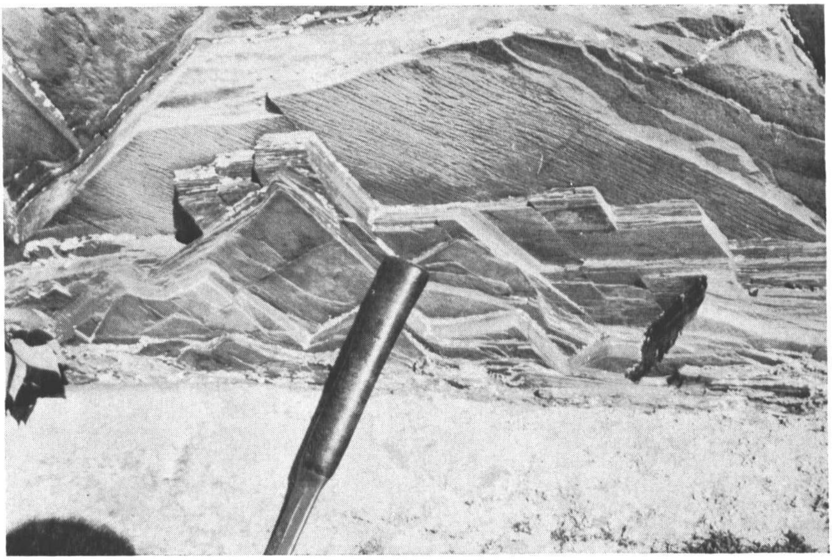


Figure 15. Underside of Pawlet Graywacke bed. Bed dips down and away from the observer. Linear markings on the base of the graywacke bed show current direction. Note the difference of joint spacing in the different lithologies.

however, rudimentary bedding is found at the top of the graywacke unit.

Near South Poultney, linear markings were found on the bottoms of three graywacke beds (Fig. 15). These marks, which are orientated in a northwest-southeast direction, probably parallel paleocurrent directions.

Thickness: The Pawlet Graywacke is the youngest metasediment within the Taconic sequence of the Pawlet quadrangle. It overlies the Middle Ordovician unconformity and is about 300 feet thick. In the northern Taconics, Dale (1899) estimated that the thickness of the unit is more than 500 feet.

Type section, age, and correlation: The Pawlet Graywacke is named for rocks exposed in Pawlet Township. Excellent outcrops of the slate, graywacke, and the basal fossiliferous slate occur approximately three-fourths of a mile west of Rock Hill. Readily accessible exposures are near the railway depot in the town of West Pawlet.

Fossils found in the basal slates are indicative of the *Climacograptus bicornis* graptolite zone, and therefore are upper Normanskill Middle Orcovician (Fig. 6). Lists of fossils identified by W. B. N. Berry from localities shown on Plate 1 follow:

Locality 1. *Glyptograptus euglyphus*, *Dicellograptus* sp.

Locality 2. *Dicellograptus sextans*, *Dicellograptus* sp., *Didymograptus sagitticaulis*

Locality 3. Biserial forms one of which is a *Hallograptus?*, the other two are *Glyptograptus teretiusculus?*—and some pieces look like parts of a dicellograptid or didymograptid

Locality 4. *Dicellograptus intortus*, *Didymograptus sagitticaulis*, *Glyptograptus teretiusculus*, *Cryptograptus tricornis*

Locality 5. *Dicellograptus sextans*, *Dicellograptus* sp., *Glyptograptus teretiusculus*, *Glyptograptus euglyphus*

Locality 6. Fossils collected every foot from base of exposure. (Shumaker, 1959, p. 67)

1. *Dicellograptus sextans*, *Dicellograptus gurleyi*, *Dicellograptus intortus*, *Climacograptus eximius*, *Dioymograptus sagitticaulis*
2. *Dicellograptus sextans*, *Dicellograptus* sp., *Climacograptus parvus*, *Dicranograptus spinifer*, *Dicranograptus ramosus*
3. *Dicellograptus divaricatus*, *Dicellograptus* sp., *Climacograptus* sp., *Orthograptus calcaratus* var. *incisus*, *Glyptograptus teretiusculus*
4. *Dicellograptus sextans*, *Dicellograptus divaricatus*, *Climacograptus parvus*, *Glyptograptus teretiusculus*, *Nemagraptus exilis*, *Nemagraptus gracilis*, *Glyptograptus euglyphus*, *Cryptograptus tricornis*, *Orthograptus calcaratus* var. *acutus*, *Climacograptus bicornis*, *Didymograptus sagitticaulis*
5. *Dicellograptus sextans*, *Dicellograptus intortus*, *Dicellograptus divaricatus* var. *salopiensis*, *Climacograptus eximius*, *Climacograptus parvus*, *Climacograptus bicornis* var. *peltifer*, *Glyptograptus teretiusculus*, *Glyptograptus euglyphus*, *Hallograptus mucronatus*, *Orthograptus calcaratus* var. *incisus*, *Didymograptus sagitticaulis*, *Dicranograptus nicholsoni* var. *diapson*
6. *Dicellograptus sextans*, *Dicellograptus intortus*, *Dicellograptus divaricatus*, *Climacograptus parvus*, *Glyptograptus teretiusculus*, *Cryptograptus tricornis*, *Hallograptus mucronata*, *Orthograptus*
7. *Climacograptus parvus*, *Dicellograptus intortus*, *Glyptograptus teretiusculus*, *Hallograptus mucronatus*, *Cryptograptus tricornis*, *Nemagraptus exilis*, *Didymograptus sagitticaulis.*
8. *Dicellograptus intortus*, *Climacograptus parvus*, *Climacograptus eximius*, *Hallograptus mucronatus*, *Cryptograptus tricornis*

9. *Dicellograptus sextans*, *Dicellograptus smithi*, *Climacograptus parvus*, *Climacograptus* sp., *Nemagraptus exilis*, *Nemagraptus gracilis*, *Glyptograptus teretiusculus*, *Dicranograptus ramosus*

The Pawlet Graywacke is considered equivalent to Fowler's Normanskill Graywacke (1950, p. 56). However, Fowler's Normanskill is considered inappropriate since it includes rocks of older and lithologies different from the type Normanskill far to the south. The Pawlet Graywacke is synonymous with the Hudson River Grit of Dale (1899), but this name is abandoned because of a variety of ages and lithologies assigned to "Hudson River" (Wilmarth, 1938, p. 990). The Austin Glen Formation of the area near Catskill, New York, (Ruedemann, 1942, p. 102) is lithologically similar and is apparently the same age as the Pawlet Graywacke. The term Austin Glen is not adopted because of the distance separating it from the Pawlet quadrangle. The Pawlet Graywacke is approximately the same age as the Ira and Walloomsac formations (MacFadyen, 1956) of the Vermont Valley sequence, and it may be equivalent to the Shaw Mountain Formation or Northfield Slate of the Eastern Vermont sequence. However, the Pawlet Graywacke (Hudson River Grit) is definitely not equivalent to some of the rocks which Dale (1899, Pl. 13) mapped as Berkshire Schist, and the graywacke does not show the facies change to his Berkshire Schist on Rupert Mountain (Dale, 1899, p. 176, 187, 192). If Dale's interpretation is correct, the graywacke should grade laterally into green schist in the vicinity of the western hills of the Pawlet quadrangle of Moosehorne, Haystack, and the Pattern hills, etc., and then grade back into typical Pawlet Graywacke in the Edgerton and Tadmer hills of the Edgerton window (Plate 3, Fig. 4). However, the graywacke does not change, and the green schist is considered to be the St. Catharine Formation rather than the Berkshire.

Permian System

Basic Dikes

Basic intrusive dikes which occur at several localities in the Pawlet quadrangle appear to be typical of the scores of small lamprophyric and diabasic dikes found in eastern New York and western Vermont. Dale (1899, Pl. 13) shows six dike localities in the Pawlet quadrangle, and seven new localities were found during this study (Plate 1). The dikes of the region have been referred to as augite camptonites, camptonites, diabases, and analcimites (Dale, 1899, p. 223).

These igneous rocks tend to weather rapidly and exposures are infrequent and poor. They are generally exposed in quarry walls or stream beds, but their presence in regions of thin drift may be indicated by float of dike rock. The weathered rock is tan-brown, aphanitic or porphyritic; phenocrysts of mafic minerals weather differentially to give the rock a pitted appearance. The most abundant minerals in the dikes are euhedral, zoned polyaugite, hornblende, and labradorite with minor amounts of analcite, pyrite, calcite, magnetite, ilmenite (?), chlorite, sericite, and apatite in a groundmass of euhedral to subhedral laths of polyaugite, hornblende, and labradorite.

One dike at the south end of Spoon Mountain is different. This rock is tan to light green, and resembles a fine-grained quartzite. On fresh surfaces, minute colorless laths of feldspar may be seen in an aphanitic gray-green groundmass. In thin section the rock is similar to the other basic dikes, except that the labradorite is highly sericitized, and the mafic minerals are less abundant and altered to chlorite.

Fowler (1950, p. 59) suggests that the basic dikes should be correlated "with the Monteregion intrusives, which may be correlated with the White Mountain magma series." The White Mountain plutonic, extrusive series have been assigned a Pre-Triassic, Late Permian age by Lyons (Lyons et. al., 1957, p. 540).

Quaternary System

The surficial deposits of the Pawlet quadrangle are continental Pleistocene sediments related to Wisconsin glaciation, and recent talus and alluvial deposits. These sediments are thickest in the major valleys which extend from Middletown Springs to Dorset and Lake St. Catharine to Rupert.

Unsorted deposits: Because of the mountainous topography, the most widespread deposits are those of recent talus. These deposits grade laterally from the hill slope into the valley alluvium. Only two extensive areas of ground moraine were found: east and west of Little Pond near the south end of Lake St. Catharine, and in the valley south of Burnham Hollow. Undoubtedly, additional extensive deposits of ground moraine are buried under the alluvium and talus.

Alluvial deposits: Most of the lowlands of the Pawlet quadrangle are filled with Wisconsin alluvium. An interesting and varied sequence of these deposits is found in the valley south of Middletown Springs. The deposits include kame moraine with foreswamp indicative of drainage reversal, incompletely developed kame terraces which extend southward in front of the recessional moraine, and a segmented sinuous esker in the center of the valley. Isolated deposits of this esker can be traced for over a mile along South Brook.

Southward from The Cobble, extensive high-level outwash terraces, shown on the topographic map (Pl. 1), probably coalesce with outwash terraces of the Vermont Valley near Manchester (Fig. 4). Equally as impressive, but not nearly as extensive, are the terraced outwash deposits which occur near Wells and Blossom Corners south of Lake St. Catharine. More limited deposits of outwash occur in the Indian valley, which are continuous with deposits near Rupert in the White Creek valley.

Lacustrine deposits: One minor occurrence of massive gray, lake clay was found along Flower Brook just east of the Village of Pawlet. This small deposit is probably not related to the lake sequence of the Champlain Valley.

SEDIMENTATION

The sedimentation of the Taconic rocks has not been studied in detail, and the writer is not sufficiently acquainted with the lithologies of adjoining areas to reconstruct a picture of the detailed regional sedimentary environment. However, a few suggestions will be made both about the depositional history of the sediments of the Pawlet quadrangle and about relationships of the Taconic sediments to those in adjoining areas.

St. Catharine and West Castleton Formations

The impure slates or phyllites of the lower St. Catharine Formation and the upper slates and phyllites of this unit were fine-grained, well-sorted sediments. The clays were deposited from suspension, and fine-grained sands were deposited by traction currents below effective wave base. The coarse sands of the Zion Hill and Mudd Pond (Zen, 1959, p. 1) quartzites are shallower water deposits. The presence of such coarse clastics suggests that, at certain times during the deposition of the middle and upper portions of the St. Catharine Formation, the sea floor was near or above effective wave base. The shape of the coarse deposits of the Zion Hill Conglomerate implies that this member was deposited as a series of bars on a fine sandy bottom. The limestone "pebbles" of the Castleton Conglomerate might be interpreted as having been transported by strong currents in shallow water, or that a short period of desiccation and fracturing occurred. However, this member may occur as a single continuous bed, or as a series of pod-like lenses associated with limy slates. It is more likely, therefore, that the Castleton Conglomerate was deposited as a discontinuous limestone below effective wave base, and that this material was disturbed during a violent storm or by a severe shock wave. Lochman (1956) considers that the source of the sediments was from both the west and east, and that faunas similar to those found within the Taconic rocks are indicative of an "intermediate" depositional environment (between miogeosynclinal depths 0–50′ and eugeosynclinal depths 400–600′) (Lochman-Balk and Wilson, 1958, p. 315–316).

The abrupt change from purple and green muds of the St. Catharine Formation to the black muds of the West Castleton Formation was caused by an influx of organic debris beyond the oxidizing potential of the depositional environment. The homogeneous character of the black slates indicates a continuation of weak currents and deposition of the mud from suspension. However, pebble conglomerates occurring within the West Castleton Formation to the north (Zen, 1959, p. 2) and quartzites, which occur to the southwest (Potter, 1959, p. 1658), suggest that the currents were stronger and the water was shallower in adjoining western areas. It is inferred from the occurrences of an unconformity above the West Castleton that uplift caused restriction of circulation, perhaps by silting, which brought about the accumulation of organic debris with the muds.

It would appear, then, that Lower Cambrian sedimentation occurred in relatively stable conditions below effective wave base. Occasional brief periods of shallowing brought about accumulation of coarse deposits, but it was not until the latter part of the Lower Cambrian that uplift became general, and the bottom emerged to become a low-lying land mass.

Hatch Hill, Poultney, and Indian River Formations

Hatch Hill quartzites were deposited as sand-size quartz, feldspar, and carbonate fragments which had been transported by moderately strong traction currents. As exposed to the north and west of the Pawlet quadrangle, the quartzites are cross bedded, and have scour markings which indicate local strong traction currents in shallow water. The sands and black muds were probably deposited in shallow water along the erosion surface as discontinuous channel deposits in a marine environment. Overlying ribbon limestones (Schaghticoke lithology) "show delicate cross bedding" (Zen, 1959, p. 8) and are indicative of basin deepening and deposition below any effective wave base. The Poultney Formation was deposited as mud and fine-grained, smooth-bedded sand even farther below wave base, in a depositional environment similar to that of the St.

Catharine Formation. However, the limestone conglomerates (Deepkill) of the Poultney Formation are not indigenous, and they were probably transported downslope to the site of deposition where they were rapidly dumped. Angular shape, variation of grain size, and lithologic heterogeneity are characteristic of such deposits. These carbonates generally occur along the western margin of the Pawlet quadrangle and in the quadrangles to the west. The abundant current markings in the units exposed in these adjoining quadrangles suggest a western source area for the carbonates and distributive currents. An increase in clean sand content of the Poultney Formation in the most easterly exposures may indicate an eastern source area.

Homogeneous argillite of the overlying Indian River Slate suggests relatively quiet water deposition. The red color of these sediments is a reflection of deposition in a highly oxidizing environment, perhaps even subareal. Black mud interfingering with the red and green argillite most probably reflects an interfingering of sedimentary phases in the geosyncline. Graptolites found within the black slates indicate that the depositional site of the mud was marine. The brief erosion interval above the Indian River Slate and within the Middle Ordovician sediments, is a further indication of the tectonic instability which marks the close of the Ordovician.

Thus Upper Cambrian sedimentation was initiated by shallow water deposition, followed by prolonged deep water accumulation in the lower Ordovician, and terminated by uplift in the Middle Ordovician. Generally, the sediments were similar to those deposited in the Lower Cambrian, and the general emergence at the close of the cycle was rapid and of brief duration. The sharp angularity of the overlying Pawlet Graywacke indicates that, although the uplift was brief, it was vigorous.

Pawlet Graywacke

Deposition of Pawlet Graywacke sediments marks the end of smooth traction transportation and sorted sedimentation. Graywacke interbedded with mud is generally characteristic of transport by turbidity currents, tectonic instability, and a rapidly subsiding basin. Deposition must have occurred sporadically and rapidly, resulting in unsorted sediments, interposed with periods of quiescent black mud accumulation. The occurrence of unoriented slate particles in the graywacke suggests that the eastern unstable region was a source area. The stable Adirondack massif of igneous and high-grade metamorphic rocks is not considered a likely source for the slate particles of the graywacke.

It is extremely important to note that if the environment of deposition of the late Middle Ordovician is interpreted correctly and the late Ordovician is marked by tectonic activity and uplift, then the green phyllites which Dale mapped as Berkshire Schist (1899, Pl. 13) are completely out of phase. Rather, these rocks are in phase with the Lower Cambrian St. Catharine Formation.

Regional Applications

The character of the Cambrian and Lower Ordovician sediments suggests that they were deposited, for the most part, below effective wave base from suspension or by smooth traction currents. These sediments and the inferred ecologic habitat of the fossil assemblages are indicative of a depositional environment which was neither miogeosynclinal nor eugeosynclinal, but "intermediate" in character (Lochman-Balk, Wilson, 1958, p. 315). Two depositional sites have been suggested as appropriate for the development of an "intermediate" environment. A suitable environment would be a basin on the Champlain carbonate shelf, which fringes the Adirondack massif. The other favorable area would be seaward to the east of the Champlain shelf (Fig. 1) between the cratonic shelf and the Magog eugeosyncline (Kay, 1951).

If the Taconic rocks were sediments deposited within the elongate basin on the Champlain shelf, facies changes should be apparent in both the carbonate and pelitic sequences near the margin of the intracratonic basin. On the other hand, if the Taconic sediments were deposited between the cratonic shelf and the eugeosyncline, the Taconic sequence must have been thrust into its present position on the shelf facies.

The western margin of the Taconics is bounded by a series of thrust faults (Walcott, 1888), and complementary facies changes are not apparent in the carbonates or argillites. The facies changes along the eastern margin of the intracratonic basin have been interpreted by some workers to be hidden under Upper Ordovician rocks mapped, for the most part, as Berkshire Schist. According to Dale (1899, Pl. 13; 1912, Pl. 1), the eastern two-thirds of the Pawlet quadrangle is composed of Berkshire Schist. The proposed facies changes in the Cambrian and Lower Ordovician units are logical if an adequate amount of Berkshire Schist is present to cover the transitional and interfingering lithologies. However, the rocks which Dale (1899, 1912) mapped as Berkshire Schist in the Pawlet quadrangle are now mapped as typical lithologies of the slate belt (Plates 1 and 2). Designation of the rocks within the eastern portion of the Bird Mountain slice as Cambrian Taconic units is certainly still open to question, but the rocks exposed within the Edgerton window (Fig. 4) are definitely the same as Cambrian and Ordovician rocks found to the west. Therefore, the argillites of the Taconics and the carbonates of the Vermont Valley sequence are mapped to within two or three miles of one another. Neither sequence shows even the slightest tendency toward the necessary gradational facies changes.

All of the available paleontologic evidence (Lochman, 1956; Zen, 1959; Thompson, 1952) and the position and character of the unconformities indicate that the two sequences are time equivalent.

Part of the objection to the proposed facies changes might be eliminated if age assignments of certain units were incorrect, and if more comparable lithologies were equated. Only two units readily lend themselves to such readjustment; the Nickwacket of the Valley sequence and the St. Catharine of the Taconics (Fig. 5). Both units might be correlated and considered to be Lower Cambrian. The obstacle to this minor readjustment is that it requires erosional truncation of the upper 2000 feet of the Lower Cambrian carbonates and quartzites across the two or three miles which separate the sequences, for the Lower Cambrian carbonate crops out all along the Valley but is completely absent in the Taconics. Furthermore, the truncated portion of the section must occur only under the eastern Bird Mountain slice (Fig. 4 and Pl. 1). This is not readily acceptable, inasmuch as neither sequence has an angular unconformity at the top of the Lower Cambrian rocks. Also, such a solution requires great erosion of the marginal quartzites and carbonates and the central deposits of the basin, without erosion of the adjacent shelf areas.

