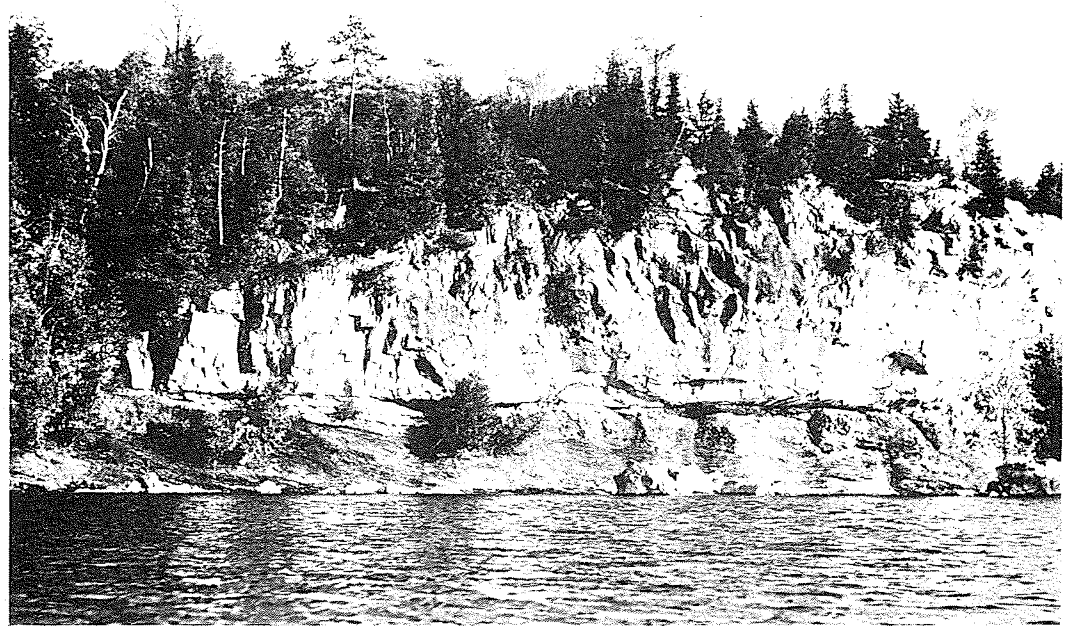
New York State GEOLOGICAL ASSOCIATION Guidebook to Field Excursions



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COVER PHOTOGRAPH:

Champlain Thrust at Long Rock Point, Burlington, Vermont. Lower Cambrian Dunham Dolomite overlies the Middle Ordovician Iberville Formation.

Photo courtesy of Vermont Geological Survey

GUIDEBOOK

to

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at the

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Stockton G. Barnett, Editor

Host:

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State University College

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PREFACE

The papers presented in this Guidebook bring together many of the approaches that have recently been utilized in geological investigations in northeastern New York and northwestern Vermont. The variety of trips offered certainly demonstrates that a large number of worthwhile projects have been carried out in this area in recent years. Fortunately, many of the people involved in these investigations are represented in this Guidebook.

Due to the short time available for preparation of this Guidebook authors were not able to see proofs of their articles.

The Editor wishes to express his appreciation to Miss Muriel Burdeau of the secretarial staff of the State University College at Plattsburgh for typing the final copy of many of the Guidebook articles.

Mr. Sherwood Keyser of the Office of College Publications arranged for the printing of this volume and provided valuable technical advice.

Stockton G. Barnett
Editor

CONTENTS

Trip A

SEDIMENTARY CHARACTERISTICS AND TECTONIC DEFORMATION OF MIDDLE AND UPPER ORDOVICIAN SHALES OF NORTHWESTERN VERMONT NORTH OF MALLETTS BAY

by

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INTRODUCTION

The central lowland of the Champlain Valley is underlain by Cambrian and Ordovician sedimentary rocks, bordered on the west by the Adirondack Mountains of Precambrian crystalline rock upon which Cambrian sandstone lies unconformably, and against which sedimentary rocks have been dropped along normal faults. The lowland is bordered on the east by low-angle thrust faults on which massive dolomite, quartzite, and limestone, as old as Lower Cambrian, from the east over-rode weaker Ordovician shale and limestone. The westernmost thrusts, the Highgate Springs thrust in the north, and the overlapping Champlain thrust in the south, trace an irregular line a few feet to 3 1/2 miles inland from the east shore of Lake Champlain. For most of the distance between Burlington and the Canadian border, the high line of bluffs marking the trace of the Champlain Thrust are composed of the massive, Lower Cambrian Dunham dolomite.

The shales, youngest rocks of the autochthonous lowland sequence, outcrop on most of the islands in Vermont, and the mainland between the thrusts and the lake. Although exposures on almost continuous shore-line bluffs are excellent, there are few outcrops inland because of glacial cover and low resistance of the shales to weathering. Fossils are rare in the older calcareous shale (Stony Point) and absent in the younger non-calcareous shale (Iberville). The lithic sequence was established almost entirely on structural criteria. Where it can be found, the Hathaway submarine slide breccia structurally overlies the Iberville.

DESCRIPTION OF FORMATIONS

Glens Falls Limestone

Kay (1937, p. 262-263) named the lower Glens Falls the Larrabee member, found it to be 72 feet thick on the Lake Champlain shore in the northwestern part of South Hero Township, Vermont, and to be composed there of thin-bedded, somewhat shaly limestone. Fisher (1968, p. 27) has found the Larrabee member to be 20 to 30 feet thick in the vicinity of Chazy, N.Y., and to be coarse-grained, medium- to thick-bedded light gray limestone full of fossil debris (brachiopods, crinoids, pelecypods, and trilobites).

The upper Glens Falls was named the Shoreham member by Kay (1937, p. 264-265), and described as the zone of Cryptolithus tesselatus Green, a distinctive trilobite. He found 30 feet of the Shoreham exposed in the lakeshore in northwestern South Hero Township. Fisher (1968, p. 28) prefers to call this the Montreal limestone member, following Clark's usage for the Montreal area (1952), and has described it as medium dark gray to dark gray argillaceous limestone with shale partings, fossiliferous with trilobites, brachiopods, molluscs, and bryozoa. He estimates it to be 150-200 feet thick in Clinton County, N.Y.

Cumberland Head Formation

The "Cumberland head shales" was a term used, but not carefully defined by Cushing (1905, p. 375), referring to the interbedded shale and limestone forming a gradation between the Glens Falls and the overlying Trentonian black shales. Kay (1937, p. 274) defined it as "the argillaceous limestones and limestone-bearing black shales succeeding the lowest Sherman Fall Shoreham limestone and underlying the Stony Point black shale". He measured 145 feet on the west shore of South Hero Island, Vt., just south of the Grand Isle-South Hero town line. The lower 30 feet have 8- to 12- inch beds of gray argillaceous limestone interbedded with dark gray calcareous shale. Above that the shale is predominant, but limestone beds are abundant, 3 to 12 inches thick with undulating surfaces, interbedded with half-inch to 12-inch layers of black calcareous shale. Less than one third of the Cumberland Head has more than 50 per cent shale, and about half has more than 60 per cent limestone beds. Some units as thick as 15 feet have 80 per cent limestone beds. The proportion of shale increases gradually but not uniformly upward.

of silt and argillaceous material in harder argillaceous limestone varies greatly. Intricate, fine, current cross-bedding occurs in four thin zones, indicating currents flowing northeastward.

Above this zone rich in laminated argillaceous limestone the proportion of calcareous shale increases, and 239 feet near the top of the Stony Point is composed entirely of calcareous shale. This shale section, 1.4 miles S 37° W from Long Point, North Hero, Vt., is assumed to represent the uppermost part of the Stony Point because it lies on the nose of a long, northeastward-plunging anticline between a thick argillaceous limestone section to the southwest, and a large area of Iberville shale to the north and northeast.

In this field area it is not possible to measure the entire thickness of the Stony Point, but from piecing together several measurable sections a minimum thickness is 874 feet. The total thickness is estimated to be 1000-1500 feet, allowing for probable thicknesses that could not be measured in the middle and upper parts of the Stony Point (Hawley, 1957, p. 83). In the log of the Senigon well near Noyan, Quebec, about 4 miles north of the international boundary at Alburg, shale apparently equivalent to the Stony Point is 924 feet thick (Clark and Strachan, 1955, p. 687-689).

Iberville Formation

The Iberville formation was named by Clark (1934, p. 5) for its wide outcrop belt in Iberville County, southern Quebec, about 10 miles north of the international boundary at Alburg, Vt. Clark (1939, p. 582) estimated the Iberville to be 1000-2000 feet thick in its type area.

The base of the Iberville has a gradational contact and was chosen on the basis of lithic criteria by which it can be most easily distinguished from the Stony Point. The Stony Point is entirely calcareous shale and argillaceous limestone with occasional beds of light-olive-gray weathering, dark gray fine-grained limestone. Above the lower transition section, the Iberville is composed of interbeds of medium to dark gray noncalcareous shale (1-12 inches, usually 2-4 inches), moderate-yellowish-brown weathering, dark gray laminated dolomitic siltstone (one quarter inch to 10 inches, usually 1/2 - 1 1/2 inches), and occasionally moderate-yellowish-brown weathering, dark gray fine-grained dolomite. The most conspicuous change from the Stony Point is the appearance of the yellowish-brown weathering dolomite beds, and the

Stony Point Formation

The Stony Point shale was defined by Ruedemann (1921, p. 112-115) as "hard, splintery dark bluish-gray calcareous shale" at Stony Point, 1 1/2 miles south of Rouses Point, N.Y., on the west shore of Lake Champlain, and correlated faunally with upper Canajoharie shale of the Mohawk Valley (Middle Trentonian).

The base of the Stony Point is exposed on the lake shore 0.55 miles south of the breakwater at Gordon Landing, the eastern end of the Grand Isle-Cumberland Head ferry. Deposition was continuous from the Cumberland Head up into the Stony Point, and the contact is somewhat arbitrarily chosen where the proportion of shale increases upward, and the wavy, irregular limestone bedding of the Cumberland Head gives way upward to smooth, even limestone beds of the Stony Point. The 215 feet of Stony Point formation exposed here is interbedded dark gray calcareous shale with light-olive-gray weathering, dark gray fine-grained limestone in beds of 1 to 12 inches, about 70 per cent shale. Two units about 9 feet thick are about 80 per cent limestone beds.

The thickest and least deformed measurable section of Stony Point begins 0.6 mile north of Wilcox Bay and extends for 1.8 miles northward along the shoreline bluffs of northwestern Grand Isle (Hawley, 1957, p. 59, 87-89). In this section of 635 feet, there are a few gross vertical lithic variations which are recognizable throughout this field area. Above the lower 215 feet, as described above, the percentage of calcareous shale decreases upward. Olive-gray to light-olive-gray weathering, dark gray argillaceous limestone appears in increasing proportion through the upper 400 feet of this section, where the percentages are: argillaceous limestone, commonly silty, 66 per cent; calcareous shale, 29 per cent; fine-grained limestone beds, 5 per cent.

The argillaceous limestone commonly occurs in thin, even beds (one quarter to three quarters of an inch) with fine lighter- and darker-gray laminae, but occasional beds reach 10 inches. Thicker-bedded zones suggest cyclic deposition: from calcareous shale (1 to 4 inches) upward through 5 to 6 inches of laminated argillaceous limestone, to a 1- to 3-inch bed of fine-grained limestone; then through 4 to 5 inches of argillaceous limestone to 1 to 4 inches of calcareous shale. Where the interval between calcareous shale beds is thinner, the fine-grained limestone bed in the middle is missing. The proportion noncalcareous shale which is more brittle and more lustrous cleavage surfaces than the calcareous shale. The transition section is at least 72 feet thick at Appletree Point in northern Burlington (Hawley, 1957, p. 64), and may be as thick as 200 feet. A section from Stony Point to Iberville is almost continuously exposed, though somewhat deformed, along the lakeshore southeastward for a half mile from Kibbee Point, in northeastern South Hero Township, Vt. The base of the Iberville is defined as the first appearance of the noncalcareous shale and dolomite beds.

Iberville shale and dolomitic siltstone show remarkable rhythmic bedding. The base of each cycle is a sharp contact, sometimes a slightly scoured surface, on which a thin bed (0.25-0.75 inch) of yellowish-brown weathering, dark gray laminated dolomitic siltstone was deposited. The typical siltstone layer becomes finer-grained upward with decreasing quartz and dolomite, and increasing argillaceous and carbonaceous material, and grades into dark-gray noncalcareous thin-cleaving shale (1-4 inches). Usually at the top is an eighth to three quarters of an inch of grayish-black shale interlaminated with the dark gray. Occasionally the dolomitic siltstone may be missing at the bottom of the cycle, or the grayish-black shale laminae missing at the top. Ripple-drift cross-lamination is a common feature of the dolomitic siltstone layers. In some beds only a single storey of ripples were built, but in others down-current ripple drift continued long enough to form two, and occasionally three or four tiered beds. Current directions indicated by the ripple cross-lamination are invariably southwestward in the Iberville, in contrast to northeastward in the Stony Point.

Six thicker (5-10 inches) non-laminated graded siltstone beds with 1 mm.-long shale flakes in their lower parts are found on northeastern Burton Island, southwest of St. Albans Point. They grade finer upward, and some are laminated above the lower third. One has large (5 by 1 1/4 inches is the largest) angular shale fragments in the mid-portion. They commonly have contorted lamination in the middle, above which lamination is more marked, and they are topped with drift ripples grading upward into shale.

The thickest measurable sections of the Iberville are 732 feet, with an estimated 2200 depositional cycles, on the west side of Woods Island, and 304 feet with an estimated 1215 cycles on Clark Point, southwestern Hog Island, West Swanton, Vt. The cyclic character of the Iberville layers, the graded beds, graded laminated beds, and convolute laminae, are all characteristic of sedimentation by turbidity currents (Kuenen, 1953; Bouma, 1962, p. 48-54).

