FIGURE 1. SNAKE MOUNTAIN AND NORTHERN RED SANDROCK RANGE
Aerial view showing concentric relation to Adirondack border. Photograph by George Lathrop, Bristol, Vermont.

FIGURE 2. THE GREEN MOUNTAIN FRONT
Aerial view north over Lake Dunmore. Photograph through courtesy of W. Storrs Lee, The Middlebury College Press.

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STRATIGRAPHY AND STRUCTURE OF WEST-CENTRAL VERMONT

BY WALLACE M. CADY

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ILLUSTRATIONS

ABSTRACT

The lithologic units recognizable in the fossiliferous succession along southern Lake Champlain are structurally continuous with and traceable eastward into the "marble belt" of west-central Vermont immediately west of the Green Mountain Front. They are also traceable northward through west-central Vermont into a succession in northwestern Vermont bounded on the east and west by major thrusts, where they pass laterally northeastward into fossiliferous shales, the faunal zones of which are correlated with those along southern Lake Champlain and in the Hudson and Mohawk valleys. The Cambrian strata are traced into west-central from northwestern Vermont whereas the Ordovician correlation is with rocks in the Mohawk-Hudson-Champlain region. Certain of the Upper Cambrian strata can be correlated with established formations of this age in both of the outlying areas.

The structural pattern reflects movements dependent upon the original distribution of sedimentary facies. Interbedded Cambro-Ordovician limestones and dolomites grade westward into foreland sandstones and eastward into geosynclinal shales. Cambrian and early Ordovician sandstone tongues extend far to the east. Later Ordovician strata of the shale facies overlie the calcareous and sandy succession and are locally unconformable on it. In the Taconic Range allochthonous Cambrian strata of shale facies are superposed upon the autochthonous Ordovician beds. The rocks of the klippe were derived from a zone at least 50 miles east of the present westernmost exposures, arriving there by movements confined largely to the shale facies. Flexural folding and thrusting, effects of alternating competency and incompetency of the foreland sequence, affected the latter succession and possibly the Taconic Allochthone. Breaking of competent strata in the flexures initiated the thrusts; thrusting continued by the stripping of competent from incompetent beds. This was accompanied by counterclockwise rotation of the thrust slices around pivotal zones bordering on a foreland massif to the southwest. Thus rock cropping out in the north-south thrust slices originally extended northeast-southwest. The thrusts are concentric to the Adirondack crystallines and are cut by the Adirondack normal faults; probably they were warped during normal faulting and uplift of the Adirondacks.

INTRODUCTION

The area (Fig. 1) of this study, totaling about 500 square miles, extends northward from the north end of the Taconic Range approximately 70 miles to a point 20 miles south of the Canadian border. The Red Sandrock Range bounds it on the west (Pl. 1, fig. 1) the Green Mountain Front on the east (Pl. 1, fig. 2), making the area 5 to 10 miles wide. Northward to the Canadian border and beyond lies the established Cambro-Ordovician section of northwestern Vermont. The standard Upper Cambrian-Lower and Middle Ordovician section of the Champlain Valley is to the west and southwest. Only by solving the structural complexities of west-central Vermont can the geologist test the continuity of the strata from the Champlain Valley section to northwestern Vermont and to the "marble belt" at the west foot of the Green Mountain Front.

Therefore, the author first describes and correlates the deformed rocks between the Red Sandrock Range and the Green Mountain Front. Then the position of this area in the regional geologic setting must be established and regional interpretations are suggested.

Stratigraphic work is inhibited by structural complications, metamorphism, and lateral gradation of both primary and secondary facies. Mapping, although made difficult by the widespread drift cover, is an important aid in establishing the stratigraphic sequence. Over 70 years ago, by walking out the beds, the Rev. Augustus Wing (Seely, 1901, p. 1–8) arrived at what is essentially the stratigraphic and structural interpretation of the present paper. The author used Wing's five notebooks and checked the data in them expanding the area of Wing's study northward.

Field study was done during the spring of 1934, the summers of 1935, 1937, and 1938, and interruptedly in 1939, 1940, and 1941. Much time was devoted to proving the physical continuity of the various lithologic units by means of areal mapping on a scale of 1 inch to the mile. The area is necessarily large because stratigraphic correlation between three areas at considerable distance from each other is involved, and because extensive studies could provide more data on which to establish the correlation and the mode of genesis of the lithologic units. Details of structure and stratigraphy have been worked out in critical localities.

ACKNOWLEDGMENTS

Professor G. Marshall Kay of Columbia University has supervised field work and directed the organization of this paper. The author has also been guided by the inspiration of friends, consultation with associates, and advice of those who already have done considerable work in this area and in related studies. The teaching of Prof. Bruno M. Schmidt of Middlebury College was inspiring. The interest and co-operation of the author's family have made continuation of the work possible. Prof. A. L. Howland of Northwestern University accompanied the writer in the field and gave valuable advice during the early part of the work. The author is also indebted to professors C. H. Behre, J. R. Ball, and W. H. Haas for advice given while at Northwestern. Stimulating discussions with Prof. Walter Bucher have on many occasions aided the author to get a better grasp of the problem.

The author wishes to express his appreciation for the time and interest given by G. W. Bain, Robert Balk, Marland Billings, Josiah Bridge, G. Arthur Cooper, H. N.

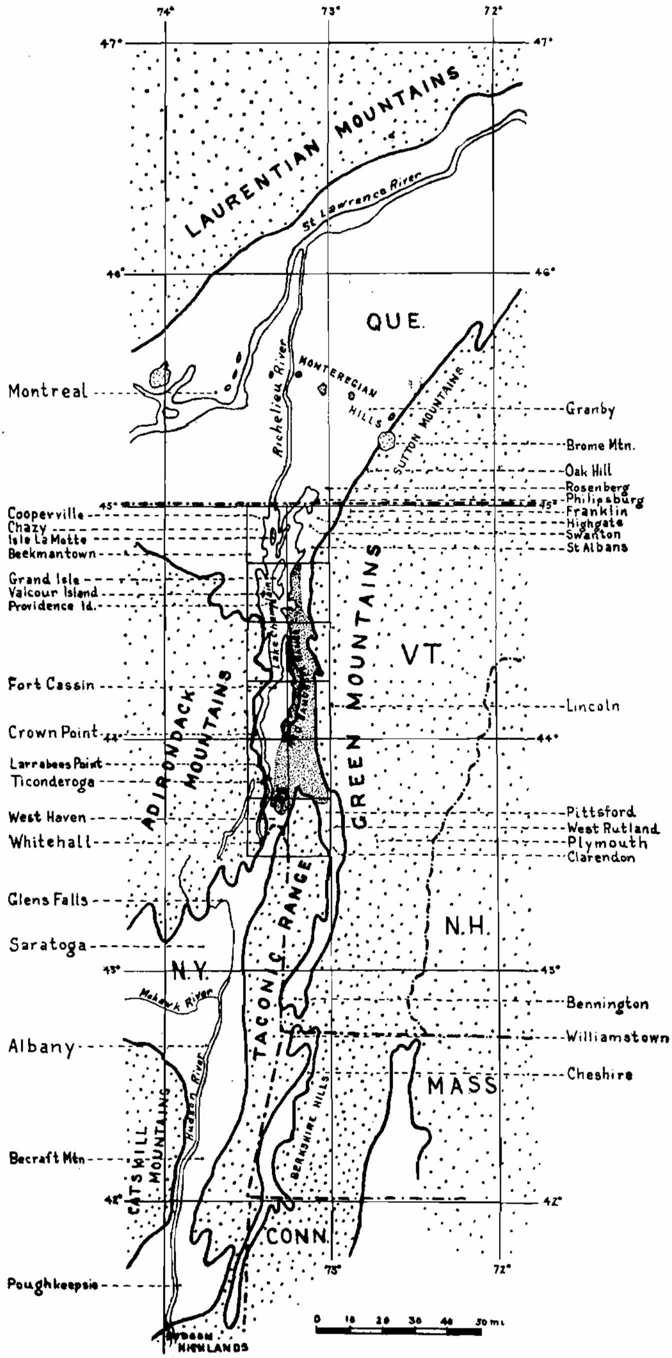


FIGURE 1.—*Geographic location of west-central Vermont*

Coryell, B. F. Howell, M. King Hubbert, Elbridge C. Jacobs, Arthur Keith, E. B. Knopf, Andrew Leith, A. K. Lobeck, and P. E. Raymond. Bain has pointed out important corrections in the author's areal mapping in the vicinity of Brandon indicated by diamond drilling records. Bridge and Coryell have kindly examined and helped in the identification of fossils from the relatively unfossiliferous eastern area. Professor Jacobs has had a former student, David Hawley, check the author's work in the vicinity of Burlington. Keith has allowed the writer to examine his numerous maps and field notes in advance of publication.

The author has collaborated with several others whose investigations have been in or adjacent to the area of west-central Vermont. To Verne Booth, Robert Cushman, E. P. Kaiser, Herbert E. Hawkes, Jr., Richard Jahns, John Rodgers, Frank Stead, William D. Stull, and Robert Wheeler he is indebted for valuable suggestions.

The aerial photograph taken in the winter of 1940 is by George Lathrop of Bristol, Vermont, without whose kind co-operation it would not have been available. Others are furnished by the courtesy of W. Storrs Lee of the Middlebury College Press.

The writer is further indebted to Helen R. Cady who assisted in the technical compilation of the manuscript.

PREVIOUS WORK

The major topographic features bounding this area, the Red Sandrock Range, the Green Mountain Front, and the Taconic Range, are developed on great thicknesses of resistant and lithologically distinctive rock; because the exposures are good, the distribution and general structural relations of the formations have been easily recognized. The Lower Cambrian age of the red sandstone underlying Snake Mountain and the low-angle thrust dipping gently eastward beneath (Pl. 6, fig 1) were recognized at an early date by tracing the critical units southward through the Red Sandrock Range from fossiliferous outcrops north of Burlington where E. Billings (1861a, p. 943–955) and Logan (1863, p. 233) had worked out the stratigraphic and structural relations of the "Red Sandrock". The Lower Cambrian age of the white quartzite (Pl. 6, fig. 2) along the Green Mountain Front becomes apparent when the rock is traced southward to the Bennington area, where Walcott (1888, p. 235) found Lower Cambrian trilobites. The slates of the north end of the Taconic Range contrast in facies with the slates and limestones of the adjoining lowlands (Dale, T. N., 1904, p. 185, Pl. 11; Ruedemann, 1909, p. 191). The overthrust nature of the Taconic formations (Keith, 1913, p. 680) is suggested by the lowland limestone striking into the Taconic rocks.

The lowland limestones between the ranges are separated into two isolated areas by an east-west anticline in Monkton township (Keith, 1923a, p. 105). The limestones north of Monkton have been differentiated on several maps from the red quartzites to the west and from certain argillaceous rocks to the east (Hitchcock *et al.*, 1861, Pl. 1; Brainerd, 1885, p. 9; Walcott, 1888, p. 346, Pl. 3; Perkins, 1910, p. 249, Pl. 40). Likewise, several maps (Hitchcock *et al.*, 1861, Pl. 1; Dana, 1877b, p. 36–37; Brainerd, 1885, p. 9; Walcott, 1888, Pl. 3, p. 346; Seely, 1910, p. 258,

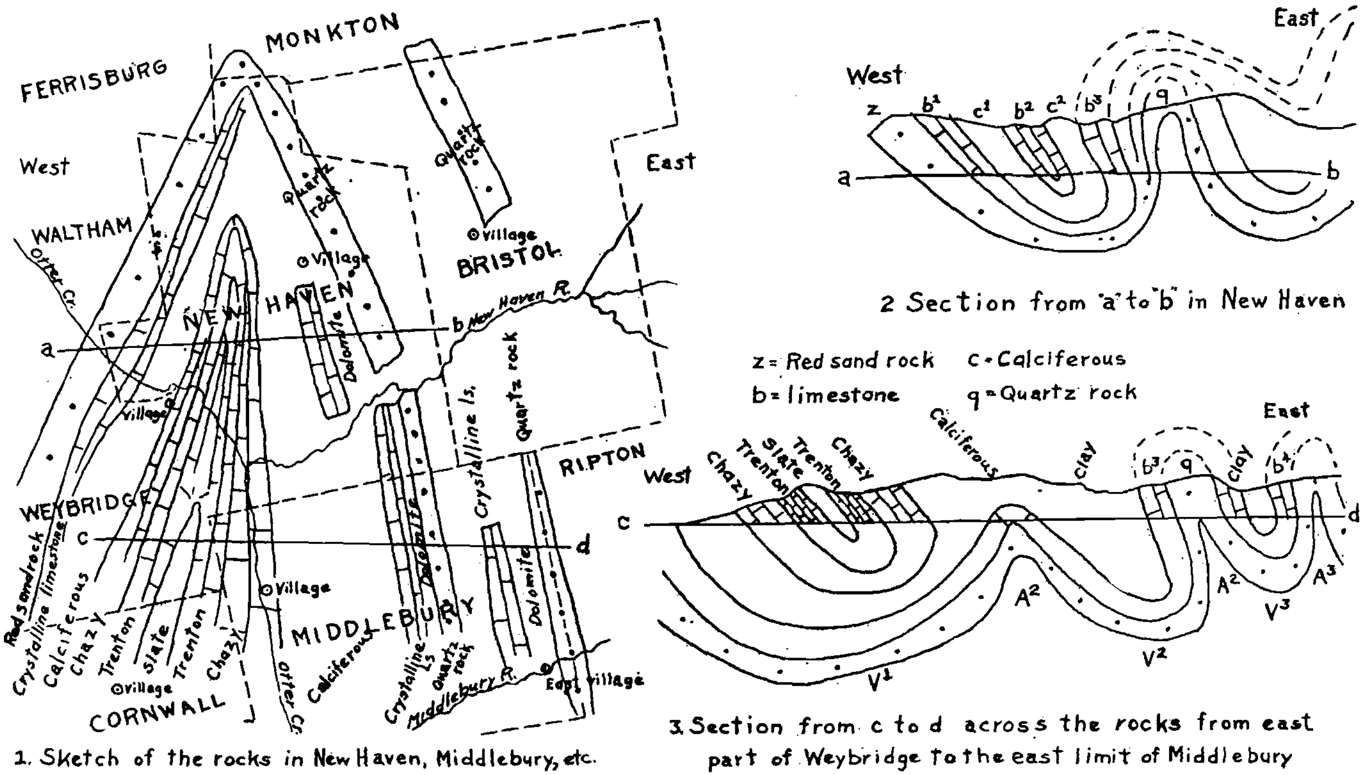


FIGURE 2.—*Wing's manuscript sketch map, and cross sections of Middlebury synclinorium*

Pl. 48) have shown the distribution of the calcareous rocks in the southern area and indicate the presence of an argillaceous band striking northward through the center of the limestones from the north end of the Taconic Range to Weybridge.

Wing first recognized the true stratigraphic and structural relations of the limestones that underlie the lowland south of Monkton between the Red Sandrock Range and the Green Mountain Front. He did not live to publish his findings, and unfortunately the brief résumé given by J. D. Dana in 1877 did not emphasize the importance of Wing's work, although the significance and results of all his findings are clearly stated.

In 1860 Wing discovered that the limestones of the southern lowland area with the quartzites of the Red Sandrock Range and Green Mountain Front stratigraphically beneath formed a southward-plunging syncline (Fig. 2). In his notebook, sometime between November 3rd, 1859, and July 30th, 1860, he made a rough cross section of this structure. In a letter dated October 9th, 1867, intended for James Hall but never finished, Wing outlined the correct stratigraphy of both limbs of the syncline. By that time Logan and Billings clarified the overthrust relation at Snake Mountain, and Wing traced easily recognizable beds (Pl. 3, fig. 2; Pl. 4, fig. 2) around the nose of the syncline. The "Calciferous"-Chazy-Trenton-"Utica" rocks of the west limb, fossils of which had been identified by Billings in Sudbury, Orwell, and Shoreham (Hunt, 1868, p. 227–228), were found traceable northward through Whiting, Cornwall, Weybridge, and New Haven, across the axis of the syncline and south through Middlebury, Salisbury, Leicester, and Brandon on the east limb.

From Wing's description of the fossil zones, lithologic units, and areal geology of the "folds" of "Calciferous" rock in Shoreham township, Vermont (Dana, 1877a, p. 342–344), grew Brainerd and Seely's (1890) classification of the "Calciferous" (Fig. 3). They (1890b, p. 5, map) showed the structure of the "folds" and described in detail the five Divisions of the "Calciferous" (p. 2–3), which later became the Little Falls dolomite and Beekmantown B, C, D, and E. They mapped the southern extension of the "great break", as they called the "Logan's Line" thrust, a little west of the Shoreham "folds". They noted several minor thrust faults in addition to the anticlines and synclines indicated by Wing. They traced the divisions of the Beekmantown north on the west limb of the syncline to Cornwall but failed to recognize the lithologic distinctions that Wing had made along the east limb (Seely, 1910, p. 258, Pl. 48). The present work reveals that Brainerd and Seely as well as Wing erroneously correlated certain portions of the Beekmantown, Chazy and possibly Black River found in the more deformed eastern outcrops (Fig. 3). Walcott (1888) established the Lower Cambrian age of the quartzites on both limbs of the syncline. Wolff (1891) and Foerste (1893) indicated Lower Cambrian limestone above the white quartzite on the east limb of the syncline and suggested that the "Calciferous", Chazy, and Trenton are there also. The classifications of Keith (1932, p. 360–361) and Bain (1934, p. 126) have indicated a sequence of lithologic units immediately beneath and including the "marble belt" of this eastern area which the present work reveals is an Upper Cambrian-Lower Ordovician succession.

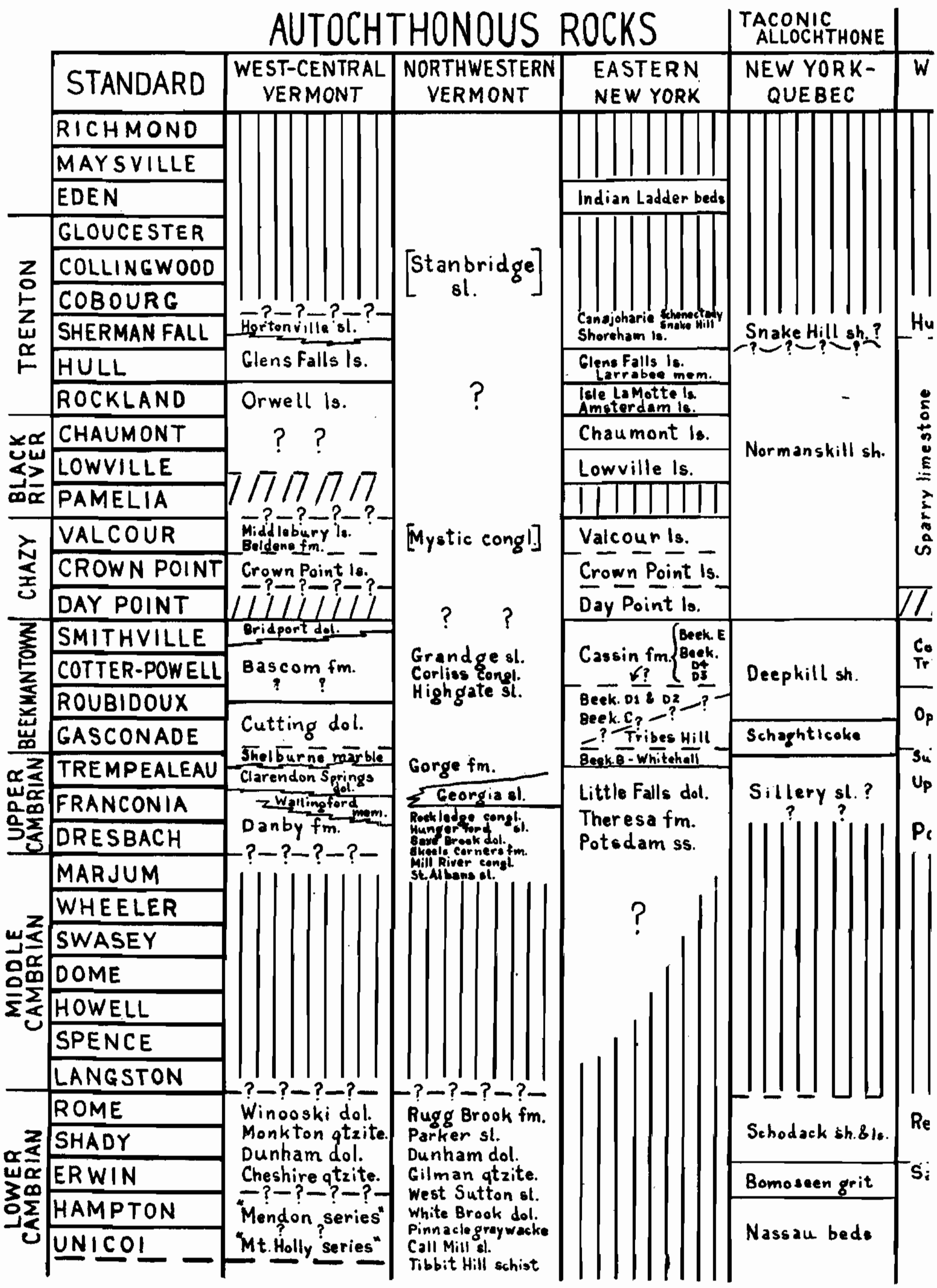
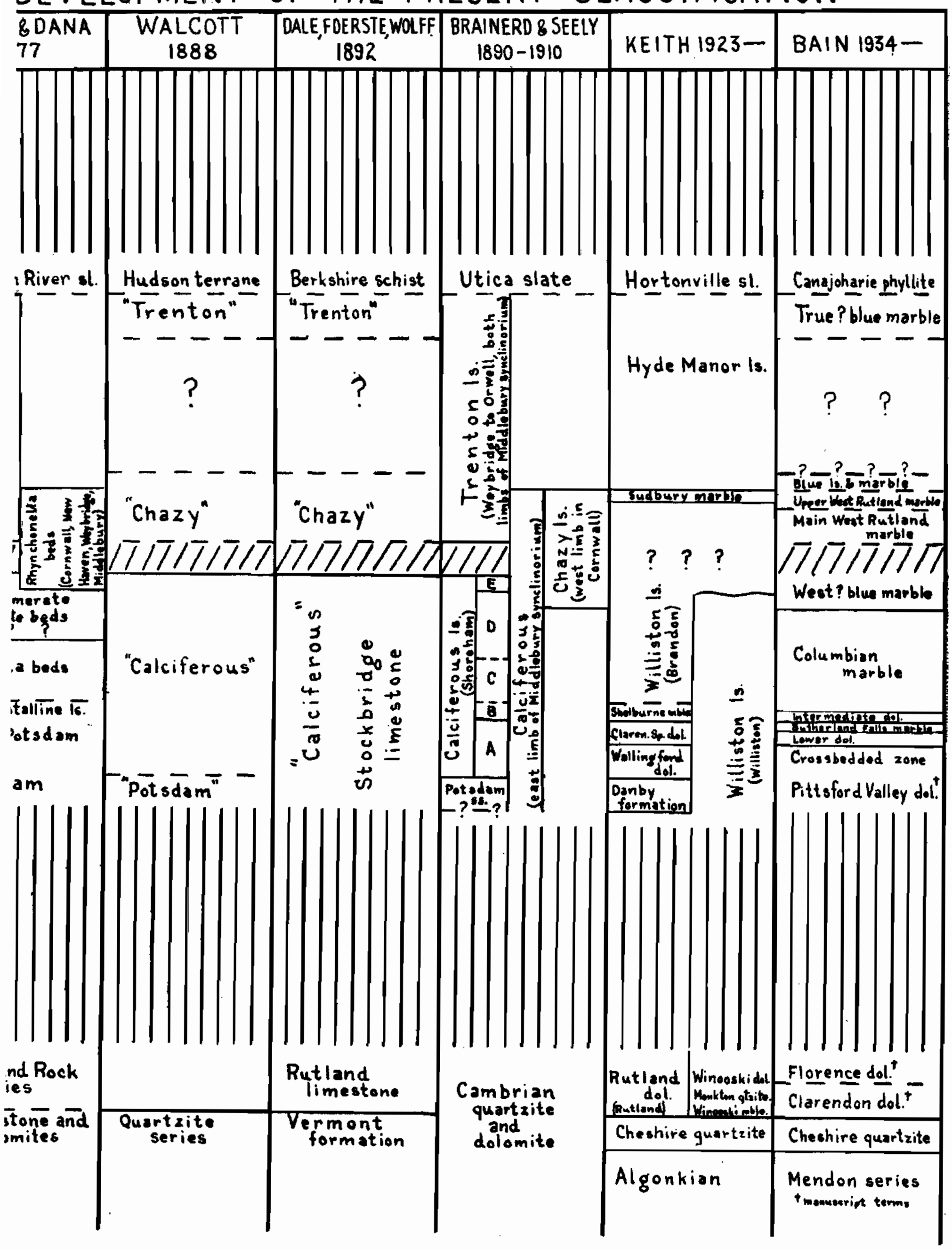


FIGURE 3.—*Stratigraphic correlation c*

DEVELOPMENT OF THE PRESENT CLASSIFICATION



...t-central Vermont and adjoining areas

STRATIGRAPHY

THE PRESENT CLASSIFICATION

Briefly the stratigraphic succession in west-central Vermont is as follows:

Correlation	Formation	Lithology	Thickness feet
Black River-Trenton	Hortonville slate	Red-brown-weathering, blue-black slate	400+
	Glens Falls limestone	Thin-bedded, dark blue-gray, rather coarsely granular limestone	115±
	Orwell limestone	Massive, closely knit, heavy ledged, light-dove-gray-weathering, rather fine-textured black limestone criss-crossed with innumerable white calcite veins	50+
Chazyan	Middlebury limestone	Buff streaked, dark blue-gray, somewhat nodular and granular, thin-bedded and incompetent partially dolomitic limestone	600±
	Beldens formation	Bright, orange-buff-weathering dolomite in beds 1 or 2 feet thick, interbedded with snow-white marbly limestone	0–700
	Weybridge member of Beldens formation	Limestone with sandy streaks about half an inch wide that weather into raised ridges about 1 inch apart more granular and darker than the intervening blue limestone	0–500
	Crown Point limestone	Lead gray, compact, massive rock that weathers to a gray surface; on dip slopes this weathered surface appears to be stippled with numerous light-buff dolomitic protuberances about a quarter to half an inch in diameter	0–150
Beekmantownian	Bridport dolomite	Beds of dolomite 1 or 2 feet thick, weathering drab, yellowish, or brown	0–500
	Bascom formation	Beds of limestone, dolomite, sandstone, quartzite, limestone breccia, and sandy calcareous shales	375±
	Cutting dolomite	Beds of dolomite 1 or 2 feet thick; thin laminated *Scolithus*-bearing sandstone at base and black chert in upper part	350±
?	Shelburne marble	White marble or dove-colored limestone, intermingled with light-gray dolomite	0–600?

Correlation	Formation	Lithology	Thickness feet
Late Cambrian	Clarendon Springs dolomite	Uniform, massive, smooth-weathering gray dolomite characterized by numerous geodes and knots of white quartz, some of which contain doubly terminated quartz crystals	50–200
	Wallingford member of Danby formation	Interbedded dolomites and irregularly weathered dolomitic sandstones	300–400
	Danby formation	Protruding, differentially weathered beds of gray quartzite 1 or 2 feet thick separated by varying thicknesses of dolomite	400–800
Early Cambrian	Winooski dolomite	Beds of dolomite 4 inches to 1 foot thick separated by thin siliceous partings which protrude as dark, slightly undulating ridges	100–800
	Monkton quartzite	Red quartzite in layers from a few inches up to 3 feet thick separated by beds of pink to gray dolomite	0–800
	Dunham dolomite	Siliceous buff-weathering dolomite containing well-rounded sand grains irregularly distributed	1700–2000
	Cheshire quartzite	Pure massive white quartzite; lower portion less massive appearing and weathers brown	1000±
	?"Mendon series"	Succession of irregularly distributed conglomerates, graywackes, dolomites, and phyllites	1000±

This succession lies on the southeastern border of the foreland or Canadian Shield area. The strata below the middle Trenton were deposited in a foreland trough paralleling the southeastern margin of the foreland. They are equivalent to geosynclinal sediments originally deposited southeast of the foreland margin and now found in the Taconic Allochthone. Middle Trenton and later strata of the autochthonous succession were deposited in the northwestward-migrating geosyncline, causing post-middle Trenton geosynclinal shales to lie on a pre-middle Trenton foreland calcareous facies in west-central Vermont. Foreland sediments deposited near the Adirondack massif remain relatively flat-lying, but those near the southeastern margin of the foreland in west-central Vermont lie in a folded belt rather persistent to the north-northeast and south-southwest of the area.

LOWER CAMBRIAN SERIES

General correlation.—The lowest fossiliferous Paleozoic horizons of western Vermont are those of *Olenellus* (Walcott, 1886, p. 13–20, 26–28; 1888, p. 233–236, 241–242; 1889, p. 29–42). A threefold faunal division of the Lower Cambrian rocks with the recognition of successive zones of *Obolella*, *Bonnia*, and *Olenellus* correlated with similar zones in the central and southern Appalachians, has been suggested (Resser and Howell, 1937, p. 199).