Additional evidence opposed to deposition of the Taconics *in situ* is shown by the juxtaposition of time equivalent Poultney Slate in the Edgerton window, and Bascom-Beldens marbles in the Pawlet and Purchase windows (Plates 1 and 2). Not only do these units lack the necessary complementary facies changes, but the Poultney Slate apparently has decreased in lime content in the Edgerton area.

Upper Middle Ordovician units in both sequences include comparable lithologies that are suitable facies equivalents. Black slates of the Ira, Pawlet, and Indian River formations may be the same phase of deposition. However, the likelihood that the Pawlet and Indian River units were deposited in place is dispelled by their abrupt disappearance at the western edge of the Eastern Bird Mountain slice. Thus, the west margin of the slice marks the termination of all Taconic units that are exposed in areas outside of this slice.

If the sedimentary data are not consistent with the interpretation of intracratonic basinal deposition, it should be determined whether these data may fulfill the requirements for deposition at a site between the Champlain shelf (miogeosyncline) and the Magog trough (eugeosyncline). A test of the reliability of placing Taconic sedimentation at this latter site is to compare the Taconic rocks with indigenous units (Eastern Vermont sequence, Fig. 1). This comparison has to be made on published descriptions of the Eastern Vermont sequence, as the writer has not studied these rocks.

The Pinney Hollow (Fig. 5) is lithologically similar to and has been assigned the same approximate age as the St. Catharine Formation (Osberg, 1952, p. 22).

Pinney Hollow Formation "1000–1500 feet":

> "Pale-green quartz-muscovite-chlorite schist and albite-quartz-muscovite-chlorite schist" (Osberg, 1952, p. 22).

The Ottauquechee Formation (Fig. 5) is comparable in lithology to parts of the St. Catharine and West Castleton formations. It has generally been assigned a Cambrian age.

Ottauquechee Formation 1400 feet:

> "Graphitic quartz-muscovite schist and quartzite, dark-green albite-epidote-carbonate-chlorite schist." (Osberg, 1952, p. 22).

Lithologies comparable to those found in the Stowe Formation are not present above the West Castleton black slate. This is expected, inasmuch as an unconformity overlies the West Castleton Formation and an undetermined thickness of Lower and Middle Cambrian rock is missing. The absence of Stowe equivalent units in the Taconic sequence can, therefore, be attributed to erosion of the shelf and marginal regions, while sediments of the Stowe Formation were being deposited in the deeper portion of the basin. It may be, however, that the lower part of the Poultney Slate is equivalent to the Stowe Formation, for the easternmost exposures of the Poultney Formation have horizons of green pinstripe phyllite near the base.

Stowe Formation 900 feet:

> ". . . in many respects similar to the Pinney Hollow. . . . Locally the rock consists of bands 1 to 2 millimeters thick of quartz and albite separated by chlorite." (Osberg, 1952, p. 65–66).

Descriptions of the Moretown (Fig. 5) are similar to the easternmost exposures of the Poultney Formation.

Moretown-Cram Hill: 9000–10,000 feet:

Moretown:

"Gray-green micaceous quartzite, quartz-muscovite-chlorite schist and greenstone schist." (Thompson, 1952, p. 40).

"Commonly called the pinstripe." (Konig, 1959, p. 17).

Cram Hill Formation:

"Quartzite and gray to black, rusty weathering phyllite or schist. . . ." (Thompson, 1952, p. 40).

"A Lower Ordovician age for its lower boundary (of the Moretown) has been proposed by most recent workers in central Vermont. Thus the Moretown ranges in age from Lower Ordovician to approximately Middle Ordovician." (Konig, 1959, p. 31).

Both the age and lithology of the Poultney Slate indicate a direct correlation between it and the Moretown and Cram Hill formations.

If the correlation of the Poultney and Moretown-Cram Hill formations is correct, then the Pawlet Graywacke and Indian River Slate are equivalent to the Cram Hill and Partridge formations.

In direct contrast to the intracratonic site, the rocks of the Eastern Vermont sequence and Taconic sequence are similar, and the sediments of the Taconics are compatible with the depositional environment of the Magog trough. The discrepancies between the sequences, disproportionate thicknesses and amounts of igneous material, may be attributed to the marginal position of the Taconics.

The above discussion is not intended to prove the existence of a Taconic klippe, but it does suggest that an environment suitable for deposition of Taconic sediments, in proper sequence, did exist just seaward of the Champlain miogeosyncline.

STRUCTURAL GEOLOGY

Detailed mapping of the Taconic metasediments is complicated by 1) large-scale folding and overturning prior to major metamorphism and folding, 2) scarcity of physical criteria associated with faulting, 3) repetition and widespread occurrence of similar lithologies throughout the section, and 4) local variations of structural features. Fortunately, as is to be expected in the green-schist metamorphic facies, bedding and bedding schistosity are present in most of the formations.

Bedding and Bedding Schistosity

Most Taconic rocks show bedding as color, composition and textural changes. However, flow cleavage often obscures bedding that is marked by changes in grain size or chemical content. Color variations, other than green and red shades in the slates, may be fairly reliable indicators of bedding, and in the Pawlet quadrangle they usually reflect changes in carbonate or organic content. The walls of slate quarries, when studied from a distance, sometimes show faint lines which follow the trend of bedding. However, care must be exercised in deciding the attitude of bedding in any unit, because color, composition, and textural changes may result from either metamorphic segregation banding, or less frequently, from deformed calcite-filled joints.

During folding, a schistosity is developed along planes of bedding by slippage between adjacent beds. This bedding schistosity is well developed in banded phyllites, banded quartzites and alternating slates and graywackes. Bedding schistosity is often crinkled by small folds and wrinkles, and the associated fracture cleavage parallels the axial planes of these crinkles. Therefore, the presence of bedding schistosity may be recognized in the Pawlet area by the presence of crinkles. In most rocks of the Pawlet quadrangle, bedding schistosity is the commonest and most readily recognizable structural feature, and can be used the same as bedding in determining the attitude of folds.

Flow Cleavage

Flow cleavage is formed by general recrystallization of platy minerals sub-parallel to axial planes of folds. Abundant closely spaced cleavage planes and thorough recrystallization of these minerals may completely obscure bedding. In the Pawlet quadrangle, flow cleavage is restricted to units that initially had a high clay content. With decrease in the amount of clay minerals or the intensity of folding, flow cleavage becomes less perfect and eventually will be replaced by fracture cleavage.

Fracture Cleavage

Axial-plane fracture cleavage is developed along shear planes and is independent of platy mineral arrangement. Usually this type of cleavage is more widely spaced than flow cleavage, and does not pervade the entire rock. Perhaps eighty percent of the cleavage found in the Pawlet quadrangle is axial plane fracture cleavage developed during the major deformation.

As is the case with flow cleavage, fracture cleavage in similar lithologies is intensified in the eastern portions of the quadrangle. Generally, the impure and pinstriped argillites have well-developed, closely spaced fracture cleavage throughout the eastern part of the quadrangle. Similar rocks have distinctly less numerous, widely spaced cleavage planes in the west. Commonly, in the pinstriped rocks which consist of interbedded argillite and quartzite stringers, the argillite has well-developed cleavage and the quartzite has no cleavage.

In general, graywackes have poorly developed and widely spaced fracture cleavage. Cleavage in the interbedded slates is well developed and is sub-parallel to the axial planes of associated folds.

Metamorphic segregation banding is common in the pinstriped and impure argillites throughout the central and eastern portions of the quadrangle. Osberg (1952, p. 87) suggested that segregation banding is caused by the growth of albite in the troughs and crests of minor folds.

Lineation

The most common linear element measured in the Taconic rocks is the trend and plunge of crinkle axes (Plate 2). Inasmuch as crinkle axes parallel contemporaneous fracture cleavage, its relative position on the related, next larger fold can be determined.

Shearing and stretching of quartzite and limestone beds on limbs of small folds form "b" tectonic lineations (Fig. 17). Slickensides and mineral streaming on bedding and bedding schistosity form "a" tectonic lineations. The angle between the trend of the slickenside and the dip of the bed approaches the amount of plunge of the genetically related fold.

Folding and Faulting

The post Taconic deformation created widespread and abundant structural and metamorphic features in Taconic rocks that effectively obscured the structural pattern of all earlier folding and faulting. However, it was found that some structural features of the early folding and faulting could be differentiated in a few critical areas, and therefore a sequence of deformation can be reconstructed.

North-south regional structural trends throughout most of Vermont were established in the late Precambrian or very early Lower Cambrian, by a major geosyncline of deposition. There was no major deformation in the Cambrian, but a broad uplift accompanied by erosion produced a regional disconformity in Middle Cambrian (Fig. 6).

Ordovician Taconic Deformation: Massive inversion of the Cambrian section, as exposed in the Bird Mountain slice, is the principal evidence of the first major, early deformation. Zen (1959) gave structural, sedimentary and stratigraphic evidence that a major portion of the Taconic sequence of the Castleton quadrangle, immediately to the north of the

Pawlet quadrangle, has been inverted. Shumaker (1959) suggested that the stratigraphic sequence of the "high Taconics" of the Pawlet quadrangle has been inverted. In the Pawlet quadrangle the formations exposed in the Bird Mountain slice are considered to be Cambrian, for the following reasons:

1. The stratigraphic sequence is similar to that of Cambrian units of the slate belt region.
2. The depositional environment of the rocks is similar to that of the Cambrian, and dissimilar to the depositional environments of the Upper Ordovician.
3. Evidence of thrust faulting is found under the Bird Mountain slice.
4. Structural evidence for inversion of the stratigraphic sequence, is found in the Pawlet and Purchase windows.
5. Stratigraphic, sedimentary, and structural evidence for similar inversion is found within the adjoining Castleton quadrangle (Zen, 1959).

The larger size and general recumbency of the Taconic fold (or folds) is borne out by the large areal extent and low dip pattern of the inverted sequence (Plate 1 or 2). Zen described the Taconic fold as:

> "The anticline is interpreted as being a recumbent fold, overturned from the east, now lying on its side with nearly horizontal axial plane and nearly north-south axis." (Zen, 1959, p. 3).

Data assembled by the writer from work done within the Pawlet quadrangle are in agreement with Zen's interpretation. The Taconic fold must have developed after deposition of the Hatch Hill (as this is the youngest recognizable unit of the inverted sequence) and before the development of the Taconic allochthone. The age is determined as post-Upper Cambrian, and pre-Upper Ordovician. The lithologies of the sediments deposited during the Middle Ordovician, and the probability that formation of the Taconic fold was closely followed by the development of the Taconic allochthone suggest an upper Middle Ordovician age.

Initially a massive segment of the Taconic fold was thrust westward from the Green Mountain site over the Champlain shelf sediments. Such thrusting presumably took the form of a detachment sheet since basement thrust faults are absent in Precambrian rocks to the east. It is also important to note in this respect that the emplacement of the Taconic rocks occurred only after the Middlebury synclinorium started to form. At that time there was a suitable slope to permit gravity

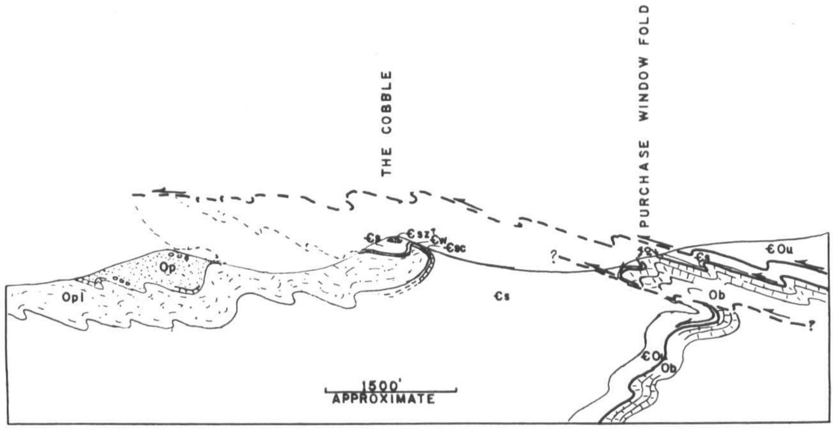


Figure 16. Purchase window structure

to drive this thin sheet westward to its present position. In the later stages of emplacement, the western portions of the allochthone slowed, and the more rapidly advancing eastern portions of the thrust sheet sheared over the frontal portion. Such major shear zones were probably localized wherever the underlying marble (and basement) made a sharp downbend such as along the line from the Purchase window through the Pawlet window. Once a major slice formed, further deformation was largely relieved along this new thrust. Eventually, the allochthone consisted of several imbricate thrust slices which overrode the early, lower portion of the allochthone (Fig. 4).

The folded sequence in the Purchase window supports the above interpretation (Cross section C, and Fig. 16). The Pawlet Graywacke, Poultney Slate, West Castleton Slate, and St. Catharine Formation are part of the overturned limb of an anticline. The inversion is strengthened by the stratigraphic sequence and graded bedding in the Pawlet Graywacke. If the structural interpretation is correct, the St. Catharine Formation overlies the Ordovician Valley marbles in the Purchase window anticline, and the Taconic thrust is a reality.

It has been suggested (Dale, 1912) that the green phyllites of The Cobble are Berkshire Schist (post Ira Slate, probably Upper Ordovician) rather than green phyllites of the St. Catharine Formation. If this suggestion is true, then the green phyllites would be younger than the Pawlet and the Poultney formations. However, correlating the green phyllites on The Cobble with Ordovician Berkshire Schist is not probable for the following reasons:

1. The green phyllites and the Pawlet Graywacke can not both overlie the Poultney Slate simultaneously.
2. Regional and local relationships preclude any sharp angular unconformity beneath the green phyllites and above the Poultney and Pawlet formations.
3. The green phyllites are folded with the Pawlet and Poultney formations. If the green phyllites are the youngest unit, then they should occur in the centers of synclines in the Edgerton window. They do not.
4. The Purchase window fold can not be a syncline because the green phyllites, together with the Pawlet, and part of the Poultney, are younger than the Bascom-Beldens marbles.
5. High-angle basement up-thrusts are *not* indicated in either the Taconic or Vermont Valley sequences. (See Fowler's 1950 map northward along strike of the Purchase-Pawlet faultline for evidence of high-angle basement fault under the Taconics.)

All of the evidence then, suggests that the Purchase fold is an anticline, and that the St. Catharine Formation is exposed on The Cobble and has a detached relationship with the underlying Vermont Valley marbles.

If this is true, the explanation for the absence of Poultney and Pawlet lithologies on the upper, normal limb of the Purchase window anticline must be that the normal limb was sheared out by the detachment fault producing the Bird Mountain slice (Cross section C of Plate 1). That is, the Purchase fold was formed and later cut by an upper slice developed in the more active eastern part of the allochthone.

The general structural pattern of the Purchase window anticline is repeated again in the Pawlet window (Cross section E). The interpretation of the structure shown in the cross section is supported by:

1. The anticlinal outcrop patterns of the Ordovician marbles and the overlying undifferentiated slate and St. Catharine formations.
2. The complementary syncline occurs to the west in the Poultney, Indian River, and Pawlet formations. Therefore, the west limb of the Pawlet anticline is an inverted sequence, and the St. Catharine Formation overlies the Bascom-Beldens marbles.
3. Cleavage-bedding relationships, three-fourths of a mile north of Spruce Top, also indicate overturning of the west limb of the Pawlet window anticline.
4. If the green phyllites bounding the marbles of the window are

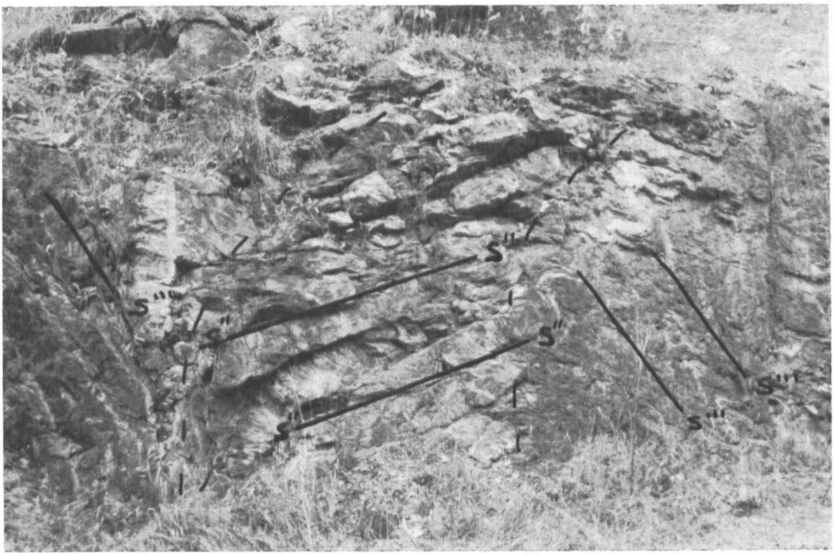


Figure 17. Sheared quartzite beds of the Poultney Formation near Haystack Mountain. Bedding dips steeply to the west. Shear surfaces (s'') and regional cleavage (s''').

Ordovician Berkshire Schist, then the Middle Ordovician Poultney, Indian River, and Pawlet formations on the west should dip under the green phyllites and reappear at the contact of the green phyllites and the Lower Ordovician, Bascom-Beldens marbles. They do not.

Additional evidence of the complexity of the allochthone is found along the western margin of the Edgerton window (Fig. 4 and Plate 2). Here, Cambrian and Ordovician formations dip under Cambrian units of the western Bird Mountain slice. This is well exposed along Route 30 near the base of Haystack mountain. Middle Ordovician graptolites were found in black slates of the Pawlet Graywacke at the base of the mountain on its southeastern side. Because the outcrop of the graywacke was limited to about four square feet, the area could not be differentiated on the map and was included in the Poultney Slate (Plate 2). The purple and green phyllites and slates within the western Bird Mountain slice, directly above the Poultney rocks, are mapped as St. Catharine Formation. The fold at this exposure along Route 30 is an anticline with Cambrian units overlying Ordovician rocks; therefore, the thrust is present. The distortion and fold pattern near the fault is shown in Figure 17. Horizontal shear surfaces (Fig. 17) are presumably related to thrust faulting, and the east-dipping cleavage is related to later, post Taconic metamorphism and folding (Figs. 18 and 19).

The Cambrian rocks, which occur above the thrust in the western Bird Mountain slice, can be traced northward into the Castleton quadrangle where they are exposed above marbles of the Vermont Valley sequence (Fig. 4).

The eastern and western portions of the Bird Mountain slice can be traced to within less than a mile of one another in the Castleton quadrangle (Fig. 4). The structural similarity of the two portions of the slice is shown in the Pawlet quadrangle by geomorphic and geologic relationships (Plate 1). Generally, the hills are formed by rocks of the slice, whereas the valleys are windows. The low-dip outcrop pattern and the sheet-like pattern of the Bird Mountain slice is distinctly different from the folded pattern of the underlying units. All Taconic metasediments were closely folded by the post-Taconic deformation. However, these folds are readily mappable only where the stratigraphic sequence is distinctive, as on the upper limb of the Taconic fold (Plate 2).