Hathaway Formation

The Hathaway formation, named for Hathaway Point on southeastern St. Albans Point, Vt. (Hawley, 1957, p. 68), designates argillite and bedded radiolarian chert, commonly intensely deformed, with included small fragments to large blocks of quartz sandstone, coarse graywacke, dolomite, limestone, and chert. Some fragments strongly resemble dolomite and dolomitic siltstone beds of the underlying Iberville, but the coarse sandstone, chert and graywacke are unlike any strata in the autochthonous formations of the Champlain lowland. Where the Hathaway and Iberville are in contact or close proximity, there is marked disparity in intensity and nature of their deformation. The Hathaway appears to have deformed by flowage without the development of good cleavage, commonly with disintegration of less mobile beds into blocks and boulders. The Iberville has undergone much less intense folding and faulting,of a type normally associated with the regional structure. For these reasons, the Hathaway is inferred to be a submarine slide breccia initially deformed while its muddy constituents were still soft.

The best accessible exposures of the Hathaway are on Hathaway Point, and extending north for 1200 feet from Beans Point on the east shore of the lake, in northwestern Milton Township, Vt. As fate would have it, the most impressive and extensive exposures of the Hathaway are on Butler Island, between St. Albans and North Hero, accessible only by boat. Almost all of Butler Island is composed of the Hathaway, which is usually a mashed, streaky light and dark gray argillite with inclusions of dolomite, dolomitic siltstone, and occasionally black chert and graywacke, from 1 by 2 to 8 by 24 inches. On the southeast side of Butler Island are found the largest inclusions in the Hathaway: blocks of dolomitic siltstone up to 3 by 20 feet, and coarse-grained graywacke up to 15 by 50 feet. Argillite foliation wraps around these blocks, and around innumerable smaller pebbles and boulders. Hawley has described in detail these and other localities (1957, p. 68-75).

SUMMARY OF DEPOSITIONAL HISTORY

The fossiliferous limestones of the Glens Falls and older formations in this area indicate rather shallow, clear-water carbonate deposition, often in an environment of considerable wave and current turbulence (reefs, coarse calcarenites, and cross-bedding in the upper Chazyan). In the Cumberland Head formation fossils are much scarcer and there is a transition from the shallow water carbonate environment to a muddier, deeper water depositional environment. The lower two hundred feet of the Stony Point is 70 per cent calcareous shale, and the next 400 feet is laminated argillaceous limestone (66%) interbedded with calcareous shale (29%) and hard, purer fine-grained limestone (5%) in a somewhat cyclic pattern. Current cross-lamination indicates flow toward the northeast. The complete absence of primary structures associated with shallow water, and the fine lamination of the argillaceous limestone, and the paucity of fossils, suggest a deeper, quieter, muddier depositional environment.

Through the lower hundred feet (or more) of the Iberville, a marked change in the character of the rock appears with dolomite replacing limestone as the hard, fine-grained interbeds, and noncalcareous shale replacing the calcareous shale of the Stony Point. At some unknown distance above the base, a section of at least 730 feet shows cyclic interbedding of noncalcareous shale and graded, laminated dolomitic siltstone commonly with current cross-lamination. The currents flowed toward the southwest. This suggests the changed character of the rock is at least partly the result of a change from a westward source of sediment (for the Stony Point), to an eastward source for the Iberville, and that turbidity currents dominated the depositional character of the Iberville. Uplift of deep sea bottom east of the Champlain Valley in late Mohawkian and early Cincinnatian time could have provided the new source of sediment and the westward slope down which the turbidity currents flowed. Some simultaneous deepening of the Champlain Valley region also occurred.

The Hathaway formation, composed of argillite and bedded radiolarian chert, chaotically deformed, with included masses of limestone, dolomite, dolomitic quartz siltstone and sandstone, coarse graywacke, and chert, is interpreted as a submarine slide breccia. Some of the types of inclusions, particularly the graywacke and chert, are unknown in autochthonous underlying formations of the Champlain Valley, nor in regions to the south and west. The slide (or slides?) seem to have come from the east, down the slope suggested by the direction of flow of turbidity currents which deposited sediment in the Iberville. The Taconic orogeny was occurring at this time, and some believe that the major thrusts of western and northwestern Vermont accompanied this orogeny. If this be true, thrust fault escarpments on the sea bottom to the east of the Champlain Valley could account for the slides and the assemblage of inclusions in the Hathaway. Earthquakes associated with the Taconic orogeny may have triggered the turbidity currents of the Iberville.

TECTONIC DEFORMATION

The shales are complexly folded and sheared, with fold axes trending a little east of north in the southern part of the area, and swinging more toward the northeast (N 20° - 30° E) in the north. Although elongate narrow belts of intense deformation parallel fold trends, separated by broader belts of more gentle folding, general intensity of deformation increases toward the Champlain and Highgate Springs thrusts. In areas underlain by shale, particularly in North Hero and Alburg, the topographic "grain" of long, low hills accurately reflects the trends of fold axes. From Grand Isle northward the smaller folds plunge northward and southward, but the pattern of structural elements and formational boundaries indicates the northeastward plunge is more prevalent and perhaps a bit steeper. The area might be visualized as having northeastward trending folds imposed on an eastward regional dip, though there are many individual exceptions to this generalized picture.

Fracture cleavage is nearly everywhere present in the more argillaceous beds of the Stony Point and Iberville formations. The term is used here as defined by Swanson (1941, p. 1247), "the structure is due to closely spaced planes of parting a certain small distance apart," and "as a rule it is possible to see that the rock between the planes of parting . . . has no structure parallel to them, or at most any parallel structure is confined to a thin film along the parting planes." In these shales, cleavage planes are more closely spaced in belts of intense folding, and, under the same structural conditions, they are more closely spaced in more argillaceous beds than in more calcareous beds. Fracture cleavage plates in the argillaceous limestone of the Stony Point formation commonly range from one half inch to 5 inches thick. Fracture cleavage in calcareous shale is finer, and in the noncalcareous shale of the Iberville the planes are so close as to resemble flow cleavage (Swanson, 1941, p. 1246), but in thin section cut perpendicular to the finest cleavage, it is seen to be composed of somewhat irregular and discontinuous joint-like fractures 0.02 to 0.06 mm. apart. Bedding displacements of 0.01 to 0.04 mm. occur along the cleavage planes (Hawley, 1957, p. 82).

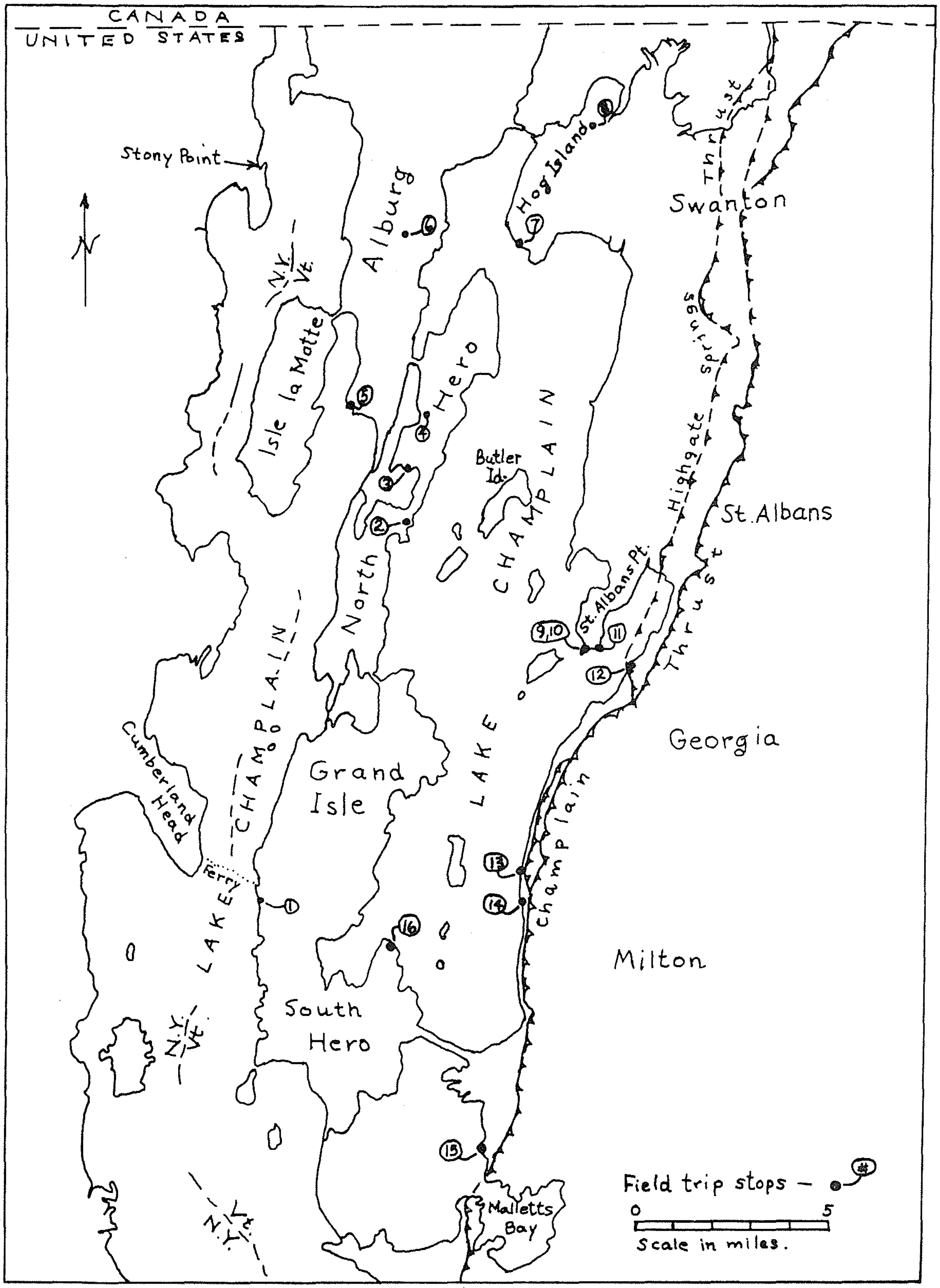

Figure 1. Trip A Stops

Although innumerable faults cut the shales, only a few displace them enough to juxtapose different formations. On most faults the rock of both walls is so similar that only minor displacement can be assumed. Block faulting typical of the western and southern Champlain Valley is distinct only in the older Trenton, Chazy, and Canadian formations of western South Hero, where Kay and his former students have mapped them (personal communication). Shear along bedding surfaces, cleavage surfaces, and at varying angles to both is very common. In more intensely folded belts, multiple shears occur along crests and troughs of folds. The bearing of slickensides is remarkably constant, regardless of the attitude or type of surface on which movement occurred. Of 119 slickensides bearings measured in this area, only three lay outside the arc between N 25 W and N 85 W (Hawley, 1957, p. 81).

FIELD TRIP STOPS

The best exposures of the shales and limestone are along the lakeshore bluffs. During the spring months the lake may be as much as five feet higher than normal, and many of these exposures may be inaccessible. They are listed for those who use the guidebook in the future, but some may have to be omitted on the initial field trip. THE STOPS ARE ON PRIVATE LAND. PERMISSION HAS BEEN OBTAINED FOR THE STOPS WE WILL VISIT. THOSE WHO MAY WISH TO VISIT THESE LOCALITIES IN THE FUTURE SHOULD GAIN PERMISSION FOR EACH VISIT. GIVE GEOLOGY A GOOD NAME BY BEING VERY THOUGHTFUL.

Stop 1. West shore of South Hero Island, extending for one mile southward from the breakwater at Gordon Landing. The lower 215 feet of the Stony Point formation is exposed between the breakwater and the top of the Cumberland Head formation, 2900 feet to the south. In the next 2300 feet of shoreline, the upper 145 feet of the Cumberland Head formation is exposed. These sections are described in the text article. The south end of this section is cut off by a right lateral wrench fault striking N 59° W, dipping 79° NE. South of the fault the interbedded limestone and shale (about 79% ls., 30% sh.) have been mapped as the Shoreham member of the Glens Falls formation (Erwin, 1957) on the basis of lithology and the presence of *Cryptolithus*.

Stop 2. Road cut on east side of U.S. 2 halfway between City Bay (North Hero Beach roadside park) and Carrying Place. This outcrop shows the interbedded laminated argillaceous limestone and calcareous shale typical of the middle section of the Stony

Point formation. It lies close to the axis of a major, northeastward plunging anticline.

<u>Stop 3.</u> Middle point on north side of Carry Bay, North Hero, 2000 feet east of Blockhouse Point. Typical Iberville cyclic bedding is exposed for about 1500 feet along this shore, extending eastward from the place where the access road meets the shore. From west to east are: an asymmetrical syncline, an asymmetrical anticline , and to the east of a covered interval is the east, overturned limb of a large syncline. These folds are in the axial area of a large, northeastward plunging, overturned syncline. Relationships of cleavage to bedding, axial surfaces, and direction of plunge are well shown. Small-scale current cross-lamination on some beds indicates southwestward flow.

<u>Stop 4.</u> Quarry in Iberville (mislabelled "gravel pit" on No. Hero Quad. map), 1.6 miles S 10° E from east end of North Hero-Alburg bridge. The beds are almost flat-lying, and only about 15 feet of section is exposed, but it is typical cyclic deposition, and the details are well shown, as described in the text.

<u>Stop 5.</u> Unnamed promontory 1300 feet WNW of Coon Point, south Alburg. Upper-middle Stony Point beds measuring 258 feet (Hawley, 1957, p. 60, 89-91), with the base of the section on the southern tip of the point. It is composed of: olive-gray weathering, dark-gray argillaceous limestone, frequently silty, with light olive-gray weathering bands and laminae, in units of 3 inches to 40 feet, making up 96 percent of the section; light olive-gray weathering, dark-gray fine-grained limestone in beds of half an inch to 12 inches, making up 4% of the section. There are four thin (1/4 inch to 1 1/4 inch) beds of medium light-gray weathering, medium gray fine-grained crystalline limestone. Pyrite concretions are common. Trilobites (<u>Triarthrus becki</u>) are fairly common on a few bedding surfaces, and a few unidentifiable graptolite fragments were found. This stop is relatively inaccessible, may be cut off by high water, and may have to be omitted.

<u>Stop 6.</u> A small quarry, 3.3 miles north of Alburg-North Hero bridge on the west side of highway U.S. 2. Stony Point argillaceous limestone-rich section as described for Stop 5, including a fair scattering of <u>Triarthrus becki</u>.