Formations.—The Lower Cambrian lithologic units have been variously named, described, and defined by Hitchcock *et al.*, (1861, p. 326–394); Emerson (1892;1917, p. 32–34); Keith (1923a, p. 106–110, 126–129, 1932, p. 369–372, 395–396), Clark (1934, p. 9, 10; 1936b, p. 144–147), and Howell (1938, p. 99–101).

The stratigraphic order here adopted is based upon the reinterpretation of a contact within the Rosenberg slice that was mapped (Keith, 1933, p. 53, Fig. 13) as a line of thrust faulting along the east side of Malletts Bay (Pl. 10) but now is considered a normal sedimentary contact between the Dunham dolomite and the overlying Monkton quartzite. The Winooski dolomite was thought to have been repeated here, but under the present interpretation, as the type Winooski is above the Monkton, a new name must be introduced for the beds beneath the Monkton. Inasmuch as the Mallett facies is observed north of Mallets Bay only and grades into typical Dunham south of Burlington, it is not advisable to expand the term Mallett. The author adopted the term Dunham, already applied to a corresponding dolomite horizon in the Oak Hill slice (Fig. 5) of southern Quebec (Clark, 1934, p. 9, 10; 1936b, p. 146–147), after an examination of the type locality and its southern extension in Vermont. Thus the term Dunham dolomite is added to this part of the succession, and Mallett is a member or facies type within the Dunham. North of Malletts Bay in Milton township the Dunham lies beneath the Parker slate which occupies about the same stratigraphic position as the Monkton but disappears southward without coming in contact with it much as the Monkton that overlies the Dunham at Malletts Bay, disappears northward (Fig. 4). This leaves the Dunham on the west slopes of Diamond Hill in Milton township directly overlaid by a sandy dolomite in line of strike with the type Winooski farther south. The Dunham and Winooski beds are also in contact where the Monkton pinches out to the south in Pittsford township and are called the Rutland dolomite. The Rutland lies on the Cheshire quartzite at the base of the fossiliferous Cambrian.

CHESHIRE QUARTZITE (Emerson, 1892): the Cheshire (Pl. 2, fig. 1) is typically a pure, massive white quartzite. In many places relatively small rectangular blocks are wedged out by frost action. The lower portion of the formation weathers brown, is much less massive and more argillaceous, is mottled gray on the fresh surface, and in most places is schistose. A single bed of the massive white rock may be 10 feet thick. The white quartzite unit is about 350 feet thick and above it is a zone of interbedded quartzite and dolomite that grades into the overlying Dunham dolomite through a vertical distance of about 50 feet. The thickness of the schistose brown quartzite is indeterminate, chiefly because the contact with the underlying "Mendon series" is obscure at all points where a measurable section is available. It is probably not less than 500 feet thick. Keith (1923a, p. 126) has indicated a thickness of at least 1000 feet for the whole Cheshire formation at Bristol.

The Cheshire quartzite extends far to the north and south of the area being the most extensive and well-defined stratigraphic unit in the Cambro-Ordovician section of western Vermont. South of the type locality at Cheshire, Massachusetts, the Poughquag quartzite of western Connecticut and southeastern New York corresponds to the description of the Cheshire both lithologically (Dana, 1872, p. 250–256) and faunally (Dwight, 1887, p. 27–32; Walcott, 1888, p. 236). In the Oak Hill slice (Fig. 5) of southern Quebec the Gilman quartzite (Clark, 1931, p. 225–226; 1934, p. 9, 10; 1936b, p. 144–146) according to Clark differs in appearance from the typical Cheshire. The writer

FIGURE 1. CHESHIRE QUARTZITE
Contact of white quartzite to west on "brown" quartzite 1 mile north of East Middlebury village.

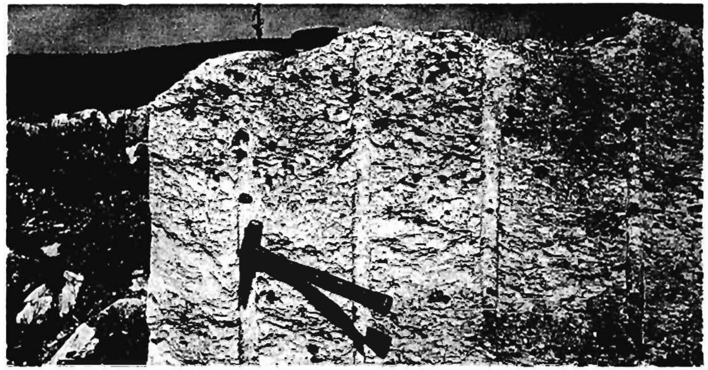


FIGURE 2. DUNHAM DOLOMITE
In quarry 1 1/8 miles NE. of East Monkton village.

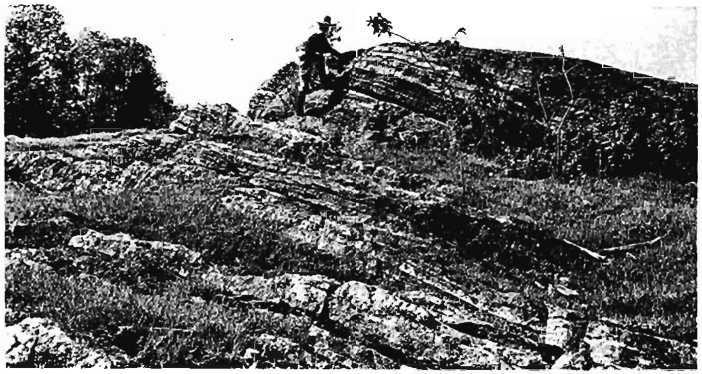


FIGURE 3. MONKTON QUARTZITE
Dipping east on eastern slope of Snake Mountain 2 miles WNW. of Waybridge village.

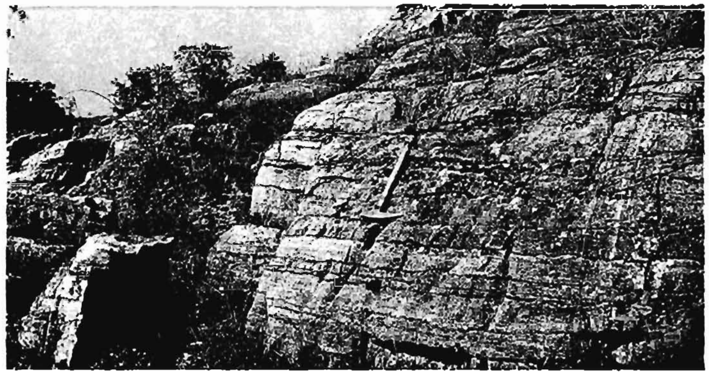


FIGURE 4. WINOOSKI DOLOMITE
1 3/4 miles SSW. of village of Barnumtown in Monkton township.

LOWER CAMBRIAN STRATA

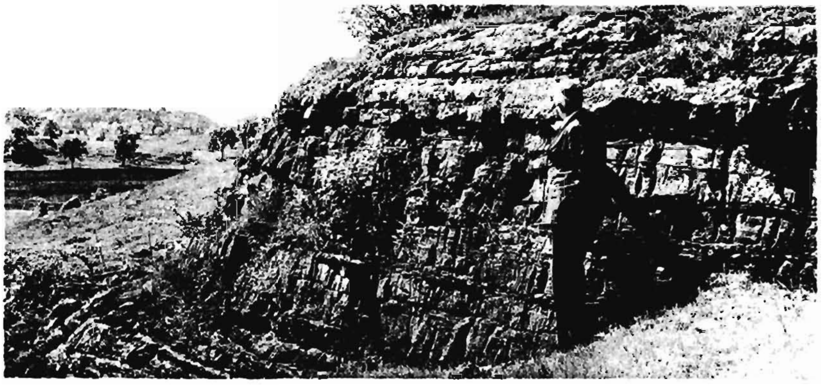


FIGURE 1. DANBY FORMATION
At axis of Middlebury synclinorium 3 miles N. of New Haven village. Pick point is beneath quartzite bed.

FIGURE 2. SHELBURNE MARBLE
On east limb of Middlebury synclinorium 1¼ miles S. of New Haven village.

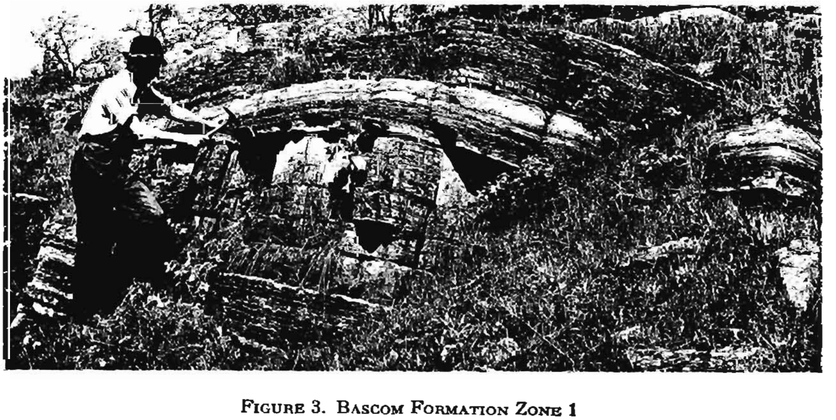


FIGURE 3. BASCOM FORMATION ZONE 1
1 mile NW. of Cornwall village.

UPPER CAMBRIAN AND BEEKMANTOWN STRATA

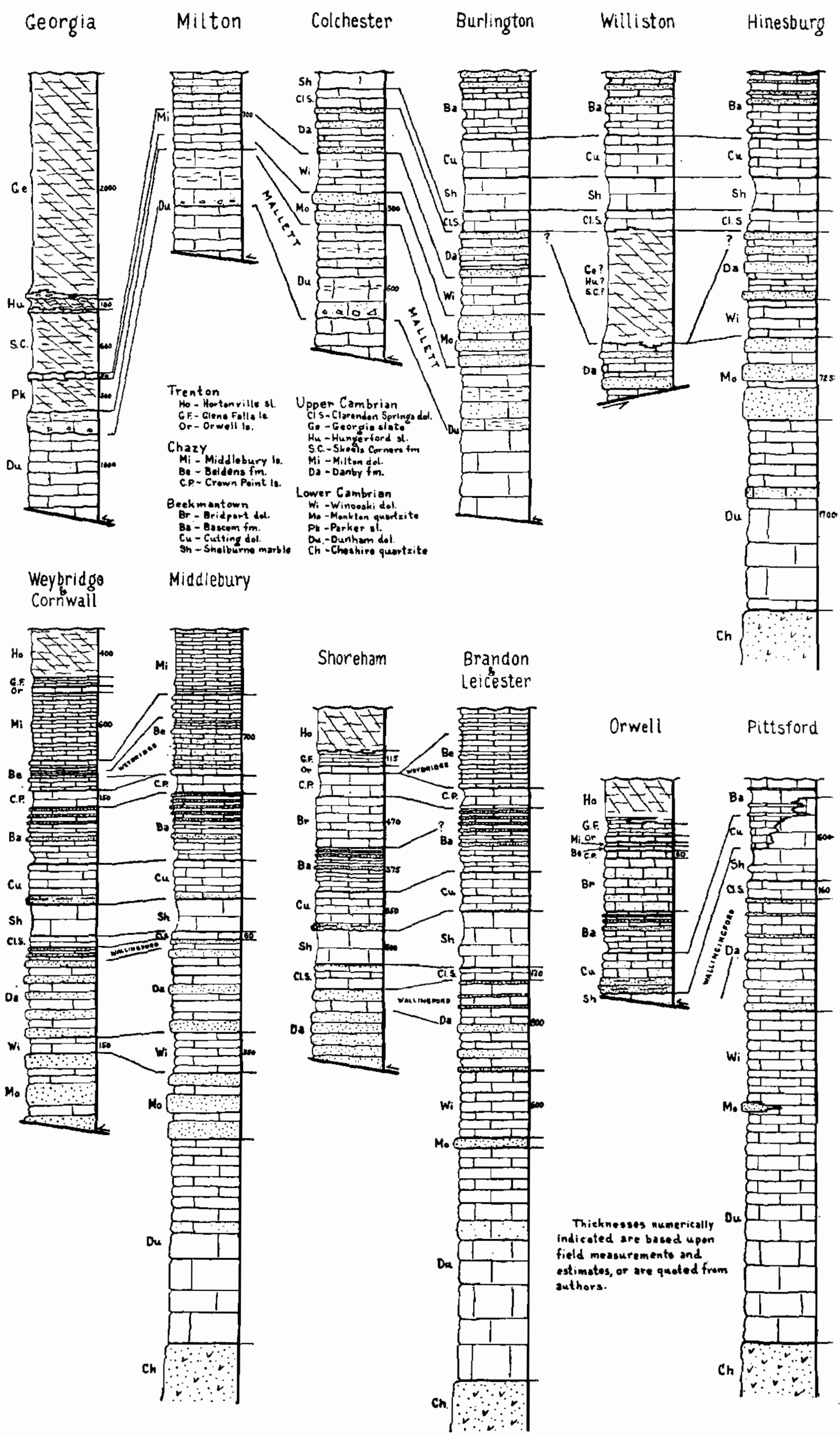


FIGURE 4.—*Columnar sections of strata in west-central Vermont*

visited outcrops of this formation at Milton village in northern Vermont where E. C. Jacobs identified the rock as the Brigham Hill graywacke, which he had previously named and correlated (1935, p. 85) with the Gilman quartzite of Quebec. It is lithologically similar to the lower brown Cheshire of west-central Vermont. The white quartzite seems to offlap the brown quartzite toward the south.

Fossils are scarce in the Cheshire, and none has been found by the author. Walcott (1888, p. 235) found *Olenellus*, *Nothozoe*, and *Hyolithes* near Bennington, Vermont, and with Seely (1910, p. 307) discovered them on Sunset Hill, a little west of Lake Dunmore, and thus demonstrated the Lower Cambrian age of the quartzite. Previously, "*Lingula*" had been reported at Rockville in Starksboro township (Hitchcock *et al.*, 1861, p. 356) and "*Scolithus*" by several authors at scattered localities.

Foye (1919, p. 85) and Keith (1923a, p. 127; 1932, p. 395–396) identified a basal conglomerate in the Cheshire quartzite, and Emerson (1917, p. 33) indicated a conglomerate unconformable on a pre-Cambrian gneiss and continuous upward into the Cheshire quartzite of western Massachusetts. Foye noted an unconformity beneath the basal conglomerate at the Falls of Lana on Sucker Brook, a little east of Lake Dunmore. The latter locality is the only one in the area that shows definite angular relations between the Cheshire and the underlying "Mendon series" (Whittle, 1894a, p. 408–414); it is without doubt an angular unconformity. The author has noted abundant opalescent blue quartz grains and pebbles in this basal conglomerate.

The basal conglomerate and unconformity, together with the fact that the Cheshire quartzite includes the lowest fossiliferous Paleozoic horizon in the region, have suggested to most authors a major systemic break directly beneath the quartzite. However, the Cheshire strata are not at the base of the unbroken sedimentary succession described in the Oak Hill slice of southern Quebec (Clark, 1931, p. 225–226; 1934, p. 9, 10; 1936b, p. 146) and northwestern Vermont (Booth, 1938, p. 1869); and the field relations, except in the immediate vicinity of the outcrop at the Falls of Lana, are rather obscure. Commonly the contact is so strongly affected by deformation or hidden by similarity of the facies involved that it cannot be located. Whether or not the Cheshire overlies a systemic break in west-central Vermont seems an open question.

DUNHAM DOLOMITE (Clark, 1934): The Dunham is mainly a siliceous buff-weathering dolomite containing well-rounded sand grains irregularly distributed (Pl. 2, fig. 2). The basal two thirds of the formation, overlying the interbedded quartzites and dolomites at the top of the Cheshire, includes 1100 feet of massively bedded dolomite that weathers buff; it is pink and cream mottled or buff to gray on the fresh surface. At certain localities the beds are thinner and may be separated by thin partings of red shale and sand distributed along deeply undulating bedding planes. In the upper part of this massive zone are thick beds of smooth "round weathering" gray dolomite, which have a sandpaperlike surface, produced by the weathering out of abundant, well-rounded sand grains. Above these sandy beds is 600 feet of thinner-bedded, locally cross-bedded dolomites that contain slightly wavy, but not deeply undulating thin siliceous partings every 6 to 8 inches. The lowermost of these thin beds are buff-colored on the fresh surface and the partings are black, but near the top of the formation the dolomite is pink whereas the partings are distinctly reddish. No definite horizon separates the Dunham from the overlying Monkton quartzite, but for mapping purposes they have been roughly differentiated where quartzite beds over 1 foot thick are separated by less than 25 feet of dolomite.

THE MALLETT (Keith, 1923) MEMBER (new revision): This member occupies the upper part of the Dunham formation in the Rosenberg slice (Fig. 5) north of Burlington. It is typically of sandy, smooth, "round weathering" gray dolomite as observed in the upper middle part of the Dunham in west-central Vermont, together with interbedded buff dolomite beds showing both the deeply undulating and the slightly wavy siliceous partings described in the Dunham to the south. In addition to these two lithologic types, are several beds of flaggy, dark-brown-weathering shaly dolomites bearing abundant trilobite fragments. An intraformational breccia, containing angular blocks of the varied Dunham dolomitic facies, is found in the lower part of the Mallett member as far north as St. Albans, Vermont. The blocks have a maximum diameter of about 2 feet. Fragments of jasper are also rather numerous. The matrix is sand and a red siliceous material such as is found in the partings of the mottled dolomite. The top of the Mallett north of Diamond Hill and Long Pond in Milton township (Pl. 10) occurs in most places at a rather sharply defined horizon beneath the Parker shale. North of the area studied Schuchert (1937, p. 1025, 1034) noted a distinct erosional break between Mallett and Parker, the latter containing angular pieces of the dolomite. On the west slopes of Diamond Hill the Mallett lies beneath and grades eastward and into dolomites that are in line of strike with the type Winooski south of the Lamoille River. However, farther south, at the northeast corner of Malletts Bay, it grades up through intraformational breccias into the Monkton quartzite. At Malletts Bay (Schuchert, 1937, p. 1024) the Mallett member is 800 feet thick. It does not crop out south of Malletts Bay. The author believes that it grades southward into the upper beds of the Dunham (Fig. 4).

The Dunham formation extends beyond the geographic limits of the present study, but as a well-defined lithologic unit it is much less extensive than the Cheshire. Dolomite prevails in the fossiliferous Lower Cambrain of eastern North America, but due to the vicissitudes of the overlying and interbedded clastic rocks, its distribution may be variable and poorly defined. The southward disappearance of the Monkton quartzite causes the Dunham and the Winooski to lose their identities in the Rutland dolomite (Fig. 4). Similarly, Dunham is lost for a short distance in Milton township at the gap between the north end of the Monkton and the south end of the Parker, but not for a distance sufficient to lessen its usefulness as a mapping unit north of Milton township.

The Dunham beds were noted in localities other than those of the typical "Red Sandrock" along Lake Champlain by several geologists. J. B. Perry (1868, p. 344) evidently recognized the red mottled dolomites of the "Red Sandrock Series" in Franklin township, Vermont (now known to be in a locality east of the Oak Hill thrust, Fig. 5) and traced them into southern Quebec nearly to the locality where the Dunham dolomite has been described and named by Clark (1934, p. 9). E. Billings (1872a, p. 145) has suggested that the marble beneath the slates at Swanton and St. Albans, Vermont is the same as that which overlies the "quartz rock"—Cheshire—of the Green Mountain Front. T. N. Dale (1912, p. 43) directly correlated the marbles of Swanton and those along the foot of the Green Mountain Range in Monkton and also suggested their areal continuity. Keith (1923a, p. 128) correlated the "pink mottled" facies at the base of the "Rutland" in Monkton with the pink-mottled "Winooski" marble along Lake Champlain but indicated that it immediately overlay both Cheshire and Monkton, which were assumed to be contemporaneous. Continuity between the outcrops of Dunham along Lake Champlain, those along the west foot of the Green Mountains, and those in the Oak Hill slice cannot be demonstrated, but the rocks in each area occur in comparable stratigraphic successions. Several formations above the Dunham may be traced from the localities near northern Lake Champlain to the west foot of the Green Mountains without interruption.

Hall (1847, p. 31; 1859, p. 525–529) described fossils from the Dunham dolomite in Highgate township, northwestern Vermont, including forms now classified as *Olenellus* (Resser and Howell, 1937, p. 197-198) and *Ptychoparella* (Schuchert, 1937, p. 1007)—and established the Lower Cambrain age of the formation. Forms closely related to *Olenellus* are reported from the Dunham of the Oak Hill slice in Quebec (Clark, 1936b, p. 147). *Kutorgina* and *Salterella* (Wolff, 1891, p. 334-335; Foerste, 1893, p. 441), also *Hyolithes* (Dale, T.N., 1892a, p. 516), have been reported from the Dunham equivalent southeast of west-central Vermont in the "marble valley". Walcott identified several genera collected by Perkins (1908, p. 239) in Colchester township, *Ptychoparella*, *Olenellus*, *Nisusia*, *Hyolithes*, and *Salterella;* Perkins lists these forms as having been obtained from the "Red Sandrock", so some may have come from the Monkton or Winooski, though the lithologic and locality descriptions suggest that most came from the Dunham. At Fox Hill in Milton township, near the base of the Mallett member of the Dunham, Perkins (p. 232) has reported *Salterella* and *Hyolithes*.

In west-central Vermont the writer obtained no fossils from the Dunham dolomite, but the continuity of the formation into the three outlying fossiliferous localities is fairly well established by the similarities of lithology and succession. Resser (Resser and Howell, 1937, p. 202) places the Dunham in the *Bonnia* zone of the Appalachians, correlating it with the Forteau formation of Labrador and Newfoundland (p. 201), with certain limestone boulders in the overthrust shale sequence near Quebec City (p. 201), with the Schodack formation in the Taconic overthrust sequence of southwestern Vermont and eastern New York (p. 204), and with the Vintage (p. 205) and Shady (p. 207) dolomites of the Appalachians south of New York.

MONKTON QUARTZITE (Keith, 1923): The red color in the outcrops of the type locality in Monkton township, along Lake Champlain, and in the Red Sandrock Range distinguish the formation (Pl. 2, fig. 3). Keith's original description (1923a, p. 107) in these localities is as follows:

"The Monkton quartzite has a decidedly reddish color varying from reddish-brown through brick red and purple to light shades of red, pink, buff and white. . . . Striking contrasts are formed by the white and vari-colored layers, particularly in the cliffs. . . . The formation consists almost entirely of quartzite in layers from a few inches up to three feet in thickness. A few seams and beds of reddish or purplish shale are interbedded with the quartzite . . . most of them being only a few inches thick. In the upper part of the formation a few thin layers of gray or pink dolomite form a transition into the overlying Winooski. . . . [On the broad exposures of dip slopes] . . . the minor characteristics of the formation are well exposed; ripple marks and mud cracks are very common, and there are numerous raindrop impressions and trails of animals. Cross-bedding in the quartzites is very common and there are numerous seams of coarse material which are too fine to be called conglomerate".

Between the beds of quartzite lie numerous rather thick successions of dolomite, such as those of the upper part of the underlying Dunham or overlying Winooski; these were interpreted by Keith (1933, p. 53, Fig. 13) as slices of Winooski dolomite over which the Monkton had been thrust.

The Monkton facies changes gradually eastward from the outcrops along Lake Champlain and the Red Sandrock Range due largely to the greater deformation in eastern localities. Eastward in Monkton township, and particularly from New Haven southward, it becomes a schistose rock in which the red color is apparent only in certain zones "protected" from deformation; instead of well-bedded, cliffed exposures, round, smooth, glacially scoured outcrops prevail. Nevertheless, the dolomites interbedded with the Monkton, although extremely squeezed, can be recognized by the pink or cream color on the fresh surface and the thin black to red siliceous partings which separate the individual dolomite beds.

Several exposures of tightly folded Monkton strata show good sections across vertically dipping beds at the cross-anticline in Monkton and Hinesburg townships which separates the Middlebury synclinorium to the south from the Hinesburg synclinorium to the north (Pl. 10). Here the Monkton is 725 feet thick. East of Malletts Bay Perkins (1908, p. 227) reported 450 feet of Monkton. Three hundred feet is reported (Keith, 1923a, p. 108) at the northeast corner of Malletts Bay, near the northern limits of the Monkton outcrops, and less than 4 miles northward the Monkton disappears. South of Monkton the thickness varies considerably and in Pittsford, Vermont, 30 miles to the south, it pinches out completely (Fig. 4). The Monkton appears much reduced toward the east, and in southeastern Hinesburg township most of the quartzite is absent. The westward extent is uncertain, since it has been cut out southward along the Champlain overthrust and has been eroded back to the present Red Sandrock Range. Several authors believe that the horizons of the Monkton are overlapped by the Upper Cambrian Potsdam sandstone around the Adirondacks, although very possibly some of the lower unfossiliferous Potsdam in the northern Adirondack region may be contemporaneous with the Monkton.

To the north, east, and south the Monkton grades laterally into underlying and overlying dolomites. The mode of thinning north of the northernmost outcrops of the typical Monkton, in Colchester township (Pl. 10), is obscured by the Lamoille River gravels, but on the west slope of Diamond Hill near Long Pond in Milton township a small patch of quartzite, like the Monkton, grades laterally northward into the Parker slate and southward into sandy dolomites. The quartzites in Colchester township, a little south of the Lamoille River gravels, also show lateral gradation of individual beds from quartzite to sandy dolomite. The eastward thinning and southward extinction of the Monkton takes place by interfingering with and disappearance of individual beds into the dolomitic rocks. This distribution relative to the inclosing dolomites suggests that the source of the Monkton clastics lay to the west or northwest of west-central Vermont, depending on the predeformation position of the Rosenberg slice (Fig. 6).

Few fossils have been reported from the Monkton quartzite. Perkins (1908, p. 229, 230) mentions "*Ptychoparia adamsi*" and *Olenellus* from the quartzites east of Malletts Bay. No fossils other than possible worm borings, tracks and trails, and algal forms have been reported elsewhere in the Monkton.

The present investigation indicates that the Monkton lies at about the horizon of the Parker slate. Long distance correlations (Resser, 1938, p. 6) place the Monkton as equivalent to the Antietam quartzite of the Southern Appalachians, but probably on the mistaken assumption that the Monkton and the Cheshire lie at about the same horizon. Resser (p. 6) suggests a correlation between the Monkton and the Bomoseen grit of the Taconic sequence of southwestern Vermont, basing it on the same assumption and also inadvertently giving the impression that the Taconic rocks are continuous with the Monkton instead of occurring in a separate exotic facies situated far to the east prior to thrusting (Fig. 6). The Parker slate has been placed in the *Olenellus* zone (Resser and Howell, 1937, p. 202–203), but possibly this zone, insofar as its distribution within the Lower

Cambrian is concerned, reflects a lithologic facies rather than a stratigraphic zone (Stose and Jonas 1938, p. 21-22). The gradational limits of the Monkton that have been described above seem to indicate a lithologic facies having changing vertical limits in the Lower Cambrian dolomites.

WINOOSKI (Hitchcock, 1861) DOLOMITE (new revision): This dolomite (Pl. 2 fig. 4) is indistinguishable from strata found in the upper 600 feet of the Dunham dolomite at localities where the Mallett facies is absent. As exposed immediately above the Monkton quartzite, the Winooski is pink on the fresh surface, but higher up it is less pink and the upper part is buff or even gray. The beds, 4 inches to 1 foot thick, are separated by thin siliceous partings. Immediately above the Monkton these partings are pink or red and higher up they are black and protrude as dark, slightly undulating ridges. The Winooski is distinguished from the lowest Dunham dolomite, as seen around Malletts Bay and at Swanton, Vermont, by the absence of strong mottling and also by the absence of deeply undulating siliceous partings. It is easily distinguished from the much more thickly bedded lower Dunham seen to the south.

The original trade name Winooski marble was applied to the product of a quarry in Colchester township near Malletts Bay (Pl. 10) by a quarryman whose home was in Winooski village (Hitchcock *et al.*, 1861, p. 773). Hitchcock (p. 329) adopted the name and applied it to all of the "Red and variegated Dolomites" of the "Red Sandrock Series", without specific reference to succession. In a cross section (p. 329) he designated as "Winooski limestone" certain beds lying beneath the "Red sandrock" proper that are located a little south of the Winooski River at the present site of Ethan Allen Park and its vicinity in the city of Burlington, but elsewhere (p. 774) he notes that "it helps to form the summit of Snake Mountain, in Addison". The former is at an outcrop of the Dunham dolomite and the latter reference involves a dolomite interbedded in the Monkton. Keith (1923a, p. 108) stated that the name was "given for the fine exposures along Winooski River in Burlington" and (1932, p. 370) that "the original quarry was on the north bank of the Winooski River at Burlington. . . ." Schuchert (1937, p. 1023) stated that "the type locality for the geological formation is along the Winooski River and at Winooski Falls, north of Burlington." Because the localities near Malletts Bay, in the vicinity of Ethan Allen Park, and along the lower Winooski River are in a distinctly lower stratigraphic zone, separated from that at Winooski Falls by the Monkton quartzite, the author revises the term Winooski and restricts it to the dolomite beds at Winooski Falls or Winooski village in Colchester township.