The emplacement of the Taconic allochthone is generally considered to have concluded by the Later Ordovician (Fowler, 1950). Undisturbed autochthonous pre-Upper Silurian rocks rest upon folded and faulted Cambro-Ordovician metasediments in the southern and central Hudson River Valley.

Zen (1959, p. 5) reports the occurrence of Taconic blocks in autochthonous black slates near the western periphery of the allochthone. These black slates are probably upper Middle Ordovician, Trenton. Presumably the Taconic allochthone moved into place over Trenton formations. Inasmuch as the Indian River Slate and the Pawlet Graywacke are approximately Trenton age, these two formations may have been deposited just prior to the emplacement of the allochthone. If Zen (1959, p. 5) is correct in his interpretation that the "Taconic sequence, perhaps as soft unconsolidated material, is believed to have moved into a Trenton sea in which black mud was depositing," then the Pawlet and Indian River formations may have been deposited during the development of the allochthone. The depositional environment of the Indian River Formation, the angular unconformity, and the abrupt change in source and character of the Pawlet Graywacke strongly suggest tectonic activity during Trenton time.

Post-Taconic Deformation: The major structural feature of southwestern Vermont is the south-plunging Middlebury synclinorium (Fig. 1). This large downfold, which is unsymmetrical and overturned toward

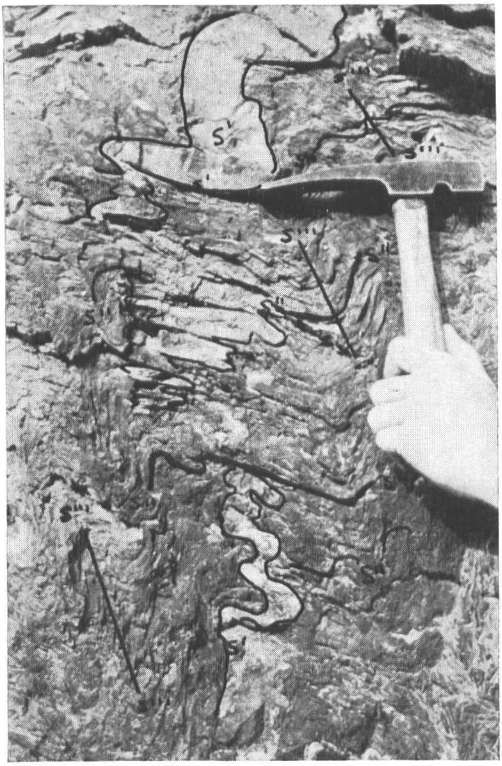


Figure 18. Same outcrop as Figure 17. Bedding (s′), crinkled shear surface (s″), regional fracture cleavage (s‴).

the west, was initiated during the Ordovician, Taconic deformation. However, subsequent post-Taconic folding and metamorphism effectively intensified the synclinorium, and the resulting structural features in the Taconic rocks transect and tend to obscure any early structural patterns.

Folds and axial plane fracture cleavage associated with this last major deformation are abundant in Taconic rocks (Plate 1). The largest of these folds in the Pawlet quadrangle are present as long narrow, north-south trending structures that are approximately three-fourths of a mile wide. In the commercial slate belt, axial planes and the cleavage of the major synclines and anticlines dip 60 or 70 degrees to the east. Farther east, the folds are progressively further overturned, so that along the eastern edge of the quadrangle they are nearly recumbent. The largest of the secondary folds is approximately 100 to 150 feet across. Folds in all

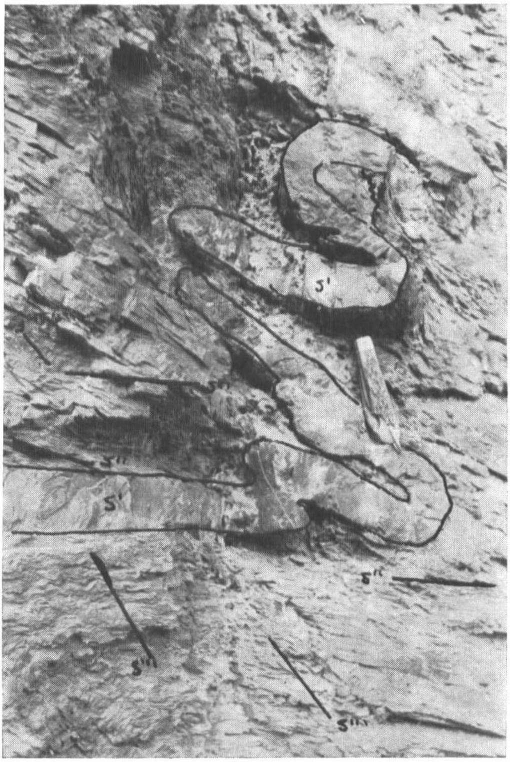
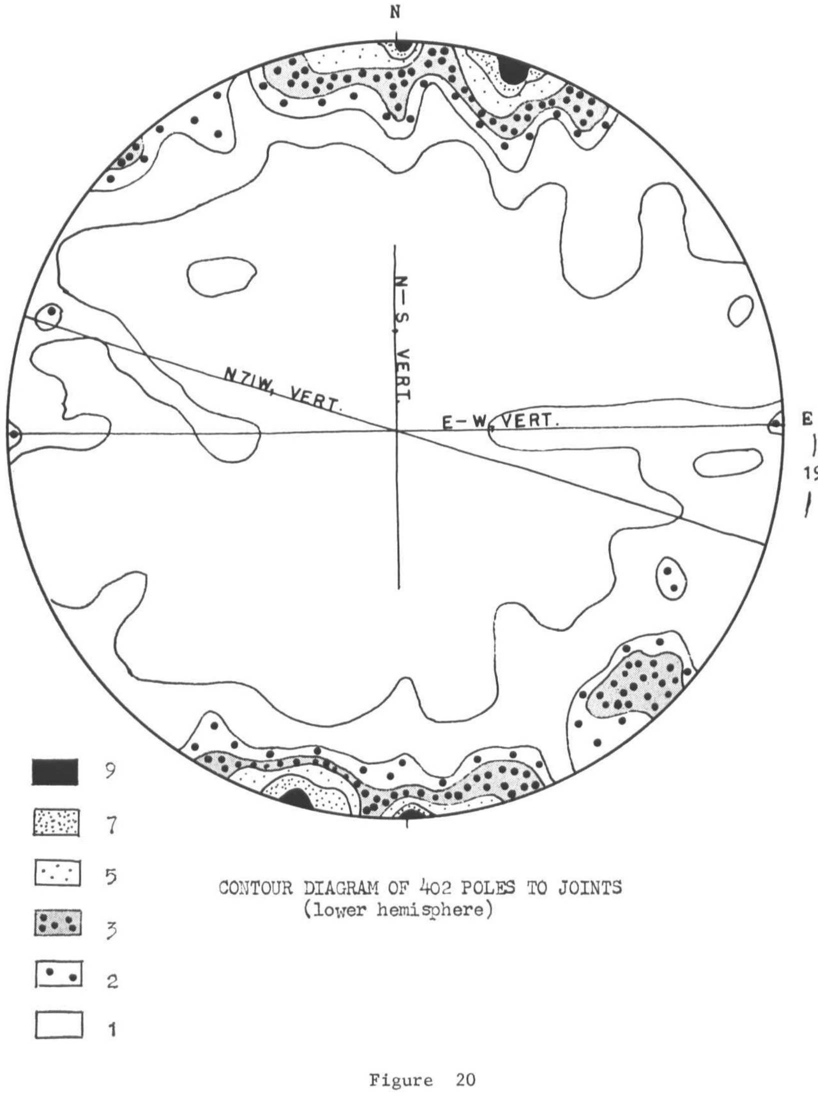


Figure 19. Same outcrop as Figure 17. Bedding (s′), crinkled shear surface (s″), regional fracture cleavage (s‴).

commercial slates are nearly isoclinal, but in other lithologies they are usually unsymmetrical. The most abundant folds are those which range from crinkle size to a foot or so across. These minor folds and the accompanying axial plane fracture cleavage reflect the trend and general symmetry of the larger folds, although the minor folds and crinkles plunge in both directions along the strike of the larger structure.

Aside from bedding schistosity, fracture cleavage is the commonest structural feature and is paralled or sub-parallel to the axial planes of these folds. Well-developed flow cleavage is restricted to the commercial slates and fine-grained schists.

High-angle reverse faults are found in association with the post-Taconic deformation. Generally they displace the overturned limbs of folds. These faults are probably quite common, but they are not often



CONTOUR DIAGRAM OF 402 POLES TO JOINTS
(lower hemisphere)

Figure 20

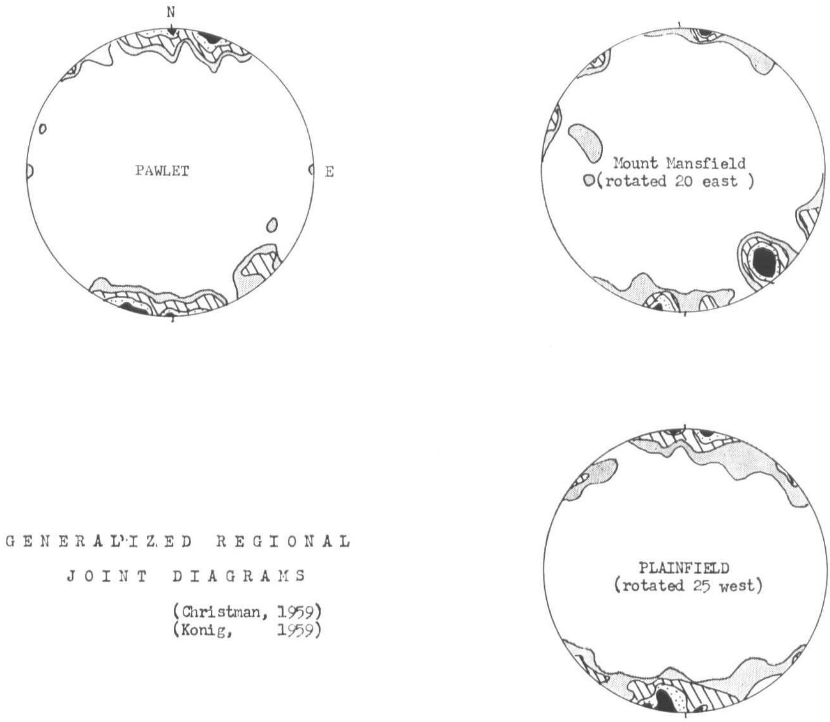


Figure 21

recognized because the similarity of the rocks conceals the effects of displacements.

The exact age of the post-Taconic deformation can not be determined in the Pawlet quadrangle. Exposures of non-metamorphosed, slightly folded Upper Silurian-Lower Devonian rocks above metamorphosed, well-folded Cambrian and Ordovician formations at Becraft Mountain, some 85 miles to the south, suggest an upper limit of late Silurian for the post-Taconic deformation. Parallelism of the structural pattern of the northern Taconic range with the Green Mountain anticlinorium suggests contemporaneous deformation with the anticlinorium.

Joints

The last structural feature imposed upon the Taconic rocks was a well-developed regional joint pattern. The attitude of some four hundred

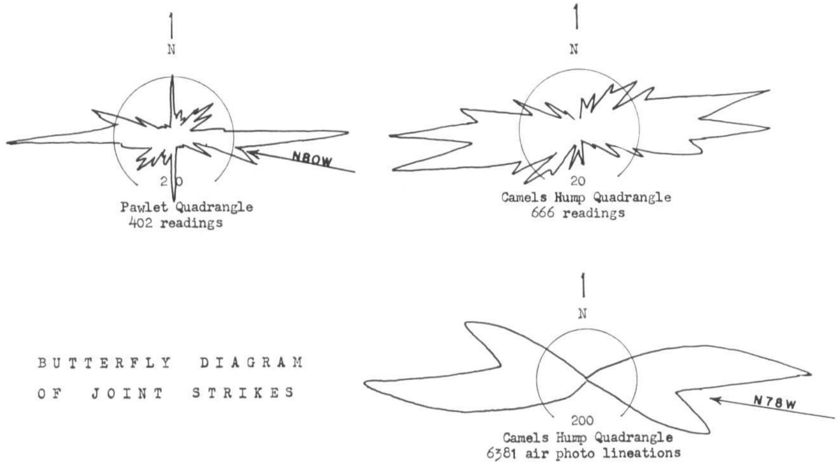


Figure 22

joints was measured during the field work. The poles to joint planes were plotted on the lower hemisphere of an equal area stereographic net (Fig. 20).

At least two prominent joint sets are developed: a vertical east-west and a north-70-west compound "shear" set, and a north-south strike set. The north-south set is weakly represented in Figure 20 because of the limited number of readings taken. This joint direction is largely concealed at most outcrops by its parallelism to the strike of bedding, bedding schistosity and cleavage.

The similarity of joint patterns for the Pawlet, Camels Hump, Mount Mansfield, and Plainfield quadrangles (Figs. 21 & 22)[1] suggest a common origin for the joints east, west, and along the axis of the Green Mountain anticlinorium. Furthermore, the apparent fixed angular relationships of the various joint sets to the trend of the structures in these quadrangles again implies that the joints are related to the forces which caused these folds (Secor, 1959). Thus, it is inferred from these data that the major folding is post Ordovician, because the rocks east of the Green Mountains are at least late Silurian to early Devonian in age. By comparing joint patterns in the metasediments and plutonic rocks of the New Hampshire

[1] In Figure 21, the regional strike of rocks has been rotated the stated amount to bring them into a sub-parallel arrangement with the strike of the Pawlet quadrangle rocks.

series Konig suggested (personal communication) that the joints of the Plainfield quadrangle are post-Silurian and probably Devonian. If the joints in the quadrangles of Figures 21 & 22 are related, then those joints seen in the Taconic rocks may be Devonian.

Surprisingly, the joint patterns of the Lower and Middle Paleozoic rocks of the Hudson Valley as shown by Parker (1942), are quite different from those observed in southwestern Vermont and adjacent eastern New York. Further work certainly must be undertaken to establish the precise relationship of joints here to those on and east of the Green Mountains; additional investigation should fix the relationship of the joints in equivalent age rocks farther south and west in New York.

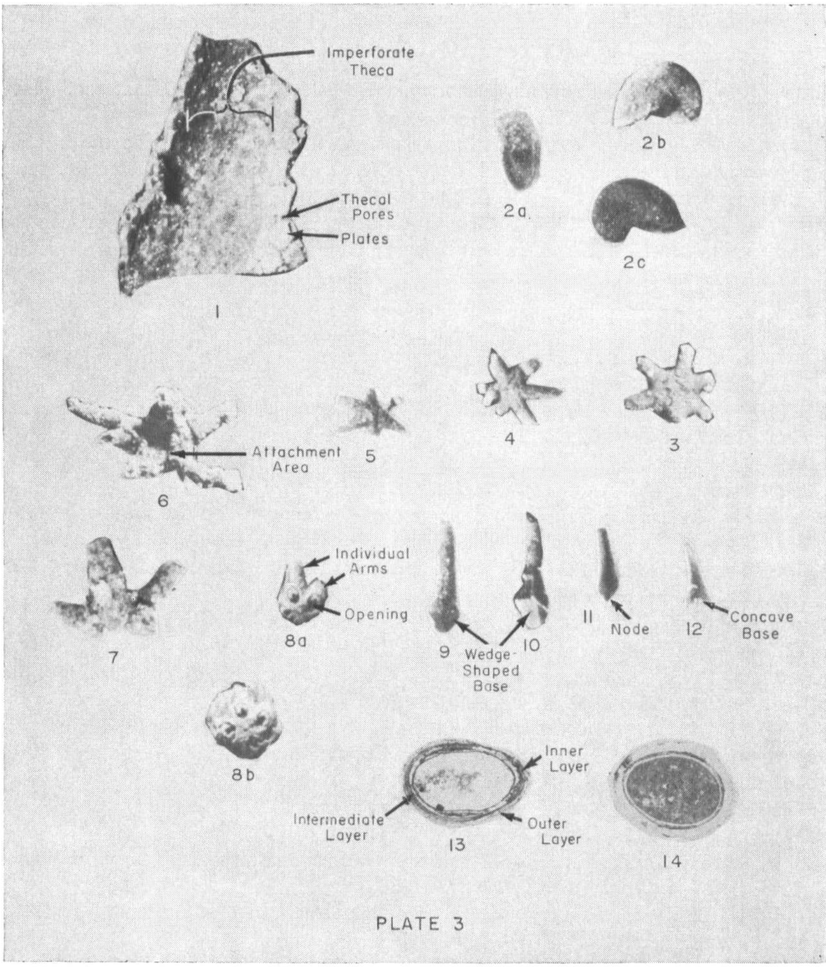


PLATE 3

1. Echinoderm? fragment. × 20, locality 7.
2. *Pelagiella* sp. a. Apertural view. b,c, top and basal views × 6, locality 7.

3–10. Hexactinellid (?) sponge spicules.

3–5. Central disc and radiating arms. × 10, localities 7 and 8.

6. Central disc with 5 radiating arms attached. × 10, locality 7.
7. Central disc with 3 radiating arms attached. × 10, locality 8.
8. a,b Central disc and 4 radiating arms, basal view showing openings into hollow arms. × 15, locality 7.

9,10. Individual arms, base is wedge-shaped in area of attachment to central disc and adjacent arms. × 10, locality 7.

11–12. Individual sponge spicules or Echinoderm spines. × 10, locality 7.

13–14. Incerta sedis (probably individual spicules), section view. × 100, locality 7. See also Fig. 11.

REFERENCES CITED

BAIN, G. W., 1938, The central Vermont marble belt: New England Intercollegiate Geol. Ass. 34th Ann. Meeting, 23 p.

BERRY, W. B. N., 1959, Graptolite faunas of the northern part of the Taconic area, in New England Intercollegiate Geological Conference: Guidebook 51st Ann. Meeting, p. 61–62, Table 1.

BILLINGS, M. P. (et. al.), 1952, Geology of the Appalachian highlands of east-central New York, southern Vermont, and southern New Hampshire: Geol. Soc. America Guidebook for field trips in New England, 71 p.

BUCHER, W. H., 1957, Taconic klippe—a stratigraphic-structural problem: Geol. Soc. America Bull., v. 68, p. 657–674.

CADY, W. M., 1945, Stratigraphy and structure of west-central Vermont: Geol. Soc. America Bull., v. 56, p. 515–588.

CHRISTMAN, R. A., 1959, Geology of the Mount Mansfield quadrangle, Vermont: Vt. Geol. Survey Bull., no. 12, 75 p.

DALE, T. N., 1899, The slate belt of eastern New York and western Vermont: U. S. Geol. Survey, 14th Ann. Rept., p. 163–307.

—— 1912, The commercial marbles of western Vermont: U. S. Geol. Survey Bull., 521, 170 p.

FENNEMAN, NEVIN M., 1938, Physiography of eastern United States: McGraw Hill Book Co., New York and London, 714 p.

FOWLER, PHILLIP, 1950, Stratigraphy and structure of the Castleton area, Vermont: Vt. Geol. Survey Bull., no. 2, 83 p.

KAISER, E. P., 1945, Northern end of the Taconic thrust sheet in western Vermont: Geol. Soc. America Bull., v. 56, p. 1079–1098.

KAY, G. M., 1951, North American geosynclines: Geol. Soc. Memoir 48, 143 p.

KEITH, ARTHUR, 1912, New evidence on the Taconic question (abstract): Geol. Soc. America Bull., v. 23, p. 720–721.