<u>Stop 7.</u> Upper Iberville beds on the south and west sides of Clark Point, at the south end of Hog Island (Vt. Hwy. 78 crosses

Hog Island), west Swanton. A 304-foot section has been measured here (Hawley, 1957, p. 91-92). It is 97.8 per cent thin-cleaving noncalcareous shale (1 to 3 inch beds) with thin laminated dolomitic siltstone (1/4 inch to 2 inch beds, occasionally thicker) and homogeneous fine-grained dark gray dolomite, weathering yellowish-brown, in beds up to 14 inches, at intervals of 3 to 50 feet. Cyclic deposition is prominent here, with an estimated 1215 cycles in this section. The thin dolomitic siltstone beds commonly show cross-lamination and ripple drift, with southwestward currents indicated.

Stop 8. Upper Iberville beds in quarry (mislabelled "sand and gravel pit" on East Alburg Quad. map) 1800 feet north of Vt. Hwy 78 and 600 feet west of Campbell Road, northern Hog Island, west Swanton. The rock is similar to that described for Stop 7. The quarry exposes an overturned anticline, thrust faulted on the upper, eastern limb, with adjacent syncline immediately westward, also faulted.

Stop 9. Southernmost tip of St. Albans Point, on property of Camp Kill Kare. Northeastward plunging asymmetrical anticline with linked small syncline northwest of it, in Iberville noncalcareous and calcareous shale with dolomitic interbeds.

Stop 10. Between Camp Kill Kare's access road and the lake, about halfway between the private cottages and the Camp buildings. There are 31 feet of white weathering, grayish-black chert in beds of 2 to 6 inches, dipping steeply (69°) southeastward on the southeast flank of the anticline at Stop 9. Structurally overlying the chert beds is black siliceous argillite in which bedding is not apparent because of its irregular, chippy foliation. The argillite contains rounded pebbles (avg. 1 by 2 inches) of gray dolomite and fragments of chert. Some graptolites were found in the argillite, but smearing precluded identification. This is part of the Hathaway formation. It is likely that the chert beds here represents a larger mass involved in a submarine slide.

Stop 11. Hathaway Point, at the south end of St. Albans Point. This is the type locality for the Hathaway formation. It has a matrix of pale-greenish-yellow weathering rock seen on a polished surface to be composed of small, irregular, curdled masses of greenish-gray to olive-gray argillite. Streamed and isoclinally folded in the matrix is black siliceous argillite similar to that associated with the chert beds at Stop 10. "Floating" in the matrix are small masses of grayish-black radiolarian chert which are commonly angular, as well as masses of bedded chert measurable in tens of feet. Fragments of dolomite and dolomitic siltstone occur in the western part of the Hathaway point exposure. Numerous slickensided tectonic shears are present in a variety of orientations. One 40-foot wedge between shears is composed of isoclinally folded calcareous and noncalcareous shale with occasional boudinaged masses of fine-grained limestone, resembling the transition beds at the base of the Iberville. Both of the islands east of Hathaway Point, in the middle of the bay, are composed of chaotically deformed argillite and chert. It is assumed that St. Albans Bay may lie over a deep synclinorium.

Stop 12. Lime Rock Point, on the southeast side of St. Albans Bay. At the base of the bluff composed of the Beldens (Upper Canadian) crystalline limestone with buff-weathering dolomitic beds, there is a dramatic exposure of the Highgate Springs overthrust; Lower Ordovician Beldens Limestone over Upper Ordovician Iberville calcareous and noncalcareous shale with occasional beds of yellowish-brown weathering fine-grained dolomite and silty dolomite. At the base of the high, steep bluff about one half mile to the east is the Champlain overthrust, on which the lower Cambrian Dunham dolomite is thrust westward over the Beldens. South of Lime Rock Point the Highgate Springs thrust slice is overlapped by the Champlain thrust for two and a half miles. It reappears for four miles, and then disappears again under the Champlain thrust, southeast of Beans Point. This is as far south as the Highgate Springs slice can be traced.

Stop 13. Beans Point, east shore of lake in northwest Milton. The Hathaway crops out intermittently for 1200 feet north from Beans Point. This is in a zone of intense deformation close to the Highgate Springs thrust, the trace of which is covered, probably about 600 feet back from the shore. The base of the steep bluffs 2000 feet back from the shore marks the trace of the Champlain fault, on which lower Cambrian Dunham dolomite has been thrust over Beldens crystalline limestone and dolomite of the Highgate Springs slice.

The Hathaway is composed of boulders and fragments "floating" in mashed argillite. The argillite is mottled olive gray to dark greenish gray to greenish black. On a polished surface cut perpendicular to foliation the mottled colors are seen to represent original bedding which has been folded most intricately, and sheared with no development of slickensides or breccia. The small-scale shearing has completely healed, and some minute fold crests merge into the adjacent bed, a streaming of one bed into the next with no sharp boundary. Included in the argillite are rounded fragments of moderate-yellowish-brown weathering, dark gray fine-grained dolomite and cross-laminated dolomitic siltstone, sub-angular to rounded, up to 4 by 7 by 20 inches in size. The long axes of the boulders are approximately parallel, plunging about 55° toward S 45° E. Foliation causes the argillite to split into irregular tapered chips. Thirty-six feet of cover separates the north end of the Hathaway outcrop from cyclic-bedded upper Iberville which lies overturned, dipping 46° northeastward.

Stop 14. Camp Watson Point, 3/4 mile south of Beans Point (Stop 13). The core of a large, overturned syncline is exposed on the point, plunging 18° toward N 56° E. The overturned limb, dipping 29° southeastward, is exposed for 200 feet or more along the shore to the south. The rock is lower Iberville transition, with interbedded calcareous and noncalcareous shale, argillaceous limestone, and silty laminated dolomite.

Stop 15. Clay Point, between Malletts Bay and the Lamoille River, east shore of lake. THIS PROPERTY IS POSTED, AND PERMISSION MUST BE OBTAINED. In the transition beds in the lower Iberville (interbedded calcareous and noncalcareous shale, with argillaceous limestone, argillaceous dolomite, fine-grained dolomite, and silty-laminated dolomite with current cross-bedding) there is a small, overturned anticline cut by small thrust faults. The relationship of cleavage to bedding, plunge of the fold, identification of tops by cross-bedding, and the faulting make this a worthwhile stop for a structural geology class.

Stop 16. From Kibbee Point (northeastern South Hero) southeastward along the shore for 2500 feet, is exposed the transition from Stony Point to Iberville formations. With a few minor rumples the dip is southeastward all the way to a deep gully and small bay which separate a steep bluff-point to the east from the shore northwestward to Kibbee Point. This bluff, 2800 feet SE of Kibbee Point is composed of Stony Point argillaceous limestone and calcareous shale, overturned and dipping 55° southeastward. Thus, the gully conceals the faulted core of an overturned syncline. The fault is very likely a thrust, east side up.

West of the gully is Iberville, about 90% finely cleaved noncalcareous shale, with interbedded silty cross-laminated dolomite. Northwest from here to Kibbee Point the proportion of noncalcareous shale and silty laminated dolomitic interbeds decrease and the proportion of calcareous shale increases. About 220 feet southeast of Kibbee Point the southeastward-dipping beds are massive calcareous shale (Stony Point fm.). About 900 feet south of Kibbee Point on its west shore the Stony Point beds still lower in the section are predominantly argillaceous limestone, interbedded with calcareous shale.

REFERENCES CITED

Bouma, A.H., 1962, Sedimentology of Some Flysch Deposits: Amsterdam/New York, Elsevier Publishing Co., 168 p.

Clark, T.H., 1934, Structure and stratigraphy of southern Quebec: Geol. Soc. America Bull., v. 45, p. 1-20

------ 1939, The St. Lawrence lowlands of Quebec, Pt. 1 of Canadian extension of the interior basin of the United States: Geologie der Erde, v. 1, p. 580-588, Berlin, Gebruder Borntraeger, 643 p.

Clark, T.H., and Strachan, Isles, 1955, Log of the Senigon well, southern Quebec: Geol. Soc. America Bull., v. 66, p. 685-698

Cushing, H.P., 1905, Geology of the northern Adirondack region: N.Y. State Mus. Bull. 95, p. 271-453

Erwin, Robert B., 1957, The geology of the limestone of Isle La Motte and South Hero Island, Vermont: Vermont Geol. Survey Bull. 9, 94 p.

Fisher, Donald W., 1968, Geology of the Plattsburgh and Rouses Point, New York-Vermont, Quadrangles: N.Y. State Mus. and Science Service, Map and Chart Ser. No. 10, 51 p.

Hawley, David, 1957, Ordovician shales and submarine slide breccias of northern Champlain Valley in Vermont: Geol. Soc. America Bull., v. 68, p. 55-94

Kay, Marshall, 1937, Stratigraphy of the Trenton group: Geol. Soc. America Bull., v. 48, p. 233-302

Kuenen, P.H., 1953, Significant features of graded bedding: Am. Assoc. Petroleum Geologists Bull., v. 37, p. 1044-1066

Ruedemann, Rudolf, 1921, Paleontologic contributions from New York State Museum: N.Y. State Mus. Bulls. 227, 228, p. 63-130

Swanson, C.O., 1941, Flow cleavage in folded beds: Geol. Soc. America Bull., v. 52, p. 1245-1264

Trip B

RECENT SEDIMENTATION AND WATER PROPERTIES, LAKE CHAMPLAIN

by

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University of Vermont

INTRODUCTION

Purpose

Within the last few years a considerable amount of research has been done on Lake Champlain. The purpose of this trip is to show what type of work is being carried out on the lake and to report some of the findings.

Acknowledgements

We would like to thank Dr. Milton Potash who contributed data on general limnology, Dr. Philip W. Cook who provided information on the phytoplankton, and Dr. W. Philip Wagner who read the manuscript. Many graduate students at the University of Vermont have contributed data as follows: Sander Sundberg is acknowledged for chemical information and Carl Pagel for data on bottom animals. Thomas Legge is credited for zooplankton data and seasonal aspects of the lake. Richard Clement contributed the data used in compiling the Malletts Bay portion of the sediment map of Lake Champlain and Peter Townsend supplied the heavy mineral analyses. Much of the information on iron-manganese concretions is credited to David G. Johnson.

The work upon which this research was based was supported in part by funds provided by the U. S. Department of Interior as authorized under the Water Resources Research Act of 1964, Public Law 88-379.

PRESENT LAKE CHAMPLAIN

Lake Champlain is approximately 110 miles long and has a maximum width of approximately 12 miles, measured from the Little Ausable River, N.Y., to the shore of Malletts Bay, Vt. It has a mean elevation of 92.5 feet above sea level and a water surface area of 437 square miles (gross area 490 square miles). The lake has a maximum depth of 400 feet near Thompson's Point and a mean depth of approximately 65 feet.

The lake drains northward through the Richelieu River into the St. Lawrence River. The mean discharge of the 8,234 square miles of the Champlain basin at Chambly, Quebec, is 10,900 cubic feet per second. At its southern end, Lake Champlain is connected, via several locks, with the Hudson River through the Champlain Canal.

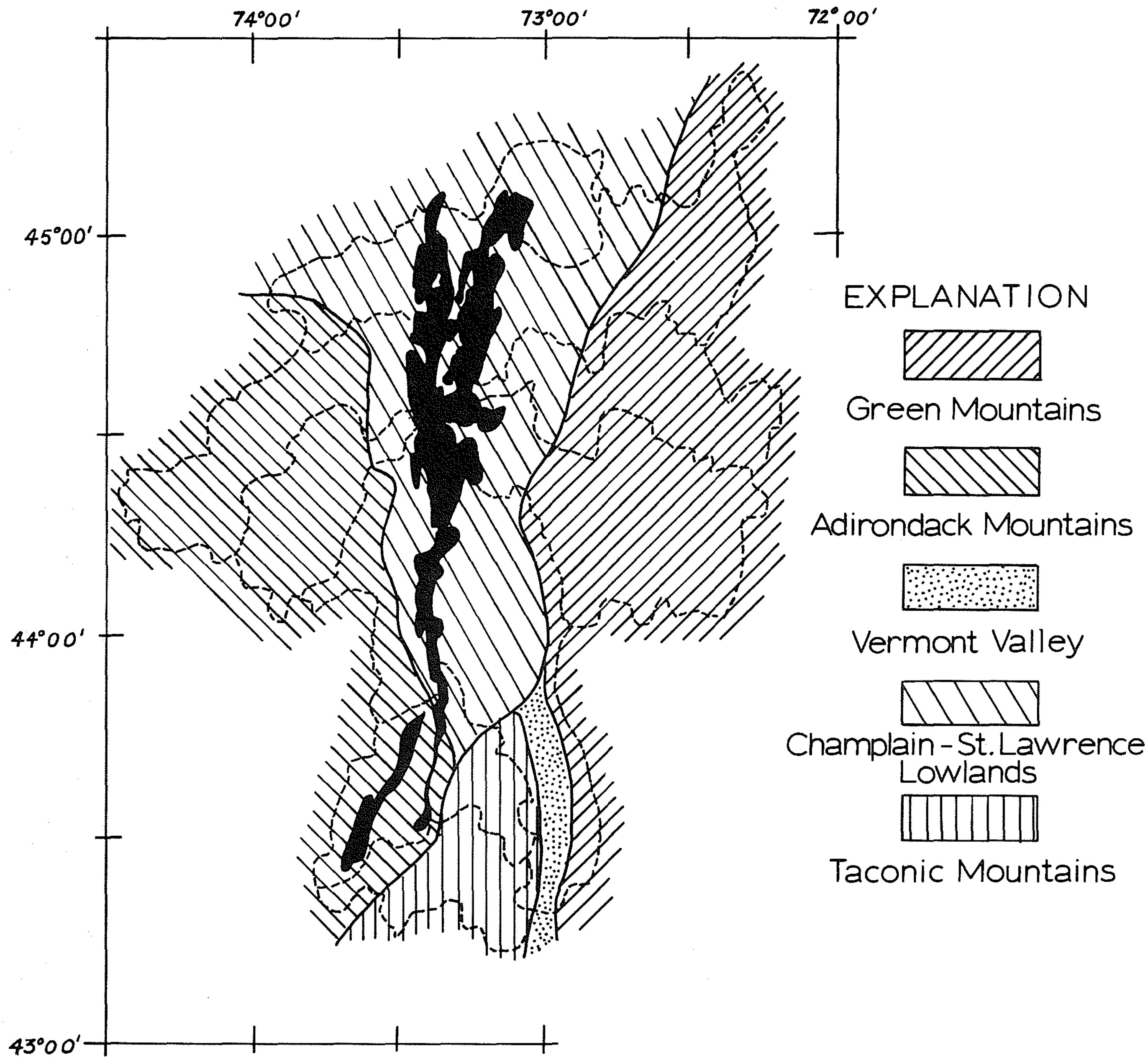

Figure 1 Morphological regions of the Champlain drainage basin. The dashed lines designate drainage sub-basins.