The thickness of the Winooski dolomite is difficult to determine, because the faunal break that must exist between the top of the formation and the beds of Upper Cambrian age is not accompanied by a recognized lithologic break. The top of the Winooski has never been defined. Superficially viewed, an unbroken succession from Lower to Upper Cambrian occurs. For mapping purposes, the top of the formation is placed beneath the lowest beds of typical Upper Cambrian Danby quartzite. Likewise, the lower limit of the Winooski is indistinct. Keith (1923a, p. 109) noted that, "at the base of the formation the dolomite is interbedded with quartzite layers for a thickness as great as 50 feet, forming a transition into the underlying Monkton quartzite". This lower contact is mapped, as in the case of the Dunham-Monkton contact, through points where, passing down in the section, quartzite beds over 1 foot thick and separated by less than 25 feet begin to appear. Keith (1923a, p. 109) gave a thickness of less than 100 feet on the east slope of Snake Mountain; the author estimates 100 to 150 feet. Measurements from the east limb of the Middlebury synclinorium in about the same latitude are around 350 feet. Southward the formation thickens to about 600 feet at Brandon and at least 800 feet at Pittsford (Fig. 4). The Winooski is approximately 350 feet thick north of Snake Mountain on the east slope of Buck Mountain. From this point north through Burlington the thickness appears to vary but is generally about 350 feet. Good measurable sections of the Winooski can be obtained in very few places as the formation is either covered by Pleistocene till and lake beds that lie between the harder quartzite ridges or is involved in tight folding. Southward increase in the thickness is very apparent, although not accurately measurable, and appears to complement the southward thinning of the underlying Monkton as might be expected along such a gradational contact. Similarly, in Hinesburg township, an eastward increase in the thickness of the dolomite with a corresponding diminution of the Monkton is discernible.

Correlation of the Winooski depends entirely upon its intergradational relation with the underlying Monkton quartzites, since no fossils have yet been reported. The vertical extent of the Lower Cambrain in this formation is indeterminate without fossils. It may include all the strata that show the thin dark siliceous partings separating the dolomite beds and possibly may extend even higher.

The northern limits of the typical Winooski facies lie nearly at the north boundary of Colchester township northeast of Malletts Bay. Rather sandy dolomites which weather salmon lie north of this point in Milton township but are separated from the typical Winooski outcrops by about a 4-mile wide belt of the Lamoille River gravels. This sandy dolomite facies grades down into the Mallett dolomite south and west of Diamond Hill in Milton township, between the north end of the Monkton quartzite and the south end of the Parker slate (Fig. 4).

The salmon-weathering sandy dolomite and an associated gray-weathering variation occur at several points north of Milton in northwestern Vermont. They are in rather close association with the Parker slate. In most places the slate lies below the salmon dolomite, separated from it by a more or less distinct erosional break, but at some points the slate seems interbedded with the dolomite. The dolomite in Milton township forms the western outcrops of Keith's (1923a, p. 112–114) Milton dolomite. Farther north this facies appears to include the Rugg Brook formation defined by Howell (1938) as well as the beds that Schuchert (1937, p. 1013, 1025, 1035, 1036, 1037–1038) referred to as the "dolomite at the top of the Parker" and as "overthrust Mallett" (p. 1030 1035, 1036). Howell (1938, p. 101) indicated that it is impossible to determine whether the Rugg Brook is of Lower or Middle Cambrian age, inasmuch as stratigraphic breaks both above and below it separate it from fossiliferous beds. Dolomites interbedded with the upper part of the Parker at certain localities in northern Milton township are by definition Lower Cambrian. It is about as difficult to find the top of the Lower Cambrian dolomites in the "Milton" terrane as it is to locate the top of the typical Winooski in west-central Vermont. Northeast and southeast of Roods Pond in Milton township (Pl. 10) an eastward cessation of the salmon-weathering dolomites is rather apparent; a break between sandy dolomites to the west and less sandy dolomites to the east suggests a rather sharp separation.

Both in west-central and northwestern Vermont the upper limit of the Lower Cambrian is undetermined. Possibly the upper part of the Winooski, as mapped in west-central Vermont, may include Upper Cambrian beds, and it is not inconceivable that some Middle Cambrian horizons may be present.

Distribution and genetic relations.—Several lithologic units or facies are included in the Lower Cambrian of west-central Vermont. None is a very satisfactory stratigraphic unit inasmuch as each grades laterally into other units and contains fossils characteristic of facies rather than stratigraphic zones. Fossils are well preserved only in the more clastic rocks. The whole of the Lower Cambrian, about 5000 feet thick, is in general characterized by the trilobite genus *Olenellus.* The lower Cambrian is pervaded by a dolomite facies in zones which were out of range of the distribution of clastics. The Monkton quartzite, the highest sandstone deposit in the Lower Cambrian, thins northeastward, eastward, and southeastward from a center that must have been near the site of the present Adirondacks. The Parker slate and its apparent stratigraphic equivalents in the Oak Hill slice (Booth, 1938, p. 1869) are found in about the same stratigraphic position as the Monkton, but thickening north and east of it, suggesting a source of material diametrically opposite to that of the Monkton. The outcrop of the Cheshire quartzite is so nearly linear from north to south that it is difficult to determine whether this formation thickens to the east or to the west. Certain white quartzites, only 25 to 100 feet thick, found on the east side of the Green Mountain axis at Plymouth (Perry, E.L., 1929, p. 17–18) possibly are the Cheshire (Keith, 1932, p. 405). Most of the reworked (Schuchert, 1937, p. 1024) clean sands, such as those of the Cheshire, found in the Cambro-Ordovician above the Cheshire thicken toward a western source. Toward the northeast, in the Oak Hill slice, the clean white Cheshire facies thins and its position is occupied by a more argillaceous Cheshire equivalent. Possibly the Cheshire, also, thickens westward. Cambrain sediments, thought to be the stratigraphic equivalent of the "Mendon series" which is locally unconformable beneath the Cheshire in west-central Vermont, lie conformably beneath rocks equivalent to the Cheshire in the Oak Hill slice. The "Mendon series" and the pre-Cheshire members of the Oak Hill succession are largely an unreworked argillaceous facies but contain considerable sandstone, quartzose conglomerate, and graywacke. Booth (1939, p. 13) has suggested that the boulders in the conglomerate may have come from a western source.

UPPER CAMBRIAN SERIES

General correlation.—Sandstone beds of Upper Cambrian age (Walcott, 1886, p. 20–45), described and defined as the "Potsdam sandstone" (Emmons, 1842, p. 102; 1855, p. 83; Hitchcock *et al.*, 1861, p. 265), lie at the base of the stratigraphic section along Lake Champlain and rest on the pre-Cambrian of the Adirondacks. The sandy beds grade without apparent stratigraphic break into the overlying dolomites which have therefore been included with the sandstones in the Upper Cambrian (Cushing, 1908, p. 169-170). In northwestern Vermont, a little north of the area, dolomites, very sandy in part, occupy a stratigraphic position similar to that of the "Potsdam" sandstone and its overlying dolomites and have been named and described as the "Milton dolomite". Fossils (Schuchert, 1937, p. 1046) reported elsewhere from the Upper Cambrian, particularly *Lingulella acuminata* and *Cryptozoön* have been obtained from localities in the "Milton" terrane. South of the "Milton' terrane, in west-central and southwestern Vermont along the "marble belt", Walcott (1888, p. 240-241) mapped a "Potsdam" terrane.

A threefold division of the Upper Cambrian along the eastern border of the Adirondacks and also in northwestern Vermont, based on the three primary Upper Cambrian divisions of the Upper Mississippi Valley,—Dresbach, Franconia, and Trempealeau—has been suggested. Rodgers (1937 p. 1575–1576) called Franconia certain transition beds between the sandstones and overlying dolomites at the Adirondack border near Whitehall, N. Y.; using lithology and stratigraphic position, he correlated the underlying horizons with the Dresbach and those overlying with the Trempealeau. Wheeler (1941a, 1941b, 1941c, 1941d) indicated the Franconia and Trempealeau at somewhat higher levels than did Rodgers. Howell (1939b, p. 1964) correlated certain shales, sandy dolomites, conglomerates, and dolomites of the St. Albans, Vermont, area, which appear to be lateral equivalents of at least a part of the rocks of the "Milton" terrane, with the Dresbach and Trempealeau.

The present study indicates a rather continuous exposure of the Upper Cambrian between the "Milton" terrane of northwestern Vermont and the "Potsdam" and associated dolomites that lie on the pre-Cambrian near Lake Champlain in New York and western Vermont. Probably most of these beds are actually Upper Cambrian. However, (Cushing, 1905, p. 359-360) the unfossiliferous lower beds of the Potsdam that border the northern Adirondacks may be older.

Formations.—The formation names (Keith, 1932, p. 396-397) applied in the Upper Cambrian succession were originally considered to designate Lower Cambrian strata. The present work indicates that Keith's three divisions—Danby, Wallingford, and Clarendon Springs—are about the equivalent of the thus far established fossiliferous Upper Cambrian along Lake Champlain.

The Danby formation is expanded for mapping purposes to include the Wallingford dolomite or transition beds above the original Danby, making Wallingford a member of the Danby formation. The term Potsdam is not used in west-central Vermont, because most of the outcrops resemble Keith's Danby. The term Clarendon Springs, applied locally by Keith to the mappable dolomite unit between the sandy dolomite formation and the overlying Shelburne marbles and nondolomitic limestones, seems more suitable in west-central Vermont than Little Falls (Cushing, 1908, p. 170) applied to the dolomites between the Potsdam and the overlying limestone beds along Lake Champlain and in the Mohawk Valley.

DANBY FORMATION (Keith, 1932; new revision): Protruding, differentially weathered beds of gray sandstone 1 or 2 feet thick are separated by 10 to 12 feet of dolomite (Pl. 3, fig. 1). These are the only sandstones in the dominantly dolomitic succession between the Monkton quartzite and the Shelburne marble. The lower sandstone beds are actually white quartzites that form massive, clean cut blocky outcrops between the less prominent dolomites. These beds pass upward into the less common, more irregularly weathered, more dolomitic and less quartzitic sandstones which with the interbedded dolomites constitute the Wallingford member.

WALLINGFORD (Keith, 1932) MEMBER (new revision): This member has been described as "Dark iron-grey magnesian limstone, usually in beds one or two feet in thickness, more or less siliceous, in some beds even approaching a sandstone" (Brainerd and Seely, 1890b, p. 2). The uppermost sandy beds of the Wallingford in most places are spaced farther apart and are thinner than those at the base. Throughout the Danby formation the sandy laminae show abundant cross-bedding that is very useful in determining the attitude of the beds.

The Danby formation, including the beds between the lowest and the highest sandstones, is about 800 feet thick at Leicester. At other localities along a north-south line smaller thicknesses were observed. Typically the Wallingford includes a little less than half of the Danby formation or between 300 and 400 feet. At Leicester (P. 10) the Wallingford member is 300 feet thick; a little north of the New Haven River, where the typical Danby facies may be thicker than anywhere else, the Wallingford is only 60 feet thick.

In northwestern Vermont Keith (1932, p. 372) indicated a thickness of 700 feet for the beds of the "Milton" terrane. Strata synchronous with some of the dolomites above or below the Danby may be included in this terrane. South of the area of study, in the Otter Creek Valley and the "marble belt", Bain (1938, p. 8, 10) suggested a thickness of 1050 feet for beds approximately equivalent to the Danby formation. At Whitehall, New York, west and south of the area included here, Rodgers (1937, p. 1575) pointed out that, "the thickness of the whole formation approaches 400 feet, although close to the Adirondacks the thickness may be considerably less." Walcott (1891, p. 345) has noted a 70-foot transition facies above the sandstones and beneath the overlying dolomites at this locality, comparable to that of the transition beds in the Wallingford member and, as defined here, identical with them.

The sandy beds of the Danby are thicker and more numerous toward the west, grading laterally into and interfingering with a dominantly dolomitic succession to the east (Fig. 4). In a north-south direction the sandstones show no major change in distribution and bulk. The inference is that the source of these clastics lay in a general uplift to the west. At Lake Champlain and to the west the clastics occur as a shore facies directly on the pre-Cambrian crystalline rocks.

Fossils useful in correlation have not been found within the area. At two localities (Pl. 10) half a mile northeast of Hinesburg village and 2 3/4 miles east of Middlebury village, some of the quartzite beds show on their under-surfaces several small depressions less than half an inch deep and half an inch across, which appear to be the sandy capping that filled in over the internal mold of the beak region of a brachiopod. The rocks equivalent to the Wallingford at Whitehall contain the Ironton fauna (Rodgers, 1937, p. 1575) found at the base of the Franconia of Wisconsin. The upper half of the Danby at Whitehall is at least in part, therefore, of Franconia age. Rodgers suggests that the lower portion "is presumably Dresbach". The Danby is correlated in west-central Vermont by lithologic comparisons and by the tracing of the beds into outlying localities.

Clarendon Springs Dolomite (Keith, 1932): This formation is a rather uniform, massive, smooth-weathering gray dolomite characterized by numerous geodes and knots of white quartz. Some of the geodes contain doubly terminated quartz crystals. On the fresh surface the dolomite is iron gray. At some localities sandy beds may be discerned in the upper part of the dolomites and "near the top large irregular masses of impure black chert" are found, particularly in the western part of the area, "which when the calcareous matter is dissolved out by long exposure, often appears fibrous or scoriaceous" (Brainerd and Seely, 1890b, p. 2). The latter may grade up into massive light-gray dolomites varying in thickness and lateral distribution.

Keith included in this formation all of the dolomitic beds between the uppermost sandy strata of the Wallingford member of the Danby formation and the lowermost calcite marbles and nondolomitic limestones of the overlying Shelburne marble. In several of the more fossiliferous localities along the western side of the area the dolomite includes horizons both above and below a sandstone (Rodgers, 1937, p. 1576–1577). Rodgers indicates a stratigraphic break above the dark-gray dolomites containing the quartz knots and suggests that the sandstone is basal to certain of the higher strata. Such a break has not been recognized in the eastern part of the area. It should be pointed out that the upper strata of the dolomite facies grade laterally into the limestones and marbles at several places and thus are contemporaneous with the limestones and marbles.

The Upper Cambrian dolomites of west-central Vermont represent a succession averaging about 180 feet thick. Keith (1932, p. 397) gave a thickness of 100 to 200 feet for the Clarendon Springs formation. Bain (1931, p. 509) found it to be about 160 feet thick south of Brandon. The author had determined thicknesses of 120 feet at Brandon, 80 feet at Middlebury, and 45 feet at New Haven. Walcott (1891, p. 345) gave 250 feet for these beds at Whitehall, New York. About 230 feet of the dolomites occurs at Shoreham. At Highgate Center in northwestern Vermont Schuchert (1937, p. 1052) noted 220 feet of equivalent strata. Thus the Clarendon Springs thins eastward somewhat.

Rodgers (1937, p. 1576) correlated all but the uppermost portion of the Clarendon Springs, comprising more than four fifths of its thickness and including all the iron-gray dolomites that contain the quartz knots, with the Little Falls formation on the Mohawk Valley in New York. The author has observed formations in the same position as and lithologically similar to the Clarendon Springs dolomite in northwestern Vermont and southern Quebec, in the Philipsburg and Rosenberg slices (Fig. 5).

Several authors have reported the occurrence of the latter dolomites. McGerrigle (1931, p. 185) named the Rock River dolomite at Philipsburg, Quebec, previously described by Logan (1863, p. 276). Its correlation with rocks equivalent to the Clarendon Springs has been recognized by Brainerd and Seely (1890b, p. 23) and Bradley (1923, p. 329–330). A similar dolomite was recognized at an early date in the Rosenberg slice at the gorge of the Missisquoi River below Highgate Center (Emmons, 1855, p. 81–82). Raymond (1924a, p. 199) reported and described late Upper Cambrian fossils from portions of this dolomite, and Schuchert (1933, p. 375–377) named it the Gorge formation. Howell (1939b, p. 1964) places the Gorge in the Trempealeau. At localities in Williston township, Vermont, along a meridian south of Essex Junction, the Clarendon Springs lies west of (Hitchcock, 1861, p. 376, 416) and probably stratigraphically above slates and conglomerates which farther north along the strike in Georgia township are known (Howell, 1939a, p. 1964) to include beds of Dresbach age. The slates lie stratigraphically above beds that crop out for 1 or 2 miles west of Williston village and are lithologically similar to the sandy and cherty facies of the "Milton" terrane.

Distribution and genetic relations.—The Upper Cambrian succession is over 900 feet thick in the east of the area. It thins westward to about 700 feet at the meridian of Lake Champlain and pinches out farther westward in the Adirondack region. Clastics are more abundant in the lower two thirds of the succession and thicken westward. The Upper Cambrian succession is traceable, with little lateral change except reduced sand content, far south in the "marble belt". The dolomites crop out in the Philipsburg slice to the north (Marcou, 1861, p. 243; McGerrigle, 1931,

p. 185). In a northeasterly direction all of the Upper Cambrian strata pass laterally into the dolomite typical of the "Milton" terrane. These dolomites, largely Upper Cambrian, found in the northern part of the Rosenberg slice (Schuchert, 1937, p. 1046) and in the Oak Hill slice (Booth, 1938, p. 1869), grade laterally northeastward into and interfinger with Upper Cambrian slates (Howell, 1939b, p. 1964) which in extreme northeastern localities form a thick succession, probably much thicker than the predominantly calcareous succession to the southwest. Possibly the Upper Cambrian argillaceous sediments are from a source diametrically opposite to that of the sandstones, much as in the Lower Cambrian.

The slate succession, unlike the calcareous succession found farther south in the Rosenberg slice, is visibly discontinuous vertically and is broken by several "disconformities" (Howell, 1939b, p. 1964). Breccias associated with these "disconformities" (Schuchert, 1937, p. 1014–1015, 1050–1052; Raymond, 1937, p. 1138–1140) have a shaly or sandy dolomitic matrix, with limestone pebbles apparently derived from limestone bioherms. Southward these stratigraphic breaks disappear in the conformable and predominantly dolomitic succession of the "Milton" terrane. Northward they disappear in a comparable succession near the northern tip of the Rosenberg slice. Slates varying from Lower Cambrian through Middle and Upper Cambrian to Ordovician are found in the intervening region. The higher unconformities cut down through older ones. At the lowest of three such breaks found beneath breccias in the Upper Cambrian, complete removal (Howell, 1938, p. 100) before deposition of the breccia of all but a small patch of Middle Cambrian shale, 10 miles long from north to south, occurred in such a way that the shale cannot be traced laterally proving or disproving its original lateral gradation into the dolomites. At the western edge of the Rosenberg slice the shale terrane is not more than 20 miles in north to south extent, but it is more extensive in eastern exposures. The region of the shale terrane was apparently a site of alternating deposition and erosion, which during Cambro-Ordovician times projected northwestward in a salient roughly symmetrical to the axis of the great northwest salient of the orogenic belt (Keith, 1923b, p. 309, Pl. 4; 1932, p. 363–364) found in northwestern New England and southern Quebec.

The typical complete cycle of sedimentation found here is in the sequence: sandy dolomite—breccia—shale. At the borders of the shale salient, thick dolomites are present and breccias and shale are thin. Toward the focus of the salient, however, dolomites are absent and limestone bioherms occur. They are the source of the limestone pebbles in the breccias. The bioherms are less prevalent deep in the argillaceous salient and may account for a reduction in the amount of breccias in that direction.

A typical succession of events that could account for such a distribution of sediments, beginning at an episode of cessation of clastic deposition, would be as follows.

(1) Dolomite deposited with some frosted sand grains.

(2) Salient begins to rise; dolomite at margin of salient and slate closer to focus of salient, which which is brought near to surface of sea; bioherms develop; influx of water-borne sand from west and deposition with dolomite on outer margin of salient.

(3) Sandy dolomite and bioherms rise above wave base; dolomite reworked and bioherms brecciated by waves; limestone breccia with sandy dolomite matrix is formed.

(4) Salient rises above sea level as northwest-plunging nose, but wave erosion continues and breccia zone settles through the margin of the dolomite facies on to underlying shales and migrates shoreward toward the focus of the salient, at least the marginal portions of which are reduced to wave base; argillaceous clastics from salient spread southwestward, northwestward, and northward, gradationally overlapping breccias and dolomites in these directions.

(5) Marginal area of salient reduced to wave base; only the interior portions subjected to subaerial erosion; some of the northwestern portion possibly isolated as islands.

(6) Salient and foreland margin begin to sink; argillaceous clastics gradationally offlap toward focus of salient; water-borne sands gradationally offlap toward foreland source; dolomite deposited with some frosted sand grains in widening seaway gradationally overlaps clastics.

ORDOVICIAN-BEEKMANTOWN "GROUP"

General correlation.—The "Calciferous sandstone", which as originally described included all the strata between the Potsdam sandstone and middle Chazy beds (Emmons, 1842, p. 118–122, 310–315), has since been variously restricted (Hall, 1847, p. 14–36; Cushing, 1908, p. 169-170; Wheeler, 1941a, 1941b; 1941c; 1941d) and is termed "Beekmantown". The lower Beekmantown *Ophileta complanata* (Wheeler, 1941a, 1941b; 1941c; 1941d,) or *Helicotoma uniangulata* (Ulrich, 1911, p. 631, 639), the middle Beekmantown *Lecanospira compacta* (Dana, 1877a, p. 342-344; Brainerd and Seeley, 1890b, p. 6, 7, 10, 14) and the upper Beekmantown Cotter (Wheeler, 1941a; 1941b) and *Eurystomites kelloggi* (Brainard and Seely, p. 17) faunas[1] have been variously reported in west-central Vermont in beds not far from and bordering on Lake Champlain. Wing (Dana, 1877a, p. 405–409, 415-416) correlated the deformed limestones of Beekmantown age that lie in the eastern part of the Middlebury synclinorium of west-central Vermont with the fossiliferous Beekmantown along the lake. Beekmantown strata are also reported in the Hinesburg synclinorium (Schuchert, 1937, p. 1078), in the folded belt south of west-central Vermont (Wolff, 1891, p. 336; Walcott, 1888, p. 239; Dwight, 1880, p. 51-52; 1884, p. 249–259; Prindle and Knopf, 1932, p. 273), and in the Philipsburg slice to the north (Brainerd and Seely, 1890b, p. 23; Bradley, 1923, p. 329–334).

Beekmantown horizons have been recognized in the shale facies that crop out at the north end of the Rosenberg slice in northwestern Vermont (Raymond, 1937, p. 1133–1135) and in the Taconic Allochthone near the city of Quebec (Raymond, 1913, p. 29-30; 1914a, p. 523) and in the vicinity of the Taconic Range (Ruedemann, 1902, p. 559).

Formations.—The mappable units of the Beekmantown group of west-central Vermont correspond with few exceptions to "Divisions B, C, D, E" of Brainerd and Seely (1890b, p. 2-3). They are here called Shelburne marble, Cutting dolomite, Bascom formation, and Bridport dolomite, respectively.

The term Shelburne marble is used in preference to other published designations applicable to about the same stratigraphic zone because it defines a good mapping unit. Good exposures of the

[1] Five faunas have been described from rocks in eastern North America equivalent to the Beekmantown group of some authors. These faunas are as follows from older to younger: *Helicotoma uniangulata* (Gasconade-Chepultepec) fauna (Butts, 1926, p. 91; 1936, p. 19); *Ophileta complanata* (Tribes Hill) fauna (Bassler, 1919, p. 99); *Lecanospira compacta* (Roubidoux) fauna (Butts, 1926, p. 93-94); *Ceratopea keithi* or Cotter fauna (Ulrich, 1911, p. 665-666; Butts, 1926, p. 98), and a fauna characterized by *Eurystomites kelloggi* (Whitfield, 1886, p. 293–345; 1890, p. 25–39; 1897, p. 177–184; Ulrich, 1911, p. 668),—the Fort Cassin fauna. The *Helicotoma uniangulata* and *Ophlileta complanata* faunas are found in formations that are generally considered to be identical in stratigraphic position.

Cutting dolomite are found on the eastern dip slope of Cutting Hill in southeastern Shoreham township (Pl. 10). Two miles north along the strike in this "Eastern Shoreham" section (Brainerd and Seely, 1890b, p. 1–3) the Bascom formation crops out at a locality designated by Wing (Dana, 1877a p. 343) as "Bascom's Ledge". The Bridport dolomite is named for wide exposures on the hills in the southeastern part of the town of Bridport. The term "Cassin formation" (Cushing, 1905 p. 362–364), originally defined to include Beekmantown D3, D4, and E beds, is not used in west-central Vermont because not only is it a poor mapping unit, but typical Beekmantown E beds are not present at its type locality at Fort Cassin. These formations, as well as those adjoining in subjacent or superjacent groups, are readily distinguishable as mappable units throughout most of the area because of an alternation of dolomitic and nondolomitic facies, which begins with the change from the dolomite of the Clarendon Springs to the nondolomitic Shelburne. Also, the Bascom formation contains abundant limestones which separate it lithologically from the underlying and overlying dolomites. Inasmuch as the dolomitization is to some extent randomly distributed and increases toward the west in the beds included in lower "Division D", these formations are distinctly lithologic units with only general time significance. The change from a dolomitic facies on the west to pure nondolomitic limestone on the east has been indicated for areas farther south in the Appalachian region (Ulrich, 1911, p. 652–653; Bassler, 1919, p. 96; Butts, 1928, p. 374; Rodgers 1937, p. 1577). The Beekmantown rocks of west-central Vermont seem to be located about midway in such a transition from dolomite to limestone.

Shelburne Marble: "The formation is composed almost entirely of marble, always light colored, and for the most part white [Pl. 3, fig. 2]. Other colors are light buff or cream, bluish-white, and various beds are mottled with cream and white or blue and white. A few thin beds of light grey dolomite are found in various parts of the formation" (Keith, 1923a, p. 116).

Toward the west it becomes a

"dove-colored limestone, intermingled with light grey dolomite, in massive beds; sometimes for a thickness of twelve or fifteen feet no planes of stratification are discernible. In the lower beds, and in those just above the middle, the dolomite predominates; the middle and upper beds are nearly pure limestone; other beds show, on their weathered surfaces, raised reticulating lines of grey dolomite" (Brainerd and Seely, 1890b, p. 2).

In its southern extension the Shelburne marble is also divided by a dolomite bed into two nearly pure marble bands. The lower marble

"and the lower ten to forty feet of the ... [upper marble] ... deposit have slightly discontinuous irregular dolomite veins. These do not cross bedding planes but vary in orientation and position within a bed. The dolomite weathers to lighter color than the calcite and appears as minute chains on surfaces" (Bain, 1934, p. 127).

The dolomite that separates the upper and lower marbles is made up "of slightly silicified massive beds averaging three feet thick" (1931, p. 509). These dolomites are rather sandy in places. The minute dolomitic "chaining" and "curdling", that forms a network on the weathered surfaces of both the limestones and marbles is the most characteristic feature of the Shelburne and is very useful in identifying it where the rock is not metamorphosed to a distinctive white marble.

The white-weathering calcite marble zone along the marble belt south of Brandon averages from 500 to 600 feet thick (Bain, 1931, p. 509, 523; 1938, p. 10). Here the Shelburne probably extends upward through stratigraphic zones which grade laterally into higher Beekmantown strata farther north. It thins visibly north of Brandon, but due to intricate folding and unfavorable conditions of exposure the amount or mode of thinning at most localities is unknown. At one locality a little north of the New Haven River (Pl. 10) the lower marble thins out almost to extinction along the strike. Here probably secondary dolomitization is indicated where a quartzite bed strikes north from a central position within beds of pure marble into a similar position in extensions of the same strata which are dolomitized. The nondolomitic limestones are between 275 and 300 feet thick on the west limb of the Middlebury synclinorium in eastern Shoreham township. Still farther west, at points north and south of Shoreham village, and particularly on the western slope of Mutton Hill, the limestones almost completely pinch out locally along the strike. At Mutton Hill a zone of chert beneath the limestones and one of sandstone above do not approach any closer to each other where the limestones disappear. Therefore it is probably, as at the locality north of the New Haven River, that their disappearance is produced by secondary dolomitization. Along Lake Champlain in the region of Fort Ticonderoga the nondolomitic limestones are absent (Ulrich and Cushing, 1910, p. 98). This is probably due to secondary dolomitization also, and thus the stratigraphic zone may not be missing.

Fossils reported from the Shelburne indicate little concerning its stratigraphic position. Brainerd and Seely (1890a, p. 504; 1890b, p. 6) reported the straight cephalopod *Orthoceras* (?) *primigenium* (Vanuxem) and mentioned a new species of marine algae *Cryptozoön steeli* from the pure nondolomitic limestones in the Shelburne. *Orthoceras* (?) *primigenium* was first described (Vanuxem, 1842, p. 36; Hall, 1847, p. 13) from beds of the "Calciferous" in the Mohawk Valley, New York, now included in the Tribes Hill formation (Ulrich and Cushing, 1910). Whitfield (1889, p. 56) reported and described "*Orthoceras primigenium*" from middle Beekmantown beds in the Champlain Valley. Ruedemann (1906, p. 504) has indicated that this form may have been misidentified by both Hall and Whitfield. *Cryptozoön steeli* (Seely) (Seely, 1906, p. 161-162, 166–168) is likewise a poor guide fossil.