—— 1913, Further discoveries in the Taconic mountains (abstract): Geol. Soc. America Bull., v. 24, p. 680.

KEITH, ARTHUR, 1932, Stratigraphy and structure of northwestern Vermont: Wash. Acad. Sci. Jour., v. 22, p. 357–379, 393–406.

KONIG, R. H., 1959, Geology of the Plainfield quadrangle, Vermont: Vt. Geol. Survey Bull., no. 16, 86p.

LARRABEE, D. M., 1939–40, The colored slates of Vermont and New York: Eng. and Min. Jour., v. 140, no. 12, p. 47–53; vol. 141, no. 1, p. 48–52.

LOCHMAN, CHRISTINA, 1956, Stratigraphy, paleontology and paleogeography of the *Elliptocephala asaphoides* strata in Cambridge and Hoosick quadrangles, New York: Geol. Soc. America Bull., v. 67, p. 1331–1396.

LOCHMAN-BALK, CHRISTINA and WILSON, J. L., 1958, Cambrian biostratigraphy in North America: Jour. Paleontology, v. 32, p. 312–350.

LYONS, J. B. (and others), 1957, Lead-alpha ages of some New Hampshire granites: American Jour. Sci., v. 225, p. 527–546.

MACFADYEN, J. A., 1956, The geology of the Bennington area, Vermont: Vt. Geol. Survey Bull., no. 7, 71 p.

OSBERG, P. H., 1952, The Green Mountain anticlinorium in the vicinity of Rochester and East Middlebury, Vermont: Vt. Geol. Survey Bull., no. 5, 127 p.

Parker, J. M., 1942, Regional systematic jointing in slightly deformed sedimentary rocks: Geol. Soc. America Bull., v. 53, p. 381–408.

Pond, A. M., 1927–28, The peneplanes of the Taconic Mountains in Vermont: Vt. State Geologist 16th Rept., p. 292–314.

Potter, D. B., 1959, Stratigraphy and structure in the central Taconic region, New York (abstract): Geol. Soc. America Bull., v. 70, p. 1658.

Price, J. K., 1956, A new approach to the Rensselaer graywacke problem: unpublished Master's thesis, Geology Department, Cornell University.

Pumpelly, R., Wolff, J. F. and Dale, T. N., 1894, Geology of the Green Mountains in Massachusetts: U. S. Geol. Survey Mon. 23, 206 p.

Rosenfeld, J. L., Thompson, J. B., Jr., Zen, E., 1958, Data on coexistent muscovite and paragonite (abstract): Geol. Soc. America Bull., v. 69, p. 1637.

Ruedemann, R., 1909, Types of inliers in New York: New York State, Mus. Bull. 133, p. 164–193.

——, and Cushing, H. P., 1914, Geology of Saratoga Springs and vicinity: New York State, Mus. Bull. 169, 177 p.

—— 1942, Geology of the Catskill and Kaaterskill quadrangles: Part 1, Cambrian and Ordovician geology of the Catskill quadrangle: New York State, Mus. Bull. 331, 188 p.

Secor, D. T., 1959, Joints of the Camels Hump area, Vermont: unpublished Master's thesis, Geology Department, Cornell University.

Shumaker, R. C., 1959, Pawlet quadrangle, in New England Intercollegiate Geological Conference: Guidebook 51st Ann. Meeting, p. 59–60.

Theokritoff, G., 1957, Use of the term "Schodack formation" in Washington County, New York (abstract): Geol. Soc. America Bull., v. 68, p. 1804–1805.

—— 1959, Stratigraphy and structure of the Taconic sequence on the Thorn Hill and Granville quadrangles, in New England Intercollegiate Geological Conference: Guidebook 51st Ann. Meeting, p. 53–57.

Thompson, J. B., Jr., 1952, Southern Vermont, in Guidebook for field trips in New England: sponsored by the Geol. Soc. America, p. 14–23.

—— 1959, Stratigraphy and structure in the Vermont Valley and the eastern Taconics between Clarendon and Dorset, in New England Intercollegiate Geological Conference: Guidebook 51st Ann. Meeting, p. 71–85.

Walcott, C. D., 1888, The Taconic system of Emmons: American Jour. Sci., 3rd ser., v. 35, p. 229–242, 307–327, 394–401.

Zen, E-an, 1956, Stratigraphy and structure of the north end of the Taconic range, Vermont: Geol. Soc. America Bull., v. 67, p. 1829–1830.

—— 1959, Stratigraphy and structure at the north end of the Taconic range and adjacent areas, in New England Intercollegiate Geological Conference: Guidebook 51st Ann. Meeting, p. 1–16.

—— 1964, Stratigraphy and structure of a portion of the Castleton quadrangle, Vermont: Vt. Geol. Survey Bull., no. 25, 70 p.

BEDROCK GEOLOGY OF THE PAWLET QUADRANGLE, VERMONT

PART II
EASTERN PORTION

By

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VERMONT GEOLOGICAL SURVEY
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DEPARTMENT OF WATER RESOURCES
MONTPELIER, VERMONT

BULLETIN NO. 30 1967

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BEDROCK GEOLOGY OF THE PAWLET QUADRANGLE, VERMONT

PART II
EASTERN PORTION

By

JAMES B. THOMPSON, JR.*

ABSTRACT

Precambrian gneisses of the Mount Holly Complex are overlain with profound unconformity by Lower Cambrian conglomerate, graywacke and orthoquartzite. These strata are succeeded by about 3,500 feet of Cambrian and Lower Ordovician carbonates, entirely dolomite in the lower part, but with interbedded calcite marbles in the Upper Cambrian and Lower Ordovician. Early in Middle Ordovician time high-angle faulting occurred, the Tinmouth Disturbance, accompanied or closely followed by volcanic activity recorded by the Baker Brook Volcanics (new name). The Tinmouth Disturbance was followed by an extended period of erosion so that Middle Ordovician phyllites and interbedded limestones of the Ira Formation rest unconformably on an erosion surface that had bared, locally, the Precambrian of the upthrown blocks.

The Taconic Allochthone overlies, with structural discordance, the rocks of the Champlain Valley sequence just described, and was probably emplaced, from an easterly source, during the late Middle Ordovician. Phyllites and schists of the Taconic Sequence form the highest summits in the area. Dark, albitic phyllites with some interbedded carbonates and quartzites are described herein as the Netop Formation (new name), and are interpreted as underlying, stratigraphically, the green phyllites of the St. Catharine Formation.

Later the rocks of both sequences were folded and faulted with overthrusting toward the west. The latest deformational event recognized is the emplacement of the Dorset Mountain nappe, interpreted as a gravity slide off the front of the rising Green Mountain anticlinorium, possibly during the Acadian orogeny.

The entire area has been subjected to a regional metamorphism that overlapped in part the later stages of deformation. The mineral assem-

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blages are typical of the biotite zone or upper green schist facies. Post-metamorphic mafic dikes are correlated tentatively with the White Mountain Magma Series of probable Mesozoic age.

INTRODUCTION

General Statement

The eastern portion of the Pawlet quadrangle (Part I, Fig. 1) lies at the western margin of an area, extending east into New Hampshire, in which the writer has been carrying on detailed field investigations since 1947. The field work in the Pawlet quadrangle was done sporadically between 1951 and 1965, but mainly in 1951–52 and 1958–59. Nearly all of the mapping was done using aerial photographs providing stereoscopic coverage at a scale of 1:20,000, supplemented by the use of an aneroid barometer and by pacing where necessary. Transfer of these data to the Pawlet quadrangle at 1:62,500 has introduced an unfortunate, but unavoidable loss of accuracy and detail. The topographic survey was done by "reconnaissance methods" in 1894 and is woefully inadequate. The road net and buildings are fairly well located but features of topography and drainage may be as much as one-half mile out of place—notably in the swampy areas along the Tinmouth Channel and in the high country on Dorset, Tinmouth, and Woodlawn mountains. For this reason the geologic features as shown on Plates 1 and 2 are not everywhere consistent with the topography. To have made it so would have distorted the boundaries in an unreal and misleading manner. In such instances, however, the geologic features are correctly located with respect to nearby roads and houses.

Acknowledgments

The costs of the field studies were met partly by funds provided by the Department of Geological Sciences, Harvard University (formerly Division of Geological Sciences), and by the Vermont Geological Survey. The writer has benefited greatly from field conferences and exchanges of ideas with workers in nearby areas, notably W. F. Brace, W. M. Cady, P. C. Hewitt, J. A. MacFadyen, P. H. Osberg, R. C. Shumaker, G. Theokritoff and E-an Zen. Assistance in the field was provided by G. J. F. MacDonald and S. M. Ornstein, 1951; E-an Zen, 1952; and G. A. Mairs, 1959. H. B. Whittington accompanied the writer in 1959 on visits to nearly all the fossil localities and also re-examined the older data.

History of Geologic Investigations

Early descriptions of the present area may be found in Hitchcock *et al.* (1861). Later more detailed investigations include those of C. D. Walcott (1888), J. E. Wolff (1891), A. F. Foerste (1893), and, more extensively, T. N. Dale (1891, 1894, 1902, 1912). Since the work of Dale, important contributions to the stratigraphy, structure, and economic geology have been made by A. Keith (1932, 1933), C. E. Gordon (1921, 1925) and G. W. Bain (1927, 1931, 1933a, 1933b, 1938). The modern cycle of investigation was initiated largely by W. M. Cady (1945) with contributions by E. P. Kaiser (1945) and P. Fowler (1950). Recent published reports dealing with adjacent or nearby areas include those of W. F. Brace (1953), E. Zen (1961, 1964), G. Theokritoff (1964), R. C. Shumaker (1959, 1960, and Part I, this Bulletin), P. C. Hewitt (1961), J. A. MacFadyen (1956), and P. H. Osberg (1952). Earlier summaries of the work herein presented appear in Billings, Rodgers and Thompson (1952), Thompson (1959), and on the Centennial Geologic Map of Vermont (Doll *et al.*, 1961).

Topography, Drainage and Culture

The eastern part of the Pawlet quadrangle contains many of the higher summits of the Taconic Mountains, and extends east to the main Vermont Valley. The relief is relatively strong for this part of New England, exceeding 3000 feet, and has been of considerable aid in deciphering the complex geologic structure.

Elevations along the east-central edge of the map and in the southeast corner, in the floor of the Vermont Valley, are between 700 and 800 feet above sea level. This portion of the Vermont Valley is drained by Otter Creek which flows north toward Rutland and ultimately enters Lake Champlain near Vergennes. The Tinmouth Valley, at elevations of 1000 to 1400 feet above sea level, lies 2 to 3 miles west of the Vermont Valley and is drained by Mill Brook, Baker Brook, and the Tinmouth Channel (Clarendon River), all tributaries of Otter Creek. The ridge separating these valleys includes Danby Hill (2112 feet), and Clark Mountain (2040 feet).

At its south end the Tinmouth Valley is cut off abruptly by the Dorset Mountain massif. This includes Dorset Mountain (east peak, 3804 feet; west peak 3560 feet), Netop Mountain (3040 feet), and Dorset Hill (2880 feet). Green Peak or Mount Aeolus (3185 feet) lies just to the south in the Equinox quadrangle. These peaks form a horseshoe-shaped mass about Dorset Hollow at the headwaters of the Mettawee River. The Mettawee River follows a general southwesterly, westerly, then northwesterly course across the Taconics, entering Lake Champlain at Whitehall, New York.

The Tinmouth Valley is bordered on the west by the easternmost range of the main area of the Taconic Mountains. The principal summits are Tinmouth Mountain (2847 feet), Dutch Hill (2480 feet) and Woodlawn Mountain (3100 feet). The western slope of this range is drained by tributaries of the Mettawee River.

The floors and lower slopes of the Tinmouth and Vermont valleys, and of Dorset Hollow, are utilized principally for dairy farming, and the higher summits are covered by dense forest. Intermediate elevations are characterized by scrub pasture and abandoned farms. Many former farm-houses in this belt are now vacation or retirement residences.

Marble quarrying was once a major industry along the west side of the Tinmouth Valley and on the eastern and southern slopes of the Dorset Mountain massif. The only quarries now operating, however, are on the northeast ridge of Dorset Mountain in Danby.

Sink holes and caves are not uncommon in areas underlain by calcite marbles. The more extensive and interesting of these caves are found on the eastern slopes of the Dorset Mountain massif. These and other caves have recently been described by Scott (1959).

The area mapped lies principally in Dorset, Danby and Tinmouth townships, but includes parts of Wallingford, Clarendon, Ira, and Middletown Springs. Access is provided mainly by the Rutland Railroad and by U. S. Highway 7 along the Vermont Valley, Vermont Highway 30 along the Mettawee River, and Vermont Highway 140 passing from Wallingford through Tinmouth to Middletown Springs.

STRATIGRAPHY AND LITHOLOGY

General Statement

Dorset Mountain and the eastern range of the Taconic Mountains are underlain by phyllites and fine-grained schists believed to be mainly of Cambrian age. The Vermont Valley and the Tinmouth Valley are underlain by a series of Cambrian to early Ordovician carbonate rocks. These carbonate rocks dip westward beneath the phyllites of the Taconics.

The oldest rocks in the area are Precambrian gneisses. These crop out on the ridge that separates the Vermont and Tinmouth valleys. Precambrian rocks, mainly gneisses, schists, and calc-silicate rocks are also extensively exposed to the east in the Wallingford quadrangle where they form part of the extensive area of Precambrian outcrop in the central and southern Green Mountains. The Precambrian gneisses and associated rocks are here referred to as the Mount Holly Complex, originally the Mount Holly Series of Whittle (1894). They are probably correlative with the Grenville Series of the eastern Adirondacks cropping out some 18 miles to the west.

The rocks of the Mount Holly Complex are overlain with profound unconformity by a sequence of conglomerates, graywackes, phyllites and orthoquartzites having a total thickness of about 1500 feet, and at least in part of Lower Cambrian age. These are in turn succeeded conformably by about 3500 feet of dolomitic and calcitic marbles with minor intercalations of quartzite and phyllite. The dominant rocks in the lower part of the carbonate series are dolomites and those in this upper part are calcite marbles with minor dolomites. There are few diagnostic fossils in this area, hence the dating is partly dependent on continuity of mapping into other areas, and upon regional correlations. The carbonate sequence contains Cambrian, Lower Ordovician, and possibly Middle Ordovician strata.

Middle Ordovician phyllites with minor intercalations of fossiliferous limestone and locally, near the base, metamorphosed pyroclastics, rest with angular unconformity upon all of the above rocks including the Precambrian gneisses. This unconformity is thought to record an episode of high-angle faulting, accompanied by minor volcanic activity, early in the Middle Ordovician. These phyllites and the underlying carbonates, quartzites and conglomerates constitute what has long been called the *Valley Sequence*, or, more specifically, the *Champlain Valley Sequence*.

The rocks of the Champlain Valley Sequence are in turn overlain, with what is believed to be a structural discordance, by a variety of phyllites with minor intercalated carbonates and quartzites. These crop out west of the Tinmouth Valley, and on Dorset Mountain, and form a part of what is known as the *Taconic Sequence*. Though non-fossiliferous here, these rocks are believed, by correlation with the western Taconics and with the areas east of the Green Mountains, to be of Cambrian age. If so they constitute an eastern portion of the Taconic Allochthone (Cady, 1945).

All of the above rocks have been metamorphosed to at least the green schist facies. The youngest rocks of the area are post-metamorphic mafic dikes. These are probably correlative with the White Mountain Magma Series (Billings, 1956), now believed to be of early Mesozoic age on the basis of radiometric ages summarized by Toulmin (1961). Dale (1912) reports that dikes cutting the marbles in Dorset and Danby include augite camptonite and olivine camptonite.

Mount Holly Complex

Precambrian gneisses crop out in the valley of Mill Brook and on the lower northern slopes of Dorset Mountain. More extensive exposures are found on the north ridge and east slopes of Clark Mountain, and in irregular patches between Clark Mountain and the northeast corner of the quadrangle. The rocks are mainly chloritic gneisses and feldspathic schists cut by sheared and brecciated pegmatite. All of these rock types are characterized by blue opalescent quartz (probably caused by the abundant tiny rutile needles visible under highest magnification), by extensive sericitization and saussuritization of the feldspars, and by chloritization of the ferromagnesian minerals. The Clark Mountain belt extends into the adjacent parts of the Wallingford and Rutland quadrangles where there are vitreous quartzites, mica schists, and also coarse marbles and calc-silicate rocks. These rock types have not been observed in the Pawlet quadrangle.

The above rocks are similar in all respects to common rock types found in the Green Mountains in the central and eastern parts of the Wallingford quadrangle. The Precambrian rocks in the Green Mountains belong to the Mount Holly Series of Whittle (1894), and the rocks in the Pawlet quadrangle are almost certainly correlative with them. We refer them here to the Mount Holly Complex following Doll *et al.*, (1961). There are also strong resemblances to the Precambrian rocks of the southeastern Adirondacks as described by Walton and de Waard (1963). The nearest outcrops of the latter are in the Fort Ann quadrangle near Comstock, New York, approximately 18 miles west of the Clark Mountain exposures.

The overlying Dalton Formation contains cobbles and pebbles of quartzite and gneiss and smaller pebbles of blue quartz and feldspar derived from the quartz veins and pegmatites in the Mount Holly Complex. It is clear that the rocks of the Mount Holly Complex were metamorphosed and deformed, then deeply eroded, before the deposition of the younger formations.

Paleozoic Rocks: Champlain Valley Sequence

Dalton Formation

The Dalton Formation was first named by B. K. Emerson (1892), but was not described in any detail until some time later (Emerson, 1917).

The type area is in Dalton, Massachusetts, where it consists (Herz, 1960) of conglomerate and arkosic quartzite resting directly on the Precambrian basement. Similar rocks may be traced north along the west flank of the Green Mountains (MacFadyen, 1956; Hewitt, 1961; Thompson, 1959) into the Wallingford quadrangle, immediately to the east of the area here described.

The names Pinnacle (Clark, 1931), Nickwaket (Keith, 1932) and Mendon (Whittle, 1894) as used by other authors (Osberg, 1952; Brace, 1953; MacFadyen, 1956; Hewitt, 1961) for rocks at this position, are more or less equivalent in meaning, but the name Dalton has precedence and is more clearly applicable to the unit as mapped in the present area. It is probable that the Mendon Series, as defined by Whittle, includes strata that are mapped here with the overlying Cheshire Quartzite.

The best exposures of the Dalton Formation are on Clark Mountain and on the hills in the northeast corner of the quadrangle. It is about 250 feet thick. The basal beds are typically pebble or cobble conglomerate grading upward into graywacke with discontinuous pebbly beds and beds of pale-green chlorite schist. Near the top some of the beds are rich in rusty-weathering carbonate and approach impure dolomites in composition.

The Dalton Formation is non-fossiliferous in Vermont, but trilobite remains on Clarksburg Mountain near North Adams, Massachusetts, at the south end of the Green Mountains (Walcott, 1888, p. 235–6), are probably in quartzites correlative with the upper part of the Dalton Formation as here mapped. The writer visited the Clarksburg Mountain area in 1949, accompanied by W. F. Brace and F. Barker. The fossil locality was not relocated, but it was evident that the rocks in the vicinity were more characteristic of the Dalton Formation than of the overlying Cheshire Quartzite.