The Champlain Valley drainage basin (excluding area of the lake) is 7,744 square miles, of which about 34 percent is in New York to the west and 66 percent in Vermont and Quebec to the east and north. It may be divided into five morphological regions (Fig. 1). These include the Green Mountains, Adirondack Mountains, Vermont Valley, Champlain-St. Lawrence lowlands, and the Taconic Mountains. Twelve sub-basins have been recognized within the Champlain Valley drainage basin: Chazy, Saranac, Ausable, Bouquet, Lake George and Mettawee in New York; the Pike in Quebec; and the Missisquoi, Lamoille, Winooski, Otter, and Poultney in Vermont. The Missisquoi and Chazy also drain portions of Quebec (Hunt, Townsend and Boardman, 1968). Six tributaries, each with a catchment area greater than 500 square miles, drain 52 percent of the entire drainage basin. These are the Otter Creek, Winooski, Lamoille, Missisquoi, Saranac, and Ausable. As a first approximation the mean annual discharge into the lake from the drainage basin is 11,900 cubic feet per second. Given a mean discharge from the lake of 10,900 cubic feet per second and an estimated lake volume of 9.12×10^{11} cubic feet (Boardman, Hunt, and Stein, 1968), the mean retention period of the lake is slightly less than three years. Some additional hydrological data have been summarized in Figure 2.

	West Side	East Side	Total Basin
Catchment area:	2,618	5,126	7,744 sq. mile
Percent of total area:	33.8	66.2	100
Mean discharge/sq. mile:	1.327	1.639	1.523 cfs/sq mile
Calculated total discharge into lake:	3,474	8,402	11,876 cfs
Percent of discharge into lake:	29	71	100

Figure 2. Provisional summarized hydrological data for Lake Champlain

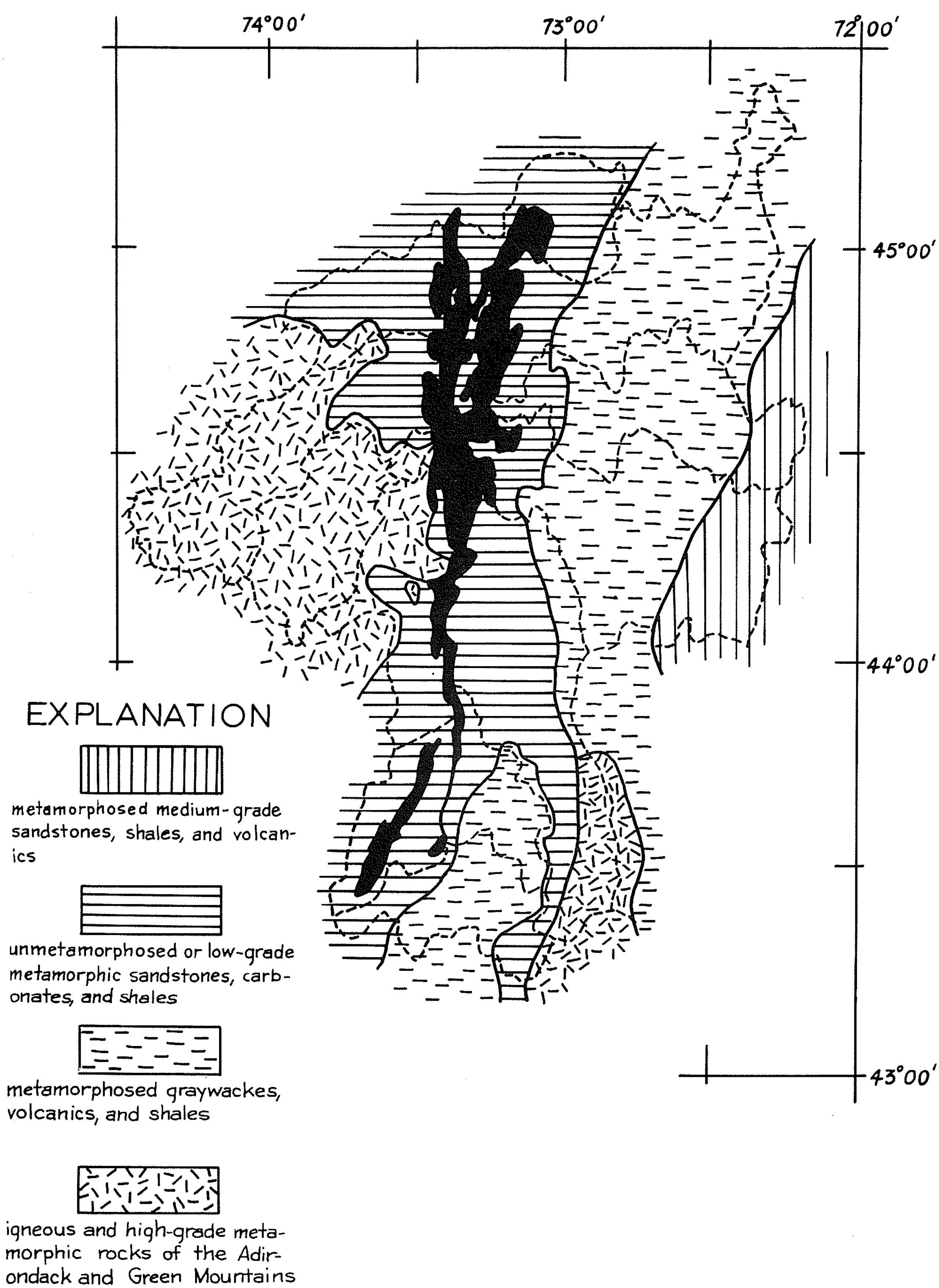

Figure 3 Major rock terrains of the Champlain drainage basins. The dashed lines designate drainage sub-basins.

The bedrock geology surrounding the lake basin is made up of a diverse assembledge of rocks (Fig. 3). High grade metamorphic rocks of the Adirondack Mountains, mantled by unmetamorphosed sandstones and carbonate rocks, border the western margin of the lake. Unmetamorphosed or low grade metamorphosed carbonates, sandstones, and shales border the eastern margin and presumably underlie a large portion of the lake proper.

GEOLOGICAL HISTORY

The recorded geologic history of the Champlain basin started in the lower Paleozoic when sediments were deposited in marine waters that invaded eastern North America. These sedimentary rocks, which consist of limestones, shales, and sandstones, form the present lake basin. Thrusting from the east during the Paleozoic brought more highly metamorphosed rocks, which define the eastern margin of the lake basin, into contact with the relatively undisturbed basin rocks. The elongate shape of the basin as well as the rapid change in the bedrock lithology across the lake suggest that faulting may have played a part in deepening the basin.

The history of the lake from the Paleozoic to the Late Pleistocene is not known, although for at least part of this interval the basin may have served as a river valley.

Possibly several times during the Pleistocene, ice occupied the valley. To date, however, no Pleistocene deposits earlier than Wisconsin have been identified in the Champlain basin. Evidence for glacial scouring may be found today in the lake's ungraded longitudinal profile and in basins more than 300 feet beneath sea level. The last lobe of ice is believed to have been present in the Champlain Valley during the Mankato substage (Stewart, 1961, p. 84). Although field studies are presently being conducted by Dr. W. Philip Wagner (see Trip D of this guidebook), the most complete study presently available of the Pleistocene history of Lake Champlain was done by Chapman (1937) and the résumé given here is taken from his work. Interpretations of the lake's Pleistocene history are based primarily upon the recognition of former lake levels, which today may be several hundred feet above sea level. The elevated shorelines are identified by finding ancient beaches, wave-cut and wave-built terraces, river deltas, and outlet channels. These features do not occur at random elevations, but form several planes which have been interpreted as ancient lake stands that formed as lake level temporarily remained stationary. Chapman recognized three water planes. Two of these end abruptly when traced northward through the Champlain Valley. The highest plane may be traced to Burlington and the middle plane to the International Boundary. The two higher planes presumably terminate because they formed in a water body which abutted against the retreating ice margin. The lake in which these planes formed has been given the name

Lake Vermont.[1] During the time when the highest plane recognized by Chapman was formed, which he called the Coveville stage, Lake Vermont drained southward through an outlet channel at Coveville, New York. The lower water plane, which also drained to the south, formed at a later time when a new, more northerly outlet of lower elevation developed near Fort Ann. The water level of Lake Vermont dropped about a hundred feet between the Coveville and the Fort Ann stage. After the ice lobe had retreated from the St. Lawrence Valley, the water level in the Champlain Valley dropped several hundred feet and was continuous with marine water in the St. Lawrence Valley.[2] No appreciable tilting of the basin occurred (Chapman, 1937, p. 101) from the time that the Coveville plane was developed until the invasion of marine water, correlated with Two Creeks interval (Terasmae, 1959, p. 335). Some time after the inundation by marine water, however, the northern portion of the valley began to rise more rapidly relative to the southern portion. In time, the Richelieu threshold just north of the International Boundary became effective in preventing marine waters from entering the valley and the existing fresh water lake developed. Future tilting of only four-tenths of a foot per mile would allow Lake Champlain to drain southward as during Lake Vermont (Chapman, 1937, p. 122). This is only a small fraction of the tilting which has taken place since the Champlain basin was inundated by marine waters.

WATER PROPERTIES

Temperature

The major portion of Lake Champlain can be considered to be a deep cold-water mesotrophic lake. Technically, it is classed as a dimictic lake (Hutchinson, 1957). This means that it has two periods during the year when the water in the lake is of equal temperature and is mixing. These periods of mixing are alternated by periods of thermal stratification.

Thermal stratification begins to develop in June, and the stratification is well established in July and August. During mid-summer the metalimnion is at a depth of approximately 15 meters and includes the

[1]Stewart (1961, p. 105) recognized a third water plane above the two that formed in Lake Vermont which he correlates with the Quaker Springs stage of Woodworth (1905, p. 193).

[2]MacClintock and Terasmae (1960, p. 238) have suggested that the St. Lawrence Valley was subject to subaerial weathering after lake clays were deposited, but before the deposition of marine sediments. If so, Lake Vermont drained when the Fort Covington ice dam broke and the invasion of marine waters had to await a eustatic rise in sea level.

12^{o} C - 18^{o} C isotherms. The period of summer stratification is short, for the depth of the thermocline increases steadily through August and September until the fall overturn takes place in October or November. Bottom temperatures in deep water remain at or about 6^{o} C during summer, but may rise to 12^{o} C at the onset of the fall overturn.

The waters in the southern end and in the northeastern region of the lake are somewhat warmer than in the main lake, and warmer water is found in the bays along both shores.

During the winter most of the lake freezes over, and inverse thermal stratification develops, with 4^{o} C water at the bottom and 0^{o} C water under the ice. Freezing begins in the narrow southern end, in the northern end, and in the northeastern portion of the lake. The wide main body of the lake is the last to freeze. In mild winters this portion may remain open throughout the winter season. The duration and intensity of the freeze depends on the severity of the winter.

Transparency

The transparency of the lake, as measured with a Secchi disc, ranges from about 3 to 6 meters. The deeper readings are encountered in late summer when algal growth is less. The disc reading in the southern part of the lake is usually less than 1 meter. Legge (1969) measured the penetration of light in the lake in 1966 and 1967, using a submarine photometer. Ten percent of incident light was usually found at a depth of 3 meters, 5 percent at 5 meters, and 1 percent at approximately 10 meters. The level of 1 percent incident light is therefore above the level of the thermocline.

pH and Alkalinity

Champlain is an alkaline lake. The pH of surface water is above 8.0, but in the deep water the pH may get as low as 7.3.

The total alkalinity in the main lake, predominantly as bicarbonate, ranges between 38 and 46 mg/l, and averages 41 mg/l. Alkalinity values are higher in the southern end of the lake, and minimal values are found in the water in the northeastern sector. Abnormally high values are sometimes encountered at stations close to shore, modified by tributary inflow. The alkalinity at Rouses Point, near the outlet of the lake, is actually less than that of the main lake.

Major Cations

The four major cations (Ca, Na, Mg, and K) have been measured in the lake with some thoroughness, and the results are summarized in Potash, Sundberg, and Henson, (1969a). In the main lake the concentrations of these four cations are ranked in descending order as Ca, Na, Mg, and K, with median values of 15.8, 3.9, 3.6, and 1.1 mg/l. In the southern part the descending rank order is Ca, Mg, Na, and K, with median values of 24.4, 5.8, 5.1, and 1.2. In flowing from the south to the central lake, the water is diminished in the concentration of all four cations, especially in magnesium. The concentrations in the northeastern region of the lake are significantly less than in the main lake. In this part of the lake the descending rank order is Ca, Na, Mg, and K, the same as for the main lake, but the median values are 13.2, 3.0, 2.9, and 1.3 respectively. It is suspected that these differences between the main lake and the northeastern portion of the lake are influenced, in part, by ground-water intrusion, while the differences between the main lake and the southern lake are a result of surface inflow.

Major Anions

The dominant anion in the lake water is the bicarbonate ion, which is mentioned under alkalinity. A few determinations have been made of the chloride and the sulphate ions. In the main lake the median concentration of sulphate is 15.4 mg/l, and of Cl, is 5.7 mg/l. The pattern for these anions is the same as for the cations; values are higher in the southern end of the lake, and lower in the northeastern part of the lake.