Rodgers (1937, p. 1576–1577) has described a "stratigraphic unit", the †Whitehall formation, in the west limb of the Middlebury synclinorium in Shoreham township and as far south as Whitehall, New York. It includes the bulk of the nondolomitic Shelburne facies and extends several feet lower to the bottom of a "basal sandstone". He reports fossils from the †Whitehall formation in Shoreham township (1937, p. 1577) that were stated by Ulrich to indicate horizons "correlative with the Strites Pond beds of the Philipsburg, Quebec sequence". Ulrich (1911, p. 631–639) has referred to a fauna in the "Little Falls dolomite" at Whitehall (presumably lowest †Whitehall) which he correlates with that of the Chepultepec dolomite of the southern Appalachians and with the Gasconade of the Ozark section,—the *Helicotoma uniangulata* fauna—lowest stratigraphically in the Ordovician. Wheeler (1941a; 1941b; 1941c; 1941d), however, correlates most of the †Whitehall with the Upper Cambrian Hoyt (Cushing *et al.*, 1910, p. 97) limestone at Saratoga, New York, and more distantly with the Trempealeau of the Upper Mississippi Valley. Pending clarification of these apparently divergent conclusions it seems advisable to follow the procedure initiated by Cushing (1908, p. 171), and since generally accepted (Wilmarth, 1938, p. 147–148), of placing the "Division B Beekmantown" beds in the lowest Ordovician.

CUTTING DOLOMITE (new name): Although similar to other Beekmantown dolomites, this formation is identified by its position (Brainerd and Seely, 1890a, p. 509–510) above the marbles and beneath the interbedded limestones and dolomites of the Bascom formation. At its base is a very thinly laminated sandstone that bears the worm boring "*Scolithus*." Nodules of black chert, found in the upper part of the formation, are also useful in identification. Brainerd and Seely (1890b, p. 2) have effectively described the Cutting at its type locality in eastern Shoreham township as follows:

"4. Magnesian limestone like No. 2, frequently containing patches of black chert	120 ft.
"3. Sandstones, sometimes pure and firm, but usually calciferous or dolomitic	70 ft.
"2. Magnesian limestone in thick beds, weathering drab	100 ft.
"1. Grey, thin bedded, fine grained, calciferous sandstone, on the edges often weathering in fine lines, forty or fifty to the inch, and resembling coarse grained wood. Weathered fragments are frequently riddled with small holes, called *Scolithus minutus* by Mr. Wing	60 ft."

In the eastern part of the area the basal sandstone of Zone 1 and the sandy dolomites of Zone 3 are not so prominent as in Shoreham township. On the east limb of the Middlebury synclinorium the dolomite passes laterally southeastward into a blue limestone that has a "coarsely curdled" surface formed by localized dolomitization. South of Brandon this "curdled" limestone may either disappear stratigraphically or pass laterally into the upper part of the Shelburne marble as exposed in Pittsford township. According to Bain (1934, p. 127) the green stripes in the marbles at Pittsford are where dolomite has "recrystallized" to form chlorite and actinolite. Possibly this phenomenon may account for a complete lithologic change in the Cutting dolomite south of Brandon.

Fossils are not visibly abundant in most of this formation, possibly because they have been destroyed by dolomitization. "*Scolithus*", however, is prominent in the basal sandstone. Wheeler, (1941c) reports the *Ophileta complanata* and *Lecanospira compacta* faunas "above the cross-bedded base" of the Cutting. Inasmuch as the Cutting forms the lower portion of a stratigraphic unit which also includes the fossiliferous limestone beds at the base of the overlying Bascom formation its correlation, distribution, and stratigraphic relations are discussed in the following paragraphs.

Bascom Formation (new name): In addition to limestone this formation contains dolomite, sandstone, quartzite, limestone breccia, and sandy calcareous shales, representing the greatest lithologic variety found in any formation of the Cambro-Ordovician succession in west-central Vermont, thus distinguishing it. The formation has been described in Shoreham township (Pl. 10) as follows (Brainerd and Seely, 1890b, p. 3):

"4. Blue limestone in thin beds, separated from each other by very thin tough slaty layers, which protrude on the weathered edges in undulating lines. The limestone often appears to be a conglomerate, the small enclosed pebbles being somewhat angular and arenaceous	100 ft.
"3. Sandy limestone in thin beds, weathering on the edges in horizontal ridges one or two inches apart, giving to the escarpments a peculiar banded appearance. A few thin beds of pure limestone are interstratified with the siliceous limestone	120 ft.
"2. Drab and brown magnesian limestone, containing also toward the middle several beds of tough sandstone	75 ft.
"1. Blue limestone in beds one or two feet thick, breaking with a flinty fracture; often with considerable dolomitic matter intermixed, giving the weathered surface a rough curdled appearance [Pl. 3, fig. 3]; becoming more and more interstratified with calciferous sandstone in thin layers, which frequently weathers to a friable ocherous rotten-stone	80 ft."

North and east of Shoreham dolomite and sandstone in the lower half of the formation decrease and limestone increases, rendering Zone 2 hardly recognizable. The sands of Zone 3 northerly become "curdled" limestones and Zone 4 becomes the thickest sandy unit of the formation. A generalized section for the Hinesburg synclinorium as developed west of Muddy Brook in South Burlington (Pl. 10) is as follows:

(4) Thin-bedded slaty quartzites and sandstones
(3) "Curdled" limestone, some bedded dolomite
(2) Buff dolomite and thin-bedded dolomitic quartzites
(1) White marble or limestone with thin sandy or argillaceous stripes.

The sandy stripe of Zone 1 is found toward the east in the Hinesburg synclinorium, whereas the argillaceous stripe, much like that of Zone 1 on the east limb of the Middlebury synclinorium, or like the ocherous stripe at Shoreham and Cornwall, is found to the west. The buff dolomite of Zone 2 is insignificant in the Hinesburg synclinorium, but the associated sandy dolomite facies, although never more than 20 feet thick, is a persistent unit. It is rather closely associated both above and below with striped limestones similar to those of Zone 1. The "curdled" limestones and interbedded dolomites of Zone 3 in the Hinesburg synclinorium grade up out of the striped limestone of Zone 2. Near the north end of the Middlebury synclinorium the banded sandy beds like those of Zone 3 in Shoreham township lie at the base of the "curdled" facies and are very much reduced in thickness. They have not been observed in the Hinesburg synclinorium. Zone 4 in the Hinesburg synclinorium is much more sandy than Zone 4 in Shoreham township. Northward in the Middlebury synclinorium, however, the increase of sand in Zone 4 is very apparent (Fig. 4) and is similar in amount and appearance to the sands of this Zone farther north. South of Brandon, in the "marble belt", the Bascom formation loses its identity to a considerable extent, due apparently to metamorphism and to change in original lithologic character (chiefly loss of sand) or possibly to actual stratigraphic extinction of some of the beds. The beds of Zone 1 appear to pass laterally into the upper part of the Shelburne marble facies in Pittsford township, the typical Cutting facies being absent here. Zone 2 is probably represented by buff dolomite bands that crop out a little west of the Pittsford Valley quarries. The interbedded "curdled" and dolomite beds that extend west to the blue marbles at this latitude are like those of Zone 3. The less dolomitic blue marbles of the western quarry belt in Pittsford have a dark graphitic, siliceous stripe which may be traced northward into the sandy beds of Zone 4 west of Brandon.

The Bascom formation overlies the Cutting dolomite without apparent stratigraphic break throughout most of west-central Vermont. At most places the contact has dolomite below and limestone above. The Bascom is itself overlain by the Bridport dolomite without apparent break. The Bridport is absent from the east limb of the Middlebury synclinorium, where the Bascom facies is directly overlain by the Chazy group. The sandstones of Zone 4 of the Bascom are the youngest beds present at the surface in the Hinesburg synclinorium. Brainerd and Seely (1890b, p. 3) have indicated a thickness of 375 feet for the Bascom in Shoreham township. Intricate folding on the east limb of the Middlebury synclinorium and in the Hinesburg synclinorium has made measurement of thickness impracticable. However, it is probably about the same as in Shoreham township, although Zone 4 appears to be somewhat thicker, possibly at the expense of the overlying Bridport dolomite facies. Rodgers (1937, p. 1579) has suggested a stratigraphic break between Zones 2 and 3 in Shoreham township. The "basal sandstone", which he has reported in Zone 3, appears to grade laterally into "curdled" limestone beds in an easterly and northerly direction. The striped limestones and marbles, which in these directions supplant the dolomite in Zone 2, are not readily distinguished from the "curdled" limestone of Zone 3 in the eastern and northern localities. Consequently, a stratigraphic break between the upper and lower Bascom cannot be detected so readily in the Hinesburg synclinorium or in the eastern part of the Middlebury synclinorium.

Fragmented fossils, more or less poorly preserved, have been discovered at several well-distributed localities in both the Hinesburg and Middlebury synclinoria. They occur, or are more readily observed, in the limestone beds, where, after selective dolomitization of the shell, the "latent image" has weathered out as a slightly raised buff pattern on the surface. Rarely are fossils found that can be freed from the matrix, and they are rarely recognizable on the fresh surface. In Shoreham township fossils are most common in Zones 1 and 4. In Zone 1 are straight or slightly curved cephalopods 1 or 2 inches long, associated with less abundant gastropods that appear in section to be plani-spiral with the preserved portion less than 1 inch in diameter. Zone 4 also contains various gastropods and cephalopods and more or less plentiful trilobite fragments. In eastern localities, where limestones are abundant, the small straight or slightly curved cephalopods are found in Zones 2 and 3 also. A few coiled cephalopods have been observed in the upper beds in Shoreham township but not in the highly deformed areas to the east and northeast. Very close examination of even the eastern rocks reveals numerous minute gastropods, brachiopods, sponge spicules, and crinoid fragments. At a locality 1¼ miles southwest of New Haven Junction (Pl. 10) two specimens of what appears to be a *Cryptozoön* were found in Zone 3.

Wing (Dana, 1877a, p. 342–344), Brainerd and Seely (1890b, p. 6, 7, 10, 14), and Whitfield (1890, p. 28) recognized faunal differences between the upper and lower Bascom. From the beds of Zone 1 in Shoreham township Wing and Brainerd and Seely report what is probably *Lecanospira compacta* (Salter). From Zones 3 and 4 Wing reports *Isoteloides whitfieldi* Raymond, *Hystricurus conicus* (Billings), *Maclurites matutinus* (Hall), and *M. sordidus* (Hall). Brainerd and Seely also report *Isoteloides whitfieldi* as well as *Maclurites affinis* (Billings) and *Schroederoceras eatoni* (Whitfield) in Shoreham township, all apparently from Zones 3 and 4.

The Lecanospiras from Zone 1 were referred to as both "*Ophileta compacta*" and "*Ophileta complanata*" by Wing (Dana, 1877a, p. 342) and as "*Ophileta complanata*" by Brainerd and Seely (1890b, p. 6). Inasmuch as the fossils reported by Wing and by Brainerd and Seely were probably identified by Billings and Whitfield respectively, both of whom had figured and described in their publications only the depressed spired gastropod, *Lecanospira compacta* (Salter), the specimens were probably *Lecanospira compacta* (Salter) and not the low-conical spired gastropod *Ophileta complanata* Vanuxem, a form which up to that time had been inadequately described and figured (Bridge, 1930, p. 198–

199, 203–205). Elsewhere in North America the *Lecanospira compacta* fauna is reported from zones stratigraphically above the beds carrying *Ophileta complanata* (Bassler, 1919, p. 96–103) or from zones above that carrying the *Helicotoma uniangulata* fauna (Butts, 1926, p. 91, 93), both of lower Beekmantown age.The *Lecanospira compacta* fauna lies beneath the *Ceratopea keithi* or Cotter fauna (Ulrich, 1911, p. 665–666; Butts, 1926, p. 98; 1936, p. 26–27) and a probably still higher faunal zone characterized by *Eurystomites kelloggi* (Brainerd and Seely, 1890b, p. 17) of upper Beekmantown age.

Thus the lower half of the Bascom formation and parts of the Cutting dolomite (Wheeler, 1941c) are probably in the *Lecanospira* zone. Wheeler (1941a; 1941b) indicates that the zone of the Cotter fauna is present, apparently in the Bascom. Brainerd and Seely (p. 20–22) recognized Zones 3 and 4 of the Bascom formation at Fort Cassin near the mouth of Otter Creek on Lake Champlain in beds that carry *Eurystomites kelloggi*. The upper half of the Bascom is therefore within a distinct faunal zone of widespread significance in the upper Beekmantown. It will be discussed later in connection with the Bridport dolomite, with which it appears to be conformable.

Bridport Dolomite (new name): This formation is identified primarily by its stratigraphic position (Brainerd and Seely, 1890a, p. 509–510). Beneath it is the trilobite-bearing sandy or shaly limestone at the top of the Bascom formation; above it, at most places in west-central Vermont, are the *Maclurites magnus*-bearing beds of the middle Chazy. Brainerd and Seely (1890b, p. 6) described this formation as a "fine-grained magnesian limestone in beds one or two feet in thickness, weathering drab, yellowish or brown. Occasionally pure limestone layers occur, which are fossiliferous, and rarely thin layers of slate." Where the overlying and underlying beds are covered the Bridport dolomites may be confused with the somewhat similar dolomites of the Beldens formation that occur in the Chazy above the *Maclurites* beds, in areas east of Shoreham. Inasmuch as the Bridport pinches out east of eastern Shoreham township (Brainerd and Seely, 1890b, p. 6) these two dolomites rarely crop out in the same vicinity (Pl. 10). A knowledge of its distribution is thus helpful in the identification of the Bridport. North and south of Shoreham the Bridport crops out over considerable areas, but it is seen nowhere east of the meridian of eastern Shoreham township. In eastern Shoreham Brainerd and Seely (1890b, p. 3) found it to be 470 feet thick. To the north (p. 20) and south it apparently has about the same thickness. The contact between the Bascom and the Bridport appears to mark merely a change from limestone to dolomite unaccompanied by a distinct stratigraphic break (Fig. 4). However, the upper contact of the Bridport, is in most places a distinct depositional break, above which may lie any of the formations of the Chazy or Black River-Trenton that crop out in the area occupied by the Bridport.

Fossils are not readily found in the Bridport dolomite. A few obscure forms were observed in southeastern Bridport township, all in limestone layers. Brainerd and Seely (1890b, p. 7) have reported *Bucania tripla* Whitfield, *Turritospira confusa* (Whitfield), and *Isochilina seelyi* (Whitfield) in Shoreham township. Whitfield (1890, p. 38–39) described *Bathyurus glandicephalus* in the same locality. This fauna is also found on Providence Island in Lake Champlain (Brainerd and Seely, 1890b, p. 20), where it overlies directly the upper Bascom beds that carry *Eury-*

stomites kelloggi, *Plethospira cassina*, and other fossils (p. 17) characteristic of the famous Fort Cassin fauna. The Bridport also overlies rocks that carry the Fort Cassin fauna at Thompson's Point in southwest Charlotte township (p. 14–16) Brainerd and Seely (p. 20–22) pointed out that the beds in the isolated outcrops at Fort Cassin are probably equivalent to Zones 3 and 4 of the Bascom. Whitfield (1890, p. 28) and Cushing (1905, p. 362–364) indicated that the upper half of the Bascom and the whole of the Bridport comprise a distinct faunal zone as well as a uniquely distributed stratigraphic unit. The Bridport dolomite is therefore correlated, along with horizons as low as the bottom of the sandstone at the base of Zone 3 of the Bascom, with the "Cassin" formation of Cushing.

The Cassin has been more or less closely correlated with several formations in widely distributed localities of eastern North America. The closest homotaxial correlation specified in the literature is with the Smithville formation of the Ozark region (McKnight, 1935, p. 26–28), which carries *Eurystomites kelloggi*, *Plethospira cassina*, and other species found at Fort Cassin (Ulrich, 1911, p. 668–670). The Beekmantown beds at Ogdensburg, New York, occurring above the *Lecanospira compacta* zone, have a fauna Ulrich considers as probably below that of Fort Cassin (Cushing, 1916, p. 48). The upper Ogdensburg formation has two species of *Eccyliopterus* also reported from the Cotter formation of Missouri (Weller and St. Clair, 1928, p. 84). The Cotter lies stratigraphically below the Smithville (McKnight, 1935, p. 27) and above the Roubidoux with its *Lecanospira* fauna (Bridge, 1930, p. 126–128). Ulrich (1911, p. 653–655) has reported a situation similar to that of the Ogdensburg in the Beekmantown of Maryland, stating that the third division of the Beekmantown of that locality, Bassler's (1919, p. 103–104) *Ceratopea* zone, "probably . . . represents an earlier facies of the . . . [Bascom] . . . fauna than the one found at Fort Cassin, Vermont". Butts (1936, p. 27, 29, 32) reported that the beds with the Smithville fauna (Ulrich, p. 660) overlie those of the Cotter fauna in central Pennsylvania and Virginia. Neither the zone of the Cotter fauna, or *Ceratopea keithi* as in Alabama (Butts, 1926, p. 98), nor the combined *Ceratopea* and *Turritoma* zones as in Maryland (Bassler, 1919, p. 103–105) has been recognized in the Champlain Valley.

The rocks of the Philipsburg slice (Fig. 5) in southern Quebec also contain the *Eurystomites kelloggi* fauna (Ami, 1896, p. 123–125, 130–131). Here the Beekmantown lithologic units are in part comparable (Brainerd and Seely, 1890b, p. 23; Bradley, 1923, p. 328–334) to those of west-central Vermont, but above the Bascom Zone 4 facies, probably represented by the Solomons Corner formation near Philipsburg, the beds are largely limestones. These higher strata, in a stratigraphic position similar to that of the Bridport at Shoreham, appear also to have the *Eurystomites kelloggi* fauna, suggesting the faunal unity of all of the beds above the middle Bascom, including the Bridport, which is not a unit of widespread stratigraphic significance. Authors have used the term "Beekmantown E" (the Bridport) to designate the age of the highest beds of Beekmantown equivalents at localities outside of the Champlain Valley area, simply on their stratigraphic position. Actually the "Beekmantown E" dolomite has neither stratigraphic position nor a diagnostic fauna but is a lithologic unit, possibly of secondary nature, which may change laterally and cut across stratigraphic horizons. Lithologic similarity (Ulrich, 1911, p. 654–655), as well as stratigraphic position, has been used in correlations with the Bridport. Inasmuch as lower and middle Beekmantown beds grade laterally eastward from dolomite to limestones in west-central Vermont and as this also appears to take place between west-central Vermont and the Philipsburg slice in rocks of upper Beekmantown age, lithologic correlation with the Bridport is probably unwarranted.

Distribution and genetic relations.—Strata bearing the probable lowest Beekmantown *Helicotoma uniangulata* and *Ophileta complanata* faunas have been variously reported from the Mohawk Valley (Vanuxem, 1842, p. 36; Cleland, 1900; 1903) southwest of west-central Vermont, as well as in the southern Champlain Valley (Ulrich, 1911, p. 631–639; Wheeler, 1941a; 1941b; 1941c; 1941d). A sandy facies is particularly abundant in the western part of the area and thins eastward, indicating a western source. It is absent from the "marble belt" to the southeast. Somewhat west of the area, at Thompsons Point, Vermont, on Lake Champlain, a rather coarse dolomitic breccia is associated with the sands. The thickest section of the sandstones observed by the author is at Beekmantown, New York, northwest of the area of west-central Vermont. The distribution of this sand suggests a source of clastics centered in the region of the present Adirondacks.

Although the *Lecanospira compacta* zone has not been faunally well established in west-central Vermont, new evidence and a reconsideration of earlier fossil discoveries indicates that it possibly includes the upper part of the Cutting dolomite and Zones 1 and 2 of the Bascom formation; this amounts to as much as 300 feet, an average thickness for most of the area. Sands scattered vertically in the *Lecanospira* zone have about the same horizontal distribution as those in the lower Beekmantown. The *Lecanospira* zone is reported west of west-central Vermont at Beekmantown, New York, the type locality of the "group",—northeast of the Adirondacks (Whitfield, 1889, p. 41–42; Ruedemann, 1906, p. 419, 443, 496; Butts, 1936, p. 27). Here the *Scolithus*-bearing basal Beekmantown sandstone is apparently the only other fossiliferous zone of the "group" represented (Ulrich, 1938, p. 23). Northwest of the Adirondacks (Cushing, 1916, p. 37–48) the *Lecanospira* zone thins (less than 100 feet), and lower zones of the Beekmantown group are not established, the original easternmost deposits of the *Lecanospira* zone in Vermont have been eroded so that the limits are indeterminate, but the trend from west to east would indicate reduced dolimitization, a decrease in sand, and increased thickness in this direction. At the north end of the Rosenberg slice a stratigraphically equivalent slate succession takes the place of the Beekmantown limestones (Schuchert, 1937, p. 1070–1075). This succession contains limestone pebbles with "several specimens of small and not well-preserved, closely septate cephalopods, similar to those occurring in the limestone near Brandon" (p. 1075), a limestone the author placed in the *Lecanospira* zone. Eastward in the Rosenberg slice the slate terrane broadens, and in the Oak Hill slice the Ordovician as well as the Upper Cambrian may be represented in the higher parts of the slate succession. Like the Lower and Upper Cambrian argillaceous sediments those of the Beekmantown appear to have an eastern source. The zone of *Ceratopea keithi* or the Cotter fauna is not well established in west-central Vermont.

In Shoreham township and along Lake Champlain the zone of *Eurystomites kelloggi* is 690 feet thick (Brainerd and Seely, 1890b, p. 3). On the east limb of the Middlebury synclinorium it would seem that it must thin to about 200 feet with the loss of the Bridport dolomite, but if Zone 4 of the Bascom is here merely a lateral gradational facies of the Bridport facies in Shoreham, as seems possible, thinning is slight. In the Philipsburg slice this zone is about 1000 feet (Logan, 1863, p. 844–845; McGerrigle, 1930, p. 185) thick. The sandstones at the base of the zone in Shoreham township thin toward the east as well as toward the north and south, indicating a source of clastics centered approximately at the site of the present Adirondacks. The zone as a whole probably pinches out west of the Adirondacks (Cushing, 1919, p. 48–50). Within Zone 4 of the Bascom formation (Brainerd and Seely, 1890b, p. 6) are beds of slaty calcareous sandstones, which thicken and become sandier toward the east (Fig. 4); and to the north in the Hinesburg synclinorium sand is increased so that Zone 4 is a thin-bedded quartzite. At a meridian about 4 miles west of these latter outcrops, west of the Champlain thrust in southwestern Charlotte township (Brainerd and Seely, 1890b, p. 14–17) and at Providence Island (p. 17–20), the equivalent horizons contain limestone and lack the slaty calcareous sandstones and quartzites. It is inferred that the original east-west distance between the contrasting facies has been reduced, since deposition, by movements on the Champlain overthrust (Fig. 6). This is the lowest horizon at which sands from an eastern source have been found. Possibly they represent reworked sands originally coming from the west.

ORDOVICIAN-CHAZY GROUP

General correlation.—Hall (1847, p. 14–36) expanded the Chazy limestone, which as originally described (Emmons, 1842, p. 107, 315–317) contained only the middle or *Maclurites magnus* zone, to include the lower or *Hebertella exfoliata* zone and at least a part of the upper or *Camarotoechia plena* zone. These three fossil zones are approximately equivalent (Raymond, 1906, p. 573–574) to three lithologic divisions found along Lake Champlain variously designated as A, B, and C (Brainerd and Seely, 1888, p. 323–330) or Day Point, Crown Point, and Valcour (Cushing, 1905, p. 368). Elkanah Billings (Hunt, 1868, p. 227–228), Wing (Dana, 1877a, p. 341, 342, 343, 344, 345, 415), and Brainerd (1891, p. 299, 300) have reported Chazy strata of somewhat different facies east of the lake. Chazy fossils are reported from rocks of this latter facies at several localities both to the north (Logan, 1863, p. 272–274, 280–281, 854–858; Brainerd, 1891, p. 298; Ells, 1896, p. 29J) and to the south (Billings, E., 1872b, p. 133; Dana, 1877a, p. 337–339) of west-central Vermont. Strata in west-central Vermont designated as lower Chazy (Brainerd, 1891, p. 299) appear actually to be the uppermost Bascom beds of the Beekmantown group.

Formations.—Inasmuch as the Day Point beds thin or pinch out, north of west-central Vermont, and inasmuch as the Valcour undergoes considerable change in lithology and obliteration of fossils eastward from Lake Champlain, Crown Point limestone is the only term used along Lake Champlain appropriate in the eastern area. The terms Beldens formation, Weybridge member of the Beldens, and Middlebury limestone are here applied to beds probably mainly equivalent to the Valcour.

CROWN POINT LIMESTONE (Cushing, 1905): The limestone is lead gray, compact, massive, weathering to gray; on dip slopes this weathered surface appears to be stippled with numerous light-buff dolomitic protuberances about one quarter to half an inch in diameter. The dolomitic patches may be elongated and anastomosing giving the surface a "curdled" rather than a stippled appearance (Pl. 4, fig. 1). Where fossils are present the dolomitization follows the outline of the section of the fossil shell on the weathered surface, producing the preservation noted in the Beekmantown limestones. In exposures cutting across the bedding the buff stippling is observed to be produced by thickening and thinning of faint buff dolomitic stripes or partings parallel to the bedding; these may be 5 inches apart. One- to 2-foot beds of buff dolomite commonly occur within the limestone, particularly near the op and bottom, but dolomitization is not sufficient to blot out its dominant lime-

FIGURE 1. CROWN POINT LIMESTONE
1 mile NNW. of Cornwall village.

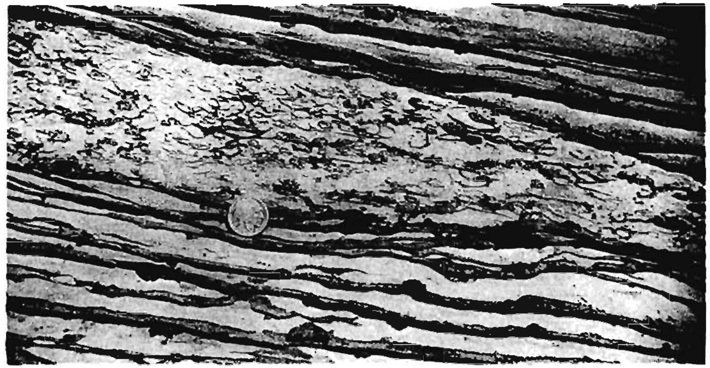


FIGURE 2. WEYBRIDGE MEMBER OF BELDENS FORMATION
Close-up view of channel filled with fossil fragments, 1⅝ miles S. of New Haven Junction.

FIGURE 3. BELDENS FORMATION
Eastward-dipping inverted contact with Middlebury limestone, below falls of Otter Creek at Middlebury village. (*See* Seely, 1910, p. 30, Pl. 49.)

FIGURE 4. MIDDLEBURY LIMESTONE
¼ mile E. of Center in Weybridge township.

CHAZY STRATA

FIGURE 1. ORWELL LIMESTONE
¾ mile SSW. of Buck Mountain, Waltham township.

FIGURE 2. GLENS FALLS LIMESTONE
2 miles SSW. of Orwell village.

TRENTON STRATA

stone features. In the deformed rocks of the eastern localities, except where shielded from deformation, the formation is rather difficult to distinguish. Where it can be traced out of a protected zone the dolomitic stippling changes into very distinct elongated buff streaks.

The large depressed-spired gastropod, *Maclurites magnus* (Le Sueur), its operculum, and a small spherical algal form most like *Girvanella ocellata* (Seely) are very useful in field identification of the Crown Point limestone. The weathered sections of *M. magnus* may be confused with those of other species of *Maclurites* above the Chazy, but if a closer examination of the outcrop reveals the usually abundant small algae the presence of the Crown Point is established. *M. magnus* is 3 to 4 inches in diameter with about three rapidly expanding whorls visible. *G. ocellata* is about half an inch in diameter and shows, megascopically, numerous thin concentric layers. With extreme deformation, species of *Maclurites* and *Girvanella* are completely obliterated, but the presence of the former may be still indicated by the conical operculae which apparently were rather resistant. In the eastern localities, however, even the operculae are rarely seen. The buff dolomite stippling suggests the presence of *Girvanella* at many localities. At several places this stippling actually appears to have been produced by the dolomitization of indistinct species of *Girvanella*.

The Crown Point overlies the upper Beekmantown Bridport dolomite at the meridian of Shoreham and west (Fig. 4). It lies on the Bascom formation and is overlain gradationally by the Beldens formation east of the meridian of Cornwall. At most localities, particularly the eastern ones, the base of the typical Crown Point facies is thought to be a reddish buff, locally sandy dolomite about 5 to 10 feet thick. At the north end of the Middlebury synclinorium some beds of Chazy rather than Beekmantown aspect are found immediately beneath the reddish buff dolomite. Several feet of the Crown Point next above the dolomite bed are streaked with buff to brown dolomite separated by blue-weathering limestone. Where the reddish-buff dolomite bed is indistinct, it is difficult to distinguish the lowest Crown Point from the Bascom. Close examination, however, reveals that the dark streak of the lower Crown Point is strictly dolomitic, whereas that of the Bascom is siliceous. In spite of this distinction and of the presence of the reddish-buff dolomite between them, the succession from the Bascom to the overlying Crown Point, in eastern localities, gives the impression of grading upward from more clastic to less clastic rocks. A much clearer lithologic break separates the Bridport from the Crown Point in the western localities.