Cheshire Quartzite

The Cheshire Quartzite was named by B. K. Emerson (1892) for distinctive exposures near Cheshire, Massachusetts, but again not described in detail until later (Emerson, 1917; Herz, 1960). The original name will be retained here for simplicity though the less restrictive term "formation" might well be substituted for "quartzite." Massive, vitreous orthoquartzites in beds that may be as much as 50 feet thick are, however, the dominant and most distinctive rock types, particularly in southern Vermont and western Massachusetts. The topmost bed, where identifiable as such, is a massive quartzite. Interbedded carbonaceous phyllites, quartz-mica schists, schistose quartzites, and feldspathic quartzites are increasingly abundant toward the base but comprise less than a third of the total thickness of the Cheshire.

In the Pawlet quadrangle the Cheshire Quartzite crops out in a narrow zone along the east side of the Tinmouth Valley between Baker Brook and the north boundary of the area. Because of faulting there is nowhere in this area an unbroken section through the Cheshire, although as much as 500 feet may be demonstrated locally. It is probably much thicker; roughly on strike to the north, between Rutland and Center Rutland, the Cheshire is about 1,000 feet thick. This is not inconsistent with other estimates in nearby areas (Osberg, 1952; Brace, 1953; Thompson, 1959).

The Cheshire contains no fossils in the Pawlet quadrangle, but to the east, on Green Hill, southeast of Wallingford village, there are abundant *Scolithus* tubes. Fossils indicating an Early Cambrian age have been reported from near Bennington, Vermont (Walcott, 1888), and from west of Lake Dunmore (see Cady, 1945, p. 528). The fossils on Clarksburg Mountain, referred to above, are in basal Cheshire if not in the Dalton. An Early Cambrian age for the Cheshire is thus reasonably well established.

Dunham (Rutland) Dolomite

The Dunham Dolomite (Clark, 1934) takes its name from a locality in southern Quebec near the Vermont boundary. It underlies much of the lower eastern slope of the Tinmouth Valley, but exposures are poor owing to the more resistant formations above and below. A nearly complete section may be seen, however, east of Chippenhook, about 1.5 miles north of the northeast corner of the map area (Thompson, 1959, stop 2). Other exposures of the Dunham are found midway along the eastern edge of the map in the Vermont Valley, and near Emerald Lake, in the southeast corner. There are also two isolated patches on Clark Mountain.

The formation is typically a yellow or buff-weathering dolomite in which the bedding is marked by thin siliceous partings, usually less than a quarter of an inch thick, and spaced at intervals of from a few inches to a foot or more. Some of these partings appear to be recrystallized chert, others are schistose and possibly derived from shaly material. In the upper part of the formation the dolomites are more varied and there are beds that weather gray or dark gray and locally contain abundant rounded sand grains.

The Dunham in the Tinmouth Valley is about 900 feet thick and is known to be of Early Cambrian age from fossils found east of Chippenhook, and at several localities in the northern part of Rutland (Dale, 1891, Wolff, 1891; Foerste, 1893). These have been recently restudied by George Theokritoff (personal communication, 1966) and are still regarded as Lower Cambrian. The name Rutland Dolomite as used by Keith (1923), formerly the Rutland *Limestone* (Wolff, 1891), is essentially synonymous (see below in discussion of Danby Formation), and should perhaps take precedence.

Monkton Quartzite

The Monkton Quartzite was named by Keith (1923) with reference to the conspicuous exposures of red quartzite near Monkton, Vermont. The name is misleading in that even in the type locality the quartzites, though conspicuous, constitute but a minor part of the formation. The principal rock type is dolomite, the various beds weathering orange, orange-red, yellow, buff, or dark gray. These comprise about three quarters of the formation. The rest is made up mainly of quartzite, in beds from a few inches to 30 or more feet thick. Other rock types include sandy dolomite, green phyllite and black, carbonaceous phyllite. Phyllite beds are minor and rarely more than a foot or two thick. The quartzites commonly stand out in bold outcrops giving a misleading impression of their relative abundance.

The quartzites of the Monkton may be distinguished from those in the Cheshire and from those in the younger Danby formation in that abundant feldspar as well as quartz clasts are present, and in the greater abundance of matrix material, largely chlorite and mica (both muscovite and biotite). The red Monkton quartzites in northern Vermont owe their color to finely divided hematite. In the Pawlet quadrangle, by contrast, they are green or gray-green, but commonly contain magnetite octahedra or spangles of hematite indicating that they were probably once red sandstones that have since lost their red color through metamorphism. Cross-bedding is characteristic of both the quartzites and the sandy dolomites, and many bedding surfaces show ripple marks or mud cracks.

The Monkton Quartzite in many places contains a distinctive bed of dolomitic sandstone near the base. This rock is characterized by giant cross-beds of striking appearance, accentuated by a porous weathered surface. This bed averages 8 or 12 feet thick and in places, as southeast of Tinmouth village, may be followed for over half a mile on strike. It is possible that all the occurrences of this rock are the same bed, but this cannot be proved.

The rocks here mapped as Monkton are about 600 feet thick, but the boundary with the overlying Winooski is somewhat arbitrary. Fossils were reported by Perkins (1908) from near Mallets Bay on Lake Champlain, and more recent discoveries are reported by Kindle and Tasch (1948), Tasch (1949), and Shaw (1962). These localities have recently been restudied by Theokritoff and are apparently late Early Cambrian (G. Theokritoff, personal communication, 1966). In northern Vermont the Monkton appears (Cady, 1945, p. 531) to grade laterally into the Parker Slate. According to Shaw (1958) the Parker contains both lower and Middle Cambrian faunas.

Winooski Dolomite

The Winooski Dolomite (Hitchcock *et al.*, 1861), as revised by Cady (1945), is much like the underlying Monkton in the rock types present, but the quartzite and phyllite beds are typically thinner and are much less abundant. The mapping of a boundary such as that between the Monkton and Winooski is often quite subjective, particularly in areas of sparse exposure, though an attempt was made to adhere as closely as possible to the criteria followed by Cady (1945, p. 532). As mapped in the Pawlet quadrangle, the Winooski is about 600 feet thick. The most conspicuous quartzites in the Winooski are actually in the top of the formation close to the base of the overlying Danby Formation. Had these quartzites occurred near the base they would have been mapped as Monkton.

No fossils have yet been found in rocks mapped as Winooski Dolomite nor in the Rugg Brook Dolomite (Shaw, 1958, p. 535–6), as it has been called in northern Vermont. A Middle Cambrian age is likely, however, for the reasons summarized by Shaw. Specifically, the Rugg Brook overlies the Parker Slate (Keith, 1932) which is in part Middle Cambrian, and it underlies rocks containing a *Cedaria* fauna of Dresbachian age.

Danby Formation

The Danby Formation was originally defined by Keith (1932) with reference to exposures in the Vermont Valley near Danby Village. Unfortunately Keith published no maps and the locations of typical outcrops were not given, hence it has never been wholly clear what Keith's original Danby included. However, in comparing Keith's descriptions with the writer's observations in Danby and Wallingford it appears that the rocks referred by Keith to the Danby are not in general correlative with the Danby as redefined by Cady (1945), but belong to the Monkton Quartzite and to the Winooski Dolomite as described above. It also appears likely that Keith's Wallingford Formation is the unit that contains the strata assigned by Cady to the Danby. That this is so is partly confirmed by an unpublished manuscript map of the Proctor-Brandon area, prepared in about 1932 by Arthur Keith (probably with the aid of A. C. Swinnerton). An abridged version of this map was later published by Keith (1933) but with the key boundaries omitted. Comparison of Keith's manuscript map with the maps of Cady (1945) and Fowler (1950) shows a clear inconsistency between Keith's usage and that of the later workers, as shown in Table 1. Cady's definitions are clear and are now well established, whereas Keith's original ones were inadequate and never supported by published maps. For these reasons we have regarded the definitions of Cady as superseding those of Keith. The name Danby, as used by Keith, would have to be abandoned in favor of the prior names Monkton and Winooski, but as redefined by Cady it is a useful term. Happily there are excellent exposures of the Danby, as now defined, in the gorge of Mill Brook where it passes through Danby village. Either these, or exposures a few miles south on the steep east-facing slopes below the marble quarries on Dorset Mountain, may be regarded as typical.

The base of the Danby Formation, as here mapped, is marked by vitreous orthoquartzites interbedded with gray weathering dolomites that are commonly slightly calcitic. West of the Tinmouth Channel, at the extreme north edge of the map, a basal Danby quartzite penetrates what were apparently solution pits in the underlying Winooski Dolomite. This, coupled with a marked change in the character of both the quartzites and the dolomites as compared with those in the underlying Monkton and Winooski, suggests a hiatus in the sedimentation followed by a distinct change in depositional environment.

The quartzites of the Danby are from a few inches to two or three feet thick, though thicker beds have been observed outside the present map area, at localities east of Rutland, and to the south on Mount Aeolus (Green Peak). They are most abundant and thickest in the lower part of the formation. The typical beds are vitreous orthoquartzites breaking into angular blocks. Most are white like those of the Cheshire, but some have dark bands of carbonaceous material. Associated types are dolomitic quartzites and sandy dolomites, all conspicuously cross-bedded. The cross-bedding in the gorge at Danby has been studied by Sahakian (1963) and indicates a dominant southeasterly transport direction.

Gray, commonly calcitic dolomites constitute between two-thirds and three-quarters of the Danby. Suspended sand grains are not uncommon

TABLE 1: Key to stratigraphic nomenclature, Champlain Valley Sequence, eastern limb of Middlebury synclinorium.

Probable Age	Keith (1932)	Cady (1945); Cady and Zen (1960)	Fowler (1950)	Present Usage
Middle Ordovician	Ira	Hortonville	Hortonville	Ira
		Glens Falls	Whipple	
		Orwell		
	Williston	Middlebury	Middlebury	Baker Brook (= Middlebury ?)
Lower Ordovician		Chipman / Beldens / Weybridge / Burchards	Beldens	Bascom (and Beldens)
			Burchards	
		Bascom	Bascom	
		Cutting	?	
	(W) Shelburne (E)	(W) Shelburne (E)	Boardman	Shelburne
Upper Cambrian	Clarendon Springs	Clarendon Springs	Clarendon Springs	Clarendon Springs
	Wallingford	Danby	Danby	Danby
Middle Cambrian	Danby	Winooski	Winooski	Winooski
Lower Cambrian		Monkton	Monkton	Monkton
	Rutland	Dunham	Dunham	Dunham (Rutland)
	Cheshire	Cheshire	Cheshire	Cheshire
	Moosalamoo	Mendon	Mendon	
	Forestdale			
	Nickwaket			Dalton

and faint cross-bedding may be visible even in quite pure dolomite. Locally, particularly in the Vermont Valley near Danby, there are small pods and lenses of white calcite marble.

Fragmentary fossils have been reported by Cady (1945, p. 530) near Middlebury, Vermont, but are not diagnostic. The Danby, however, has been traced by Cady (1945) around the north end of the Middlebury synclinorium (see Doll *et al.*, 1961) to exposures at Shoreham, Vermont, that are virtually identical to the Potsdam sandstone as exposed just to the south near Whitehall, New York (Rodgers, 1937), and Fort Ann, New York (Flower, 1964). According to Flower (1964, p. 156) the Potsdam at Fort Ann carries a Dresbachian fauna fifteen to twenty feet above its base and Franconian faunas at several higher levels. All available evidence therefore indicates that the Danby is Upper Cambrian. The boundary with the overlying Clarendon Springs Formation is somewhat arbitrary but the Danby as here mapped is about 150 feet thick.

Clarendon Springs Formation

The Clarendon Springs Dolomite (Keith, 1932) is here re-defined as the Clarendon Springs Formation to include two higher units mapped by Cady (1945), Fowler (1950), Thompson (1954), and others, as members of the Shelburne Marble. These are the Sutherland Falls Marble (Hitchcock, *et al.*, 1861; Bain, 1931, 1938) and the Intermediate Dolomite (Bain, 1931, 1938; see also Fowler, 1950; and Cady, 1945, p. 540). Mapping by the writer in the Pawlet, Wallingford, and Castleton quadrangles has shown that the conspicuous concentrations of chert, or of quartz recrystallized therefrom, that are characteristic of the upper part of the Clarendon Springs, as defined by Cady (1945) in the Shoreham area, are not characteristic of the Clarendon Springs as it has been mapped in the more eastern areas, but are characteristic and diagnostic of the Intermediate Dolomite of Bain (1931). Fowler (1950, p. 24–25) was apparently aware of this but Zen's more recent mapping is locally inconsistent through failure to take this into account (Zen, 1964, p. 33–55). Cady has also made the observation (1945, p. 540–541) that the Sutherland Falls marble passes laterally into dolomite as it is traced northward along the east limb of the Middlebury synclinorium. Cady, on re-examination of his data, (written communication, 1966) finds that the cherty dolomites on the east limb of the Middlebury synclinorium are in the Intermediate Dolomite, not in the rocks shown as Clarendon Springs on his (1945) map, and agrees that the upper Clarendon Springs of the west limb of the synclinorium (including the Shoreham section) is possibly correlative with the Intermediate Dolomite of the marble belt.

The Clarendon Springs is therefore here revised as the Clarendon Springs *Formation* and extended upward to include the Intermediate Dolomite of Bain (1931). The former Clarendon Springs Dolomite of the marble belt becomes here the lower dolomite member of the revised Clarendon Springs Formation. Lithologically, it resembles the carbonates of the underlying Danby Formation, but contains only minor, relatively thin beds of quartzite. Cross-bedding is conspicuous, as in the Danby, particularly in the lower part. Chert masses, or quartz knots recrystallized therefrom, are present but on a lesser scale than in the higher dolomites. This member is about 200 feet thick.

The Sutherland Falls Member takes its name from the Sutherland Falls quarry in Proctor (the village of Proctor was formerly called Sutherland Falls). It is typically a gray-streaked, white calcite marble that was formerly quarried for building and ornamental stone but is now little worked. Dolomite curdling is not uncommon. Near the contacts with the underlying and overlying dolomites, calcite marble can be seen passing laterally into dolomite at several localities. However, at no point in the present map-area can the member be shown to disappear entirely. The maximum observed thickness is about 100 feet.

The upper dolomite member (formerly the Intermediate Dolomite of the Shelburne) is a light gray, calcitic dolomite characterized by large irregular masses of quartz that are believed to be recrystallized chert. Bedding is less conspicuous than in the lower dolomites of the Clarendon Springs and quartzite beds are rare. The upper dolomite member is about 200 feet thick.

The Clarendon Springs, according to Cady (1945) and Welby (1961), is correlative with the Ticonderoga Dolomite at Whitehall, New York, as defined by Rodgers (1955 ms., published in part by *Welby*, 1961, p. 232–234, see also p. 28–39), and with the rocks mapped by Flower (1964) as Dewey Bridge and as "Whitehall" (Billings *et al.*, 1952, p. 35–36; Welby, 1961, p. 28–51; Flower, 1964) near Fort Ann, New York. The faunas in the upper part are Upper Cambrian, Trempealeauian, but the lower part is probably Franconian (Welby, 1961, p. 39; Flower, 1964, p. 157). The Hoyt Limestone of the Saratoga region (Fisher and Hanson, 1951) and the limestone lenses reported by Rodgers (Welby, 1961, p. 39) in the Ticonderoga quadrangle probably correlate in a general way with the Sutherland Falls.

Shelburne Marble

The Shelburne Marble (Keith, 1923) is named for exposures in Shelburne, Vermont, near Lake Champlain. It is here the typical "Vermont

Marble" of commerce and as such has been so widely distributed that it scarcely needs description. It is a medium to coarsely crystalline calcite marble, and may be pure white or may have gray or greenish streaks. It shows less dolomitic curdling than the marble of the Sutherland Falls Member of the Clarendon Springs, but is otherwise similar. The major quarries in Brandon, Florence, Proctor, Clarendon, Wallingford, Danby, and Dorset, and possibly those in West Rutland, are in this formation. The Sutherland Falls and some of the younger marbles have also been quarried, but the Shelburne has been the main producer. It is about 250 feet thick where not obviously thickened or thinned tectonically.

The Shelburne contains fossils near Shoreham (Cady, 1945; Flower, 1964) indicating that it is correlative with the Gasconade of the Mississippi Valley. Whether this is regarded as highest Cambrian or lowest Ordovician, or both, depends on one's prejudices concerning the definitions of that boundary (see Fisher, 1962; Whittington and Williams, 1964). According to Welby (1961) the Shelburne is correlative with the Whitehall at Whitehall, New York (Rodgers, 1937; Welby, 1961), and with the Baldwin Corners of the Fort Ann area, as mapped by Flower (1964).

Bascom Formation

The rocks here mapped as Bascom Formation include not only the Bascom Formation of Cady (1945) but also rocks resembling the Beldens Member of the Chipman Formation (Cady and Zen, 1960). The Cutting Dolomite of Cady (1945) is absent south of Brandon where it passes laterally into rocks indistinguishable from the lower Bascom (Zen, 1964, p. 34). In the present map-area the "Beldens" types overlie "Bascom" types where relations are clear, but the distinction could not be maintained except in areas of good outcrop and simple structure, hence was abandoned.

The top of the Shelburne Marble is marked in good exposures (in many of the quarries) by a fine-grained, yellow-weathering dolomite several inches to a foot thick, that is succeeded by blue or blue-gray calcite marbles, locally with dolomite mottling and containing interbeds of gray, yellow, or orange-weathering dolomite. The dolomites are from several inches to a few feet thick and are more abundant high in the section ("Beldens type"). Minor interbeds of calcareous sandstone and black phyllite occur about midway. The principal exposures are on the lower slopes of Dorset Mountain and in Dorset Hollow, but the thickness cannot be estimated here for structural reasons. As much as 450 feet of Bascom, however, can be demonstrated locally.

Many outcrops show indistinct forms that appear to be dolomitized fossils, but only crinoid columnals are recognizable. Diagnostic faunas occur, however, on the west limb of the Middlebury synclinorium (Cady, 1945). The Cutting (probably equivalent to the lower Bascom of the Pawlet quadrangle) is apparently equivalent to the upper Gasconade or lower Roubidoux of the Mississippi Valley (Cady, 1945; Welby, 1961; Flower, 1964), and the Bascom and Chipman formations apparently include much of the remaining Lower Ordovician (Welby, 1961; Flower, 1964).

Baker Brook Volcanics
(New Name)

A narrow belt of rocks believed to be metavolcanics has been mapped in the eastern part of the area for a distance of more than 3.5 miles. It may be traced about 0.7 miles farther northeast, in the Wallingford quadrangle, into the pastures west of the South Wallingford quarries. These rocks have formerly been referred by the writer either to the "Whipple Marble" (in Billings *et al.*, 1952) or to the Ira Formation (Thompson, 1959). They always underlie the typical rocks of these formations, however, and are of sufficient paleogeographic interest to warrant a name of their own. The rocks are well exposed at a falls in Baker Brook in the north part of Danby and on the hills nearby.

At first sight these rocks resemble lithologic types found in the Mount Holly owing to their feldspar content and strongly schistose or even gneissic texture. This is due, however, to their bulk composition, and there are in fact no minerals present, indicative of metamorphic facies, that are not found equally well crystallized, though in lesser abundance, in either the overlying or underlying Paleozoic formations. The principal rock types are quartz-muscovite-biotite schist with feldspar augen, mainly plagioclase, and a greenstone schist containing epidote, albite, chlorite, ankerite, and a green actinolitic amphibole. These are probably metamorphic derivatives of intermediate to mafic pyroclastics. Other rock types associated with the volcanics are quartzite and coal-black, carbonaceous phyllite.