Dissolved Oxygen

The concentration of oxygen dissolved in the lake water is one of the more significant parameters measured in lakes; it is essential for respiration for all animals and most plants, it facilitates the decomposition of organic matter in the lake, and it serves as an index for the general quality of the lake water. The major sources of oxygen dissolved in the water are from exchange with the atmosphere and from photosynthesis by the plants in the lake. Oxygen is lost through respiration, decomposition, and increased temperature. The crucial test is the amount of oxygen in the deep water below the thermocline. In the deep water there is no source of new oxygen, and the supply that is there when stratification begins must last for the entire summer until the fall overturn mixes the water and carries down a new supply.

The trophic standard of a lake is sometimes measured by the concentration of dissolved oxygen in the deep water. In an oligotrophic lake the amount of organic material in the deep water during the period of summer stratification is of such small magnitude that oxygen consumed by decomposition has little effect on the concentration of oxygen in deep water. In a eutrophic lake, however, decomposition in deep water is great enough to reduce significantly the concentration of oxygen.

The main body of Lake Champlain is considered oligotrophic to meso-

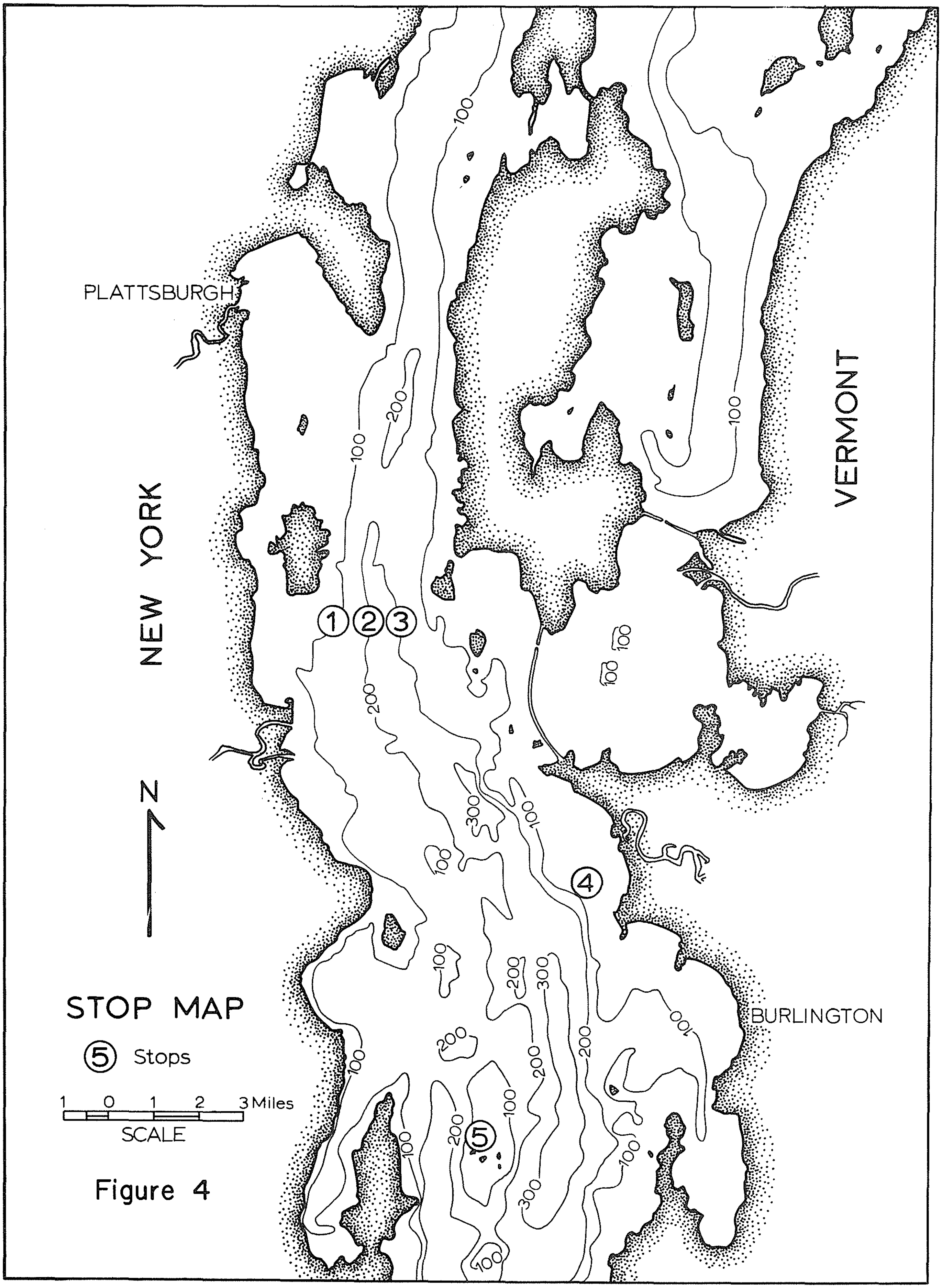
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Figure 4

trophic by the oxygen standard. The lake water was more than 90 percent saturated in April, 1967, after the break-up of the ice cover. The oxygen in deep water from August through October was slightly less than 80 percent of saturation.

In some sheltered areas of the lake, for example, Malletts Bay, deep-water oxygen may be reduced to less than 1 percent of saturation (Potash, 1965; Potash and Henson, 1966; Potash, Sundberg, and Henson, 1969b). These are considered to be eutrophic areas of the lake.

SEDIMENT PROPERTIES

In 1965 a reconnaissance survey was undertaken to determine the nature of bottom sediments in Lake Champlain. This work, which is summarized here, is preliminary to future studies relating to recent sedimentation and the geological history of the lake. Grab samples have been collected (with control maintained by shore-located transits) from predetermined stations located at the intersections of a half-mile grid overlay of the lake. To date, about 1800 samples have been analyzed for grain-size distribution (Fig. 4) and sediments from selected areas have been analyzed for other properties including organic percent, clay mineralogy, light-mineral and heavy-mineral composition, percent carbonate, and several other geochemical properties.

Although the sediments show laterally gradational boundaries, they have been separated here into six general types for purposes of discussion. These six general types have been defined primarily by the physical property of grain size, although most samples within a group have other mineralogical and chemical properties in common.

Gravels

Gravel deposits make up less than 5 percent of the sediments exposed on the lake bottom. Gravels occur primarily in shallow, nearshore water and at the mouths of rivers, although gravel-sized material is also known to exist within the pebbly sandy clay deposits discussed below. At some localities the mineralogy of the gravels suggests glacial origin but at other localities, the gravels probably were formed locally from the erosion of bedrock.

Sands

Sediments defined as sands constitute about 20 percent of the lake bottom. As do the gravel deposits, the sands typically occur near shore in shallow water. They are also found at the mouths of rivers lakeward from gravel deposits. The heavy-mineral suite of the sands reflects the igneous and metamorphic terraines which outcrop on either side of the lake. Muscovite and chlorite, for example, are minerals typically found on the Vermont side; whereas, garnet, hornblende, and hypersthene characterize the heavy minerals of Lake Champlain on the New York side. The organic percent of sands (estimated through loss by ignition) is low and rarely exceeds a few percent; the carbonate percent is equally low.

Organic Muds

Muds cover about three-quarters of the lake bottom. Those which are high in organic content (5-20 percent) are referred to here as organic muds. The surface of the organic muds is a grayish brown - reddish brown hydrosol. The reduced (unexposed) sediment is dark gray in color. The organic muds are most common in deeper water (greater than 50 feet) where wave and current action are at a minimum, where fine particulate matter can accumulate, and where the oxidation of organic matter is low. Such areas occur primarily offshore, lakeward from gravel and sand deposits. Core samples have shown organic muds to be quite uniform in grain size and are generally without lamination or structures, although carbon smears and mottling do occur.

Sandy Clays

Sandy clays are found only in a few isolated areas of the lake basin today. They occur mostly on rises where erosion or non-deposition is taking place and in deeper portions of the lake basin far from shore where sedimentation rates are low. Wet, sandy clays are gray or brown to yellow-brown in color, depending upon their state of oxidation. The difference in color between the surface and the unexposed sediment is not as striking as has been found in organic muds. The water-sediment boundary is not a hydrosol and the deposit is so well-compacted that it can be penetrated with a corer only with difficulty. Sandy clays are very poorly sorted and their organic content, which averages no more than a few percent, is much lower than that found in the organic muds. An eight-food core taken on the rise north of the Four Brothers Islands penetrated clay, clay containing shale fragments, and at the bottom broken shale fragments (beneath which presumably was bedrock). Present evidence indicates that the sandy clays may be continuous with the sediments described below.

Pebbly Sandy Clays

Sediments referred to here as pebbly sandy clays have been recognized within a trough, defined approximately by the 200 foot contour interval, which follows a sinuous course along the middle of the lake from Valcour Island to the vicinity of Burlington. The sediments are extremely poorly sorted and contain material ranging from clay size to cobbles several inches in diameter. The lithology of the gravel fraction is varied and is similar to that of a glacial till. In other properties studied, the sediment resembles the sandy clays. These deposits are surrounded by organic muds which contain no gravel-sized material and virtually no sand. It does not appear likely that this sediment, or even the gravel fraction alone, could be forming today by ice or water transport from shore across the organic mud facies. This suggests that the pebbly sandy clays are not contemporaneous with the organic muds but are probably tills or ancient ice-rafted deposits.

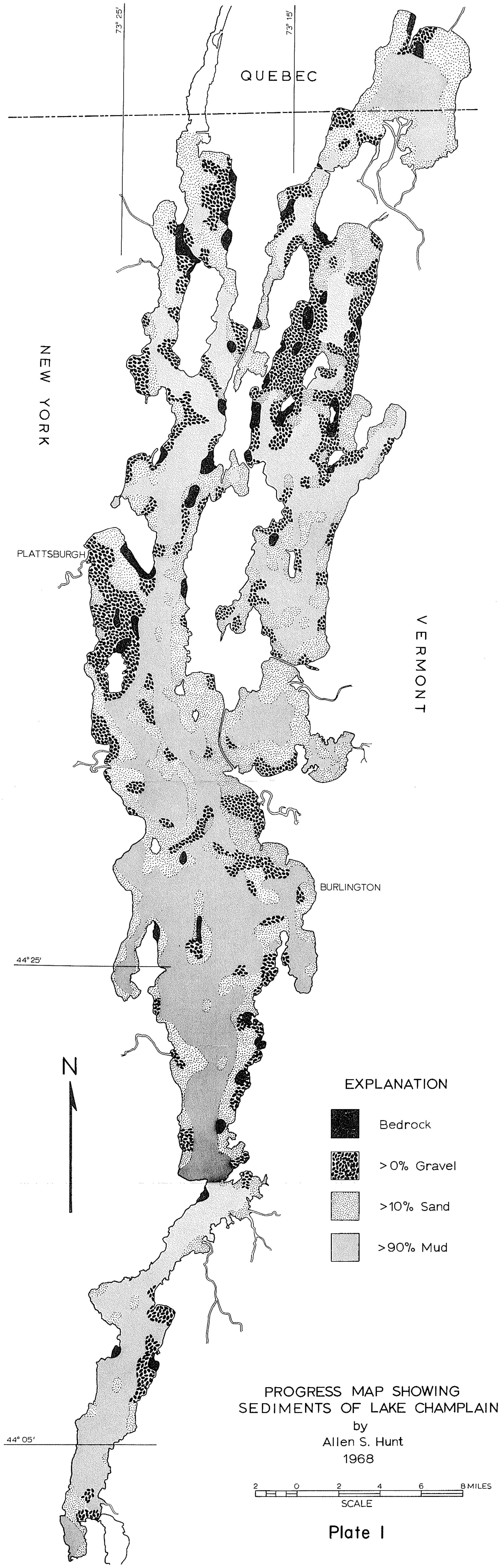
Iron-manganese Concretions

"Manganese nodules", first discovered during the 1966 field season, are now known to occur in several areas of Lake Champlain. David G. Johnson, graduate student at the University of Vermont, has been working on their origin and geochemistry. Only in the eastern part of the lake where the sedimentation rate is low are the concretions found in great concentration and with well developed concretionary structure. In other areas the concretions are mixed with a terrigenous matrix which comprises 90 percent or more of the sample. Rarely are they found at depths of more than 50 feet. At greater depths the concretions are found on slopes that adjoin shallow-water shelves. This suggests that they may have formed in shallow water and were subsequently redeposited by slumping. The shapes of the nodules range from spherical, to reniform, to discoidal. In diameter they vary from a few millimeters to greater than 10 centimeters. If well developed they reveal a concretionary structure which consists of alternating light brown and black bands, with a sand grain or rock fragment typically forming the nucleus. Chemical analysis of several dozen nodules indicates the composition to be about 10 percent manganese (MnO) with a range of 1 percent to 30 percent, and 40 percent iron (Fe_2O_3) with a range of 10 percent to 60 percent. The algae *Cladophora* sp. commonly is found attached to one surface of the concretion, indicating the orientation of the concretion on the lake bottom. Cores taken in nodule-occurring areas show that they are limited to the top 8-10 cm. of the sediment column. Scuba divers have observed a very thin veneer of silty material covering the concretions.

BIOLOGICAL ASPECTS

Phytoplankton

The phytoplankton is dominated by diatoms and blue-green algae. *Asterionella*, *Diatoma*, *Melosira*, and *Fragilaria* are dominant genera during the spring. *Ceratium* may become the dominant organism during midsummer and the late summer-autumn plankton is characterized by the abundance of *Tabellaria*, *Gomphosphaeria*, and *Anabaena*. Overall, the phytoplankton is characteristic of a mesotrophic lake. Muenscher (1930) described the algae of the lake for 1928.



PROGRESS MAP SHOWING
SEDIMENTS OF LAKE CHAMPLAIN
by
Allen S. Hunt
1968

Plate I

Zooplankton

Muenscher (1930) conducted a survey in 1928 and enumerated 9 genera of Cladocera, 4 genera of copepods and 15 genera of rotifers. Among the Cladocera, *Bosmina*, *Daphnia*, and *Diaphanosoma* were the most abundant and widely distributed. *Diaptomus* and *Cyclops* were the only ubiquitous copepods. *Dinobryon* was found to be the most common Protozoa. Legge (1969) has described the seasonal distribution of the calanoid copepods in the lake.