The typical Crown Point facies is 50 to 60 feet thick at Orwell (Brainerd, 1891, p. 300; Raymond, 1906, p. 507) pinching out southwest of the village. To the north in Cornwall township it is 150 feet thick (Brainerd, 1891, p. 299). In the eastern localities it appears to be about 150 feet thick and is bounded both above and below by somewhat clastic rocks that thicken eastward.

At Crown Point, west of this area, the middle Chazy limestone is about 280 feet thick (Brainerd, 1891, p. 300; Raymond, 1906, p. 553–554). At an intermediate locality, in Waltham township, Vermont, it may be 400 feet thick. Clastics 23 feet thick, lie below the limestone at Crown Point and are, according to Raymond (1906,

p. 569), the tangential sands beneath the Chazy group. North of the latitude of Crown Point the middle Chazy thickens along the Adirondack border by addition of beds to the top (p. 570) to a maximum observed extent of 400 feet at Valcour Island (p. 527). North of the latitude of Chazy and Isle La Motte the typical *Maclurites magnus* beds are not exposed.

South and west of Orwell village along Lake Champlain, the Crown Point and the overlying Chazy strata pinch out along the strike in some places, particularly to the west, possibly due to non-deposition (Brainerd and Seely, 1890b, p. 13). The Crown Point beds (Bain, 1931, p. 516; 1933, p. 77), as well as some of the overlying upper Chazy limestones, are exposed to the south in the "marble belt" (Billings, 1872b, p. 133; Dana, 1877a, p. 337–339), south of the eastern part of the area of west-central Vermont. Here they form the bulk of the marble deposits at West Rutland. These marbles are apparently cut out south and east of West Rutland beneath a pre-middle Trenton erosion surface (Bain, 1934, p. 136) although a fossil "assigned to the Chazy" has been reported from a locality much farther south along the strike in southwestern Vermont (Prindle and Knopf, 1932, p. 274). West of West Rutland the Chazy rocks occur in a succession covered structurally by the Taconic Allochthone, which at its western limits in the Hudson Valley lies on Trenton beds that are unbreached by erosion. Therefore their original and possibly present continuity southwest of West Rutland toward southern Appalachian exposures cannot be determined.

BELDENS FORMATION (new name): Bright, orange-buff-weathering dolomite in beds 1 or 2 feet thick is interbedded with snow-white marbly limestone (Pl. 8, fig. 1). The sharp color contrast between these two lithologic types is distinctive. The dolomite has been described as "chamois weathering" (Gordon, 1923, p. 258) and the peculiar sharply cut reticulations of the weathered surface make it look like "thread-scored beeswax" (Wing, letter to James Hall, 1867). Gleaming outcrops can be seen in the fields southeast of Beldens (Pl. 10) and west of the highway north from Middlebury. In the lower part of the formation the white marble beds are less abundant and the rock is a duller buff or gray and less dolomitic. Here mottled blue-gray limestones rather than white marbles are interbedded. In the lower middle part of the formation, above, below, and interbedded with the Weybridge member are reddish-buff dolomites in beds 8 inches to 1 foot thick and separated by half-inch black slaty partings.

WEYBRIDGE MEMBER (new name): This member is (Pl. 4, fig. 2) the "striped stratum" (Dana, 1877a, p. 343, 344, 345) which Wing (p. 415) described as "a way-mark by which the rock can be recognized without its fossils". The typical facies as exposed at "Weybridge Upper Falls" (p. 345) 1 mile east-northeast of Weybridge village (Pl. 10), now locally known as Huntington Falls, is a limestone with sandy streaks about half an inch wide that weather into raised ridges about 1 inch apart. They are more granular and darker than the intervening blue limestone. This readily recognizable facies makes the Weybridge a very useful guide horizon. In recent rock cuts the sandy stripes are indistinguishable, but weathering quickly attacks the limestone and in less than 1 year pale-buff streaks still showing no relief reveal the siliceous material. The relief appears several years later.

Original sedimentary structures are abundant and varied in the Weybridge. Channels cutting across several of the sandy laminae are common; each is lined with one of the sandy bands. In many places the channels are filled with fossil fragments (Pl. 4, fig. 2). Cross-bedding is also very common; current ripple marks were seen at a few localities. These features provide criteria for determining the tops of the beds in the highly folded eastern rocks.

These clastic beds thicken eastward (Fig. 4). At "The Ledges" in Cornwall township Wing (Letter to James Hall, 1867) reported 40 to 50 feet. A trace exists in the northeast corner of Benson township, but in general west of the meridian of Cornwall the clastics finger out within the Beldens formation (Pl. 10). At Huntington Falls on a meridian about 1 mile east of the Cornwall section they are 43 feet thick, but on the east limb of the Middlebury synclinorium, 3 miles east of the Cornwall section, the Weybridge member thickens markedly. There this facies apparently occupies all of the Beldens formation beneath the interbedded white marbles and chamois-weathering dolomites, which at Huntington Falls is about 420 feet. In the eastern localities the estimated thickness is about 400 to 500 feet. The member is not all clastic, a greater volume of dolomite and limestone being present. The scale of the original sedimentary structures is also increased in the eastern exposures. The dark sandy stripes, particularly in the upper part of the member, are 1 to 2 inches across and the intervening blue limestone bands as much as 5 inches wide. Fore-set beds 5 feet long have been observed. Channels 2 or 3 feet deep are plentiful. With one exception the Weybridge facies has not been found north or south of west-central Vermont. The author has observed apparently Weybridge strata, although very thin, on the westward-facing hill slope east of the eastern quarry openings at West Rutland. The observed distribution of these beds suggests a source of clastics somewhwere northeast of the Rutland region and extending possibly for a considerable distance in a northeasterly direction.

The thickness of the Beldens formation ranges from 0 at the meridian of Orwell to 600 or 700 feet on the east limb of the Middlebury synclinorium (Fig. 4). Half of the eastward increase is represented by the increased thickness of the Weybridge clastics. The latter are 130 feet above the top of the Crown Point limestone at "The Ledges" in Cornwall. The intervening interbedded dolomites and limestones forming the lower quarter of the Beldens grade laterally eastward from Cornwall into the base of the Weybridge facies, with which similar dolomites and limestones are interbedded. The latter beds lie immediately above the typical Crown Point limestone facies on the east limb of the Middlebury synclinorium. The interbedded dolomites and limestones above the Weybridge member at Cornwall thicken from 90 feet at the meridian of "The Ledges" to about 500 feet at Huntington Falls. East of Huntington Falls the lower 275 feet of the interbedded series grades laterally eastward into the top of the Weybridge. The typical interbedded white marbles and chamois-weathering dolomites occupy the upper 200 to 300 feet of the interbedded series east of Huntington Falls, overlapping the Weybridge in that direction. To the west of the meridian of Huntington Falls the marbles disappear completely, partly because of decreased metamorphism but largely due to convergence and lateral gradation into the interbedded limestones and dolomites of "The Ledges".

Inasmuch as the upper Crown Point limestone and overlying beds offlap in the Adirondacks possibly the westward extinction of some of the higher Chazy beds may be similarly explained.

At West Rutland, south of west-central Vermont, the marbles above the *Maclurites* bed (Bain, 1931, p. 516) are comparable in thickness to those of the Beldens farther north but are possibly cut out to the south and east of West Rutland beneath a pre-middle Trenton erosion surface (1934, p. 126). North of west-central Vermont in the Highgate Springs slice (Fig. 5) the upper 350 to 360 feet (Logan, 1863, p. 280) of interbedded dolomites and limestones (p. 273–274, 854–858), lithologically similar to the Beldens formation (Brainerd, 1891, p. 298, 299), is exposed at the axes of anticlines or as projections from beneath the Rosenberg slice. Southwest of this northern locality, at Isle La Motte, Brainerd and Seely (1896, p. 310–312; Brainerd, 1891, p. 298) have reported a "dove limestone, with bands of magnesian limestone" in the upper Chazy, which resembles that in the Highgate Springs slice and which is similar in description and position to a "dove" limestone in the upper Chazy of the type locality at Chazy, New York, west of Isle La Motte. At Chazy the Beldens facies referred to by Brainerd and Seely as "Group C, 1–7" is 23 feet thick (1888, p. 325). The upper marble beds typical of the Beldens are poorly represented, but in the Highgate Springs slice to the east they are 200 feet thick (Logan, 1863, p. 280).

At only two localities were fossils observed by the author in the typical Beldens facies of west-central Vermont. About 1 mile southeast of Cornwall village, several straight cephalopods are preserved in a rare zone of gradation between a dove limestone and a dolomite bed. About 1 mile east of Weybridge village some gastropods, bryozoans, and crinoid stem fragments are preserved in strata locally undeformed, more argillaceous, and only slightly dolomitized. Fossil fragments are rather abundant in the sandy Weybridge facies, however. Wing reported bryozoans (Dana, 1877a, p. 345) and abundant brachiopods (Letter to James Hall, 1867) from the Weybridge. The author has observed high and medium-spired gastropods, straight cephalopods, and trilobites from this member. Transported fragments of the bryozoans and the shells of brachiopods fill the channels.

At Chazy (Raymond, 1906, p. 545) and Cooperville (p. 545–546), New York, Valcour Island (p. 531–532), Isle La Motte (Brainerd, 1891, p. 298), and at several other localities on both shores of Lake Champlain, reef deposits made up largely of *Stromatocerium* (Seely, 1904, p. 144-152) and corals (Raymond, 1924b, p. 72–76) are associated with or immediately beneath a Beldens facies in which dolomites are relatively scarce. The reef material is thus apparently distributed in the western part of the dolomitic facies in the Champlain Valley. Near the algal reefs are abundant trilobites and cephalopods and less abundant brachiopods, pelecypods, and gastropods designated by Raymond (1906, p. 550, 566) the *Glaphurus pustulatus* faunule—representing the best and practically only useful fossils obtainable from the Beldens zone. *Glaphurus pustulatus* (Walcott) is reported only from the Champlain Valley, where it is common in this zone at several localities (Raymond, 1906, p. 531–532, 545–546).

MIDDLEBURY LIMESTONE (new name): A buff-streaked, dark blue-gray, somewhat nodular and granular, thin-bedded, incompetent, partially dolomitic limestone (Pl. 4, fig. 4) crops out over a wide area west of Otter Creek at Middlebury village. This formation is well exposed in the ledges on the Middlebury College campus. The Middlebury limestone grades from the underlying Beldens formation through about 10 feet of interbedded buff dolomites and blue limestones (Pl. 4, fig. 3). At a quarry $\frac{5}{8}$ mile south of Center in Weybridge township (Pl. 10) a few feet at the top of the Middlebury show an increasing number of rather widely separated sand grains near the base of the overlying Black River-Trenton. Here this contact may be recognized by the upward change to sublithographic texture and by the abrupt appearance of the calcite veins that cut at all angles through the Black River-

Trenton beds. Calcite veins are found in the Middlebury but are not as abundant. Small flexures, fracture cleavage, and rectangular jointing are common. Repetition by folding and thinning of the formation on limbs of folds have made determination of thickness difficult. The Middlebury is estimated to be not more than 600 feet thick.

A tabulate coral, from the bed of the Otter Creek below the "Pulp Mill" covered bridge 1 mile north of Middlebury village (Pl. 10), is the only good fossil. It is, however, considerably recrystallized and identification has not been completed. The only other fossils are some segments of cystoid or crinoid stems and a few small planispiral gastropods rather poorly preserved.

At West Rutland a limestone reported by Wing from a point a short distance southwest of the village is similar in lithology and position to the Middlebury and contains (Dana, 1877a, p. 338) *Raphistoma staminium* (Hall) and the cystoid *Paleocystites tenuiradiatus* (Hall). Stratigraphically the Middlebury is equivalent to a succession above the Beldens zone and beneath the Black River-Trenton of the Highgate Springs area, which according to Logan (1863, p. 273–274, 280–281, 854) is made up of 50 to 100 feet (p. 273) of "greenish-grey calcareous fine grained sandstones . . . at the top interstratified with greenish shale" overlain by less than 60 feet (p. 274) of "blackish thin bedded shaly nodular limestones, partially magnesian". This succession carries Chazyan *Dinorthis* (*Plaesiomys*) *platys* (Billings) and *Ampyx* (*Lonchodomus*) *halli* (Billings),—the latter having been originally described at Highgate Springs (Billings, E., 1861b, p. 959–960). According to Kay (oral communication) this zone is about 300 feet thick in the Highgate Springs area. The Middlebury is apparently also equivalent in position to "Group C, 8–13" of the Chazy at Chazy, New York (Brainerd and Seely, 1888, p. 325), a succession that is 100 to 134 feet thick. This is the zone of abundant brachiopods, particularly *Camarotoechia plena* (Hall). *Camarotoechia major* Raymond, found at Valcour Island, is restricted to this zone (Raymond, 1906, p. 566–567). The pelecypod *Modiolopsis fabaformis* Raymond is restricted to the uppermost part of the *Camarotoechia plena* zone at Valcour Island (p. 567). At the latter locality cystoid and crinoid remains, particularly those of *Canadocystites emmonsi* (Hudson) and *Paleocystites tenuiradiatus* (Hall), are common in the uppermost Chazy beds (p. 518, 519, 566).

The Middlebury may be partly Black River in age. In the Ottawa Valley (Raymond, 1911 p. 189; 1916b, p. 326–327; Wilson, 1932, p. 135–146) the "dove" limestones of the Pamelia formation in the lower part of the Black River group are reported to lie between the upper Chazy *Camarotoechia plena* zone and the middle Black River Lowville. Dove limestones are not found in the Middlebury beneath the higher Black River strata in west-central Vermont, but with the lack of a good ossil record it is difficult to prove the absence of lower Black River strata in the Middlebury facies.

Distribution and genetic relations.—The Chazy overlaps southwestward in the belt of folding and overthrusting. The lowest Chazy strata are apparently restricted to the center of the northern Champlain Valley and were presumably laid down along the axis of a northeast-pitching trough.

The typical middle Chazy Crown Point limestone of the Champlain Valley ranges from 0 to 400 feet thick. It overlaps southwestward through the Champlain Valley region loosing beds at the base, but visibly offlapping along its northwestern boundary adjacent to the present Adirondacks. Along the Adirondack border it is bounded above and below and grades laterally westward into sandy beds that thicken westward, suggesting a westerly source. The original southeastward extent of the middle Chazy and its method of pinching out toward the southeast are uncertain. Premiddle Trenton erosion, evidence for which is reported in the "marble belt" a little south of west-central Vermont, may partly account for its absence in southeastern localities. It is buried by the Taconic Allochthone (Fig. 5) northeast and south-southwest. Farther south in the "marble belt" and along the Ordovician outcrop as far south as eastern Pennsylvania no Chazy crops out.

The upper Chazy Beldens formation offlaps along its northwestern boundary adjacent to the present Adirondacks. From the line of offlap extinction on the west the thickness of the Beldens increases to a maximum of 700 feet. Pre-middle Trenton removal may account for a part of the extinction at the southern limits of the outcrop, both at the south end of the lake and in the "marble belt". The actual extent of the Beldens to the southwest and northeast is obscure, due to its low structural position in those regions, where it lies beneath the Taconic Allochthone. Clastics of the Weybridge member, centered in the lower middle quarter of the Beldens, thicken from 0 at the meridian of Shoreham to a scattered distribution through a maximum of 500 feet of strata at the eastern limits of outcrop where they account for most of the increased thickness of the Beldens. They were probably derived from a near-by eastern source.

The Middlebury limestone and its probable equivalents in the uppermost Chazy along Lake Champlain and northwest of the area studied may reflect regional overlap (Raymond, 1906, p. 569). However, it probably offlaps the present Adirondacks, much as the lower formations do. The method of southeastward termination is indeterminate in the area of west-central Vermont, inasmuch as erosion has removed the rock to the east. South of West Rutland in the "marble belt" the Middlebury appears to be entirely absent, suggesting nondeposition southeast of west-central Vermont. The southwestward extent of the general overlap is difficult to determine, because the Chazy is covered in that direction by the Taconic Allochthone. It is similarly covered to the northeast in southeastern Quebec. In the Champlain Valley region the equivalents of the Middlebury limestone along the lake are about 100 feet thick. Eastward as far as the eastern limits of exposure the Middlebury becomes 600 feet thick. At certain localities southeast of the Adirondacks and in the "marble belt" the Middlebury, with the other Chazy formations, may have been completely removed by pre-middle Trenton, or earlier, erosion.

The sandy clastics of the Chazy, if not directly were probably originally from the west. The Weybridge, for example, thickens eastward, but it is rather likely that the clastics were derived from the Cambrian quartzites, the sands of which were probably from a western source. This was also suggested for certain sandstones in the upper Beekmantown. A near-by eastern source of clastics, probably for the most part within the foreland limestone-sandstone facies, seems to meet these requirements. Shale facies of Chazy age have not yet been recognized in the autochthonous succession. The Normanskill shales and slates and their equivalents, in the Taconic Allochthone, are correlated with the Chazy.

ORDOVICIAN-BLACK RIVER GROUP

General correlation and lithology.[2]—The Lowville limestone or middle Black River

[2] The original Black River limestone of northwestern New York (Vanuxem, 1842, p. 38–45) is subdivided into three formations (Kay, 1937, p. 243–247),—the Pamelia limestone (Wilson, 1932, p. 135–146), the Lowville limestone (Cushing *et al*; 1910, p. 79–84), and the Chaumont formation (Kay, 1929, p. 664, Table 7). Beds of the lower division of the overlying Trenton group in the Champlain Valley (Kay, 1937, p. 260-261) have been confused with the Chaumont, which resulted in their former classification as Black River.

of authors is reported at several localities (Emmons, 1842, p. 317–318; Hitchcock, 1861, p. 280; Dana, 1877a, p. 345; Brainerd and Seely, 1890b, p. 22; White, 1899, p. 454; Raymond, 1902, p. 21–30; Kay, 1937, p. 254, 261) in the Champlain Valley. The lower Black River Pamelia limestone is not reported here. The Chaumont or upper Black River is reported in the Champlain Valley only at Crown Point, New York (Kay, 1937, p. 254, 259).

Certain beds at Orwell, Shoreham, and Bridport have been classified as Black River (Dana, 1877a, p. 345, 413; Brainerd and Seely, 1890b, p. 4, 7; Raymond, 1906, p. 507) but are probably largely, if not wholly, lower Trenton. They comprise the Orwell limestone of the present paper. Fossils diagnostic of the Black River, other than possibly *Bathyurus extans* (Hall) (Dana, 1877a, p. 341) have not yet been reported from the Orwell north of Benson in west-central Vermont. *Tetradium cellulosum* and *Phytopsis tubulosum* are found in Benson township, adjacent to the town of Orwell on the south. Less than 10 feet of "ashen gray weathering" limestones or Lowville of authors, is near the base of the Orwell here. In eastern Shoreham and northern Cornwall townships black chert beds in the Orwell merely suggest the facies of the Chaumont. The only other outcrops of possible Black River reported in the folded belt are south of the Champlain Valley area in the lower Hudson Valley near Poughkeepsie, New York (Gordon, 1911, p. 65; Knopf, 1927, p. 439), where Clarke (1899, p. 9) reported *Tetradium cellulosum*.

Distribution and genetic relations.—The Black River of west-central Vermont possibly includes only the Lowville formation of authors, here less than 10 feet thick and probably absent in the easternmost outcrops. The Lowville thickens appreciably northwest of the belt of folding. The Lowville at these few localities, to the apparent exclusion of the Pamelia and Chaumont in the deformed southeastern belt, may probably be due to its overlap of the Pamelia about as far as the line of the present folded belt. This probable original southeastern limit of the Lowville is offlapped to some extent by the Chaumont.

ORDOVICIAN-TRENTON GROUP

General correlation[3].—The Trenton limestone of the Champlain valley, which as commonly described in the past (Emmons, 1842, p. 182–183, 277–278, 319) included only strata of Hull and lower Sherman Fall age, has been extended downward to include limestones of Rockland age (Kay, 1937, p. 260–261) and upward to include the shales of upper Sherman Fall age (Ruedemann, 1921a, p. 108–116) which, bordering the lake, are the highest strata exposed in the section. Trenton limestones were recognized in the area of deformed beds east of the Champlain thrust by Wing (Dana, 1877a, p. 340, 341, 342) and E. Billings (Hunt, 1868, p. 227), followed later by Brainerd and Seely (1890b, p. 4) and Keith (1932, p. 368). Evidence for the Trenton age of the eastern slates was originally presented by Wing (Dana, 1877a, p. 346) and Keith (1932, p. 368) indicated such a correlation.

[3] A complete section of the rocks which occupy the interval between the Black River and the Eden groups, in southeastern Ontario and northwestern New York, has been designated (Kay, 1937, p.250) as the standard section of the Trenton group. In it are included six stratigraphic units, in ascending order: Rockland (Raymond, 1914b, p. 348-349; Kay, 1937, p. 251), Hull (Raymond, 1914b, p. 348-349; Kay, 1937, p. 261), Sherman Fall (Kay, 1929, p. 664; 1937, p. 263-264), Cobourg (Raymond, 1914b, p. 345, 349; 1921, p. 1), Collingwood (Raymond, 1914b, p. 348-349; 1921, p. 1; Bassler, 1915, Pl. 2; Reudemann, 1925, p. 149; Kay, 1935, p. 585), and Gloucester (Raymond, 1916a, p. 255; Foerste, 1916, p. 62; Kay, 1935, p. 585).

Formations.—Three mappable units of Trenton age are recognized in the area.

The term Glens Falls, applied to beds of Hull and lower Sherman Fall age in the Mohawk Valley (Ruedemann, 1912, p. 22) and along Lake Champlain (1920, p. 92), is equally applicable in west-central Vermont, where faunal zones, facies, and limiting horizons are identical. Beneath the Glens Falls is a unit containing beds of Rockland age and possibly some of Lowville and Chaumont age called the Orwell limestone, after extensive outcrops in the southeastern part of that township. Keith (1932, p. 360, 369) has referred to the beds of Sherman Fall age, overlying the Glens Falls of the folded belt, as the Hortonville slate, after the village of Hortonville in the northwestern corner of Hubbardton township. Because the slates of the eastern area have yielded no fossils and inasmuch as their upper limits and relation to established stratigraphic units in the Champlain-Hudson Valley are uncertain, it is deemed advisable to call these the Hortonville slates.

Orwell Limestone (new name): Typically it is a massive, closely knit, heavy ledged, light-dove-gray-weathering, rather fine-textured black limestone cut through by innumerable white calcite veins (Pl. 5, fig. 1). It may stand out above associated rock types in gleaming, almost white ledges. *Stromatocerium rugosum* Hall, *Columnaria halli* Nicholson, and *Maclurites logani* (Salter) are fossils readily identified on the weathered surfaces of the slightly metamorphosed rocks of the westernmost outcrops. This fossil association differentiates the Orwell from *Maclurites magnus*-bearing Crown Point limestones of the Chazy group. *M. logani* alone is not readily differentiated from *M. magnus* unless it is weathered out across the plane of the coils, thus revealing its bowl-shaped section as compared with the conical section of *M. magnus*. In the absence of the dolimitization effect noted in the preservation of earlier forms, the more metamorphosed eastern rocks show few if any fossils. At a few localities the upturned edges of the beds reveal a distinctive banding parallel to the bedding, in which the bands, a few inches to 1 foot wide, are composed almost entirely of angular dark objects that are seen in less deformed strata to be chiefly fragments of brachiopod and gastropod shells with convex side up and weathered in section. Where this banding is absent the more abundant calcite veins may serve to differentiate the Orwell. Locally, particularly near the north end of the Taconic Range, the Orwell may be broken by numerous shear planes that destroy the normal, massive appearance so that it is difficult to distinguish from the nodular and incompetent Middlebury limestone beneath or from the overlying, less massively bedded Glens Falls limestone. Close examination, however, may reveal the faint-buff streaks of the Middlebury or the coarse texture characteristic of the Glens Falls. Discontinuous black chert beds are found in the Orwell at a few localities. Where present in the highly deformed rocks of eastern exposures, the chert serves as an unfailing indicator of this formation although all other characters may be obliterated. The Orwell most nearly matches the description of certain rocks reported by Perkins (1902, p. 161–164; 1904, p. 107–108) from the islands of northern Lake Champlain, which Emmons (1842, p. 110) originally termed "Black marble of Isle La Motte". Kay (1937, p. 260) included the Isle La Motte limestone in the Rockland.

Beds of Rockland age may be as much as 100 feet thick in the easternmost exposures in the Champlain Valley. The writer has measured 40 to 50 feet of the heavy ledged strata a little west of the Champlain overthrust in Ferrisburg township.

East of the thrust these strata are so often repeated by folding with thinning on the limbs or thickening on the axes of folds that its thickness is difficult to determine.

Logan (1863, p. 280) reported as much as 80 to 100 feet of the heavy ledged strata in the Highgate Springs slice. This zone apparently thins westward in the northern Champlain Valley. White (1899, p. 458) reported 35 feet on Grand Isle, in northern Lake Champlain west of the exposures of the Highgate Springs slice. At Crown Point, west of west-central Vermont, that part of the heavy ledged facies above the Lowville is from 59 (Kay, 1937, p. 254) to 66 (Raymond, 1902, p. 21–24) feet thick. Therefore, even though the possibility that certain of the lower heavy ledged beds in both eastern and western localities may be older than Rockland is taken into consideration, it still remains probable that Rockland correlatives thin considerably toward the west, at least in the northern Champlain Valley. In the southern part of the Champlain Valley, in localities adjacent to the Adirondack crystalline rocks, the Rockland beds pinch out locally. This may be due to later removal, inasmuch as at certain localities where they pinch out upper Sherman Fall clastics lie above the break, which is therefore comparable in position to erosional unconformities noted elsewhere in the vicinity of the Adirondacks (Kay, 1937, p. 264, 275-276).

Rockland equivalents south of west-central Vermont in the "marble belt" are difficult to locate inasmuch as both the heavy ledged facies and the faunal content have been obscured by deformation. The Rockland may be present in the "marble belt," but the fossils thus far reported from this belt (Walcott, 1888, p. 236–240; Wolff, 1891, p. 336; Dale, 1892a, p. 517; 1894, p. 535, 543, 544; Foerste, 1893, p. 441-442; Bain, 1931, p. 516) have a much greater range than those restricted to the established Rockland beds and seem to be particularly characteristic of Trenton faunas higher than the Rockland.

GLENS FALLS LIMESTONE (Ruedemann, 1912): A thin-bedded, dark blue-gray, rather coarsely granular limestone (Pl. 5, fig. 2) is readily recognized at many localities by its position between the Orwell limestone facies and the Hortonville slates. Prominent ledges of the massive limestone and rounded hills of sod-covered slate indicate the general limits of the less resistant Glens Falls, which at most localities underlies low, covered areas. The lower contact may be somewhat indefinite, inasmuch as the lower beds are lighter gray and more massive than those higher up, thus looking like the Orwell. Where the beds have not been extremely deformed the Glens Falls may be denoted by abundant ramose bryozoans, probably including *Pachydictya acuta* (Hall). The distinctive trilobite, *Cryptolithus tesselatus* Green, may be found after fairly diligent search even in the deformed Glens Falls beds on the east limb of the Middlebury synclinorium. The hemispherical bryozoan *Prasopora orientalis* Ulrich occurs in the same zone as *Cryptolithus tesselatus*, immediately beneath the slates, and is useful in identifying the Glens Falls, particularly at localities of less deformed beds to the west near the Champlain thrust.

Ruedemann (1921b, p. 90–94) has pointed out two distinct facies in the Glens Falls along Lake Champlain: a gray crystalline limestone beneath and a bluish-black, partly shaly limestone above. *Parastrophina hemiplicata* is common in the gray limestone (White, 1900, p. 459) and is also abundant in the Hull of southeastern Ontario and northwestern New York. *Cryptolithus tesselatus* is found in the overlying shaly limestone (White, 1899, p. 460). Kay (1937) called these divisions of the Glens Falls limestone in the Champlain Valley the Larrabee (p. 262) and Shoreham (p. 264), respectively. These members cannot be readily distinguished in the strongly folded beds east of the Champlain thrust, but probably beds of Hull and earliest Sherman Fall age are present in that region.

The Glens Falls facies is estimated to be about 115 feet thick in eastern Shoreham township. The beds are sheared and possibly repeated to some extent on minor folds, so that exact thickness cannot be determined. Determination of the thickness is further complicated by the similarity in appearance of the lowermost beds and those of the underlying Orwell.

A comparable thickness, 120 feet, is reported to the north at Highgate Springs (Kay, 1937, p. 254, 263, 266). The Glens Falls (Kay, p. 254) is 102 feet thick on Grand Isle, 85 feet thick at Crown Point, 52 feet thick at Larrabee Point, and 57 feet thick at Glens Falls, and thus apparently thins southwestward in the Champlain Valley. At one locality in southwestern Orwell township it is entirely absent, Hortonville slates directly overlying the Beekmantown. Beds of Hull age, forming the Larrabee member, probably represent about half the thickness of the Glens Falls in the overthrust rocks of west-central Vermont, or about 50 to 60 feet. The Larrabee is 72 feet thick at Grand Isle, 35 feet thick at Crown Point, 19 feet thick at Larrabees Point, and 19 feet thick at Glens Falls (Kay, 1937, p. 254). It is locally absent from the region of the present Adirondacks (p. 254) apparently because of pre-upper Sherman Fall removal. Northwest of the Adirondack axis the Hull thickens to as much as 110 feet (p. 261). The thicknesses and the stratigraphic relations of the Shoreham member are discussed in connection with the Hortonville slate, with which it probably forms a widespread stratigraphic unit.