The Baker Brook Volcanics rest upon carbonates assigned either to the Bascom Formation, or to the Shelburne Marble, or to the Clarendon Springs Formation, indicating an unconformity at the base of the volcanics. In the north part of Danby, below the volcanics, but too thin to show on Plate 1, there are thin-bedded limestones with minor beds of dolomite. At places, near the contacts with the underlying Clarendon Springs and Shelburne formations, the basal beds appear conglomeratic.

These limestones closely resemble, in part, the Middlebury Limestone (Cady, 1945; Welby, 1965) of early Middle Ordovician (Chazyan) age, but contain no fossils and the outcrops are many miles distant from the nearest dateable Middlebury (at Carver Falls on the Poultney River near Fair Haven, Vermont). The Baker Brook Volcanics have a maximum thickness of about 200 feet.

Ira (Hortonville) Formation

The Ira Slate of Keith (1932) has been re-defined by Thompson (1959) and Zen (1961, 1964) as the Ira Formation. The typical rocks of the formation have also been mapped as Hortonville Slate (Fowler, 1950) or Hortonville Formation (Doll *et al.*, 1961). The names Hortonville and Ira were both first used by Keith (1932) for rocks that are now considered equivalent by all recent workers. The two names therefore have equal priority but, as pointed out by Zen (1964), the evidence for dating the formation is best in the areas east of the Taconic Range where the name Ira was first used, hence the name Ira is in some ways on a better footing.

The Ira Formation cuts unconformably across all of the older units of the Champlain Valley Sequence. Along the west side of the Tinmouth Valley, where present, it rests on carbonates of the Bascom Formation, but cuts across the Bascom, and several of the formations underlying the Bascom, at the south end of the Tinmouth Valley. At the west base of Danby Hill it rests with angular discordance on quartzites and dolomites of the Monkton, and farther north, in the valley of Baker Brook and on Clark Mountain, it may be seen in contact, successively, with the Dunham, Cheshire, and Dalton. On the eastern and northeastern slopes of Clark Mountain and on the hills farther north it rests on the Mount Holly. On the west side of the Vermont Valley, however, in Danby (Plate 1) and in Wallingford (Thompson, 1959; Doll *et al.*, 1961) the Ira rests again on the Bascom Formation. These relationships are most easily explained by postulating the occurrence of high angle faulting followed by deep erosion of the upthrown blocks during the interval following the deposition of the Bascom and preceding the deposition of the Ira. This will be discussed more fully below in the section dealing with structure.

The dominant rock type in the Ira is a dark gray to black slate or phyllite that makes up perhaps ninety percent of the formation as here mapped. Minor variants include gray to gray-green phyllites, calcareous phyllites and phyllites with sandy laminae a few millimeters thick. Beds of gray or dark gray quartzite with well-defined sand grains to 2 mm in diameter occur sporadically in the formation but have not been mapped separately.

On the northeastern slopes of Clark Mountain and on the hills in the northeast corner of the map, where the Ira Formation rests directly on the Mount Holly Complex, or on the Dalton or the Cheshire, the basal beds are at some localities a pebbly quartzite, often rusty-weathering. Though nowhere seen to be more than a few tens of feet thick, this may be confused locally with some of the clastic rocks of the Dalton Formation or Cheshire Quartzite. Where the gross geologic relations do not make the distinction clear, the mapping of certain outcrops is therefore arbitrary.

At or near the base of the Ira there are beds of blue or blue-black limestone with minor interbeds of gray dolomite and black phyllite, commonly associated with beds that may be several tens of feet thick of gray dolomite breccia characterized by light gray fragments in a darker matrix. These carbonate rocks have been mapped farther north as the Whipple Marble (Fowler, 1950) or the Whipple Marble Member of the Ira Formation (Zen, 1964), but it is by no means certain that all of these carbonates are at the same stratigraphic position. The most extensive occurrences are clearly at the base of the Ira, but isolated patches far removed from the contacts with the older formations may be lenses at a higher stratigraphic position. For this reason these limestones have been shown on Plate 1 simply as limestones in the Ira and not given member status. Fossils are abundant at several localities in these limestones (see Plate 1), but consist mainly of crinoidal debris except at the northernmost locality of the series on the eastern slope of Clark Mountain, in Wallingford, where Foerste (1893) reported *Streptelasma* and *Strophomena*. These occurrences were re-visited in 1959 with H. B. Whittington. A coral from Foerste's northernmost locality was sent to Helen Duncan of the United States Geological Survey who reported (written communication, Feb. 25, 1960) that: "In my opinion, the beds that yielded the three collections (2329, 2462, and Foerste's northernmost locality) of solitary rugose corals cannot be older than Black River and might be considerably younger," and further that: "The specimens from Foerste's locality and in lot 2462 are recrystallized and somewhat deformed so I cannot be sure just what genus is represented. So far as I can tell, these specimens have no features that would not be present in Ordovician streptelasmatids, but comparable things occur in post-Ordovician rocks. Septal deposits are sufficiently extensive to eliminate the possibility of these corals being what is usually called *Lambeophyllum*. If the specimens are Ordovician Streptelasmas, they exhibit a stage of complexity I would expect to find in early Trenton forms." Wolff (1891, p. 336) reported crinoids and "a small branching bryozoan with large cells" from a locality near the hill road north of Chippenhook (Castleton quadrangle) at about the 1100-foot elevation. Dale (1891) also reports *Heliolites* from this locality. At a nearby outcrop in the road, subject to continuing abrasion by automobile tires, the writer has observed a horn coral, like the ones referred to above, that could be seen as recently as mid-summer, 1966. These occurrences, with the evidence summarized by Cady (1945), indicate that the Ira is Middle Ordovician and that the beds that have yielded fossils thus far are no older than Black River. The top of the Ira Formation has not been identified, but it must be at least 500 feet thick.

Paleozoic Rocks: Taconic Sequence

General Statement

The phyllites and associated rocks forming the higher elevations in the Dorset Mountain massif and in the range bordering the Tinmouth Valley on the west are here assigned to the Taconic Sequence. This assignment, however, is based on regional considerations and cannot be proved on the basis of observations in the present area. Were these regional relationships not known the two formations described below would be interpreted as overlying, in normal stratigraphic order, the rocks of the Ira Formation, hence Middle Ordovician or younger. The structurally lower of the two would, in fact, probably not have been separated from the Ira as a distinct unit, and the resulting map would have then been consistent with the alternative interpretation of Hewitt and MacFadyen (1963).

From the recent work of Zen (1961), Theokritoff (1964), Shumaker (1959, and Part I of this Bulletin), and of others farther south, it is now clear beyond any reasonable doubt that the rocks of the Vermont–New York slate belt form a coherent stratigraphic sequence, the Taconic Sequence, that must lie with structural discordance on the rocks of the Champlain Valley Sequence, and that these rocks therefore form a part of the Taconic Allochthone (Cady, 1945). On these points all recent workers, including Hewitt and MacFadyen, are apparently in complete agreement. It is also apparent from the work of Zen (1961) and Theokritoff (1964) that the western boundary of the allochthone can be located unambiguously and without difficulty. This is, alas, not true of the eastern boundary, hence we have to choose between a "minimum allochthone" hypothesis (Hewitt and MacFadyen, 1963) and a "maximum allochthone" hypothesis (Zen, 1961; Doll *et al.*, 1961; and this Bulletin)[1]. The maximum allochthone hypothesis in turn has variants depending on one's interpretation of the stratigraphic order of the non-fossiliferous rocks in the eastern Taconics, Zen (1961, 1963) taking one view, Doll, *et al.* (1961, 1963) another. The interpretation of Doll *et al.* provides for a normal rather than an inverted succession in the eastern Taconics, agreeing on this point with that of Hewitt and MacFadyen.

The name Brezee Phyllite was given by Keith (1932) to certain dark phyllites and associated quartzites and carbonate rocks along the eastern and northern slopes of the Taconic Range near Brandon, Vermont, that he believed to underlie his predominately green Stiles Phyllite. The Stiles Phyllite, shown as Bull Formation by Zen (1961) and as St. Catharine Formation by Doll *et al.* (1961), crops out on the higher slopes to the south and west. Zen (1961, 1963, 1964) rejects Keith's interpretation and prefers to regard the Brezee of Keith as an easterly reappearance, in an inverted succession, of his West Castleton Formation which, in its type areas to the west, has been shown by Zen (1961) and Theokritoff (1964) to overlie the Bull stratigraphically. Zen's interpretation requires that there be a major recumbent fold within the Taconic Allochthone in that area. Zen places the axial surface of this fold partly in his Bull Formation and partly in a chloritoid phyllite, surrounded in outcrop by the Bull, that he regards as stratigraphically older and has named the Biddie Knob Formation. Doll *et al.* (1961, 1963), on the other hand, regard the Biddie Knob as a lenticular member within the Bull, and the Brezee as underlying the Bull stratigraphically, hence distinct from the West Castleton, despite some admitted lithologic resemblances.

Shumaker (1960, and Part I, this Bulletin) has followed the interpretation of Zen, hence all Taconic rocks west of the Tinmouth Valley, and west of the fault passing through the notch between Dorset and Woodlawn mountains have been shown on Plate 1 in accord with that scheme. The phyllites and schists of the Dorset Mountain massif, on the other hand, are here interpreted as an essentially normal sequence, as shown by Doll *et al.* (1961), rather than as an inverted one. The inter-

[1] Hewitt and MacFadyen (1963) take Doll *et al.* (1961, 1963) to task for not indicating the problems in locating, in Vermont, the boundary of the Taconic allochthone. This is unfortunate, for we thought we had made it as clear as possible while still presenting a unified and internally consistent interpretation. The boundary of the klippe was in fact shown by a special symbol, used for nothing else (open rather than closed teeth on the allochthone side of the boundary), and this symbol was identified in the explanation as denoting a: "Thrust fault—Exact location uncertain in places; shown by stratigraphic and paleontologic evidence; . . ."

pretation shown in Plate 1 is therefore a "maximum allochthone" interpretation that is in some ways a compromise between that of Zen (1961) and that of Doll *et al.* (1961). That the sequence on Woodlawn Mountain could be inverted while that on Dorset Mountain is normal is quite possible in view of the large displacement on the fault that separates them. It should be emphasized, however, that the present interpretation is but another hypothesis, and that its merits relative to those of competing hypotheses cannot be decided until further information is at hand concerning the actual ages of the various rock units and the directions of stratigraphic tops within them. The sequence on Woodlawn Mountain, for example, *does* resemble that on Dorset Mountain, but then again there are differences. On Plate 1 the differences, rather than the resemblances, have been emphasized, and this, admittedly, may some day be proven incorrect. The rocks west of the line sketched above, though mapped in part by the writer, are already described in Part I of this Bulletin and therefore need no further discussion here. The Taconic rocks described below are the ones exposed in the Dorset Mountain massif and on the ridge south of East Rupert.

NETOP FORMATION
(New Name)

In view of the uncertainties outlined above, and in the interest of internal consistency between the area of Plate 1 and the type Brezee, it is best to use a local name for the structurally lower formation on Dorset Mountain. These rocks have been mapped as Berkshire Schist by Thompson (1959), but continued use of the name "Berkshire" is unwise in view of the many meanings that have been attached to it. "Dorset," "Mettawee," and "Aeolian" (from Mount Aeolus or Green Peak) have all been used locally or nearby for other rock units, hence the name Netop Formation is here proposed for the predominantly dark, albitic phyllites and associated rocks cropping out on the slopes of Dorset and Netop mountains.

The Netop Formation has been distinguished from the Ira, though not always with certainty, by being rather more quartzose and less carbonaceous, and by the presence of numerous minute porphyroblasts of albite. Albitic phyllites are not unknown in the Ira, but are more characteristic of the Netop in which they comprise locally as much as fifteen to twenty percent of the rock. [Gray or dark gray albitic phyllites are also characteristic of much of the Berkshire Schist as mapped on Mount Greylock by Pumpelly, Wolff and Dale (1894) and Herz (1960, 1961).

Though finer grained, these are much like the albite schists of the Hoosac Formation farther east (see Chang *et al.*, 1965; also Prindle and Knopf, 1932)]. Pale green, chloritic (but also highly albitic) phyllites occur in the lower part of the Netop Formation on Dorset Mountain. The principal occurrences of these have been shown separately on Plate 1.

Blue-gray limestones and dark-gray to black dolomites occur in the upper part of the Netop. The principal occurrences are shown on Plate 1, but these rocks are rarely well exposed and are probably more extensive. The topographic base-map rarely shows it, but this horizon is marked by a well defined bench or terrace, much of the way around the Dorset Mountain mass. Massive quartzites, grading into dolomitic quartzite and closely resembling the quartzites of the Hatch Hill Formation of the slate belt (Theokritoff, 1964) are associated with these carbonates at many localities. Many of the limestones are highly schistose with black phyllitic partings, but others are nearly pure carbonate. These rocks are entirely similar to the carbonates and quartzites mapped as Bellowspipe Limestone on Mount Greylock by Pumpelly, Wolff and Dale (1894), and to many of the types mapped as the Plymouth Member of the Hoosac Formation by Chang *et al.* (1965) east of the Green Mountains.

The Netop Formation is at least 400 feet thick on Dorset Mountain, but if the base is a tectonic contact, as believed, the total thickness may be much greater.

St. Catharine Formation

The St. Catharine Formation was named by Shumaker (1960) for exposures near Lake St. Catharine in the northwest part of the Pawlet quadrangle. A more complete description is given in Part I of this Bulletin, hence the statements here will apply only to its easternmost occurrences. The St. Catharine here is predominantly a green, chloritic phyllite, easily distinguished from the dark, albitic upper part of the Netop, although the contact, where well exposed, appears gradational over 100 feet or more. The St. Catharine usually stands out in bold outcrops above the less resistant Netop and forms the highest summits in the area. The lower part is albitic but less so than are typical Netop phyllites. The upper part of the St. Catharine in the Dorset Mountain mass is non-albitic and characterized by a deep green, almost blue-green, color from the presence of abundant small plates of chloritoid. These chloritoid phyllites are, though a bit coarser grained, virtually identical to the chloritoid phyllites of the Biddie Knob Formation of Zen (1961, 1964). The St. Catharine also contains rare beds of white to greenish quartzite, commonly with conspicuous grains or small pebbles of blue opalescent quartz. These quartzites are not unlike the Zion Hill Quartzite Member of the Bull Formation of Zen (1961), but are rarely more than a few tens of feet thick and cannot be traced far.

The lower, chloritoid-free part of the St. Catharine is about 300 feet thick. The chloritoid phyllites are at least 300 feet thick but their top is not exposed. The St. Catharine is co-extensive with rocks mapped by various workers in the northern Taconics as Bull, Stiles, Mettawee, and Mount Anthony (MacFadyen, 1956) or upper Mount Anthony (Hewitt, 1961). It also comprises a part of what was mapped as Berkshire by early workers, and is in all essentials much like the Greylock Schist of northwestern Massachusetts (Pumpelly, Wolff, and Dale, 1894; Herz, 1960, 1961). Strong resemblance should also be noted to parts of the Rowe Schist (Emerson, 1892) and the Pinney Hollow Formation (Chang *et al.*, 1965) east of the Green Mountains.

Age of the Netop and St. Catharine Formations

The assignment of ages to these formations is necessarily dependent on one's choice among the divergent hypotheses outlined above. The resemblance of the sequence Netop–St. Catharine to the sequence Hoosac–Pinney Hollow, as described by Chang *et al.* (1965) in the Woodstock quadrangle east of the Green Mountains, is striking. For reasons outlined by Chang *et al.* (1965), based on work in northern Vermont and southern Quebec (Cady, 1960; Osberg, 1965), it is unlikely that the Hoosac or Pinney Hollow formations can be younger than Early Cambrian.

STRUCTURE

Precambrian Structures

The rocks of the Mount Holly Complex were deformed, metamorphosed and deeply eroded before deposition of the Paleozoic sequence. However, the exposed Precambrian in the Pawlet quadrangle is too limited in areal extent to permit identification of specific Precambrian structural features, as has been possible (Brace, 1953) in the Rutland area immediately to the northeast.

Pre-Ira Structures

It has already been stated above that the base of the Ira Formation truncates the older units of the Champlain Valley Sequence at the south end of the Tinmouth Valley. Similar relationships have been shown farther north by Fowler (1950), Zen (1964), and Thompson (1959). Locally, along the Clark Mountain ridge, the Ira rests directly on the Mount Holly. A mile or less to the east, however, (see Plate 1; also Thompson, 1959) it rests again on Lower Ordovician carbonates except where separated from them by the Baker Brook Volcanics. On the lower north slopes of Dorset Mountain the Shelburne Marble and the Mount Holly Complex crop out within a few yards of each other, yet both are overlain directly, nearby, by the Ira Formation. Unless the entire Tinmouth Valley–Clark Mountain terrane is part of a large allochthonous block, for which there is no other evidence, these relations can be explained only by postulating a post-Bascom (and post-Beldens, since Beldens types are here mapped with the Bascom), pre-Ira fault. This fault may be located fairly closely on the north slopes of Dorset Mountain, as noted above, and also just east of the area of Plate 1, in the extreme south part of Clarendon, west of Otter Creek (see Thompson, 1959). The fault plane has nowhere been observed, at least knowingly, but a high-angle, normal fault seems likely. The fault surface, however, must undoubtedly have been deformed by later movements.

The hypothesis just outlined is attractive in that it explains not only the present geometric relations but also casts light on the nature of the pre-Ira, post-Bascom-Beldens disturbance. The Baker Brook Volcanics are also of interest in this context, both in that they occur in close proximity to this fault and in that their stratigraphic position suggests volcanic activity more or less contemporaneous with the faulting. The exact date of the faulting, which we shall refer to, at least in a local sense, as the *Tinmouth Disturbance*, is bracketed by the Beldens Member of the Chipman Formation (latest Canadian) and by the oldest faunas yet found in the Ira (Black River or Trenton, no older than Black River). The Baker Brook Volcanics are more closely associated with the Ira than with the Bascom (and Beldens), for these, and the underlying Shelburne, are cut out locally by the volcanics. The volcanics are possibly Chazyan, as noted above. It is also evident that the faulting must have preceded the deposition of the Ira by a time interval sufficient for the Precambrian of the upthrown block to be bared by erosion. All considered, an earliest Middle Ordovician age for the Tinmouth Disturbance seems likely, either earliest Chazyan or during a Canadian-Chazyan hiatus.

It should be noted here that these relationships are consistent with observations elsewhere in the Appalachians. The unconformity at the base of the Ira can probably be correlated with like features elsewhere in New England and New York (Dale, 1920; Bucher, 1957, p. 664). P. B.

King (1951, p. 121) has summarized evidence for similar events in the Central and Southern Appalachians. The bentonites in the Middle Ordovician of the Valley and Ridge Appalachians, furthermore, are clear evidence of expanded volcanic activity farther east.

The Taconic Thrust

The factors involved in locating the thrust at the eastern margin of the Taconic Allochthone are largely stratigraphic and regional in nature, and these have already been summarized above. No minor structures that can be unambiguously related to the emplacement of the allochthone have been identified in this area. The inferred thrust surface has been intensely deformed by later folding and thrusting, and obscured by subsequent regional metamorphism. According to Zen (1961) the allochthone, in detail a complex mass containing several slices emplaced at different times, was emplaced (at least the lower slices were) during the later stages of the mid-Ordovician (Ira) sedimentation. The regional metamorphism is clearly post-Ira and may be entirely or in part Acadian. Siluro-Devonian rocks are affected in eastern Vermont (Doll *et al.*, 1961) and in New Hampshire (Billings, 1956). Whatever the correct details may be concerning the emplacement, or even the dimensions, of the Taconic Allochthone, the rocks of the Taconic Sequence must have been deposited somewhere east of the rocks in the Vermont Valley, and probably west of the Paleozoic rocks overlying the Mount Holly Complex on the east side of the Green Mountains (Chang *et al.*, 1965).