Benthos

The shallow-water (littoral) benthos consist of the usual communities of molluscs and insect larvae. The deep-water fauna in organic silt consists of small worms, the glacial relict shrimp *Pontoporeia*, small clams, and a larval chironomid.

Relict Pleistocene Fauna

The fauna of Lake Champlain includes several species that are considered to be relicts of the Pleistocene. Most of these animals are small invertebrates associated with the cold, deeper waters of the lake. They are mainly among the Crustacea. The schizopod species *Mysis relicta* (Opossum shrimp), a form common to the Atlantic Ocean, is found. Another inhabitant is the amphipod shrimp, *Pontoporeia affinis*, which was discovered in this lake only within the last five years. Both of these animals are common in the Great Lakes, but apparently are not very abundant in Lake Champlain. According to present thought these two species were able to inhabit the Pleistocene proglacial lakes and migrated from the Baltic Sea area during the Pleistocene Epoch, using a path around the Arctic Ocean, down through the Canadian chain of lakes, through the Great Lakes, to Lake Champlain (Ricker, 1959; Henson, 1966). Lake Champlain represents a terminus for these animals. *Pontoporeia* has been recorded from a lake in the State of Maine, but it has not been found north of the St. Lawrence River east of the Ottawa River. Presumably an ice block prevented their migration into this area of the continent. There are some other animals in the lake which also are considered to be glacial relicts. Among the small crustacean zooplankton would be included *Senecella calanoides*, which was first described from one of the Finger Lakes of New York, and *Limnocalanus macrurus*.

FIELD TRIP STOPS

Stop 1. This stop, along with stops 2 and 3, will constitute a west-east traverse designed to show differences in thermal patterns, benthos, and sediment types across the lake. At stop 1, a bathythermogram will be taken to demonstrate water temperature differences with depth. A grab sample will be taken and sieved for macro-organisms, and a core sample will be collected. The sediments at this stop are organic muds which contain no gravel, less than 10 percent sand, and equal amounts of silt and clay. The mean phi size is about 8, the standard deviation 2.5 phi.

Stop 2. A bathythermogram and core will be taken and a grab sample will be sieved for benthos. The sediment at this stop is described under "sediment properties" as a pebbly sandy clay. An average sample contains 25 percent sand and gravel, 10 percent silt, and 65 percent clay. The mean grain size is 8 phi, the standard deviation 4.5 phi.

Stop 3. A bathythermogram will be taken at this station to complete the traverse profile, the benthos will be sampled, and a plankton haul will be made. Water samples will be taken from selected depths and analyzed for alkalinity, pH, and dissolved oxygen. The sediments at this stop consist of organic muds. The mean grain size is 8 phi, and the standard deviation just over 2 phi. Sand makes up less than 5 percent of the sample with silt and clay equally divided among the remaining portion.

Stop 4. This is a shallow-water stop at the mouth of the Winooski delta. Grain-size analysis has shown sediments to be about 90 percent sand, 10 percent silt, and 1 percent clay. The mean grain size is 2 phi, the standard deviation just over 1 phi. Note the absence of an interface on the sediment surface.

Stop 5. This stop is to collect sandy clay sediment. The sandy clays have much in common with the pebbly sandy clays of stop 2 in that both are poorly sorted, and both have a high percent of sand and clay with only a small percentage of silt. The pebbly sandy clay and the sandy clay may be facies of the same sediment.

REFERENCES CITED

Boardman, Craig C., Allen S. Hunt, and Daniel E. Stein, 1968, The volume of Lake Champlain. (unpubl. ms.)

Chapman, C. H., 1937, Late glacial and post-glacial history of Champlain Valley. Am. Jour. Science, 5th ser., v. 34, p. 89-124.

Henson, E. B., 1966, A review of Great Lakes benthos research. Univ. Mich., Grt. Lakes Res. Div., publ. no. 14, p. 37-54.

Hunt, A. S., P. H. Townsend, and C. C. Boardman, 1968, The Lake Champlain drainage basin. Champlain Res. Rpts., issue no. 2.

Hutchinson, G. E., 1957, A treatise on limnology, v. 1, Wiley, N.Y., 1015 pp.

Legge, Thomas N., 1969, A study of the seasonal abundance and distribution of calanoid copepods in Burlington Bay, Lake Champlain. (unpubl. ms.) 133 pp.

MacClintock, Paul, and J. Terasmae, 1960, Glacial history of Covey Hill. Jour. Geology, v. 68, p. 232-41.

Muenscher, W. C., 1930, Plankton studies of the Lake Champlain watershed. In: N. Y. Conserv. Dept. Biol. Surv. of Champlain watershed, p. 146-163.

Potash, M., 1965, A study of two bays of Lake Champlain. Univ. Mich., Grt. Lakes Res. Div., publ. no. 13 (abstract).

Potash, M. and E. B. Henson, 1966, Oxygen depletion patterns in Malletts Bay, Lake Champlain. Univ. Mich., Grt. Lakes Res. Div., publ. no. 15, p. 411-415.

Potash, M., S. Sundberg, and E. B. Henson, 1969a, Characterization of water masses of Lake Champlain. (in press).

Potash, M., S. Sundberg, and E. B. Henson, 1969b, Epilimnial oxygen reduction during fall turnover in Malletts Bay. Proc. Int. Assoc. Grt. Lakes Res., (in press).

Ricker, K. E., 1959, The origin of two glacial relict crustaceans in North America, as related to Pleistocene glaciation. Can. Jour. Zool., v. 37, p. 871-893.

Stewart, David P., 1961, The glacial geology of Vermont. Vt. Geol. Surv., bull. no. 19, 124 p.

Terasmae, J., 1959, Notes on the Champlain Sea episode in the St. Lawrence lowlands, Quebec. Science, v. 130, p. 334-335.

Woodworth, J. B., 1905, Ancient water levels of the Hudson and Champlain Valleys. N.Y. State Mus. bull. no. 84, p. 65-265.

Trip C

BEDROCK GEOLOGY OF THE SOUTHERN PORTION OF THE HINESBURG SYNCLINORIUM

by

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INTRODUCTION

Purpose

The geology of west-central Vermont has attracted the attention of geologists since the early work of Hitchcock, et al. (1861), and Logan (1863), to mention only two. During this century notable contributions have been made by Perkins (1910), Keith (1923, 1932, 1933) and, in particular, Cady (1945, 1960). The mapping of Welby (1961) west of the Champlain thrust, and Stone and Dennis (1964) in the Milton quadrangle just north of Burlington, completes the present state of knowledge of the Hinesburg synclinorium and immediate surrounding areas (Fig. 1, inset map). These works are incorporated in the Centennial Geologic Map of Vermont (Doll, et al., 1961).

Among several major problems still remaining to be resolved is a clearer understanding of the effects of the Acadian and subsequent orogenies or disturbances on the structures of west-central Vermont. Inasmuch as rocks of Upper Ordovician through Devonian age are not present in western Vermont, it becomes most difficult to evaluate the relative importance of Taconic and younger events.

Answers to this central problem with its many corollaries can hopefully be provided by detailed studies involving a combination of stratigraphic, petrologic, geochronometric, and structural approaches. It is the purpose of this trip to examine the geology of the Hinesburg synclinorium and to show how recent work may be helpful in answering these problems in the future.

Acknowledgements

It goes without saying that the results of the many studies in western Vermont and Quebec must be acknowledged in forming the background for the present and future work in the Hinesburg synclinorium and adjacent areas. Although each of these cannot be mentioned individually, under-

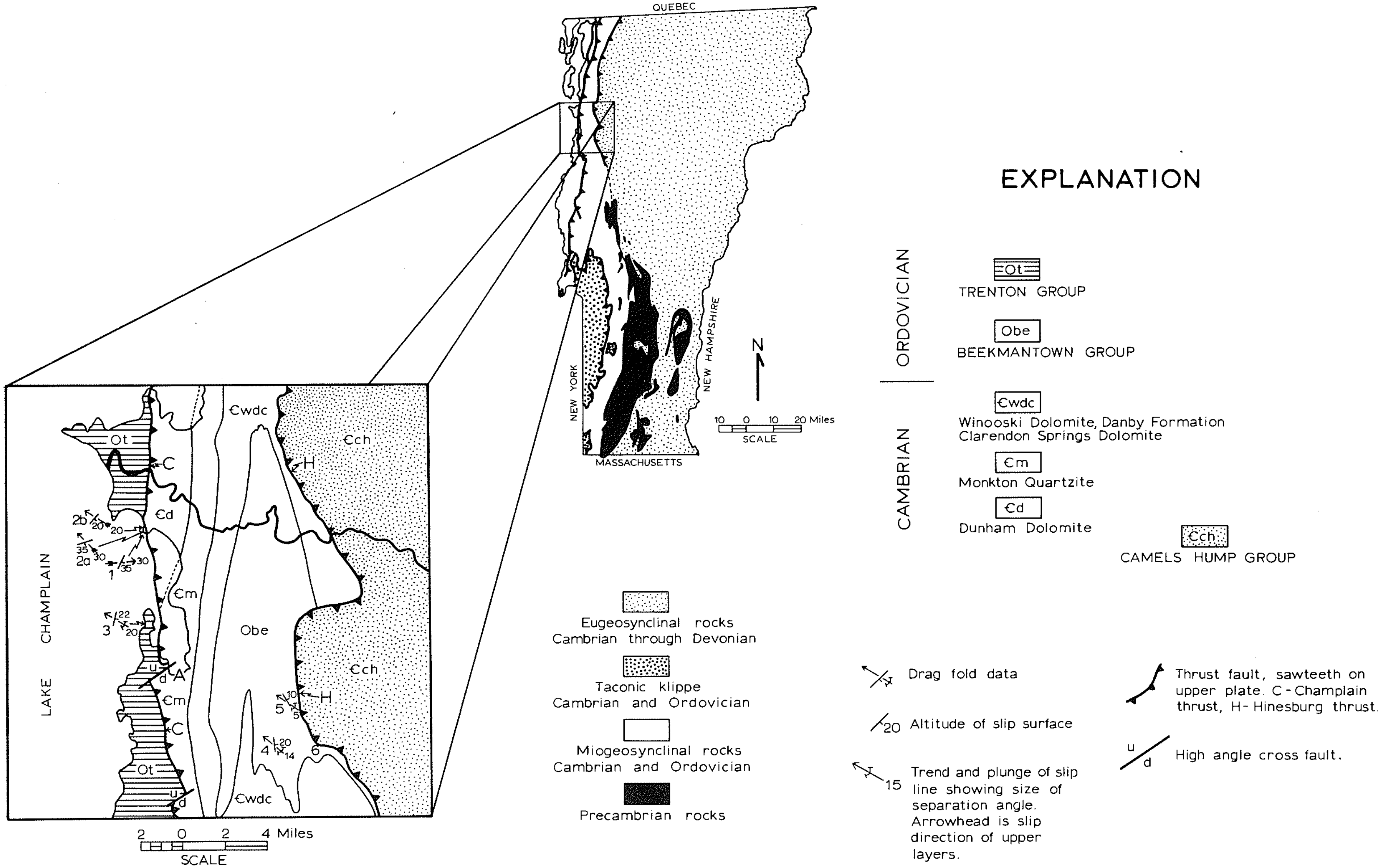


Figure 1 Generalized geologic map of Vermont and the Hinesburg synclinorium. Slip line directions at localities 1-6 are shown in more detail on figure 3.

standing of this area has been greatly assisted by syntheses of Cady (1945, 1960, 1968), Doll, et al. (1961), Rodgers (1968) and Zen (1967, 1968).

Recent work in the southern portion of the Hinesburg synclinorium has been carried out by various students at the University of Vermont in conjunction with my own work in the area. Information on the Shelburne access area is largely drawn from an unpublished research report by Charles Rubins. Quartz deformation lamellae studies have been done, in part, by Charles Rubins, John Millett, Edward Kodl, Robert Kasvinsky, Arthur Sarkisian and Evan Englund, whereas the drag fold data from localities 4 and 5 (Fig. 1) was collected by John Pratt. Data at stop 4 was provided by James Dinger. Charles Doll, State Geologist of Vermont, and E-an Zen kindly reviewed this paper and made several important suggestions. The geologic map of the Burlington quadrangle (available only to participants) was kindly provided by the Vermont Geological Survey. Mistakes in interpretation of the existing knowledge of the area are my responsibility.

REGIONAL GEOLOGY

The geology of western Vermont forms the northern extension of the Ridge and Valley provinces of the Appalachians. Carbonates and quartzites with minor amounts of shale, except in northwestern Vermont, characterize the autochthonous Cambrian and Lower Ordovician section and represent shallow-water shelf sediments of the miogeosynclinal belt. The Lower and Middle Cambrian part of this section wedges out to the west with rocks of Upper Cambrian age resting with angular discordance on the Precambrian rocks of the Adirondack dome. To the east in Vermont the Cambrian and Ordovician section thickens drastically and changes facies to graywacke and shale which represent the western part of the eugeosyncline belt of New England. Volcanic rocks, which are totally absent in the miogeosynclinal belt, are present here and become more abundant farther to the east in New Hampshire. In most places the change in sedimentary facies has been complicated by subsequent deformation, but in such areas as Milton (15 miles north of Burlington) it is relatively undeformed and appears abrupt. Rodgers (1968) has recently suggested that this change represents the eastern edge of the carbonate shelf which dropped off rapidly into the eugeosyncline to the east much like the present eastern edge of the Bahama Banks.

The upper part of the autochthonous section is composed of several thousand feet of shale with thin beds of silty dolostone and limestone, commonly graded and current rippled (Hawley, 1957), which represent the westward advance of the eugeosynclinal or transitional zone over the carbonate shelf during Middle Ordovician time (Cady, 1960; Zen, 1968). At present these rocks underlie most of the western part of the Champlain Valley whereas the older carbonate and quartzite underlie the central part of west-central Vermont (Cady, 1945).

The structure of the autochthonous part of the Champlain Valley consists of a series of east-dipping, imbricate foreland thrusts and west-facing asymmetrical or overturned folds somewhat typical of the Ridge and Valley province although the belt here is much narrower than in the central and southern Appalachians. Major culminations near Monkton and near Milton divide the belt into three synclinoria which are known from south to north as the Middlebury, Hinesburg and St. Albans synclinoria (Cady, 1945). Allochthonous rocks exist in both the Middlebury (Taconic klippen) and the St. Albans synclinoria (8 small isolated outliers) whereas no such structure has been recognized in the central Hinesburg synclinorium.