HORTONVILLE SLATE (Keith, 1932): This red-brown weathering, locally quartzitic, blue-black slate is the youngest Paleozoic formation that is exposed in west-central Vermont. Almost exclusively the slate outcrops are found in vertical cliffs and ravines. Elsewhere, particularly in open pastureland, it may be covered by a thin layer of turf but is suggested by the characteristic rather steep, but rounded and grassed over, hills. Certain more resistant, quartzose facies crop out in rather prominent hills, forming glacially polished and striated ledges. In general the bedding is obscured by the slaty cleavage. Portions of the slate, particularly near the Taconic Range, are sufficiently altered to be called phyllite.

The Hortonville slate is correlated with the Canajoharie shale of the Mohawk Valley and Lake Champlain, which lies stratigraphically above and grades up from the Glens Falls limestone (Ruedemann, 1912, p. 21–22; 1921b, p. 92–93). Keith (1932, p. 360, 369) has indicated a correlation between the Hortonville and the Snake Hill formation of eastern New York, a formation that in the opinion of Kay (1937, p. 272) "represents an eastern, more sandy facies of the Canajoharie" as exhibited in the latitude of the Mohawk Valley. This facies crops out almost continuously west of the Taconic Range from the type locality of the Snake Hill in eastern New York to the type locality of the Hortonville in west-central Vermont. Fossils have been reported only from the limestone interbeds in the base of the Hortonville (Dana, 1877a, p. 346), but a considerable fauna of lower Canajoharie age (Kay, 1937, p. 272) is found in the Snake Hill. Thus, a correlation between the Hortonville and the lower Canajoharie (Ruedemann, 1921a, p. 108–110) Cumberland Head formation along Lake Champlain (Kay, 1937, p. 274–275) is suggested. The Hortonville may also include upper Canajoharie strata (Ruedemann, 1921a, p. 110–112), equivalent to the Stony Point formation along Lake Champlain (Kay, 1937, p. 275).

Wing (Dana, 1877a, p. 343) estimated the Hortonville slate to be 300 to 400 feet thick in eastern Shoreham township. The author has checked this figure in Weybridge township near the north end of the slate and, by halving the thickness of the slate in the isoclinal syncline here estimated a thickness of about 400 feet. Inasmuch as the upper contact of the slate has been eroded away, the figure probably represents only the lower tenth of slates originally 4000 to 5000 feet thick comparable to the Canajoharie reported from the lower Mohawk Valley (Ruedemann, 1930, p. 33, 118; Kay, 1937, p. 272, 273). An interbedded zone,—in most places less than 5 feet thick—forms the gradation between the Glens Falls and the Hortonville east of the Champlain thrust in west-central Vermont. Along Lake Champlain this interbedded zone, correlated roughly with the lower Canajoharie, is 50 to 150 feet thick (Ruedemann, 1921a, p. 109; Kay, 1937, p. 275), thickening northward.

Possibly the abrupt thickening of the transition zone west of the Champlain overthrust may be ascribed to the horizontal shortening by thrusting of the original zone of lateral gradation from shales on the east to limestones on the west. Apparently the shales gradationally overlap westward on the transition zone. Probably the northward gradational overlap of upper Canajoharie shales on lower Canajoharie transition beds along the lake (Kay, 1937, p. 277) is merely a northward component of the true northwestward overlap, if the northeastward trend from the Mohawk to the Champlain Valley of the eastern limits of the transition beds (p. 275) parallels a southeastern source of the shales. The westward component of this overlap, at right angles to the Champlain Valley, is found in the Mohawk Valley (p. 271). The Canajorharie clastics thus appear to be a southeastward-thickening deposit, probably with only the lower most beds represented in west-central Vermont.

The original southeastward extent of the Canajoharie rocks is indeterminate inasmuch as the Canajoharie is eroded off and faulted in various places and hidden at southeastern localities by the tectonically higher structures of the Taconic Allochthone. Separation of the autochthonous rocks from the Allochthone has been found difficult, and thus actual post-deformation thickness of the Canajoharie clastics is hard to determine. In many places the eastern Canajoharie facies simulate some of the pseudoconformably overlying Taconic rocks. In the "marble belt" south of the area, a blue limestone about 30 feet thick (Foerste, 1893, p. 441) lies at the top of the calcareous succession that forms the southern extension of the east limb of the Middlebury synclinorium and immediately beneath the argillaceous rocks found principally in the Taconic Range to the west. According to Foerste, this limestone contains a typical Trenton fauna and inasmuch as it is dark blue, everywhere lies beneath a phyllite, and is of about the average thickness of the early Sherman Fall limestone in the Champlain and Mohawk valleys, the presence of the Shoreham member of the Sherman Fall formation is suggested. Possibly then the lower portion of the overlying phyllites is autochthonous with respect to the Taconic Allochthone and therefore of Canajoharie age. A similar relation of Trenton blue limestone at the top of the calcareous succession to overlying phyllites has been reported at other localities adjacent to and within the Taconic Range. Wing (Dana, 1877a, p. 340) reported a patch of *Cryptolithus tesselatus*-bearing limestone from the heart of the "great central belt of slates" in Hubbardton township, Vermont. Near the southwestern corner of Vermont Walcott (1888, p. 237-238) recognized Trenton fossils in limestones that grade up into the overlying phyllites by interlamination. Bain (1934, p. 126) has suggested the presence of phyllites on blue Trenton limestone at West Rutland, Vermont.

Blue limestone overlies Cambrian rocks in Clarendon township. East of West Rutland, in the Center Rutland valley, phyllites, which farther south along the same strike in Clarendon township (Wolff, 1891, p. 336) are underlain by the blue limestone, crop out near although not in actual contact with Beekmantown marbles. In southwestern Brandon township, at the northeast corner of the Taconic Range, farther north along this same general contact, phyllites lie on Beekmantown limestone. Westward across the north end of the range they lie on successively younger limestone beds and at the meridian of Hyde Manor phyllite known to be the Canajoharie equivalent is in the normal position above the *Cryptolithus tesselatus*-bearing limestone. Keith (1913, p. 680) recognized the discordance from the areal pattern of the lithologic units and located the Taconic thrust along the contact between the limestone and overlying phyllites, including the phyllites in the Taconic Allochthone.

Apparently two interpretations of the discordant relation between the eastern phyllites and the underlying calcareous succession at the north end of the Taconic Range are possible. Keith's thrust interpretation will be discussed later. Canajoharie rocks might overlap eastward unconformably on the Trenton limestones and in their eastern exposures lie upon the eroded edges of much older beds. Possibly this interpretation applies to the localities south along the "marble belt", whereas Keith's applies at the north end of the Taconic Range. At the north end of the range red and purple, and olive-green slates typical of the allochthonous facies in rather close association with the black phyllites near the limestone contact favor the thrust hypothesis. However, the author has not observed nor does the literature report such a discordance between the limestone and the phyllite at any other locality in or bordering the Taconic Allochthone. Several authors (Agar, 1932, p. 36–38; Prindle and Knopf, 1932, p. 297; Knopf, 1935, p. 208-209; Balk, 1936, p. 765–767) indicated the lack of such a discordance at various places in and bordering the central and southern Taconics. This would suggest that most of the Taconic Allochthone pseudoconformably overlies Canajoharie equivalents, making it somewhat doubtful that the discordance noted at the north end of the Taconic Range, although at a point where a thrust fault might be expected, is other than an unconformable overlap of the Canajoharie rocks on older truncated strata.

A stratigraphic break beneath the Canajoharie is well established at several localities northwest of the Taconic Range. The Canajoharie lies on Beekmantown within a small area at the Orwell-Benson line near Lake Champlain and adjacent to the Adirondack border. Similar breaks have been noted near or at the base of the Canajoharie or its equivalents at several other localities at (Clark and McGerrigle, 1936, p. 672-673; Kay, 1937, p. 264, 275-276) or east (Ruedemann, 1901, p. 546–549, 1930, p. 104–113; Kay, 1937, p. 276-277) of the meridian of the Adirondack Mountains. At the Adirondacks the Canajoharie shale gradationally overlaps northwestward upon the Denmark limestone member of the Sherman Fall formation—the non-clastic equivalent of the Canajoharie in northwestern New York (Kay, 1937, p. 267–268, Pl. 4). This break may be present also to the east of the Green Mountains (Currier and Jahns, 1941, p. 1510).

Distribution and genetic relations.—Beds of Rockland age, nowhere over 100 feet thick, overlap the Black River group in the Champlain Valley and elsewhere. They are absent in the present Adirondack region probably in part because of pre-middle Trenton removal. The Rockland beds may thicken southeastward in the folded belt, but they are apparently absent to the extreme southeast, possibly to some extent as a result of pre-middle Trenton erosion. They are not reported at any locality south of west-central Vermont in the "marble belt". Their northeastern extent in the folded belt is obscured by the overlying Taconic Allochthone, but they are probably present northeastward, much as in west-central Vermont (Fig. 5).

The strata of Hull age are distributed about as the Rockland which they overlap at certain localities bordering the Adirondacks (Kay, 1937, p. 254); like the Rockland they are absent because of pre-late Sherman Fall erosion at the meridian of the Adirondacks. The Hull is about 50 feet thick southeast of the Adirondacks in the folded belt. The southeastern limits of the Hull are uncertain, inasmuch as no diagnostic fossils have been found in the metamorphosed beds of the "marble belt". The Taconic Allochthone covers the northeastward extent of the Hull in the folded belt.

The Sherman Fall strata are more widely distributed than the Rockland and Hull, because in some localities on and southeast of the Adirondacks, particularly to the extreme southeast near the Green Mountains, early Sherman Fall erosion removed pre-Sherman Fall rocks. In west-central Vermont Sherman Fall strata may have been as much as 5000 feet thick before the erosion that produced the present topography. At northern Lake Champlain the Sherman Fall is approximately 500 feet thick (Kay, 1937, p. 274–275)—comparable to the thickness at Canajoharie (p. 269–270) on the Mohawk. Northwest of a line running through these points the Sherman Fall thins to about 200 feet. The Shoreham limestone, a member of the Sherman Fall, less than 60 feet thick (p. 254), lies beneath the great thickness of upper Sherman Fall-Canajoharie clastics to the southeast of this line. Locally the Shoreham may be thinner or absent entirely, either because of early Sherman Fall convergence or removal or both. From the northwestern margin of the folded belt northwest to the Adirondacks, the lower Canajoharie clastics in turn take on an increasing number of limestone interbeds. At the Adirondacks the Sherman Fall clastics overlap gradationally northwestward on nonclastic equivalents. The source of all clastics is to the southeast.

The Schodack, Deepkill, and Normanskill shales of the Taconic Allochthone, stratigraphic equivalents of the pre-Sherman Fall calcareous rocks, were, previous to movement, southeast of the foreland calcareous facies (Fig. 6). The Sherman Fall clastics were probably derived from these shale facies while they were still in their original position. In the pre-Sherman Fall beds of west-central Vermont the zone of lateral facies change from geosynclinal shales on the southeast to foreland limestones on the northwest happens to be at least as far east as the Green Mountain anticlinorium, embryonic forerunners of which may have acted as a barrier to the older muds. In northwestern Vermont this zone is far to the northwest of the anticlinorium and is clearly gradational; thus a continuous barrier barring the northwestward spread of earlier Cambro-Ordovician clastics seems unlikely. Instead a discontinuous barrier must have been produced by the preliminary uplifts. The final Sherman Fall uplift was southeast of the barrier and evidently was higher,—apparently a greater source of clastics than the barrier may have previously been.

SUCCEEDING GROUPS

The Eden, Maysville, and Richmond groups, comprising the Upper Ordovician series of authors, crop out north and south of the Champlain Valley, in southern Quebec, and along the lower Mohawk Valley in New York, respectively. In the latter region only the Eden (Ruedemann, 1930, p. 38) is represented by the shaly strata of the Indian Ladder beds (1925, p. 25). In southern Quebec Eden (Foerste, 1924, p. 9–11), possibly Maysville (p. 9–11), and Richmond shales are overlain by about 1000 feet of red beds (Parks, 1931, p. 27) constituting the Queenston facies (Foerste, 1924, p. 56–57). The Queenston overlaps gradationally northwestward with respect to a southeastern source on the lower Richmond deposits (Foerste, 1924, p. 56–57). Along the middle course of the Nicolet River about 75 miles north of the Canadian border, the Queenston apparently is bordered to the southeast by the Taconic Allochthone as well as by southeastward-dipping low angle thrusts, along which the red beds might be faulted off. Thus, the Upper Ordovician series is thrust faulted out of west-central Vermont, if originally deposited and not eroded. Presumably the upper Sherman Fall clastics, which are the youngest beds now exposed in west-central Vermont, may, previous to thrust faulting or erosion, have been disconformably overlain by clastics of Eden age or later.

STRUCTURAL GEOLOGY

MAJOR STRUCTURES

General setting.—The major structural features in west-central Vermont trend north-south, or parallel to Lake Champlain. They form part of the southwest limb of one of the great northwestward jutting salients of the north-northeast-trending folded mountain belt that parallels the Atlantic Coast of North America. The axis of the salient crosses the major structures a little north of the area of west-central Vermont (Fig. 5). Here the major structural trends change from northerly to northeasterly and thus parallel the St. Lawrence River (Keith, 1923b, p. 309, Pl. 4; 1932, p. 363–364). West-central Vermont is at the western edge of the deformed belt. Four types of major structures represented in or immediately adjacent to this area are: synclinoria, anticlinoria, thrust faults, and a normal fault system.

Synclinoria.—The two synclinoria (Fig. 5) lie on a common north-south axis (Pl. 10) and are separated from each other by the Monkton cross anticline, from which they pitch in opposite directions. They are bounded on the west by the Adirondack "dome" and the Champlain thrust and on the east by the Green Mountain anticlinorium and the Hinesburg-Oak Hill thrust.

MIDDLEBURY SYNCLINORIUM: This synclinorium plunges southward from the latitude of Monkton and embraces the structure of the area between Snake Mountain on its west limb (Pl. 1, fig. 1) and the Green Mountain Front on its east limb (Pl. 1, fig. 2) (Dana, 1877a, p. 414). The center of this synclinorium is covered by the Taconic Allochthone south of the latitude of Brandon. The east limb of the synclinorium may be traced more or less continuously into the "marble belt" south of this latitude. The west limb loses its identity in an area of high angle faults southwest of Orwell. Numerous minor folds that tend to reflect the shape and orientation of the major structure are found on both limbs of the synclinorium. Lines marking the intersections of the axial planes of these minor folds with the surface converge southward. On the east limb of the synclinorium the east limbs of the minor synclines dip steeply (Pl. 6, fig. 2) and are commonly overturned. Overturning of beds, except possibly in structures adjacent to thrust planes, is not as common on the west limb of the synclinorium.[4]

HINESBURG SYNCLINORIUM: This synclinorium (Bain, 1931, p. 506–507) plunges northward from the latitude of Monkton (Pl. 10) and embraces the structures of most of the area between Lake Champlain and the Green Mountain Front. The axis rises to the north in Colchester. Between Colchester and St. Albans there appears to be another cross anticline comparable with that at the latitude of Monkton, but somewhat less prominent. Most of the east limb of the Hinesburg synclinorium, except at the deep re-entrant cut by the Winooski River, is covered by the Hinesburg-Oak Hill thrust slices. The Hinesburg synclinorium is not so symmetrical as the Middlebury synclinorium, and the "folding is limited to the development of a series of moderately broad basin structures" (Bain, 1931, p. 506). The orientation of minor structures is obscured at the north end of the synclinorium

[4] Thus in both plan and cross section the convergence of axial planes of minor folds toward the younger beds of the synclinorium is evidenced. This is therefore an inclined normal synclinorium (Van Hise, 1896, p. 610), tipped over toward the west.

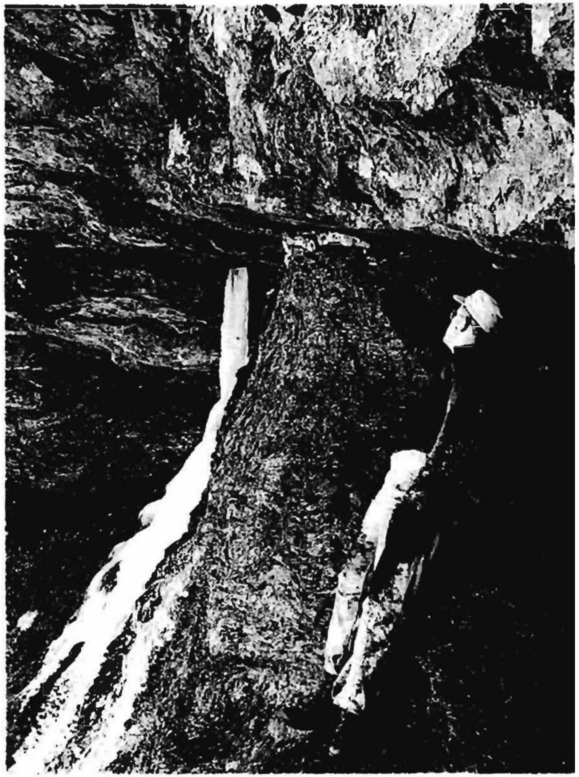


FIGURE 1. CHAMPLAIN THRUST
Beneath west limb of Middlebury synclinorium at north end of Snake Mountain, 1¾ miles ESE. of Addison village. Gently dipping Monkton formation resting above Glens Falls limestone.

FIGURE 2. EAST LIMB OF MIDDLEBURY SYNCLINORIUM
Steeply dipping Cheshire quartzite at south end of the Monkton Hills in northwestern Bristol township, 2⅛ miles NNE. of New Haven village.

MAJOR STRUCTURES

FIGURE 1. CROWN POINT LIMESTONE
Tops of beds are to east. Fracture cleavage dips more steeply east than bedding. Zone of later rupture cuts bedding and cleavage. 1¼ miles SSW. of Orwell village.

FIGURE 2. BELDENS FORMATION
Tops of beds are to west. Eastward-dipping fracture cleavage cuts beds dipping steeply west. 1 mile WNW. of Brooksville, New Haven township.

CLEAVAGE-BEDDING RELATIONS

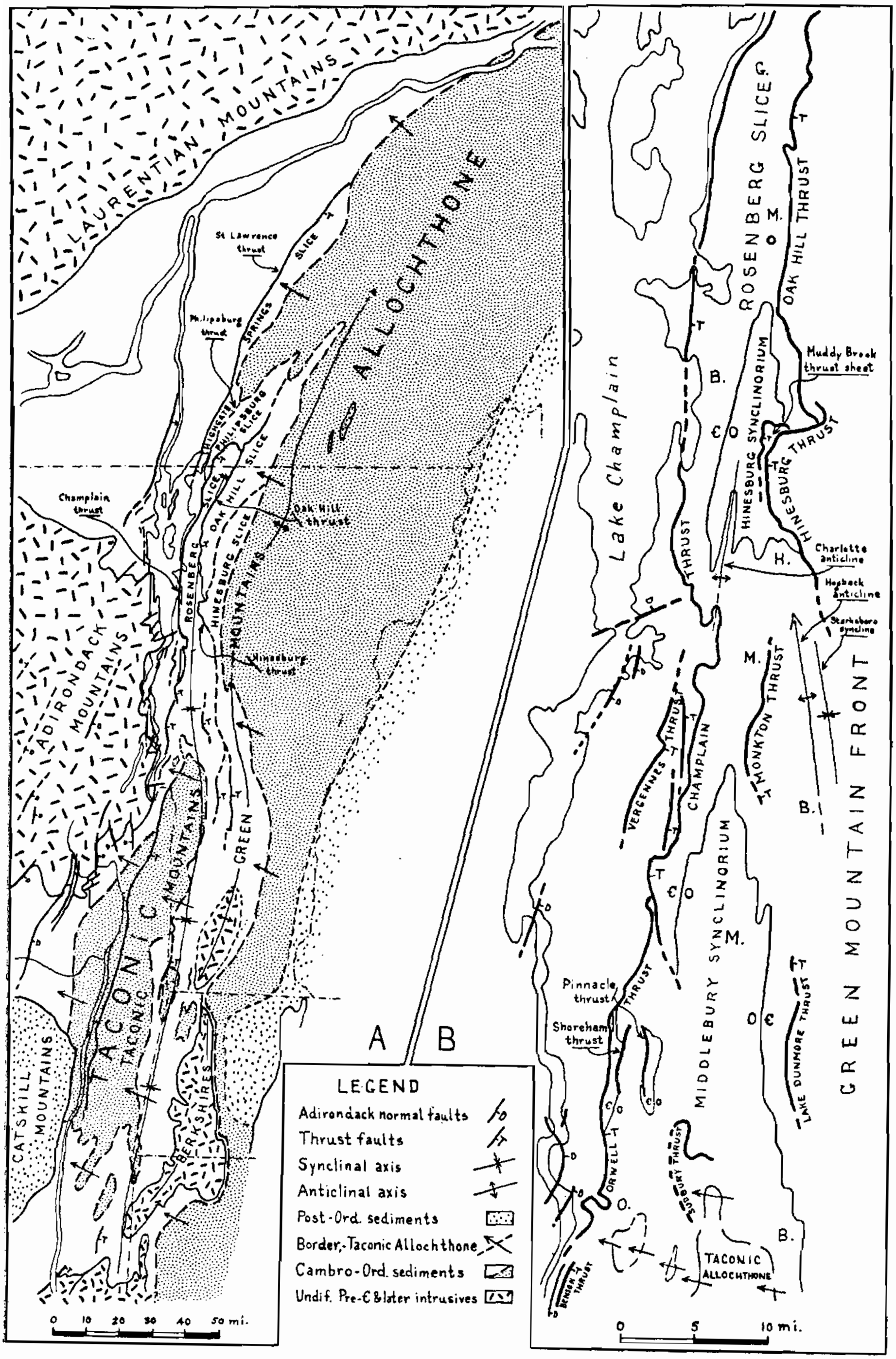


FIGURE 5.—*Regional geologic map*

(A) Western New England, eastern New York, and southern Quebec; (B) Detail of west-central Vermont by drift cover. At the northern extension of the west limb of this synclinorium, in the St. Albans area, minor folds—particularly in the argillaceous beds—are strongly overturned toward the west. The one decidedly anomalous feature in the major structure is the strong westward salient found in the strike of fold axes west of the Brownell Mountain area. The pattern of the Hinesburg thrust at Brownell Mountain suggests that this salient is the effect of drag on the tectonic zone immediately beneath the thrust slice.

Anticlinoria.—Rather pronounced structural highs are found east and west of the great synclinoria. The Adirondack "dome", a massif whose location during Paleozoic times was approximately the same as it is today, was apparently rather passive during orogenic movements.[1] The lesser folds increase in size eastward (Keith, 1923b, p. 314) from the "dome" and east of the synclinoria they form an anticlinorium.

GREEN MOUNTAIN ANTICLINORIUM: The high curvilinear ridge of the Green Mountains (Fig. 5) roughly coincides with the axis of an uplift of crystalline schists and gneisses (Adams, 1846, p. 167–168; Keith, 1932, p. 404–405) that crosses both the northern and southern borders of the state and is rather continuous for several hundred miles both to the northeast in Canada (Selwyn, 1879, p. 1–15) and to the south along the east side of the Appalachian folded belt (Keith, 1923b, p. 318). Tectonic and stratigraphic evidence suggests that the Green Mountain anticlinorium pitches northward. Whittle noted the northerly pitch exhibited by minor folds and by the areal pattern of the major structure (1894b, p. 352). Drag folds, cleavage-bedding relations, cross-bedding, and ripple marks in the northern part of the range (Currier and Jahns, 1941, p. 1507) indicate northward pitch.

Strictly speaking, the Green Mountain anticlinorium is not continuous far beyond the northern and southern borders of the State. The uplift that follows the ridge of the Green Mountains in Vermont pitches abruptly southward and disappears a little south of the State line at Williamstown, Massachusetts. North of Vermont, this uplift follows the Sutton Mountains north-northeastward and disappears about 50 miles north of the Canadian border and west of the Thetford area in Quebec. At both of these localities the ridge crest passes to another tectonic axis that lies southeast of the Green Mountain-Sutton Mountain axis. The Green Mountain-Sutton Mountain axis bulges westward markedly between the latitude of Rutland and the Canadian border. In its east flank the Green Mountain anticlinorium contains a discontinuous belt of ultrabasic intrusives (Clark, 1934, p. 12; Bain, 1936), associated with volcanics including pillow basalt, which follows all the sinuosities of the range. Southwest of Thetford in Quebec, this belt crosses over the southeastern tectonic axis which appears to pitch southwestward here (Cooke, 1937, Map 419A in pocket).

Thrust faults.—The planes of the major thrusts lie subparallel to either overlying or underlying strata or both for considerable distances. Competent strata in the overthrust blocks dip gently eastward parallel to the thrust. Incompetent strata rather commonly found beneath the thrusts are generally more disturbed. Where the rocks have been folded considerably before thrusting, the bedding may dip at an angle to the thrust plane. Commonly, the overthrust mass has moved so far to the west and northwest as to be in a region of little folding and the beds beneath the thrust trace approach parallelism with the thrust plane. Beds originally above the strata that now underlie the thrust trace have evidently been stripped off at a break parallel to the bedding. Thus the major thrusts are genetically (Billings, M. P., 1933, p. 142–143) strip thrusts.

CHAMPLAIN THRUST: This thrust (Keith, 1923a, p. 104) defines the western boundary of a continuous tectonic unit, bounded on the east and west by eastward-dipping thrust planes. T. H. Clark (1934, p. 4, Fig. 2; p. 8), following Keith's conception of the thrust sequences, recognized the slice character of the north end of this unit calling it the "Rosenberg slice" (Fig. 5). This slice includes most of the Hinesburg synclinorium and possibly the Middlebury synclinorium. The Champlain thrust trends parallel to Lake Champlain extending from a southward-pitching anticline in Cornwall township a little south of Snake Mountain northward to the Canadian border, about 3 miles north of which, near the village of Rosenberg, Quebec, it becomes obscure in a shale terrane (Clark, 1934, p. 7, 8, Fig. 3). At Snake Mountain Lower Cambrian beds of the mountain proper are thrust westward across and beyond Upper Cambrian and Beekmantown rocks of the Orwell thrust plate onto the middle Trenton limestones and shales next west and structurally continuous with those found along the lake (Pl. 6, fig. 1); the Champlain thrust apparently truncates the Orwell thrust (Pl. 10). To the north the fault lies beneath Lower Cambrian beds to a point at least as far north as the Canadian border. Between the border and the vicinity of Rosenberg it may cut higher into Upper Cambrian beds (Clark, 1934, p. 7, Fig. 7) and possibly Beekmantown or later equivalents.

The dip of the thrust plane varies from a few degrees west to about 20° east. The massive Lower Cambrian formations dip parallel or subparallel to the thrust plane. Except for a small patch of Dunham dolomite that crops out on the west slopes of the south end of Buck Mountain ridge near the Otter Creek (Pl. 10), the Monkton quartzite lies immediately above the thrust plane from Snake Mountain north to Shelburne Bay. From Shelburne Bay to the Canadian border the Dunham dolomite lies above the thrust plane.

Structures in the more massive limestones and dolomites beneath the thrust are truncated at the thrust plane in the region between the south end of Snake Mountain and Mt. Philo, and from Burlington north. At intervening localities the fault plane lies in and is roughly parallel to subhorizontal, thin-bedded, middle Trenton limestones and shales that have been correlated with the Glens Falls limestone and Hortonville slate. Between Snake Mountain and Mt. Philo, beds ranging from the upper Beekmantown Bridport dolomite to the Trenton shales are truncated by the thrust where it cuts across several north northeast—south southwest-striking minor folds and thrusts. The thrust cuts Chazy and Trenton beds at several localities north of Burlington in the belt of tightly folded limestones and shales that comprise the Highgate Springs sequence (Schuchert, 1933, p. 357–358; 1937, p. 1016–1018) found between the fault and the lake. At the Canadian border the Champlain thrust cuts Upper Cambrian to probably Trenton or younger beds in the Philipsburg slice (Logan, 1863, p. 283–285; McGerrigle, 1930, p. 184–186; Clark, 1934, p. 7, Fig. 3).

Since the Champlain thrust disappears southward, the displacement along the southern portion of the thrust must increase northward. The deep erosional re-entrants north of Snake Mountain where the Otter Creek crosses the fault and along Lewis Creek between Shellhouse Mountain and Mt. Philo indicate to some extent the magnitude of this increase. The north-northeast—south-southwest regularity of strike of beds and absence of tear structures in the massive Lower Cambrian strata of the Rosenberg slice suggest that the slice as a unit has not been appreciably distorted. The northern terminus of the Champlain thrust is masked in an extensive shale terrane, found north of Rosenberg, Quebec, that includes strata equivalent to the competent strata of the Rosenberg slice. Possibly the thrust is distributed along several planes in the shales north of Rosenberg. This zone of thrusting probably terminates far to the northeast of the Rosenberg slice proper.