Pine Hill Thrust and Chippenhook Anticline

The Pine Hill thrust was first recognized by Wolff (1891) and is thus of historic interest as one of the earlier-known features of its kind in the New England Appalachians. The name is from Pine Hill, in Proctor, several miles north of the north boundary of the Pawlet quadrangle. The fault enters the quadrangle near its northeast corner and appears to lose displacement southward. On the western slope of Clark Mountain it splits, for about a mile, into two branches, neither demanding more than a few hundred feet of displacement, and it cannot be recognized much farther south. The location of the fault is clear in most places except where it brings Ira against Ira as on the ridges north of Clark Mountain. The best exposures to show the detailed nature of the Pine Hill fault are in the southeast part of the Castleton quadrangle, near Chippenhook (Dale, 1891, 1984; Thompson, 1959, stop 2) and on Boardman Hill east of Clarendon Springs (Dale, 1891, 1894; Fowler, 1950; Brace, 1953; Zen, 1964). The actual contact may be seen at many localities on Boardman Hill, and it is clear there that the fault surface has been folded and crumpled by later deformation. A detailed map of the Boardman Hill–Chippenhook area has recently been made by the writer using a new topographic base at a scale of 1:24,000. This shows that the fault, though crumpled in detail, is otherwise not greatly deformed, and that it dips thirty degrees eastward, very nearly. Farther south, in the Pawlet quadrangle, the dip is probably steeper. The Pine Hill thrust is clearly post-Ira and older than some of the deformation, but cannot otherwise be dated.

The Chippenhook anticline was named by Fowler (1950) for exposures near Chippenhook where it is clearly displayed (Thompson, 1959, stop 2; Dale, 1891, 1894). Its axial trace may be followed south from there, well exposed nearly all the way in Cheshire Quartzite, to the valley of Baker Brook. The trace of the Pine Hill thrust is approximately paralled to it and several hundred yards to the east. On Clark Mountain there are other anticlinal axes east of the Chippenhook anticline and similar in nature. The Pine Hill thrust, and its branches there, appear to be break thrusts related to these folds. Bedding attitudes in the Tinmouth Valley carbonates, coupled with cleavage relations, indicate an easterly dip of about fifty degrees for the axial surface of the Chippenhook anticline.

Danby Folds

The sigmoid pattern made by the formations in the Vermont Valley at and near Danby is a striking feature (Plate 1; Thompson, 1959; Doll *et al.*, 1961). The anticline and its conjugate syncline plunge gently southwest beneath Dorset Mountain. The Shelburne Marble, in particular, is greatly thickened by flowage in the synclinal part of this fold and this has made the Danby area a major producer of marble. Detailed analyses of the minor structures related to these folds have been presented by G. W. Bain (1931, 1933a, 1933b, 1938).

Dorset Mountain Nappe

At many localities north and west of the Dorset Mountain mass, limestones and dolomites of the Bascom Formation can be seen resting directly upon a variety of different rocks, some younger, some older. A few yards northwest of the road corner at elevation 1620 feet on the north slopes of Dorset Mountain, the Bascom rests on Mount Holly gneisses. A short distance west it rests on Ira phyllites, and in the notch between Dorset and Woodlawn mountains it rests on phyllites of the St.

Catharine Formation (see Thompson, 1959; stops 7, 10, and 11). East of the hill road north of Dorset (Thompson, 1959, stop 11) a small isolated knoll of Bascom sits on a pavement of St. Catharine phyllite. No matter what view one takes as to the geologic age of the St. Catharine Formation, these exposures are clear evidence for a major thrust fault with a displacement of at least three miles, possibly much more. The outcrops north of Dorset Village were indeed regarded by C. E. Gordon (1921, p. 212 *et seq.*) as showing a normal stratigraphic sequence, but this view is no longer tenable in the light of the stratigraphic and structural information now at hand.

East of the locality mentioned above, near the road corner, the thrust must pass eastward into a bedding-plane thrust within the Bascom Formation. This makes it virtually impossible to locate the thrust precisely along the eastern slopes of Dorset and Netop mountains, and on Mount Aeolus (Green Peak). It is, in fact, quite possible that a discrete thrust plane does not exist in the marbles on the east side of the massif, even though one has been drawn tentatively on Plate 1, and that the movement is there taken up instead by intense flowage in the marbles. The deformation within the Bascom Formation in the Dorset Mountain area is remarkable, and in Dorset Hollow the Shelburne is also affected. The lower course of nearly every brook on Dorset Mountain that cuts through the carbonates has exposures of highly attenuated recumbent folds, many of large amplitude. The lobe of Shelburne Marble that contains the high quarries on Owl's Head in Dorset occupies the core of a large recumbent fold (Thompson, 1959; Hewitt, 1961). The spectacular structures visible in the walls of the quarries have been remarked upon by earlier investigators (Dale, 1902, 1912; Bain, 1931). An even more striking display may be seen, however, in the middle branch of the brook, southeast of Dorset Peak, that enters the Vermont Valley near North Dorset. A more accessible exposure showing a similar style of deformation in what is possibly the same tectonic zone, may be seen farther south, in the Equinox quadrangle, on the north side of the highway along the Battenkill, west of Arlington (Hewitt, 1961, Pl. 4). We shall refer to the entire overriding mass as the *Dorset Mountain nappe*, bearing in mind that although it is floored by a discrete thrust surface in many places, this may not be so everywhere.

The emplacement of the Dorset Mountain nappe is probably the latest deformational event recorded in this area. Where well exposed west of the mountain the thrust surface shows no sign of later folding or crumpling. The final uplift of the Green Mountain anticlinorium probably took place during the Acadian orogeny. Siluro-Devonian rocks dip eastward, off the Green Mountain anticlinorium, in eastern Vermont (Doll *et al.*, 1961). It is quite possible that the Dorset Mountain nappe represents a gravity slide off the Green Mountain front during the Acadian movement, but this cannot be proved.

Minor Structures

Most of the more significant minor structural observations are recorded on Plate 2. Cleavage phenomena are omnipresent in the phyllites here and in those described in Part I of this Bulletin. Dale (1896, 1902, and elsewhere) has discussed these features extensively.

The minor folds in the Dorset Mountain nappe have already been described. Minor folds elsewhere generally mimic in style the major structures with which they are associated, but not invariably. Discordant minor folds are most frequently encountered (Bain, 1931) in the calcite marbles. These have the appearance in places of having flowed like warm butter.

Boudinage occurs locally, the most striking examples occurring in the interbedded dolomite and calcite marbles of the Bascom Formation in the north part of Danby, and also farther north, in the Wallingford quadrangle, near the South Wallingford quarries.

METAMORPHISM

All of the rocks in the eastern part of the Pawlet quadrangle have undergone a low-grade regional metamorphism. The effects are most conspicuous in the phyllites and schists, giving a misleading first impression that the phyllites of the Ira Formation and in the Taconic Sequence are more metamorphosed than the carbonates and quartzites beneath. Phyllites and schists interbedded with the carbonates, however, have textures and mineral assemblages entirely consistent with those in the formations that are made almost entirely of such rocks.

Zen (1960) has made an extensive study of the regional metamorphism in the northern Taconics. The mineral assemblages reported by him from the eastern Taconics in the Castleton quadrangle, and those reported by Shumaker (Part I of this Bulletin), are not unlike those found in the eastern part of the Pawlet quadrangle. The rocks here are, however, more coarsely crystalline. Chloritoid plates in the rocks described by Zen can barely be identified as such with the aid of a hand lens, but on the east end of Dorset Mountain plates 2–4 mm in diameter may be found at several localities. Some of these rocks might well be called schists rather than phyllites. The marbles in Danby and Dorset are generally coarser than those from quarries eslewhere. The West Rutland marbles, according to Dale (1912) rarely exceed 0.5 mm in grain size, but in the Danby–Dorset area the grain size commonly exceeds 1.0 mm, locally reaching 2.5 mm.

Minerals formed by metamorphism of the carbonate rocks include talc, phlogopite (or a light brown biotite), actinolite (as rare, slender prisms that may be as much as 2.0 cm long in some of the marbles at Danby and Dorset), and tourmaline. Zen (1960) reports zoisite in the marbles at West Rutland, but it has not yet been identified here. Epidote, however, is common in the greenstones of the Baker Brook Volcanics, with chlorite, albite and a green actinolitic amphibole.

The key assemblages in the phyllites are: quartz–muscovite–paragonite–chlorite–chloritoid (in the upper part of the St. Catharine on Dorset Mountain and nearby summits); and quartz–muscovite–albite–chlorite–biotite elsewhere. The more typical assemblages, however, contain fewer minerals. Paragonite has not been found with biotite, and neither biotite nor albite has been found with chloritoid, hence these combinations are assumed incompatible in this area. Ilmenite is common and may occur as plates 1–2 mm in diameter. It is easily mistaken for biotite in hand specimens of some of the carbonaceous phyllites of the Taconic Sequence west of the Tinmouth Valley. Magnetite and (less commonly) hematite occur in the non-carbonaceous phyllites. The purple color characteristic of some of the slates and phyllites of the St. Catharine Formation in the western part of the quadrangle is not observed in the Dorset Mountain area. This is believed due in part to reaction of hematite with hydrous ferrous silicates yielding magnetite and water, and in part to coarsening of the hematite that survives. The absence here of the red color typical of the quartzites in the Monkton in areas to the north is almost certainly a related feature. Small plates of specular hematite may be seen in some hand specimens of these quartzites collected north of Danby.

Coexisting muscovite and paragonite in a chloritoid phyllite collected near the old lookout tower on Dorset Mountain have basal spacings of 9.981 A° and 9.623 A°, respectively, as reported by Rosenfeld, Thompson, and Zen (1958). Using the regression formula of Zen and Albee (1964) these give values of the ratio: Na/Na +K, of 0.12 for the muscovite and 0.96 for the paragonite, indicating a very limited mutual solubility compared with higher grade occurrences east of the Green Mountains.

The assemblages are typical of the biotite zone or upper (actinolitic) green schist facies. The metamorphism cannot be dated precisely, but is clearly younger than the emplacement of the Taconic Allochthone or the movement on the Pine Hill thrust. Fine, relict lamination is preserved in albite porphyroblasts that have been rotated. This suggests that the metamorphism overlapped some of the later deformation.

ECONOMIC GEOLOGY

Pits for sand and gravel have been opened in terraces along the west side of the Vermont Valley in Danby, and in an esker following the course of the Tinmouth Channel. Iron ores, reported (Hitchcock *et al.*, 1861, p. 821; Morrill and Chaffee, 1957) as limonite—hematite ores with a relatively high content of manganese, were once worked at several localities in Tinmouth but are no longer of importance. These are associated with deposits of "kaolin" and "ochre" and probably formed by deep weathering of ferruginous dolomite. The underlying rocks are either basal Dunham Dolomite or Winooski Dolomite.

The major mineral resource of the area is marble. The first quarries were opened in Dorset in the late eighteenth century and the quarries in Danby have been major producers in recent years. The principal quarries are in the Shelburne Marble, but there has been minor production from the Sutherland Falls Member of the Clarendon Springs Formation and from the Bascom Formation. For detailed descriptions of specific quarries the reader is referred to papers by T. N. Dale (1912) and G. W. Bain (1931, 1933a, 1933b, 1938).

REFERENCES CITED

BAIN, G. W., 1927, Geologic history of the Green Mountain Front: Vermont State Geol., 15th Report, p. 222–241.

—— 1931, Flowage Folding: Am. Jour. Sci., 5th ser., v. 22, p. 503–530.

—— 1933a, The Vermont marble belt: 16th Internat. Geol. Cong., Guidebook 1 (Eastern New York and western New England), p. 75–80.

—— 1933b, Brandon to Bennington, Vermont: 16th Internat. Geol. Cong., Guidebook 1 (Eastern New York and western New England), p. 80–87.

—— 1938, The central Vermont marble belt: New England Intercoll. Geol. Conf., Guidebook, 34th Ann. Mtg., Rutland, Vermont, 31 p.

BILLINGS, M. P., RODGERS, J., and THOMPSON, J. B., JR., 1952, Geology of the Appalachian Highlands of east-central New York, southern Vermont, and southern New Hampshire—Field trip no. 1, in Geol. Soc. Am. Guidebook for field trips in New England, p. 1–71.

BILLINGS, M. P., 1956, The Geology of New Hampshire, Part II—Bedrock geology: New Hampshire Plan. and Devel. Comm., 203 p.

BRACE, W. F., 1953, The Geology of the Rutland area, Vermont: Vermont Geol. Survey Bull. 6, 120 p.

BUCHER, W. H., 1957, Taconic Klippe: A stratigraphic-structural problem: Geol. Soc. Am. Bull., v. 68, p. 657–674.

CADY, W. M., 1945, Stratigraphy and structure of west-central Vermont: Geol. Soc. Am. Bull., v. 56, p. 515–587.

—— 1960, Stratigraphic and geo-tectonic relationships in northern Vermont and southern Quebec: Geol. Soc. America Bull., v. 71, p. 531–576.

CADY, W. M. and ZEN, E-AN, 1960, Stratigraphic relationships of the Lower Ordovician Chipman Formation in west-central Vermont: Am. Jour. Sci., v. 258, p. 728–739.

CHANG, P. H., ERN, E. H., JR., and THOMPSON, J. B., JR., 1965, Bedrock geology of the Woodstock quadrangle, Vermont: Vermont Geol. Survey Bull. 29, 65 p.

CLARK, T. H., 1931, Lowest Cambrian of southern Quebec (abs.): Geol. Soc. America Bull., v. 42, p. 225–226.

—— 1934, Structure and stratigraphy of southern Quebec: Geol. Soc. America Bull., v. 45, p. 1–20.

DALE, T. N., 1891, On the structure and age of the Stockbridge limestone in the Vermont Valley: Geol. Soc. America Bull., v. 3, p. 514–519.

—— 1894, On the structure of the ridge between the Taconic and Green Mountain ranges in Vermont: U. S. Geol. Survey, 14th Ann. Rept., Pt. 2, p. 525–549.

——1896, Structural details in the Green Mountain region and in eastern New York: U. S. Geol. Survey, 16th Ann. Rept., Pt. 1, p. 543–570.

—— 1902, Structural details in the Green Mountain region and in eastern New York (second paper): U. S. Geol. Survey Bull., 195, 22 p.

—— 1912, The commercial marbles of western Vermont: U. S. Geol. Survey Bull., 521, 170 p.

—— 1920, Local unconformity between the Berkshire schist and the Stockbridge limestone in Adams, Mass.: Am. Jour. Sci., 4th ser., v. 49, p. 369–371.

DOLL, C. G., CADY, W. M., THOMPSON, J. B., JR., and BILLINGS, M. P., 1961, Centennial Geologic Map of Vermont: Vermont Geol. Survey.

—— 1963, Reply to Zen's discussion of the Centennial Geologic Map of Vermont: Am. Jour. Sci., v. 261, p. 94–96.

EMERSON, B. K., 1892, Outlines on geology of the Green Mountain region in Massachusetts; Hawley sheet, descriptive text: U. S. Geol. Survey, Geol. Atlas, Hawley sheet, 3 p.

—— 1917, Geology of Massachusetts and Rhode Island: U. S. Geol. Survey Bull. 597, 289 p.

FISHER, D. W. and HANSON, G. F., 1951, Revisions in the geology of Saratoga Springs, New York and vicinity: Am. Jour. Sci., v. 249, p. 795–814.

FISHER, D. W., 1962, Correlation of the Ordovician rocks in New York State: N. Y. State Mus. and Sci. Service, Geol. Survey, Map and Chart Series: No. 3.

FLOWER, R. H., 1964, The nautiloid order *Ellesmeroceratida* (cephalopoda): New Mexico Bur. Mines and Min. Resources, Mem. 12, 234 p.

FOERSTE, A. P., 1893, New fossil locations in the early Paleozoics of Pennsylvania, New Jersey, and Vermont, with remarks on the close similarity of the lithologic features of these Paleozoics: Am. Jour. Sci., 3rd ser., v. 46, p. 435–444.

FOWLER, PHILLIP, 1950, Stratigraphy and structure of the Castleton area, Vermont: Vermont Geol. Survey Bull. 2, 83 p.

GORDON, C. E., 1921, Studies in the geology of western Vermont: Vermont State Geol., 12th Rept., p. 114–279.

—— 1925, Studies in the geology of western Vermont (third paper): Vermont State Geol., 14th Report, p. 218–259.

HERZ, NORMAN, 1960, Bedrock geology of the Cheshire quadrangle, Massachusetts: U. S. Geol. Survey, Geol. Quad. Map. GQ 108.

—— 1961, Bedrock geology of the North Adams quadrangle, Massachusetts–Vermont: U. S. Geol. Survey, Geol. Quad. Map. GQ 139.

HEWITT, P. C., 1961, The geology of the Equinox quadrangle and vicinity, Vermont: Vermont Geol. Survey Bull. 18, 83 p.

HEWITT, P. C., and MACFADYEN, J. A., JR., 1963, Discussion: A clarification of the MacFadyen-Hewitt position on the Taconics: Am. Jour. Sci., v. 261, p. 894–896.

HITCHCOCK, E., HITCHCOCK, E., JR., HAGER, A. D., and HITCHCOCK, CHARLES, 1861, Report on the geology of Vermont, Claremont, New Hampshire, 982 p.

KAISER, E. P., 1945, Northern end of the Taconic thrust-sheet in western Vermont: Geol. Soc. America Bull., v. 56, p. 1079–1098.

KEITH, ARTHUR, 1923, Cambrian succession of northwestern Vermont: Am. Jour. Sci., 5th ser., v. 5, p. 97–139.

—— 1932, Stratigraphy and structure of northwestern Vermont: Washington Acad. Sci. Jour., v. 22, p. 357–379, 393–406.

—— 1933, Outline of the structure and stratigraphy of northwestern Vermont: 16th Internat. Geol. Cong., Guidebook 1 [Eastern New York and western New England], p. 48–61.

KINDLE, C. H. and TASCH, PAUL, 1948, Lower Cambrian fauna of the Monkton Formation of Vermont: Canadian Field-Naturalist, v. 62, p. 133–139.

KING, P. B., 1951, The tectonics of middle North America: Princeton Univ. Press, 1951, 203 p.

MACFADYEN, J. A., JR., 1956, The geology of the Bennington area, Vermont: Vermont Geol. Survey Bull. 7, 72 p.

MORRILL, PHILIP and CHAFFEE, R. G., 1957, Vermont mines and mineral localities, Part I, Southern Vermont: Dartmouth College Museum, 43 p.

OSBERG, P. H., 1952, The Green Mountain anticlinorium in the vicinity of Rochester and East Middlebury, Vermont: Vermont Geol. Survey Bull. 5, 127 p.

—— 1965, Structural geology of the Knowlton-Richmond area, Quebec: Geol. Soc. America Bull., v. 76 p. 223–250.

PERKINS, G. H., 1908, Preliminary report on the geology of Chittenden County: Vermont State Geol., 6th Rept. p. 221–264.

PRINDLE, L. M. and KNOPF, E. B., 1932, Geology of the Taconic quadrangle: Am. Jour. Sci., 5th ser., v. 24, p. 257–302.