Although numerous thrusts of varying extent have been recognized in the Champlain Valley, the Champlain and Hinesburg thrusts are the largest and bound the three synclinoria on the west and east respectively (Fig. 1.). The Champlain thrust (Keith, 1923), first recognized as such by William Logan (1863) and hence part of Logan's line, extends for a distance of approximately 120 miles and places Lower Cambrian dolostone and quartzite (Dunham Dolomite north of Burlington, Monkton Quartzite to the south) of the Rosenberg slice (Clark, 1934) on Middle Ordovician shale to the west. North of Shelburne this fault dips eastward at less than 15^{o} but to the south it is gently folded and offset by high angle faults (Welby, 1961). The Hinesburg thrust (Keith, 1932) is the easternmost thrust in the Champlain Valley and extends for approximately 60 miles. It places the Lower Cambrian Cheshire Quartzite and the older Underhill Formation (schist and phyllite) of the western edge of the eugeosyncline (transitional sequences of Zen, 1968) on the eastern limb of the Hinesburg and St. Albans synclinoria. Although the stratigraphic throw is probably not as great as the Champlain, which is estimated at 9000 to 10,000 feet (Shaw, 1958, Stone and Dennis, 1964), the Hinesburg brings to the surface older rocks than the Champlain thrust. In contrast to the gentle eastward dip of the Champlain thrust, the Hinesburg dips both east and west at various angles and in such areas as the southern part of the Milton quadrangle definite evidence of folding of the fault surface has been found (Stone and Dennis, 1964).

Minor thrusts and high angle faults of either normal, reverse, or strike-slip displacement further complicate the geology in west-central Vermont. Such thrusts as the Muddy Brook north of the Winooski River and the Monkton south of Lewis Creek disrupt the stratigraphic continuity within the Rosenberg slice (Doll, et al., 1961) in the Champlain Valley. West of the Champlain thrust in northwestern Vermont a series of parallel thrusts (Hawley, 1957, Fisher, 1968) repeat the Ordovician section and can be interpreted as imbricate faults or possibly offshoots to the eastern master faults. Southwest of Middlebury a series of imbricate faults also mark the southern end of the Champlain thrust (Cady, 1945, Welby, 1961). Steeply dipping longitudinal and cross faults of either normal, reverse, or wrench origin further complicate the geology between the Champlain thrust and the Adirondacks (Welby, 1961, Doll, et al., 1961).

The rocks of western Vermont have been regionally metamorphosed with the grade increasing from chlorite in the Champlain Valley to biotite or higher in the Taconic and Green Mountains. The pattern of isograds accords with those of the rest of central New England in which Middle Paleozoic rocks are involved and hence is thought to be related to recrystallization during the Acadian orogeny (Thompson and Norton, 1968). However, isotopic age dates older than 400 m.y. in Quebec (Richard, 1965), along the Sutton-Green Mountain anticlinorium in northwestern Vermont, (Cady, personal communication; cited by Richard, 1965, p. 530) and in the Taconic allochthon (Harper, 1967) strongly suggest that pre-Silurian metamorphism and deformation are responsible for the fabric in that part of the Appalachians.

Igneous rocks are limited primarily to minor east-west dikes and are predominantly bostonite and lamprophyre (Welby, 1961). A small laccolith of syenite underlies Barber Hill in Charlotte (Migliore, ms; Dimon, ms). These bodies are clearly post orogenic inasmuch as they cut the folded and faulted country rocks. K-Ar determination on biotite from one such lamprophyre on Grand Isle yielded a radiometric age of 136 ± 7 MY (Zartman et al., 1967) and thus suggests a late Mesozoic age for its emplacement.

REGIONAL GEOLOGIC HISTORY

Although the interpretation varies somewhat, the geologic history of western Vermont has been well portrayed in recent papers by Zen (1967, 1968) and Cady (1960, 1967, 1968). Interested readers are referred to these and other papers (mentioned by these authors) for more information on the subject. As an aid to the reader, however, a brief summary follows of those events which are perhaps best exhibited in the Middlebury synclinorium where the Cambrian and Ordovician section is most complete.

During the Cambrian and Lower Ordovician most of western Vermont was the site of shallow-water deposition on a carbonate bank whose eastern edges dropped off steeply into the deep-water facies of the eugeosyncline. The first signs of tectonism appeared in the late Early or Middle Ordovician when, according to Zen (1967, 1968), high angle faults apparently divided the shelf into a series of grabens and horsts. Erosion ensued, accompanied by a short interval of carbonate deposition. Gradually the carbonates in the post-unconformity, Middle Ordovician section became more shaly, heralding the wave of argillaceous material that was spreading westward from the now uplifted former eugeosyncline. During this time (Zone 13 of Berry, 1960) the first segments of the Taconic allochthon were emplaced. The remaining slices (a total of six comprise the Taconics, Zen, 1967) followed during the Ordovician and were possibly in place before the formation of the three synclinoria in the autochthon

TABLE I.

COMPOSITE STRATIGRAPHIC SECTION FOR THE HINESBURG SYNCLINORIUM AND AREA WEST OF THE CHAMPLAIN THRUST AND EAST OF THE HINESBURG THRUST

(Cady, 1945; Doll, *et al.*, 1961)

Middle Ordovician

Hathaway Formation*

Missing but present north of the Lamoille River.

Iberville Formation

Noncalcareous shale, rhymthically interbedded with thin beds of silty dolomite and in the lower part with calcareous shale.

Stony Point Formation

Calcareous shale that grades upward into argillaceous limestone and rare beds of dolomite.

Cumberland Head Formation

Missing but well exposed in Grand Isle County.

Glens Falls Formation

Thin bedded, dark blue-gray, rather coarsely granular and highly fossiliferous limestone.

Orwell Limestone

Missing but present south of Charlotte.

Middlebury Limestone

Missing but present south of Charlotte.

*Only those formations encountered in the course of the trip will be described. Kindly refer to the Centennial Geologic Map of Vermont for other formation descriptions.

Lower Ordovician

Chipman Formation

Missing in the Hinesburg synclinorium but present south of Charlotte.

Bascom Formation

Beds of light bluish gray calcitic marble with laminae and thin beds of siliceous phyllite. Beds of brown-orange weathering dolomite 1 to 3 feet thick. The upper part becomes more phyllitic and is mapped separately as the Brownell Mountain Phyllite. Contact against typical Cutting dolomite is gradational.

Cutting Dolomite

Massive whitish to light grayish weathering dolostone, dark gray on the fresh surface. Sand-size quartz grains present in places especially near the base where sandy laminae are more abundant and brecciated in places. Black chert nodules are found in the upper part. Contact is sharp with sandy dolomite above and white calcitic marble of the Shelburne Formation below.

Shelburne Formation

Massive whitish gray weathering calcitic marble that is white on the fresh surface. Laminae of green phyllite present in the eastern part of the Hinesburg synclinorium. Contact is sharp with typical calcitic marble above and gray dolomite with quartz knots below.

Upper Cambrian

Clarendon Springs Dolomite

Massive, gray weathering dolomite with numerous knots of white quartz crystals. Black chert commonly found in the upper part. Contact gradational with the Danby Formation.

Danby Formation

Beds of gray sandstone interlayered with beds of dolomite 1 to 2 feet thick and sandy dolomite 10 to 12 feet thick. Sandstones are cross laminated. Beds 1 to 3 feet thick. Contact gradational.

Middle Cambrian?

Winooski Dolomite

Beds of very light gray to buff weathering dolomite, gray to light pink on the fresh surface. Thin siliceous partings characteristically separate beds of dolomite which range in thickness from 4 inches to 2 feet. Contact with the Monkton Quartzite is gradational and is placed above points where quartzite beds over 1 foot thick are separated by beds of dolomite less than 25 feet thick (Cady, 1945, p. 532).

Lower Cambrian

Monkton Quartzite

Beds of fine to coarse grained pink and brick red quartzite interlayered with minor beds of pink and red dolomite and thin laminae of red, green and purple shale. Beds of dolomite are more numerous in the lower part than the upper. Contact with the Dunham Dolomite is gradational and is placed where quartzite beds 1 foot thick are separated by beds of dolomite less than 25 feet thick (Cady, 1945, p. 532).

Dunham Dolomite

Massive, buff weathering dolomite that is pink and cream mottled or buff to gray on the fresh surface. Siliceous veins and laminae are present in places. Grains of sand size material common in places. Basal contact with the Cheshire Quartzite is gradational. Upper part of the Dunham north of Burlington is sandy and has been mapped separately as the Mallett Member.

Cheshire Quartzite

Massive, very thick bedded, white quartzite. Lower part is brown weathering and is more argillaceous and less massive. East of the Hinesburg thrust the Cheshire is more argillaceous. There the contact with the Underhill Formation is gradational and placed above "the frankly phyllitic, dark gray to greenish phyllite and below the rather characteristic mottled gray silty impure quartz schist of the Cheshire" (Stone and Dennis, 1964, p. 26).

Cambrian

Underhill Formation

Fairfield Pond Member: Predominantly green quartz - chlorite - sericite phyllite. Quartz grains common.

White Brook Member: Chiefly brown-weathered whitish, tan and gray sandy dolomite.

Pinnacle Formation

Schistose graywacke, gray to buff, with subordinate, quartz-albite-sericite-biotite-chlorite phyllite. Includes the Tibbit Hill Volcanic Member.

TABLE II

Age: System	Series	Group	Location: Saratoga Springs, New York Mohawk Valley (Fisher 1965)	West-central Vermont (Doll et. al. 1961)	West Limb of the St. Albans Synclinorium (Doll et. al. 1961) (Shaw 1958)	Lincoln Mtn.- Enosburg Falls Anticline East of Hinesburg Thrust (Doll et al 1961) (Stone and Dennis 1964)
ORDOVICIAN	Middle	Trenton	Schenectady Shale Canajoharie Shale Shoreham Limestone Larrabee Limestone	Hathaway Formation Iberville Formation Stony Point Formation Cumberland Head Formation Glens Falls Formation	Morse Line Formation	
		Black River	Amsterdam Limestone	Orwell Limestone		
		Chazy		Middlebury Limestone		
	Lower	Beekmantown	Chuctanunda Creek Dolostone Gailor Dolostone	Chipman Formation Bascom Formation Cutting Dolomite Shelburne Formation	Highgate Formation	
CAMBRIAN	Upper		Hoyt Limestone Mosherville Sandstone Thesesa Formation Potsdam Sandstone	Clarendon Springs Dolomite Danby Formation	Gorge Formation Sweetsburg Fm.: Hungerford Sl., Rockledge Congl., Skeels Corner Sl., Mill River Congl., St. Albans Sl.	
	? Middle ?					
	Lower			Winooski Dolomite Monkton Quartzite Dunham Dolomite Cheshire Quartzite	Rugg Brook Formation Parker Slate Dunham Dolomite Champlain Thrust	Dunham Dolomite Cheshire Quartzite
CAMBRIAN?				Mendon Formation	Not Exposed	Underhill Formation Pinnacle Formation
PRECAMBRIAN			Metamorphic Rocks of the Adirondack Dome	Mount Holly Complex		Mount Holly Complex

of western Vermont.

It is still not clear, particularly in the eastern part of western Vermont, to what extent the synclinoral folding and foreland thrusting continued into the Silurian or Devonian, possibly blending with some of the events in the Acadian orogeny. On the western edge of the allochthon, however, an upper age can be assigned to the Taconic orogeny. At Becraft Mountain in the Hudson Valley relatively undeformed Lower Devonian rocks (Manlius Limestone) rest with angular discordance on rocks of both the Taconic allochthon and shales of the autochthon (loc. 16 cited by Pavlides, et al., 1968, p. 69). To the east the termination of Taconic orogeny is marked by the unconformity separating the Silurian and Devonian rocks from the Cambrian and Ordovician in central New England (for series or stage assignments see Pavlides, et al., 1968). Here, however, deformation and metamorphism of the Acadian orogeny has overprinted and greatly obscured the structures of the Taconic orogeny. Herein lies the problem of dating the structural and metamorphic events in western Vermont. Specifically, how far west does Acadian deformation and metamorphism extend and, conversely, how far east does Taconian deformation and metamorphism persist? Present investigators are divided on the problem. Cady (1945, 1960) and Crosby (1963) believe that most of the foreland thrusts and folds are Acadian, whereas Clark (1934) and Richard (1965) believe similar features in Quebec are Taconian. Zen (1967, 1968) favors a pre-Acadian age but admits that Acadian movements could have modified the structures. The fact that the metamorphic isograds are not deformed in western Vermont as they are in central New England (Thompson, et al., 1968) suggests that the structures are older than Acadian metamorphism. Isotopic age dates from the Taconics (Harper, 1967, 460-445 my) and the Sutton Mountain anticlinorium in Quebec (Richard, 1965, 440-420 my) also favor a Taconian age for the deformation in western Vermont.

The degree to which post-Taconian or post-Acadian events (normal faulting and regional tilting) have further complicated the picture is not known.