HINESBURG-OAK HILL THRUSTS: The Hinesburg (Keith, 1932, p. 108, p. 364) and Oak Hill (Clark, 1934, p. 4, Fig. 2) thrusts form the western boundaries of tectonic units whose eastern limits have not been traced (Booth, 1938, p. 1869). The rocks of both the Hinesburg and Oak Hill thrust slices grade eastward into the schist and gneiss terrane of the Green Mountains. Both of these slices, so far as they have been delineated, apparently have undergone considerable displacement, as evidenced by the depth of erosional reentrants and by the outlying position of klippes. The Hinesburg and Oak Hill thrusts form the eastern boundary of the Rosenberg slice.

The terms Hinesburg and Oak Hill have been used rather interchangeably. There is, however, a major bifurcation of this thrust zone about at the latitude of the Winooski River, north of the type locality of the Hinesburg thrust. This would justify designating the part north of the Winooski River as the Oak Hill thrust extending it south across the border from its type locality in southern Quebec (Booth, 1938, p. 1869). The Oak Hill thrust passes southeastward beneath the Hinesburg thrust at Williston village (Pl. 10). The Hinesburg thrust, however, is not readily traceable northeastward into the highly deformed eastern rocks north of the Winooski River, although further investigation in that region may reveal its position there. The Oak Hill thrust passes northward into a region near Brome Mountain in southern Quebec, where from all available accounts (Logan, 1863, p. 245; Ells, 1896, p. 37, 57), the surface is underlain largely by argillaceous rocks in whose monotonously similar lithology a fault could not be readily traced. The Hinesburg thrust is traceable southward to a point 3 or 4 miles southeast of Hinesburg village.

Southward the position and extent of the Hinesburg thrust are entirely conjectural. Half a mile south of Starksboro village (Pl. 10), a slaty to phyllitic, or graywacke succession, comparable with that east of the Hinesburg thrust farther north, lies east of and probably stratigraphically beneath Cheshire quartzites that are in structural continuity with the Rosenberg slice. East of the Chesire quartzite along the Green Mountain Front and west of the main ridge of the Green Mountains farther south in Lincoln township is a dolomite very similar and probably equivalent to dolomites (Clark, 1936b, p. 143; Booth, 1938, p. 1869) occurring beneath the Cheshire correlative in the Oak Hill slice. Reconnaissance strongly suggests that no locally identifiable truncation by thrusting occurs in the succession below the Cheshire. Possibly, however, all thrust planes are folded into pseudoconformability in the eastern area and are thus not recognizable here.

The rather highly deformed quartzose slates, phyllites, and graywackes east of the Hinesburg thrust, a short distance north and east of Hinesburg village, lie with angular discordance across the east limb of the Hinesburg synclinorium, where the thrust plane truncates minor folds which are made up of beds from Lower Cambrian to Beekmantown age. The thrust plane has not been observed at any point, but the depth of the re-entrants suggests that it dips at a very low angle to the east. Non-quartzose black slates and phyllites crop out west of the quartzose rocks along the thrust front in St. George and Willston townships. These latter Upper Cambrian argillaceous rocks comprise the Muddy Brook thrust slice, which was apparently dragged up along the sole of the Hinesburg thrust. These same slates and Upper Cambrian sandy dolomites crop out in the re-entrant west of Williston village. Northwest of Williston village the quartzose rocks are thrust over a closely folded syncline of the Oak Hill slice. In this syncline are formations from Lower Cambrian to probably Upper Cambrian age (Booth, 1938, p. 1869).

In general, the rocks east of the Oak Hill thrust are less deformed and less uniform in appearance than those east of the Hinesburg thrust. The Lower Cambrian Dunham dolomite is everywhere recognizable, and at many places along the thrust front, where structures involving the Dunham are truncated at erosional re-entrants or at klippes such as Cobble Hill in Milton township (Schuchert, 1937, p. 1058), it locates the fault. Where argillaceous rocks are near the contact, the fault is much more difficult to locate, inasmuch as the eastern exposures of the Rosenberg slice are in a predominantly argillaceous terrane. The slate accompanied by conglomerate is indistinguishable from similar slate and conglomerate in the Oak Hill slice (Booth, 1938, p. 1869). The slate is exposed south "to the latitude of Burlington, where it is cut off by the Hinesburg overthrust" (Keith, 1932, p. 377; Schuchert, 1937, p. 1048). North of Georgia, Vermont, these slates "crop out nearly continuously to a point beyond the Canadian border" (Schuchert, 1937, p. 1052). The thrust plane, although unobserved, probably dips eastward at a low angle. The Oak Hill slice is distinctly different from the Rosenberg slice in that the younger beds are to the west and the strata dip steeply into the thrust plane instead of paralleling it (Clark, 1936b, p. 139, 141).

Taconic Thrust: That portion of the Taconic thrust trace here described defines the northern boundary of a klippe rather extensive in western New England and eastern New York; it includes most of the peaks of the Taconic Range and considerable area to the west. Clarification of the contact relations of this klippe may make conclusions as to its genesis possible. The gross areal relations suggest an enormous strip thrust.

In and west of the Taconic Range is a predominantly argillaceous Cambro-Ordovician succession occupying about the same stratigraphic position (Walcott, 1888, p. 241-242; Ruedemann, 1902, p. 559; 1921, p. 117-118) as calcareous beds cropping out at lower elevations to the east, west, and north. Small patches of the calcareous facies within the range (Dana, 1877a, p. 340), completely surrounded by the argillaceous facies, are exposed along the axes of anticlines. Slates and phyllites lie above limestones at several points in and adjacent to the north end of the Taconic Range (Dale, 1904; Keith, 1932, p. 399-402). Allochthonous Cambrian slates and phyllites are here superposed on the Ordovician rocks of the predominantly calcareous succession. Near the north end of the range the Taconic thrust klippe has roughly the same areal configuration as the range itself (Ruedemann, 1909, p. 188–192; Keith, 1912, p. 720-721, 1913, p. 680; 1932, p. 359, 399). A similar condition of abnormal superposition of the allochthonous facies exists at the south end of the Taconic Range

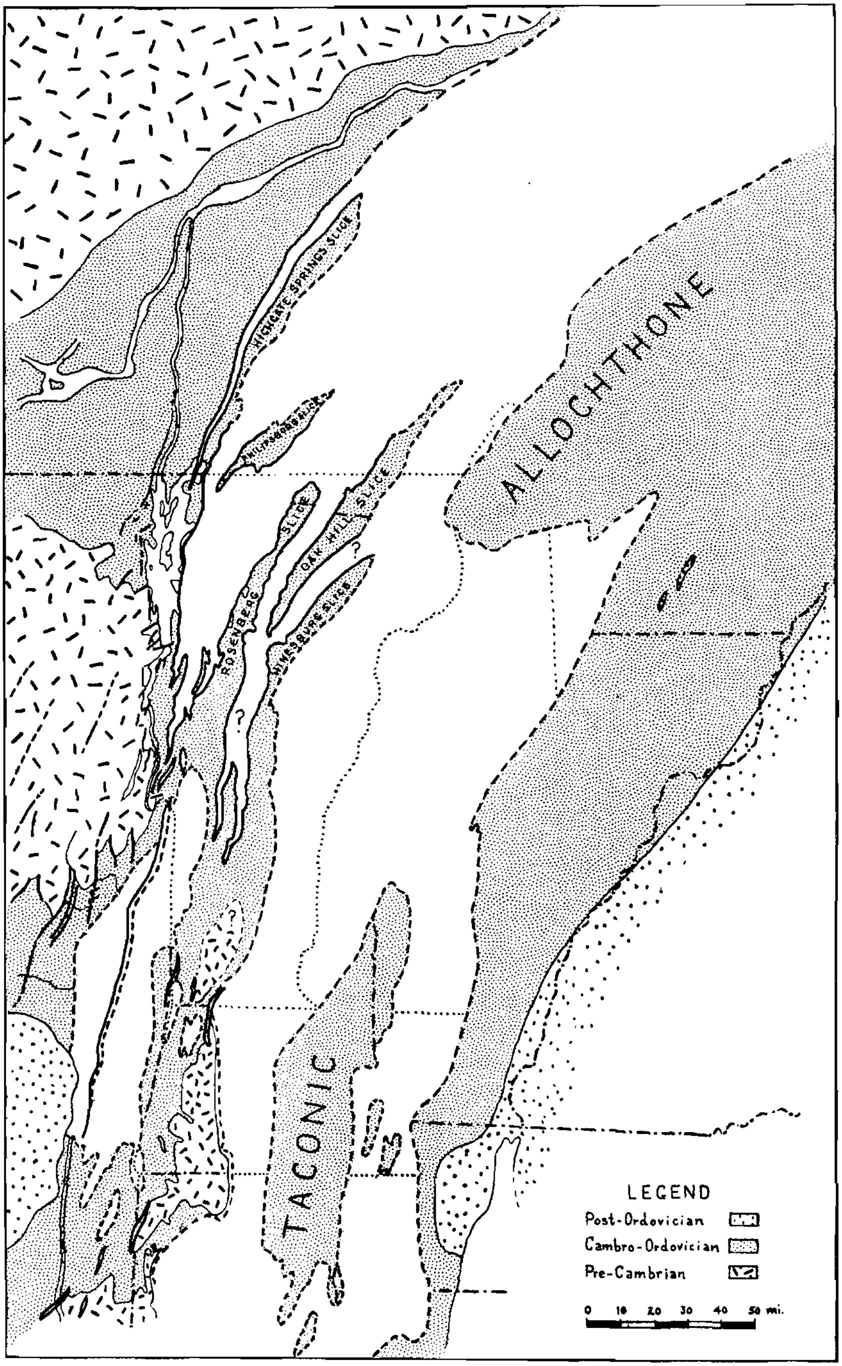


FIGURE 6.—*Palinspastic map of the region of western New England, eastern New York, and southern Quebec*

(Knopf, 1935, p. 208-209). The Taconic klippe is an erosional remnant of the once much larger Taconic Allochthone (Fig. 6).

The trace of the Taconic thrust has not been completely established. The contact between the black slate terrane in the lower slopes of the Taconic Range and the valley limestone at the north foot of the range is poorly exposed; and although in structural details it yields inconclusive evidence of thrusting, areal mapping has shown that it truncates (Pl. 10) the middle Trenton Glens Falls limestone at the meridian of Sudbury village down to the upper Beekmantown Bascom formation at the meridian of Brandon village. Keith (1913, p. 680) recognized this discordance and remarked that "an overthrust of the slates seems the only competent explanation of their relations". Possibly, however, the thrust is masked in the phyllites higher up on the range. There is evidence, already cited, in the Rutland region on the eastern edge of the Taconic Range, that certain phyllites are autochthonous and probably equivalent to the upper middle Trenton slates, suggesting that some of the lowermost slates of the range are Trenton and that the discordance along the north end of the slate terrane is an angular unconformity within the autochthonous succession.

Very likely the Taconic thrust extends far beyond the Taconic Range. Logan (1861, p. 379–382) first recognized the necessity for the "great break" from the field relations near the city of Quebec, where the allochthonous rocks, comparable to those in the Taconic region, lie adjacent to and tectonically upon the autochthonous foreland succession. This allochthonous succession may be traced southwestward (Ells, 1896, map in portfolio; Parks, 1931, Map 112 in pocket) from Quebec into the "Granby slice" (Clark, 1934, p. 4, 8, 14) southeast of the St. Lawernce River. Rather than a slice, however, the southward-jutting terrane of Sillery shales (Fig. 5) appears to be a remnant of the Taconic Allochthone (Fig. 6), which before erosion was continuous from the great klippe in the Taconic Range (Kay, 1937, p. 286, Pl. 5). The allochthonous rock at Granby is surrounded by a black shale terrane comparable with that around the base of the Taconic Range; most of this terrane is probably of upper middle to late Ordovician age (Ells, 1896, p. 17, 25–28; Foerste, 1924, p. 9–11, 56-57; Clark, 1934, p. 4, 6).

The Taconic Allochthone must have been transported from the southeast although a root zone has not been established there. In west-central Vermont a traverse can be made from the crystalline Adirondacks eastward at least to the Green Mountain Front without crossing the Allochthone, a single thrust fault, or a break of any kind. Undoubtedly the original site of the rocks of the Taconic Allochthone must have been somewhere east of the Green Mountain Front. Since the Green Mountains are underlain by a predominantly metamorphic terrane, a thrust fault might be difficult to recognize. Keith (1932, p. 404) states that

"The sole of the overthrust passes upward into the air along the east side of the Taconic Range and does not come down again until far to the east in the middle of the Green Mountains. There, rocks of the same sort as those of the Taconic Range are found resting on the Plymouth marble. . . ."[5]

Hawkes, however, pointed out (1941, p. 661–663) that there is no structural evidence of thrusting, such as truncation, at the contact between the marbles at Plymouth and

[5] On the east flank of the main Green Mountain ridge north and south of Plymouth, Vermont (Perry, E.L., 1927, p. 160–162; 1929, p 1–64; Dale, T.N., 1915, p. 22–34), in an eastward-dipping succession, are the easternmost exposures of rock of the Champlain Valley type calcareous facies. Keith (1932, p. 404-405) correlated these beds with the quartzites, dolomites, and limestones west of the Green Mountains, with the implication that before erosion they were connected across the arch of the Green Mountain anticlinorium.

overlying argillaceous rocks and that the change is gradational rather than sharp. W. White (oral communication) suggested that the Taconic thrust actually has no root, the Allochthone having moved northwestward as a detached unit after becoming separated from the southeastern rocks. Possibly that is true in Vermont, although quite as likely the thrust is masked in the deformed southeastern terrane. Much further investigation of this problem is necessary.

SUDBURY NAPPE: Adjacent to the north end of the Taconic klippe is the Sudbury nappe, found in the belt of autochthonous limestone between Sudbury and Whiting (Pl. 10). This is a true nappe according to the definition of Haug (Collet, 1935, p. 9), though small in comparison with the typical nappes of the Alps, inasmuch as it is a recumbent anticline part of whose reversed limb has disappeared through stretching. The core is massive Chazy marble and dolomite and the envelope less massive Chazy limestone. The root zone is exposed in the extreme northeastern foothills of the Taconic Range in northeastern Sudbury township. A gap in the envelope on the reversed limb of the nappe is readily recognized in northern Sudbury township. Here the marbles at the core are stretched westward and the limestones of the envelope break away from an inverted succession that includes both Chazy and Trenton beds visibly connected with the root zone. Thus the core rocks truncate the uppermost Chazy and Trenton of the inverted succession. This truncation is the Sudbury thrust, genetically a stretch thrust (Willis, 1893, p. 223). The Sudbury nappe is unique in that outcropping beds within the nappe strike north-northwest (Dale, 1904, p. 188) rather than north, which suggests greater displacement at the north end.

The present Sudbury nappe is a small remnant of the original. The root zone exposed at Middlebury (Pl. 4, fig. 3) and as far north as northeastern Weybridge is a fair indication of the minimum northward extension of the nappe structure, although the nappe proper is eroded away. At Sudbury and south-eastward simple south-southeastward-pitching folds occur all across the limestone belt in the vicinity of Hyde Manor. Here they are probably in the upper limb of the nappe not far from its southern terminus.

The comparable patterns of the Sudbury nappe and Taconic thrust seem to indicate a genetic relationship. The limestones on and east of the eastward-dipping upper limb of the nappe strike southwest rather than south-southeast as is characteristic of the east limb of the Middlebury synclinorium, suggesting drag produced by the once overriding Taconic Allochthone.

Normal fault system.—A north-south-trending belt of normal faults lies in and around the Adirondack region at the western border of the area (Fig. 5). A set of longitudinal faults downthrown to the east-southeast strikes north-northeast, separating fault blocks which tilt downward to the west-northwest (Swinnerton, 1932, p. 414; Brainerd and Seely, 1890b, p. 12). A transverse set, which strikes roughly east-northeast at a considerable angle to the longitudinal set, dies out northeastward a short distance east of Lake Champlain. The larger transverse faults cut the longitudinal faults (Rodgers, 1937, p. 1584; Quinn, 1933, p. 121) and are downthrown to the north-northwest (Brainerd and Seely, 1890b, p. 15, 16; Rodgers, 1937, p. 1583–1585; Quinn, 1933, p. 121).

A rather extensive longitudinal fault crosses the southern edge of the area north of

Wilcox Hill in Benson township (Pl. 10). South-southwest from this point, it parallels Lake Champlain for several miles and dies out to the north-northeast in the middle Ordovician-Trenton shale, increasing in displacement toward the southwest. West of Wilcox Hill the shales exposed in the bottom of a small valley are faulted down to the east against Lower Ordovician-Beekmantown strata exposed on the west wall of the valley, a stratigraphic throw of 500 to 1000 feet. This normal fault truncates the Benson thrust exposed along the east wall of the valley.

Two transverse high-angle faults have been recognized, about 1 mile west of Orwell village (Pl. 10) at a point 1/4 mile north of East Creek (Brainerd and Seely, 1890b, p. 7) where they cut across the formations and structures along the westward-facing escarpment of the Orwell thrust. They are roughly parallel, striking west-northwest, and apparently down-thrown to the north. The fault planes or zones are covered, but the Beekmantown-Trenton contact in the overthrust block is offset along a line approximately coincident with a road passing westward down the thrust scarp and along another line about 1/8 mile to the south. Heavy cover to the west, at the foot of the scarp, makes it impossible to determine whether the faults continue down into the middle Trenton shales beneath the overthrust Beekmantown dolomite, disturbing the thrust plane. Normal faults are reported elsewhere only in the zone beneath the Orwell and Champlain thrust planes, exposed west of the thrust scarp. Probably the faults west of Orwell are some rare upward extensions of the normal faults, here cutting the Orwell thrust plane.

The major faults of the Adirondack system do not intersect the major thrust faults farther east. The border of the crystalline Adirondacks, which at most places lies along one or more of these normal faults, parallels the Orwell and Champlain thrusts from Orwell northward nearly as far as Burlington, the various bends in one system being matched by those in the other (Pl. 1, fig. 1). Lake Champlain lies in the intervening area, largely on the soft middle Trenton shales and limestones, which are close to the Adirondacks on the western sides of the westward-tilted fault blocks. These two fault zones may have been separated partially by the eastward retreat during erosion of the Champlain thrust scarp; the present trace of the Champlain thrust scarp, except to the south, is consistently about the same distance east of the exposures showing Adirondack normal faulting. The Adirondack faults seem to be rather closely related to a similarly oriented joint system in the Adirondack crystalline basement and are apparently limited in distribution to a tectonic zone in the immediate vicinity of the crystallines. Possibly the Champlain thrust and associated structures are in a zone of movement concentrically above this zone of strong normal faulting that lies in and peripheral to the Adirondack massif. The close relation of the normal faulting to the Adirondack rocks suggests that it has been associated genetically with their domal uplift. Brainerd and Seely (1890b, p. 12), pointed out that the normal fault blocks along Lake Champlain are tilted down to the north as well as westward. The two largest transverse faults found along Lake Champlain are also downthrown to the north. Megathlin (1938, p. 106, 119) has indicated a similar westward tilt and a southward decrease in fault displacement south of the Adirondacks along the Mohawk River, which (p. 119) he attributed to "forces which affected essentially the Adirondack region. . . ." He pointed out (p. 106) that the faults may disappear here "both by actual decrease of throw and by passage into shale which is probably taking up some of the displacements by adjustments within itself." Thus these faults apparently die out in stratigraphic as well as in tectonic horizons farther removed radially from those of the crystalline massif from which, presumably, they were propagated. They die out upward as well as laterally, and their time of movement was possibly later than that of several of the higher stratigraphic horizons that flank the Adirondacks and are unaffected by normal faulting. Likewise they are probably later than the major thrusts flanking the Adirondacks at some distance to the east.

LESSER STRUCTURES

General relations.—Many of the lesser structures are symmetrically disposed with respect to major structures, particularly the minor folds found on the limbs of the Middlebury synclinorium. However, in the vicinity of thrust faults, the regular pattern of the folds is disturbed, and odd-appearing anticlines and synclines are prominent. Most of the smaller thrusts, all of which are eastward-dipping, are of the break thrust type (Willis, 1893, p. 223) produced simply by the rupture of the steeply dipping or overturned western limb of an anticline. A few are subsequent shear thrusts (Billings, M.P., 1933, p. 141–143) that truncate folds and have been produced by movement along thrust planes developed later than and somewhat independently of movements causing the folds. Some of the more extensive of the lesser thrusts are strip thrusts (p. 142-143) like the major thrusts.

Detailed description.—Several such structures illustrate their genesis rather well and are here described in detail (Fig. 5).

Benson Thrust: A strip thrust is traceable intermittently southward through Benson township (Pl. 10), possibly into the township of West Haven. Reconnaissance work indicated that at all points where the fault can be definitely located upper Beekmantown or, in one place, older beds form a fault line scarp, which faces westward over lowland middle Trenton shales. The upper Beekmantown has apparently been stripped from lower strata to the east and now lies to the west on the shales from which higher zones have been removed in the same manner.

The thrust contact crosses a road half a mile north-northeast of Wilcox Hill. Here upper Beekmantown Bridport dolomite dips gently eastward on middle Trenton limestones and shales. Northeast of this road is a deep, narrow valley which opens out to the south-southwest. The valley has been formed along one of the high angle longitudinal faults of the Adirondack normal fault system downthrown to the east-southeast. The northwest wall of the valley is made up entirely of a subhorizontal succession of Beekmantown sandstones and dolomites that are continuous downward into the Upper Cambrian beds flanking the Adrinondack crystallines along Lake Champlain; these beds are not affected by thrusting. The floor and lower part of the southeast wall is underlain by Trenton slates. The upper Beekmantown Bridport dolomite crops out in thrust position above the slates higher on the southeast wall. Projected northwestward against the northwest wall the thrust plane meets the unbroken Beekmantown succession in the upthrown block of the high angle fault, against which it has been downthrown. The high angle fault can be traced south-southwest between the Benson thrust and the lake for nearly 8 miles.

Orwell Thrust and Anticline: The Orwell thrust (Brainerd and Seely, 1890b, p. 5, 7) (Pl. 10) is above subhorizontally distributed Trenton shales and is overlain by and truncates a folded competent succession, Upper Cambrian to Trenton in age. The west limb of the Orwell anticline (p. 7, 9) is being removed by eastward retreat of the Orwell thrust scarp. Apparently this anticline and related structures were developed before movement along the Orwell thrust; thus the under surface of the Orwell thrust plate is that of a subsequent shear thrust, although the upper surface of the footwall block onto which it has been transported most closely approximates that of a strip thrust.

SHOREHAM THRUST, SYNCLINE, AND ANTICLINE: At Shoreham village (Pl. 10) younger beds are thrust over older (Brainerd and Seely, 1890b, p. 3–9) along a flat subsequent shear thrust. Here the Shoreman syncline and the Shoreham anticline adjoining it on the east are truncated at the Shoreham thrust, along which these folds have been transported bodily westward into such a position that Trenton strata at the axis of the syncline lie on Upper Cambrian beds near the axis of the Orwell anticline. The northern continuation of the synclinal axis in the footwall block is exposed near the east foot of Mutton Hill at a meridian about 5/8 miles east of that of Shoreham village; thus a thrust displacement of about 3000 feet is indicated.

PINNACLE THRUST AND LEMON FAIR SYNCLINE: East of the Shoreham anticline along the Lemon Fair River (Pl. 10) is a rather broad, sinuous syncline the east limb of which is truncated by an overthrust succession of eastward-dipping competent strata (Bainerd and Seely, 1890b, p. 8, 9) comprising the west limb of the Middlebury synclinorium. The truncating fault is a break thrust dipping eastward subparallel to the overlying competent strata at the meridian of the Pinnacle in eastern Shoreham (p. 5).

LAKE DUNMORE THRUST AND SYNCLINE: Lake Dunmore, located in eastern Salisbury and Leicester townships (Pl. 1, fig. 2; Pl. 10) "seems to lie in a southward pitching syncline of quartzite, with overlying dolomite, the dolomite having been largely eroded" (Dale, T.N., 1905, p. 47). At the Falls of Lana or "cascades" east of the lake the quartzites dip very steeply, nearly vertically, to the west. The beds on the east limb are more or less continuous along the strike with those of the Lower Cambrian of the Green Mountain Front. At Sunset Hill and on hills north and southwest of the lake the quartzite and underlying phyllites dip 30° ± E. Beneath the quartzite and phyllite and subparallel to them is a rather steeply inclined break thrust that is traceable from the latitude of Bryant Mountain in Salisbury township southward at least as far as a hill southeast of Mud Pond in Leicester township. The northern and southern ends of this thrust are covered. All of the low area immediately to the west is underlain by Lower Cambrian dolomite beneath the thrust plane. This dolomite is continuous along the strike for many miles, both to the north and south of Lake Dunmore, at the west foot of the Green Mountain Front (Pl. 10). Inasmuch as the dolomite is not appreciably narrowed or cut out throughout this belt, possible northward or southward continuations of the Lake Dunmore thrust along the Green Mountain Front in the nature of a border fault must be of smaller displacement than at the lake and certainly do not approach the magnitude of the Champlain, Hinesburg, or Oak Hill thrusts. The Pine Hill thrust (Dale, T.N., 1892a, p. 514–519) is a similar but much more extensive displacement in the marble valley region southeast of west-central Vermont.

HOGBACK ANTICLINE AND STARKSBORO SYNCLINE: These two adjoining folds pitch north, and in northeastern Monkton township (Pl. 10) the Lower Cambrian quartzite of the anticline disappears beneath the surface, which there is everywhere formed on overlying Lower Cambrian dolomite.

"The quartzite is typically exposed throughout the Hogback and South Mountain areas, comprising the front range or western flank of the Green Mountain range. . . . The relationship of the quartzite to the Cambrian dolomite in the Starksboro valley is well shown by the structure at the southeast end of Hogback Mountains where the dolomite overlies the quartzite, and in the bed of the New Haven River above Ackworth, where the structure of the quartzite is decidedly synclinal. . . ." [Dale, N.C., 1921, p. 51–52]. Vertically dipping beds of Cambrian quartzite and dolomite south of Starksboro and easterly dipping beds on Hogback Mountain, classify the structure as an asymmetrical syncline . . ." [1919, p. 198]. At the north end of the Hogback a low hill, situated north of the road . . . , is of hard grayish quartzyte, while the main range south is brownish and distinctly stratified, with dip 70° to 75° to the eastward. North of this low quartzite hill, the quartzite extends on northward in narrow masses, and is directly overlaid on the northeast by dolomite and bounded by dolomite around the the whole northern end. The dolomite on the west was not seen in immediate contact with the quartzyte; but at the nearest point . . . it seemed to dip east at a high angle, while a few rods to the west it had a decided westward dip . . . [Dana, 1877a, p. 409]. . . . [This] "pitching anticlinal structure at the north end of the Hogback bears evidence for anticlinal structure only for the north end of the west limb" . . . [Dale, N.C., 1919, p. 198].

South of the north end of the Hogback Mountains along the Green Mountain Front the contact between the quartzites of the escarpment and the dolomites next west in the valley is covered by talus, lake beds, swamps, delta deposits, or morainal material. Wing (Dana, 1877a, p. 410) interpreted this contact as the normal sedimentary contact between inverted quartzites on the west limb of the Hogback Mountains anticlinal and the stratigraphically higher valley dolomites to the west, much as at the extreme north end of the range. N.C. Dale (1919, p. 198) did not recognize inversion of the strata on the west side of the Hogback Mountains proper and stated that "the structure at the valley range contact would seem to be.... in the nature of a hinge fault". If the Green Mountain Front south of the north end of the Hogback Mountains is a fault-line scarp, presumably of a steeply inclined thrust, the heave of this fault is insignificant as compared with that of the other thrusts, inasmuch as the dolomite belt next west is at no point completely cut out. The simpler interpretation is that beneath the float along the west foot of the Hogback Mountains and at comparable localities of the Green Mountain Front farther south is a normal sedimentary contact between quartzites to the east and dolomites to the west, the escarpment being produced solely by differential weathering and erosion of the steeply dipping quartzite and dolomite beds. Apparently this is Keith's (1932, p. 395) interpretation.

MONKTON THRUST: This thrust (Keith, 1932, p. 359, 364) is traceable by truncation of beds and juxtaposition and abnormal superposition of widely separated stratigraphic units. It extends from a southward-pitching anticline 1 mile northeast of New Haven village northward to a northward-pitching anticline at Monkton Ridge village (Pl. 10). The massive gray facies and also a brown, rather schistose facies of the Lower Cambrian Cheshire quartzite, found in most of the hilly area trending southward through the center of Monkton township into northwestern Bristol township, overlie the thrust plane. Along the thrust between the latitude of Monkton village and the north boundary of New Haven township the red Lower Cambrian Monkton quartzite crops out immediately west of the Cheshire. The great stratigraphic separation of these two formations is proved by an examination of the northward-plunging anticline, north of Monkton Ridge village. Here the Cheshire crops out at the core of the fold and is separated from 725 feet of Monkton, well-exposed on the west limb, by 1760 feet of dolomite. This dolomite is evidently completely cut out by the Monkton thrust. The actual truncation is less apparent in the field, however, because Monkton Pond lies over the zone of truncation near the north end of the thrust and Cedar Swamp covers it near the southern terminus. In northern New Haven township, south of the Monkton area, the rocks immediately west of the thrust are covered. Possibly here the Cheshire is thrust on the Winooski dolomite, stratigraphically above the Monkton. This appears to be a break thrust with at least 2500 feet of stratigraphic throw.