PUMPELLY, R., WOLFF, J. E., DALE, T. N., 1894, Geology of the Green Mountains in Massachusetts: U. S. Geol. Survey Mon. 23, 206 p.

RODGERS, JOHN, 1937, Stratigraphy and structure in the upper Champlain Valley: Geol. Soc. America Bull., v. 48, p. 1573–1588.

ROSENFELD, J. L., THOMPSON, J. B., JR., and ZEN, E-AN, 1958, Data on coexistent muscovite and paragonite (abs.): Geol. Soc. America Bull., v. 69, p. 1637.

SAHAKIAN, A. S., 1963, Paleocurrent study of the Potsdam Sandstone of New York and Quebec: Doctoral thesis, Harvard University, 118 p.

SCOTT, JOHN, 1959, Caves in Vermont: Killooleet, Hancock, Vermont, 45 p.

SHAW, A. B., 1958, Stratigraphy and structure of the St. Albans area, northwestern Vermont: Geol. Soc. America Bull., v. 69, p. 519–567.

—— 1962, Paleontology of northwestern Vermont. IX. Fauna of the Monkton Quartzite. Jour. Paleontology, v. 36, p. 322–345.

SHUMAKER, R. C., 1959, Pawlet quadrangle, Trip G, Part 2: New England Intercoll. Geol. Conf., Guidebook 51st Ann. Mtg., Rutland, Vermont, p. 59–60.

—— 1960, Geology of the Pawlet quandrangle, Vermont: Doctoral thesis, Cornell Univ., 109 p.

TASCH, PAUL, 1949, A new fossil locality in the Lower Cambrian Monkton Formation of Vermont: Canadian Field-Naturalist, v. 13, p. 210–211.

THEOKRITOFF, GEORGE, 1964, Taconic stratigraphy in northern Washington County, New York: Geol. Soc. America Bull., v. 75, p. 171–190.

THOMPSON, J. B., JR., 1959, Stratigraphy and structure in the Vermont Valley and the eastern Taconics between Clarendon and Dorset, Trip H: New England Intercoll. Geol. Conf., Guidebook, 51st Ann Mtg., Rutland, Vermont, p. 71–87.

TOULMIN, PRIESTLY, 3RD, 1961, Geological significance of lead-alpha and isotopic age determinations of "alkalic" rocks of New England: Geol. Soc. America Bull., v. 72, p. 775–780.

WALCOTT, C. D., 1888, The Taconic system of Emmons, and the use of the name Taconic in geologic nomenclature: Am. Jour. Sci., v. 35., p. 229–242, 307–327, 394–401.

WALTON, M. S. and DE WAARD, D., 1963, Orogenic evolution of the Precambrian in the Adirondack Highlands, a new synthesis. Proc. Kon. Ned. Akad. Wetensch., Amsterdam, B, 66, p. 98–106.

WELBY, C. W., 1961, Bedrock geology of the Central Champlain Valley of Vermont: Vermont Geol. Survey, Bull. 14, 296 p.

—— 1965, A new Ordovician species of *Lichenaria* from Vermont: Jour. Paleontology, v. 39, p. 288–289.

WHITTINGTON, H. B. and WILLIAMS, ALWYN, 1964, The Ordovician Period: Quart. Jour. Geol. Soc. London, v. 1205, p. 241–254.

WHITTLE, C. L., 1894, The occurrence of Algonkian rocks in Vermont and the evidence for their subdivision: Jour. Geol., v. 2, p. 396–429.

WOLFF, J. E., 1891, On the Lower Cambrian age of the Stockbridge limestone: Geol. Soc. America Bull., v. 2, p. 331–337.

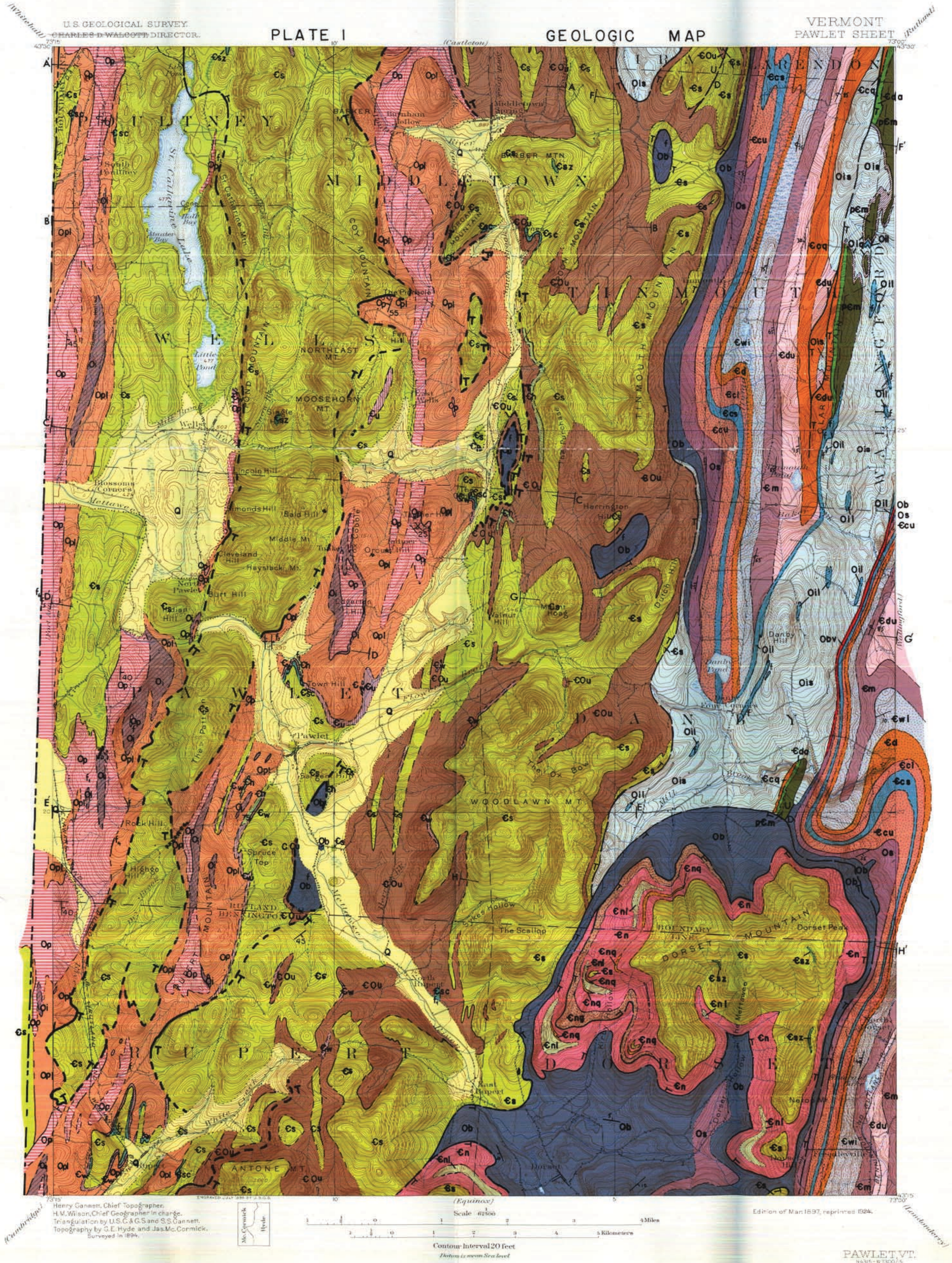
ZEN, E-AN, 1960, Metamorphism of Lower Paleozoic rocks in the vicinity of the Taconic Range in west-central Vermont: Am. Mineral., v. 45, p. 129–175.

—— 1961, Stratigraphy and structure at the north end of the Taconic Range in west-central Vermont: Geol. Soc. America Bull., v. 72, p. 293–338.

—— 1963, Age and classification of some Taconic stratigraphic units on the Centennial Geologic Map of Vermont: A discussion: Am. Jour. Sci., v. 261, p. 92–94.

—— 1964, Stratigraphy and structure of a portion of the Castleton quadrangle, Vermont: Vermont Geol. Survey Bull. 25, 70 p.

Zen, E-AN and Albee, A. L., 1964, Coexistent muscovite and paragonite in pelitic schists: Am. Mineral, v. 49, p. 904–925.



EXPLANATION

QUARTERNARY — Q — Thickest Alluvium and Outwash.

IGNEOUS ROCKS

PERMIAN OR YOUNGER — Mafic Dikes

STRATIFIED ROCKS

TACONIC SEQUENCE

MIDDLE ORDOVICIAN
- Op — Pawlet Graywacke (and black slate).
- Oi — Indian River Slate.

LOWER ORDOVICIAN
- Opl — Poultney Slate

CAMBRIAN OR ORDOVICIAN
- €Ou — Undifferentiated black slates (mainly Hatch Hill and West Castleton Formations of Taconic Sequence, but may include some of Ira Formation of Champlain Valley Sequence).

UPPER CAMBRIAN
- €h — Hatch Hill Formation
- €u — West Castleton and Hatch Hill Formations, undivided.

MIDDLE CAMBRIAN
- €w — West Castleton Formation

LOWER CAMBRIAN
- €s — St. Catharine Formation: €s; including Zion Hill Quartzite, €sz; and Castleton Conglomerate, €sc.
- €n — Netop Formation: €n; including quartzite, €nq; areas with abundant carbonates, €nl; and green albitic phyllite, €ng.

PRECAMBRIAN

CHAMPLAIN VALLEY SEQUENCE
- Ois — Ira Formation: slate, Ois; limestone, Oil; pebbly quartzite, Oiq.
- Obv — Baker Brook Volcanics.
- Ob — Bascom Formation (and Beldens Fm.).
- Os — Shelburne Marble.
- €cu — Clarendon Springs Fm.: upper dolomite €cu; Sutherland Falls Member, €cs; lower dolomite, €cl.
- €d — Danby Formation
- €wi — Winooski Dolomite.
- €m — Monkton Quartzite
- €du — Dunham (Rutland) Dolomite.
- €cq — Cheshire Quartzite.
- €da — Dalton Formation.
- p€m — Mount Holly Complex

FOSSIL LOCALITIES

f_2 Numbers, if present, are for reference in text.

CONTACTS

Good control.

Fair control.

Inferred.

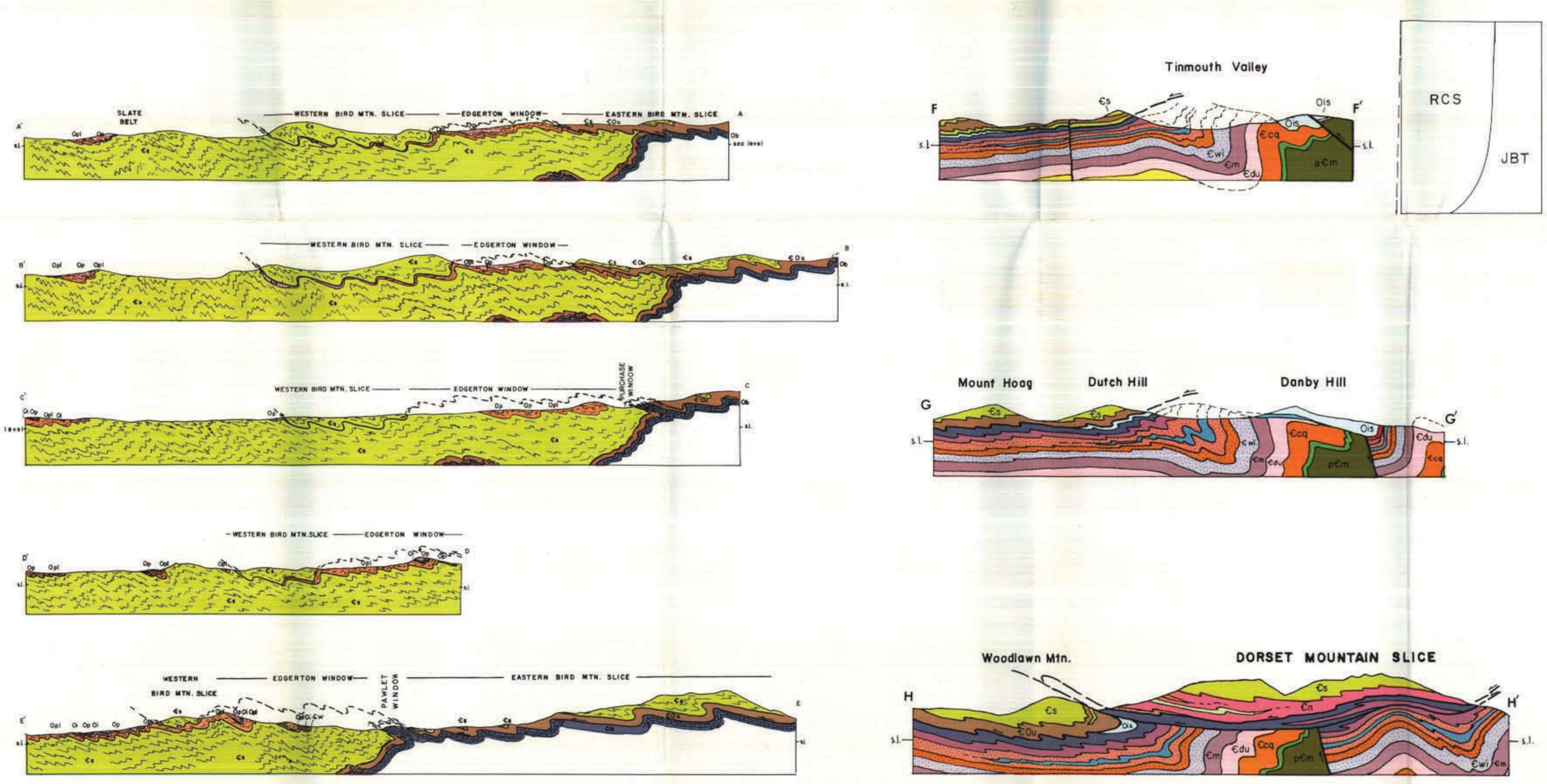
Thrust fault (T on upper plate, dashes and dots as above).

High-angle fault (U on upthrown side).

SYMBOLS

45 Strike and dip of bedding

65 Strike and dip of overturned bedding



Key to mapping: Geology by R. C. Shumaker, 1957–59; and J. B. Thompson, Jr., 1951–65, but mainly 1951–52 and 1958–59. Assistants: J. B. Rather and H. L. Thung (R.C.S.), G. A. Mairs, G. J. F. MacDonald, S. M. Ornstein, and E-an Zen (J. B. T.).

VERMONT GEOLOGICAL SURVEY
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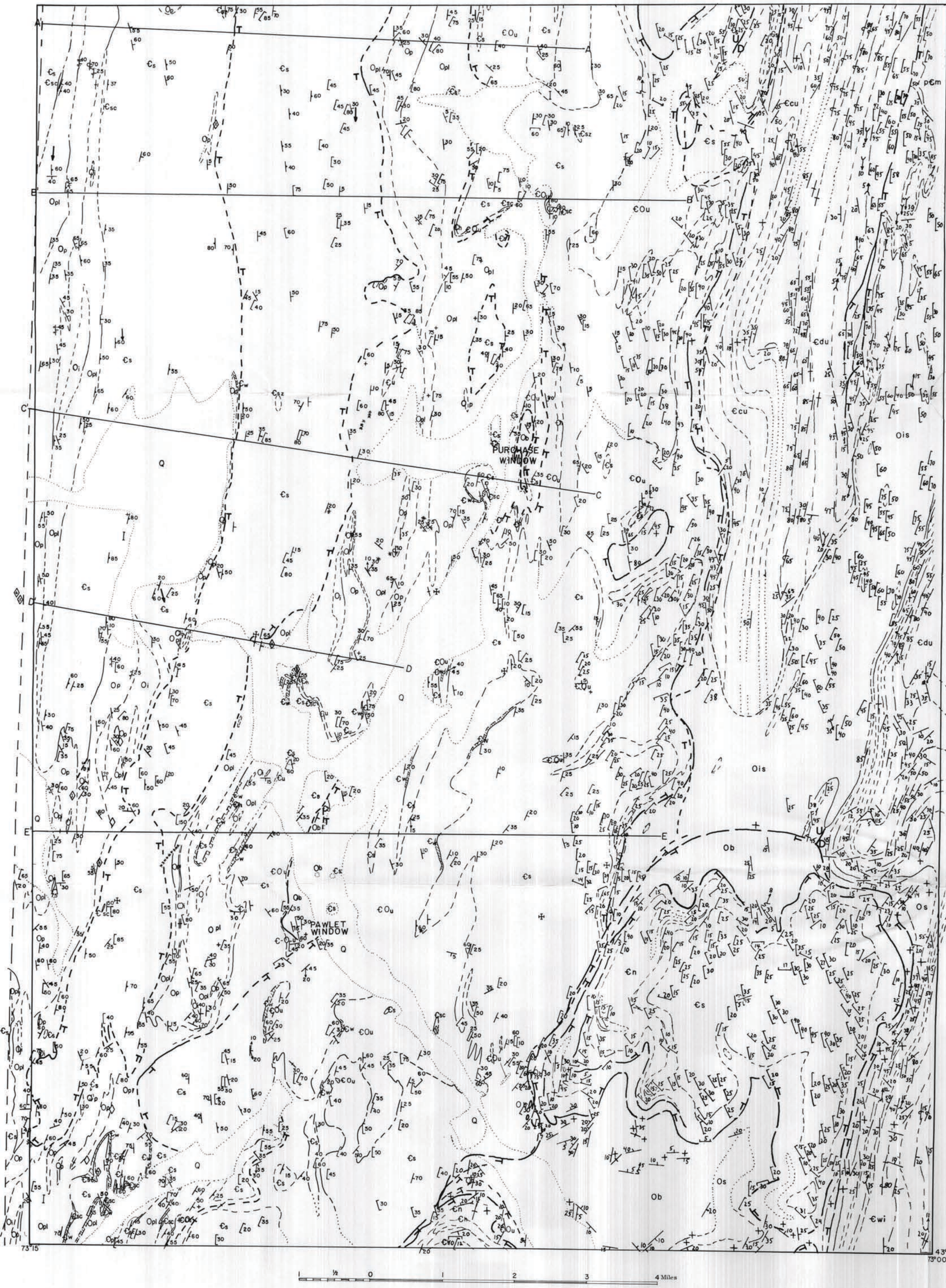
TECTONIC MAP OF THE PAWLET QUADRANGLE, VERMONT

PLATE 2



EXPLANATION

STRIKE AND DIP OF BEDDING AND BEDDING SCHISTOSITY

Inclined Vertical Horizontal

STRIKE AND DIP OF GRADED BEDDING

Normal Overturned

STRIKE AND DIP OF FLOW CLEAVAGE

Inclined Vertical Horizontal

STRIKE AND DIP OF FRACTURE CLEAVAGE

Inclined Vertical Horizontal

BEARING AND PLUNGE OF MINOR FOLD OR CRINKLE AXIS

CONTACT

Solid where definite, dashed where projected and dotted where inferred

THRUST FAULT

T On upper plate. Solid where definite, dashed where projected and dotted where inferred

HIGH-ANGLE FAULT

U on upthrown side, solid where definite, dashed where projected and dotted where inferred

BEARING OF GLACIAL STRIATIONS

FOSSIL LOCALITY

FORMATION SYMBOLS AS SHOWN ON PLATE I

By R. C. Shumaker, 1957-59;
and J. B. Thompson, Jr., 1951-65

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