MUDDY BROOK THRUST: This thrust sheet, containing Upper Cambrian slates and conglomerates, thinly underlies the present erosion surface in a narrow belt west of the Hinesburg thrust in St. George and Williston townships, near Muddy Brook (Pl. 10). It has been so deeply eroded that it is cut by numerous fensters through which Beekmantown rocks and minor folds of the Hinesburg synclinorium are exposed. These folds are truncated by the thrust. The Upper Cambrian slates are readily distinguished from the extremely quartzose rocks of the Hinesburg slice. Structures are practically indistinguishable in the Muddy Brook thrust sheet because of the homogeneity of the slates. The slates probably form a smeared out portion of the incompetent Upper Cambrian slate from the east limb of the Hinesburg synclinorium that has been dragged across the younger beds in the zone of thrusting beneath the Hinesburg slice.

CHARLOTTE ANTICLINE: A fold divides the south end of the Hinesburg synclinorium into two parts (Pl. 10). It strikes north-northeast coincident with the pattern of a re-entrant in the Champlain thrust east of Mt. Philo. This suggests that the folding that produced the Charlotte anticline may have caused some warping of the Champlain thrust plane.

VERGENNES THRUST: This thrust (Whitfield, 1886, p. 299) may be traced from north of the village of Addison, near the Addison-Panton town line, north-northeastward through the city of Vergennes and the village of Ferrisburg (Pl. 10). Although it is found west of the area, it is noteworthy being a very good example of a thrust whose displacement decreases south-southwestward toward relatively undisturbed horizontal beds rather rigidly connected with the crystalline basement at the border of the Adirondacks. In Addison township a broad anticline, at the axis of which upper Beekmantown beds are exposed, is the only sign of distrubance. Toward the north at Vergennes, the upper Beekmantown beds lie in thrust position on middle Trenton shales that flank the anticline farther south. North of Ferrisburg village, where the fault passes into the middle Trenton shale terrane, the relations are obscure. Reconnaissance revealed a similar line of faulting extending from the northern slopes of Snake Mountain north to Ferrisburg and lying between and roughly parallel to the Vergennes and Champlain thrusts. This thrust may be the northern extension of the Orwell thrust, truncated for a short distance by the Champlain thrust at Snake Mountain.

Summary statement.—The folds and thrust faults here described reflect to a varying degree the probable genesis of the regional structure. The folds in Orwell, Shoreham, and Bridport townships parallel the structures of the Middlebury synclinorium and some of them are truncated by the Orwell thrust. They are earlier than thrusting and probably of about the same generation as the folds of the synclinorium. The Vergennes, Orwell, and Benson thrust plates are much alike, being attached rather firmly to the Adirondack crystalline basement at their southern ends and showing increased displacement, particularly by strip thrusting, to the north. The three thrusts are essentially en echelon. All three disappear north-northeastward in the Trenton shale. The Vergennes thrust and the north end of the Orwell thrust are particularly significant because they trend north-northeastward beneath the Champlain thrust plane,—good structural evidence for the westward counterclockwise rotation of the Rosenberg slice. Presumably all of the thrusts developed from preexisting northeast-striking folds. Some folds may have been formed after thrusting as suggested by the agreement between the pattern of the trace of the Champlain thrust and the pattern of shallow folds in the Hinesburg synclinorium.

STRUCTURAL DETAILS

General scope.—Certain details of the geologic structure have such a constant relation to tectonic and stratigraphic zones that a brief reference to them is desirable. Typical examples of these small structures in western New England and eastern New York were described by Dale (1892b, p. 317–319; 1896, p. 549–570; 1899, p. 199–217; 1902). Slaty cleavage, schistosity, fracture cleavage, and drag folds in the absence of continuous outcrops and such original structures as cross-bedding, help determine the structural position of rocks of isolated outcrops (Pl. 7). This method (Leith, 1923, p. 176–185; Nevin, 1936, p. 75, 80, 176, 181) is particularly helpful in establishing tops of the beds of the rather highly deformed sedimentary rocks in the eastern part of the area.

Foliation.—Practically all of the argillaceous rocks east of the Champlain thrust show more or less well-defined slaty cleavage. Toward the eastern part of the area slaty cleavage passes into schistosity and the rocks are phyllitic. The best examples of slaty cleavage are in the middle Ordovician Hortonville slate found along the axis of the Middlebury synclinorium. The Upper Cambrian succession in northwestern Vermont includes slates similar in general appearance. The Parker slate of the Lower Cambrian of that region has a good slaty cleavage.

The foliation is subparallel to the axial planes of folds and is regional in aspect with a general northerly strike; it dips at varying angles to the east. The foliation planes are probably not planes of shear;—Fairbairn (1935) says of comparable foliation in the Oak Hill slice (p. 594) that (p. 600) the "spatial relation to other minerals" of the micas "suggests a minimum of differential movement . . . and indicates that they owe their present parallel arrangement simply to dimensional orientation imposed by the deforming forces."

Flow structure.—Flow structure is particularly well developed in marble interbedded with the more competent dolomites in the Beldens formation (Pl. 8, fig. 1). Whereas the dolomite beds are heavily jointed, the adjoining marbles, although deformed as much as the dolomite, are not ruptured. Lines of flow (such as are produced by dimensionally oriented particles paralleling the flow direction) if present are at most places indistinguishable from bedding. Where the dolomite beds are relatively thin or deformation is greater the dolomite is broken and pinched off into "eyes" completely surrounded by marble. This phenomenon is common at the north end of the Taconic Range, where the rather thinly attenuated and overturned Beldens formation forms the greater part of the Sudbury nappe. The Beldens has here, after reversal, been reduced to about one third its original thickness by flattening, with resultant extension in the direction of the bedding planes causing the dolomite strata to separate into the typical "eyes". Where limestone is thinly streaked with more competent material flow structure is apparent. The upper part of the Lower Ordovician Bascom formation is a limestone with rather thin sandy laminae not large enough to have produced competent folds yet sufficiently different in facies from the limestone to be differentiated from it and to reflect by their distribution a striking flowage pattern (Pl. 8, fig. 2). Here the thin competent layers are broken into fragments arranged en echelon. These fragments are cut by fracture cleavage and isolated ones are separated from each other at fracture cleavage planes. These megascopic shear planes do not necessarily extend into the limestone; thus the separation of the isolated fragments is strictly a flowage phenomenon. Rather continuous flowage fold patterns may be observed in the walls of Shelburne marble quarries. Secondary folds within the Shelburne facies are oriented on the larger competent folds in which the marble occurs in a direction opposite to that of the normal type of drag fold (Bain, 1931, p. 503–530). Possibly the two types of folds originated independently and are related to two different episodes of deformation.

The direction of plastic flow is fairly well shown by the orientation of the long axes of deformed planispiral gastropods found in the marbles and limestones at several localities. Observed in steeply dipping beds on the limbs of folds, they are stretched in a nearly vertical direction in the plane of the bedding, subparallel to the axial plane of the fold and at a right angle to the north-south-striking fold axes. A few were seen in gently folded or horizontal beds that were stretched horizontally in a north-south direction parallel to the regional structure. Probably this is only an apparent stretch parallel to the median axis of elongation of the deformed limestone and conditioned by the subhorizontal position of the gastropod disk in the structure.

Flexures.—Drag folds vary from microscopic size in the Hortonville slates and phyllites up to almost major proportions (Pl. 9, fig. 1) in the thick-bedded competent Cheshire quartzite. They may grade imperceptibly into flow structures where competent and incompetent facies blend, flexural folding and flowage folding becoming indistinguishable. The two types of folding commonly occur in adjacent strata of contrasting competent and incompetent facies (Pl. 8, fig. 2).

In limestone and marble beds nearer the top of the calcareous facies, particularly the incompetent Middlebury limestone, flexural folds occur in the form of small wavy flexures grading into fracture cleavage (Pl. 9, fig. 2). Larger flexures are not so apparent in the Middlebury except near the contact with the underlying Beldens formation (Pl. 9, fig. 4) where the competent dolomites in the Beldens (Pl. 9, fig. 3) evidently controlled the folding of the adjoining limestones.

Ruptures.—East-west tension joints are found throughout the area. Basic

FIGURE 1. FLOWAGE OF MARBLE BETWEEN BRITTLE DOLOMITE BEDS
Beldens formation, 3 miles SSW. of New Haven Junction.

FIGURE 2. FLOWAGE OF LIMESTONE REFLECTED BY THIN SANDY LAMINAE
Bascom Zone 4 limestone on east limb of Middlebury synclinorium 1 mile NE. of West Salisbury village.

FLOW STRUCTURE

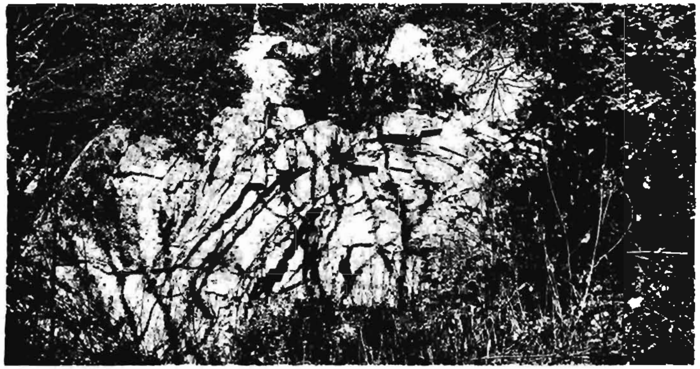


FIGURE 1. SYMMETRICAL ANTICLINE
In Cheshire quartzite in hills of northwestern Bristol township, 2 1/8 miles NNE. of New Haven village.

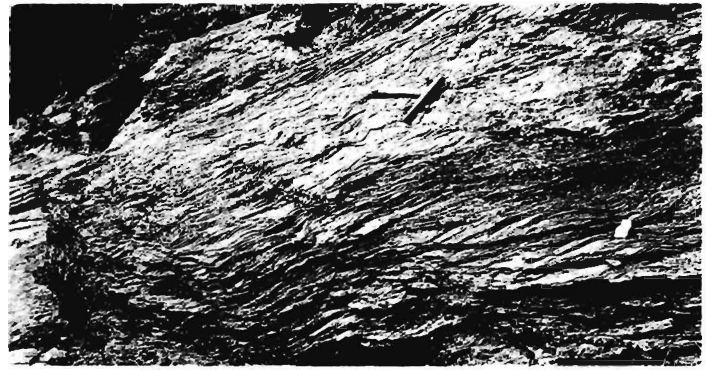


FIGURE 2. WAVY FLEXURES PASSING INTO FRACTURE CLEAVAGE
Middlebury limestone, 1 mile NNW. of Middlebury village.

FIGURE 3. ASYMMETRICAL ANTICLINE
In the Beldens formation, north bank of Otter Creek, 5/8 mile WNW. of Brooksville, New Haven township.

FIGURE 4. RECUMBENT FOLD
In Middlebury limestone near contact with underlying Beldens formation, 1 5/8 miles SSE. of Weybridge village.

FOLDS

dikes, similar to those reported by Kemp and Marsters (1893) along Lake Champlain, intrude them at several places. In the western part of the area occurs an equally prominent north-south set of tension joints. These longitudinal joints are particularly well developed on Snake Mountain and Buck Mountain in the Red Sandrock Range, where, paralleling the fold axes, they strike north-northwest, having been apparently swept out of line by the westward pivotal movement of the Rosenberg slice around the south end of the Champlain thrust in Cornwall. Joints at intermediate positions seem to reflect local strains. Gash fractures, in most places filled with calcite or quartz that form gash veins, are abundant. The Orwell limestone shows a network of cross veins in joints (Pl. 5, fig. 1).

MECHANISM OF DEFORMATION

The Middlebury and Hinesburg synclinoria are remnants of a major structural downfold, rather persistent on the southeastern border of the foreland in the region of the folded belt of and south of west-central Vermont (Fig. 5). The axis of this downfold originally trended northeast but thrusting made it strike north. The thrust slices apparently rotated counterclockwise around pivotal zones to the south where closely attached to the eastern edge of the Adirondack crystalline massif from which the north-northeast-striking thrusts spring en echelon. The major thrusts bound the two sides of the synclinorial tract where their position was conditioned respectively by the competent, gently eastward dipping, strutlike succession of the west limb and by the west limb of the Green Mountain anticlinorium to the east. Overthrusting was particularly favored in northern Vermont by a succession incompetent above and below competent center (Keith, 1923b, p. 317). The major thrusts extend to the north end of the competent facies about at the Canadian border; farther north thrust movements may have been distributed through the incompetent rocks on numerous planes. Folding of both the argillaceous rocks to the northeast and of the competent sandstones and dolomites to the southwest preceded thrusting. The competent successions tended to move forward as tabular masses undeformed within themselves; but the argillaceous rocks were folded, the slices moved ahead of the shales, and tear faulting along the boundary between the facies at the north end of the slices probably initiated thrusting. As deformation continued, this thrusting proceeded by tearing southward along the limbs of pre-existing folds oriented en echelon with respect to the foreland margin. This tearing stopped when it reached the rigid crystalline basement. The counterclockwise sweep of the slices did not stop, however, as is indicated by the position of the Rosenberg slice, which trends north-northwest athwart the en echelon folds and faults in the footwall block near Vergennes.

Throughout most of the Champlain thrust the competent Lower Cambrian quartzites and dolomites of the west limbs of the synclinoria dip gently east, subparallel to the thrust plane. Thus the initial break, where these strata were separated from the underlying, was accomplished by "tectonic stripping" of younger beds from older beds below. Toward the north end of the thrust, in northwestern Vermont, the Dunham dolomite was stripped from the Cheshire quartzite equivalent, which, where it crops out in the Oak Hill slice a little to the east, is an incompetent, rather argillaceous rock. Toward the south end of the thrust in west-central Vermont, where the Cheshire is very competent, the Monkton quartzite was stripped from less competent Dunham dolomite. The beds originally above the incompetent rocks exposed west of the present fault-line scarp were removed largely by tectonic stripping; incompetent Trenton limestones and shales underlie much of the area between the Champlain thrust and the lake. Thrusting would be opposed by a minimum of friction in such a facies.

The present fault line of the Hinesburg and Oak Hill thrusts along the east side of the synclinoria is much lower stratigraphically in the footwall than is that of the Champlain thrust. Upper Cambrian and possibly Lower Ordovician beds rather than Trenton strata underlie the area west of the Oak Hill thrust. They are mainly incompetent shales and slates, from which higher beds probably have been removed by strip thrusting. Inasmuch as the youngest (probably Upper Cambrain) beds in the Oak Hill slice are followed to the east by older (Lower Cambrian) beds, all apparently dipping into the thrust plane, probably the rocks in the present escarpment were originally separated from the substratum along either a shear or a break thrust plane. At erosional re-entrants, as the one near Williston, folds in the Oak Hill succession are truncated at the Oak Hill thrust, indicating a subsequent shear fault. Here the thrust also truncates competent folds in the footwall, but north of this point as far as it can be traced, the hanging wall rests on an incompetent footwall originally formed by strip thrusting. The Oak Hill slice seems to have been so tightly folded before thrusting that it is doubtful if a tabular mass, such as that of the gently eastward dipping Dunham or Monkton strata lying above and parallel to the Champlain thrust plane, was available in the zone of the Oak Hill slice at the time of thrusting. The hanging wall of the Hinesburg thrust, also produced by subsequent shear thrusting, truncates competent folds in the footwall Rosenberg slice. The imcompetent beds of the thin Muddy Brook thrust sheet lying beneath the Hinesburg thrust plane apparently were dragged out of an incompetent stratum truncated by the Hinesburg thrust farther east.

The mechanics of transport of the Taconic Allochthone are obscure but certain facts are suggestive: At the latitude of Rutland the westernmost exposures of the allochthonous facies are about 20 miles west of the easternmost exposures of calcareous rocks known to be of the autochthonous facies. At the latitude of Albany corresponding exposures are nearly 50 miles apart, requiring a minimum transport of that distance, with probably at least 25 miles more. Here allochthonous rocks are distributed over the whole 50 mile distance, requiring that the Allochthone be at least 50 miles wide. The allochthonous rocks exposed in the Taconic klippe and in the remnant of the Taconic Allochthone at Granby, Quebec, and northward are apparently right side up and within themselves rather simple structurally although transported far. They lack the necessary strength to be thrust out as a rigid long column, and it seems likely that the forces moving the Allochthone arose within it much as in tar moving on a sloping road surface or in continental glacial ice moving out over flat ground from a center of accumulation.[6] The Allochthone might have moved as a

[6] Hubbert (1937, p. 1498-1499) has indicated that to make a scale model of a comparable zone of crustal shortening 1 meter wide the material used to represent the rock must have the strength of vaseline, a cube of which larger than 3.3 centimeters to the side could not support its own weight.

detached mass or remained connected to a root zone. The inability of several investigators to locate a root zone suggests detached movements at points in Vermont. With possibly a few exceptions the Allochthone moved west-northwestward in a zone confined almost entirely to the various argillaceous facies. Autochthonous Middle Ordovician shales, slates, and phyllites fringing the Allochtone and lying immediately above a predominantly calcareous succession may have served as an incompetent lubricating facies forming a favorable sole for movement.

The problems of the genesis of the regional structure in western New England are somewhat comparable to those of the Alps. The various thrusts bordering the thrust slices are analogous to the Alpine foreland thrusts. The Sudbury thrust is developed by the stretching of the overturned limb of an anticline as in the typical foreland nappes. The other thrusts, such as the Champlain and Hinesburg-Oak Hill thrusts, do not appear to be associated with a nappe structure, yet are developed in the predominantly calcareous, typical foreland facies bordering on the Adirondack crystalline massif. The Taconic thrust lies beneath a typical succession of geosynclinal clastics once located in a trough southeast of the foreland; the Taconic Allochthone is comparable to the Prealpine nappes. Whether the foreland thrusts and the genetically associated folds were produced at the time of emplacement of the Taconic Allochthone or later seems a point of major issue. According to either interpretation the foreland thrust slices have been shingled up against the flanks of the Adirondack crystalline massif, as an effect of the drag of an overriding shale facies. Discreet movemnet of the Taconic Allochtone at the time of emplacement or later general movement of superincumbent shale facies, both autochthonous and in the Allochthone, could have effected such a drag.

Differential vertical movements seem to have determined the final tectonic pattern in this region. The pitching structures of the Middlebury and Hinesburg synclinoria probably were formed at the time of uplift of the Adirondack Mountains. The Monkton cross anticline between the synclinoria appears to be an eastern extesion of this uplift. The doming of the Adirondacks produced the concentric pattern of the Champlain thrust with respect to the Adirondack border. It took place by adjustments probably along pre-existing joints or faults within the crystallines; these structures determined the pattern of the normal fault system that extends into the immediately overlying Paleozoics.

DIASTROPHIC RELATIONS

GENERAL STATEMENT

In the foregoing sections possible sequences of major diastrophic events that may have operated to produce the present tectonic pattern have been implied. These events are summarized and the critical and also controversial data available both in the area of the present study and in adjacent regions are evaluated.

TACONIC ALLOCHTHONE

Post-middle Trenton—pre-late Silurian is the closest dating of the emplacement of the Taconic Allochthone possible after study of its contact relations in west-central Vermont and at Becraft Mountain, New York (Schuchert and Longwell, 1932, p.

317–321; Kay, 1937, p. 287-288, Pl. 5). In west-central Vermont the Taconic Allocththone has been described as truncating the foreland folds (Keith, 1913, p. 680), although this is possibly a misinterpretation. The youngest strata included in these folds are middle Trenton. The unconformity found above the Allochthone at Becraft Mountain can be traced southwestward in New York and New Jersey, where it lies above the autochthone and below the early Silurian Shawangunk conglomerate (Schuchert and Longwell, 1932, p. 307, 310-311). Farther southwest in Pennsylvania, according to Willard and Cleaves (1939, p. 1165–1198), it is found below a thin conglomerate correlated with the upper Maysville "Bald Eagle member of the Juniata formation". At points in Quebec south of the St. Lawrence, where the Richmond—Queenston shale, equivalent to the Juniata formation, is adjacent to the Taconic Allochthone (Ells, 1896 map in portfolio; 1900, map in portfolio; Parks, 1931, map in pocket), details of the contact relations have not been established (Parks, p. 28, 38).

FORELAND THRUSTS AND ASSOCIATED FOLDS

Foreland thrusts in Quebec and similar thrusts in west-central Vermont, where genetically related to folding, are of the same type as, and are possibly continuous to the south with, thrusts that cut both the Taconic Allochthone and Middle Devonian beds (Schuchert and Longwell, 1932, p. 324) in the Hudson Valley region. In describing the Becraft Mountain locality in the Hudson Valley Schuchert and Longwell remarked that "there is no suggestion that the second Paleozoic deformation in this region dies out to the north. If the later formations were present to record its effects, no doubt this disturbance would be recognized much farther north than Becraft Mountain. . . ." (p. 320-321). Remarking on the tectonic history of the region they stated: "It is not inconsistent with the facts to postulate that the movement was part of the Acadian orogeny in southeastern Canada" (p. 324). The deep downfolds of Silurian (Clark, 1936a) beds (Ells, 1896, map in portfolio; Clark, 1934, p. 12–13) on the east flank of the Green Mountain anticlinorium in southern Quebec are good evidence of post-Silurian folding in the Green Mountain region. In regard to the dating of the major folding and associated thrusts in western New England it is of particular significance that these movements are genetically related (Balk, 1927, p. 39–96; Barth, 1936, p. 832; Bain, 1940, p. 1989) to granitization effects and granitic intrusions that increase in amount southeastward and are "generally considered to be of Devonian age" (Clark, 1934, p. 16; Cooke, 1937, p. 85). Keith (1923b, p. 322) has suggested a genetic relation between the granitic intrusives and the formation of the arcuate pattern of the Green Mountains, pointing out that "there is a considerable concentration of them in southern Maine and the White Mountains of New Hampshire back of the northern Vermont salient".

ADIRONDACK NORMAL FAULTS

Although the Adirondack normal faults are not reported to cut beds higher than Ordovician (Trenton) possibly they may be much younger, simply dying out in displacement (Brainerd and Seely, 1890b, p. 12; Megathlin, 1938, p. 106, 119) radially upward away from the crystallines in Trenton shales. Basic dikes found along joints associated with the normal faults have a helium ratio considered to indicate their late Ordovician or Silurian age (Urry, 1936, p. 1218, 1224-1225). The joints, which seem to have originated in the pre-Cambrian (Balk, 1932, p. 68), evidently were loci of movement and intrusion in at least one and probably at several later episodes in geologic history. The normal faults are coincident in trend and position with a stratigraphic axis developed along an arch that appeared as early as the middle Ordovician-Trenton (Kay, 1937, p. 288). They are more recent than thrusting in west-central Vermont.

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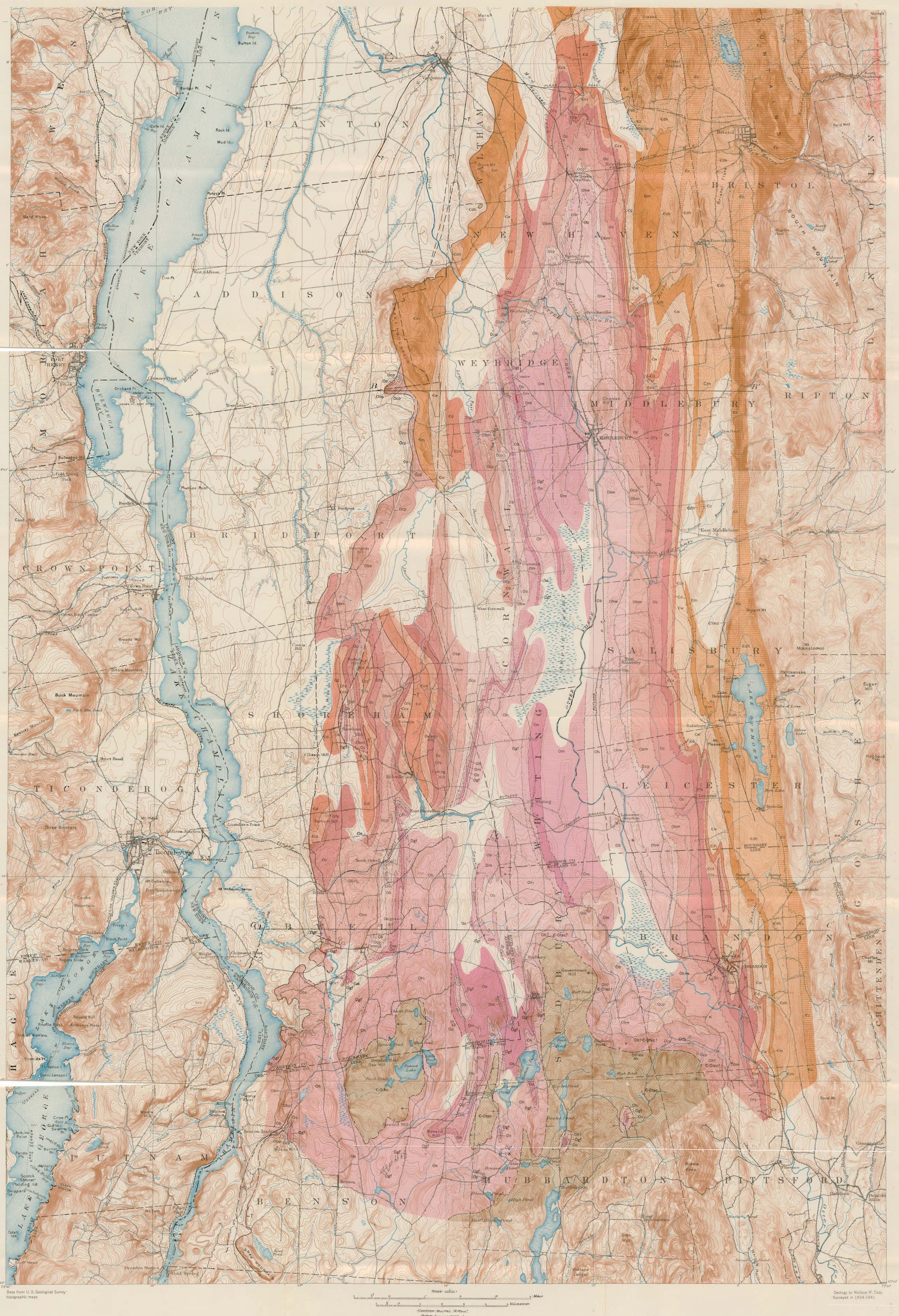
EXPLANATION
SEDIMENTARY ROCKS
Undifferentiated
Є-Otac
(allochthonous grits, slates, phyllites, quartzites, and graywackes)
CAMBRO-ORD.
Black River and Trenton groups
Oh
Hortonville slate
(red-brown weathering blue-black slate)
UNCONFORMITY
Ogf
Glens Falls limestone
(thin-bedded, dark-blue-gray limestone)
Oo
Orwell limestone
(heavy ledged, dove gray weathering, fine textured)
UNCONFORMITY
Chazy group
Om
Middlebury limestone
(thin-bedded, buff-streaked, dark-blue-gray limestone)
Ob
Beldens formation
(interbedded buff dolomite and white marbly limestone)
Obw
Weybridge member of Beldens formation
(limestone with thin raised sandy ridges 1 inch apart)
Ocp
Crown Point limestone
(lead gray, massive, buff stippled and streaked)
UNCONFORMITY
Beekmantown "group"
Obp
Bridport dolomite
(drab, yellowish, or brown massive dolomite)
Obm
Bascom formation
(limestone, dolomite, sandstone, quartzite, limestone breccia, sandy calcareous shales)
Oc
Cutting dolomite
(sandstone at base, black chert in upper part)
O?s
Shelburne marble
(white marble or dove-colored limestone, light-gray dolomite)
ORDOVICIAN
Upper Cambrian series
Єcs
Clarendon Springs dolomite
(massive, smooth-weathering gray dolomite containing masses of white quartz)
Єus
Georgia sl. (?), Rockledge congl., Hungerford sl., Skeels Corners formation (?)
(dark slates and limestone pebble congl.)
Єd
Danby formation
(interbedded gray quartzite and dolomite)
Lower Cambrian series
Єw
Winooski dolomite
(four to twelve inch beds separated by thin dark siliceous partings)
Єm
Monkton quartzite
Єp
Parker slate
Єdh
Dunham dolomite
Єc
Cheshire quartzite
Є?md
CAMBRIAN
IGNEOUS ROCKS
Basic dikes
Thrust fault
T., overthrust side
Normal fault
D., downthrown side
K
F
Direction of movement on thrust faults bordering on klippes and fensters
BEEKMANTOWN
PLATTSBURG
CUMBERLAND BAY
GRAND ISLE
SOUTH HERO
GEORGIA
MILTON
PERU
AUSABLE
CHESTERFIELD
COLCHESTER
MALLETTS BAY
CORLEAR BAY
SOUTH BURLINGTON
WILLISTON
RICHMOND
WILLSBORO
SHELBURNE
ST. GEORGE
HINESBURG
CHARLOTTE
ESSEX
WHALLON BAY
MONKTON
STARKSBORO
FERRISBURG
LAKE CHAMPLAIN
HUNTINGTON
73°30'
44°45'
73°00'



Base from U. S. Geological Survey topographic maps

Geology by Wallace M. Cady. Surveyed in 1934-1941

AREAL GEOLOGY AND STRUCTURE OF WEST-CENTRAL VERMONT