GEOLOGY OF THE HINESBURG SYNCLINORIUM

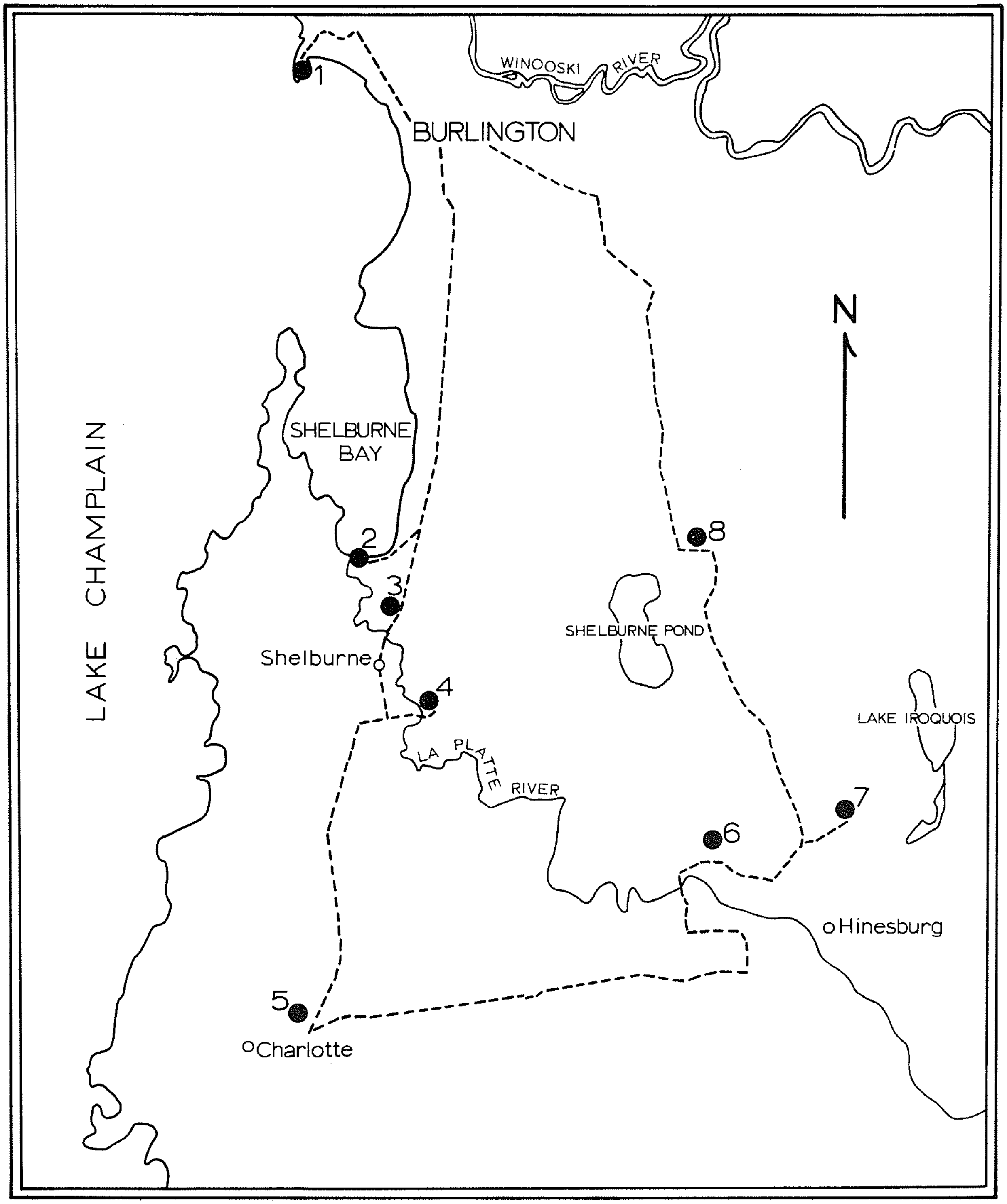
The Hinesburg synclinorium is separated from the Middlebury and St. Albans synclinoria by the Monkton cross anticline (Cady, 1945, p. 562) and the Milton culmination respectively. As seen on the geologic map of Vermont the smaller anticlines and synclines of the southern part are not mirrored by the single synclinal hinge of the northern part. This may be due to poor exposure to the north but it is probably a result of the configuration of the synclinorium and the fact that the eastern part is covered by the upper plate of the Hinesburg thrust. Minor faults, such as the Muddy Brook to the north and the Shelburne Pond to the south, partially disrupt the fold continuity of the synclinorium. These relations are well shown on the Geologic Map of Vermont (Doll, et al., 1961) and generalized in figure 1.

A composite stratigraphic section and correlation chart for the area is shown in Table 1 and Table 2 respectively. The rocks east of the Hinesburg thrust form the base of the composite section and mainly consist of metagraywacke and phyllite of the Pinnacle and Underhill formations. Isolated slivers of the Cheshire Quartzite (more argillaceous than to the west) appear just east of the Hinesburg thrust. Between the Winooski River and the Canadian border, outcrops of the Cheshire Quartzite form part of the upper plate of the thrust. The rocks in the synclinorium itself range in age from Lower Cambrian to Lower Ordovician (Dunham Dolomite through the Bascom Formation) and are characterized chiefly by carbonates which are predominantly dolomitic in the Cambrian and calcitic in the Lower Ordovician.

Quartzite beds occur in the Monkton Quartzite and the Danby Formation, and shale or phyllite is present as thin beds in the Monkton, Shelburne, and Bascom formations. The rocks west of the Champlain thrust are predominantly shales of the Stony Point and Iberville Formations and represent the argillaceous wedge that advanced westward on the carbonate shelf of western Vermont during the Middle Ordovician.

Current Work

In order to clarify the chronologic problems in the tectonic history of western Vermont, a long term project of detailed mapping and fabric analyses has been undertaken in the Hinesburg synclinorium and areas to the west, south and east of it. During the last four years students in Field Geology at the University of Vermont have been remapping parts of the synclinorium near Charlotte and Hinesburg. Although these studies are just a beginning in the long-term effort of refining the tectonic history of the area, it is already apparent that the structural relations in such areas as along the Champlain thrust and in the central part of the Hinesburg synclinorium are different from those shown on existing maps. For example, at least two generations of minor folds have been delineated in the southern part of the synclinorium and beneath the Champlain thrust. Although the styles of these folds differ in the shale and marble, the youngest generation of folds beneath the Champlain thrust shows the same northwesterly slip direction as does the older generation of folds in the Hinesburg synclinorium. The axial surfaces of these folds are in turn folded about northward-plunging axes of the younger generation. This suggests that the youngest movements on the Champlain thrust were followed by subsequent folding in the eastern part of the Hinesburg synclinorium. Furthermore, analysis of fractures, such as described at stop 2, indicates a sequence of events that can be related to the Champlain thrust. The inferred orientation of the principal planes of stress, however, does not



STOP MAP

----- Route

● Stops

1 0 1 2 3 4 Miles

SCALE

Figure 2

coincide in symmetry with other localities on the Champlain thrust. Elaborating and clarifying these relationships, as well as answering questions to other related problems, will hopefully culminate in a sequence of structural events for western Vermont that can be related to the orogenic history of the region.

STOP DESCRIPTIONS

General

The trip will consist of eight stops which are located on figure 2 and plate 1. The geologic relations along the route of the trip can be followed on figure 1 and plate 1.

Stop 1. Champlain thrust at Long Rock Point, Burlington - (Note: The Episcopal Diocesan Center has been kind enough to allow us to visit this locality. Please do not litter.) This locality is perhaps one of the finest exposures of the Champlain thrust in Vermont and Canada. Here the Dunham Dolomite (Conners facies) of Lower Cambrian age overlies the Iberville Formation of Middle Ordovician age. The thrust contact is sharp and marked by a thin discontinuous zone of breccia in which angular clasts of dolostone are embedded in a highly contorted matrix of shale. Slivers, several feet thick, of limestone are found along the fault and may represent pieces of the Beekmantown Group (Beldens Member of the Chipman Formation?). The undersurface of the Dunham Dolomite along the thrust is grooved by fault mullions which plunge 15^{o} to the southeast (Fig. 3 diagram 1 and 2a). The average southeasward dip of the thrust is 10^{o}.

A variety of minor structures are found in the Iberville Formation whereas joints are the only structures in the Dunham Dolomite. Faults and joints are oriented in a number of attitudes in the shale, but they have not been analyzed as yet. Many of these fractures are filled with calcite and grooved with slickensides. The minor folds in the shale are numerous, and are easily grouped into two age generations. The early folds have a well developed slaty cleavage which offsets the bedding and attests to differential flow parallel to the axial surface. Although only a few fold hinges are present, the slaty cleavage forms the dominant layering and is concordant to the undersurface of the Dunham Dolomite.

The younger generation consists of asymmetrical drag folds with short gently curved hinges and rather open profiles. The axial surface is rarely marked by cleavage but when it is developed, fracture cleavage, which is commonly filled with calcite, is typical. These folds deform the slaty cleavage of the older generation and hence are younger in age. The orientation of 59 axes with their sense of rotation is shown in diagrams 1, 2a and 2b of figure 3.

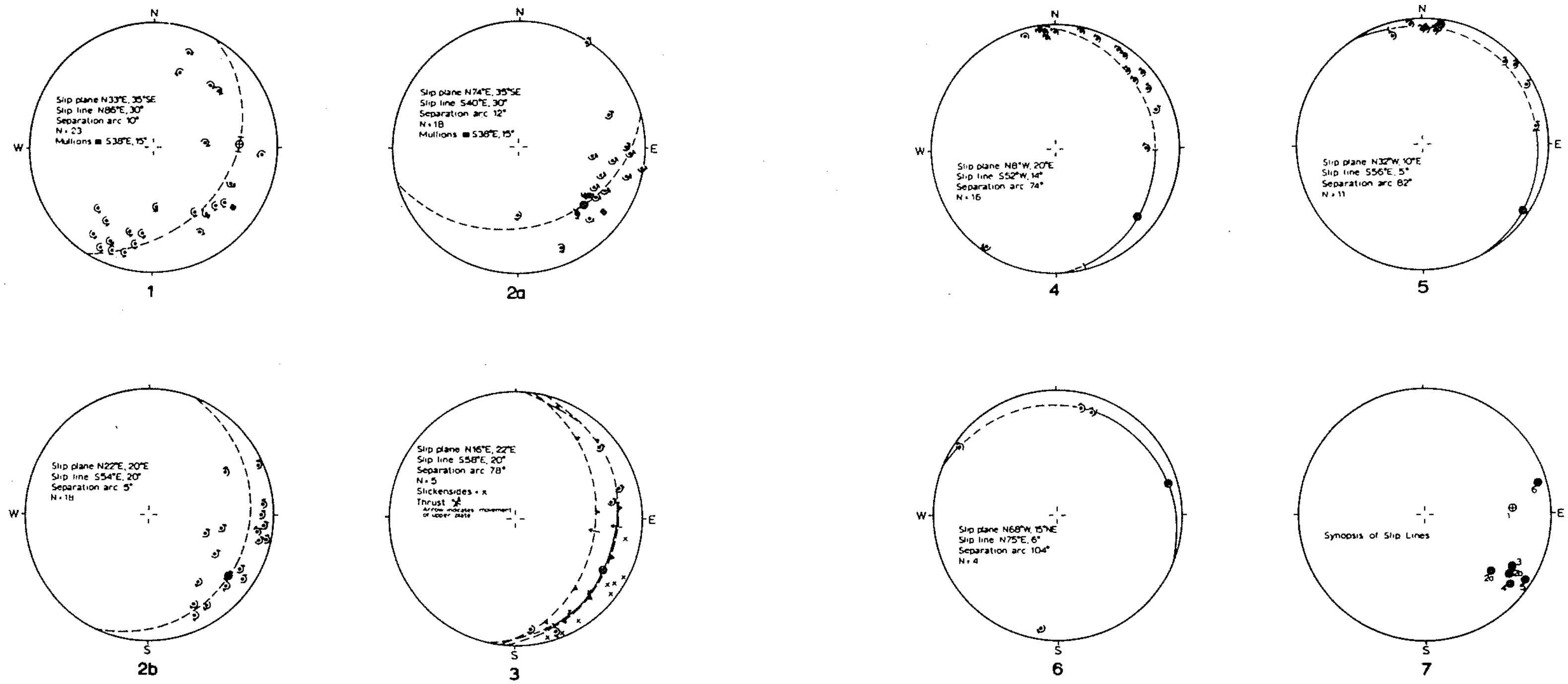


Figure 3 Slip line orientations determined from drag folds. Data shown on lower hemisphere equal area projection. Solid dots represent fold axes. Semicircular arrows show sense of rotation of asymmetrical drag folds. Dashed great circles are slip surfaces each of which contains a slip direction shown by a circle with solid dot (movement up of upper beds) or a circle with cross (movement down of upper beds). Numbers correspond to location shown on figure 1.

Slip line orientations The drag folds of the younger generation can be used to determine a slip surface and direction according to the method described by Hansen (1967, p. 390-397). The analysis assumes that these folds are in the truest sense drag folds and have formed by movement of one layer or zone over another. In this locality the drag folds in the shale presumably were formed by Dunham Dolomite moving over the Iberville Formation. Because the shale is highly anisotropic, consisting of thin layers of compact shale separated by thinner layers of extremely fissile shale, the folds are confined to zones several feet in thickness and are disharmonic in profile.

A slip plane and line were determined for three separate localities at Rock Point (Fig. 3, diagrams 1, 2a, 2b). For each place the 18 to 23 fold axes with their respective senses of rotation were plotted on a lower hemisphere equal area net. The great circle that best approximates the spatial distribution of axes defines the slip plane. In all the diagrams of figure 3, clockwise folds cluster along one part of the slip plane whereas counterclockwise folds cluster along the other. The bisector of the arc separating the two groups of folds is the slip line. The location of clockwise and counterclockwise arrows on either side of the separation arc determines whether the upper layers moved up or down along the slip line. In diagrams 2a and 2b (Fig. 3) the upper layers moved to the northwest along a line striking N 40 W for 2a, and N 54 W for 2b. In contrast the upper layers moved downward along a line striking N 86 E for the southern part of the Champlain thrust at Long Rock Point. (diagram 1, Fig. 3).

Discussion of Results The kinematic basis for this analysis has been worked out in such geologic environments as tundra and sod slides, glacial lake clays, lava flows and metamorphic rocks of all grades (Hansen, 1967; Hansen, *et al*., 1967; Howard, 1968). It can be shown that the separation arc contains the movement direction of slip line and that the drag folds which locate this direction are a product of one movement event. One can further deduce the likely position of the principal axes of stress (σ_1, σ_2, σ_3)[1] by analogy to slip systems. σ_2 would lie in the slip plane perpendicular to the slip direction. σ_1 and σ_3 would define the deformation plane which is perpendicular to the slip plane and parallels the slip direction.

σ_1 would be oriented approximately 45° from the slip plane in a direction permitted by the sense of shear indicated by the drag folds. The slip line would then parallel the direction of maximum resolved shear stress.

[1] In all subsequent discussions σ_1 is the maximum compressive stress, σ_2 is the intermediate compressive stress and σ_3 is the minimum compressive stress.

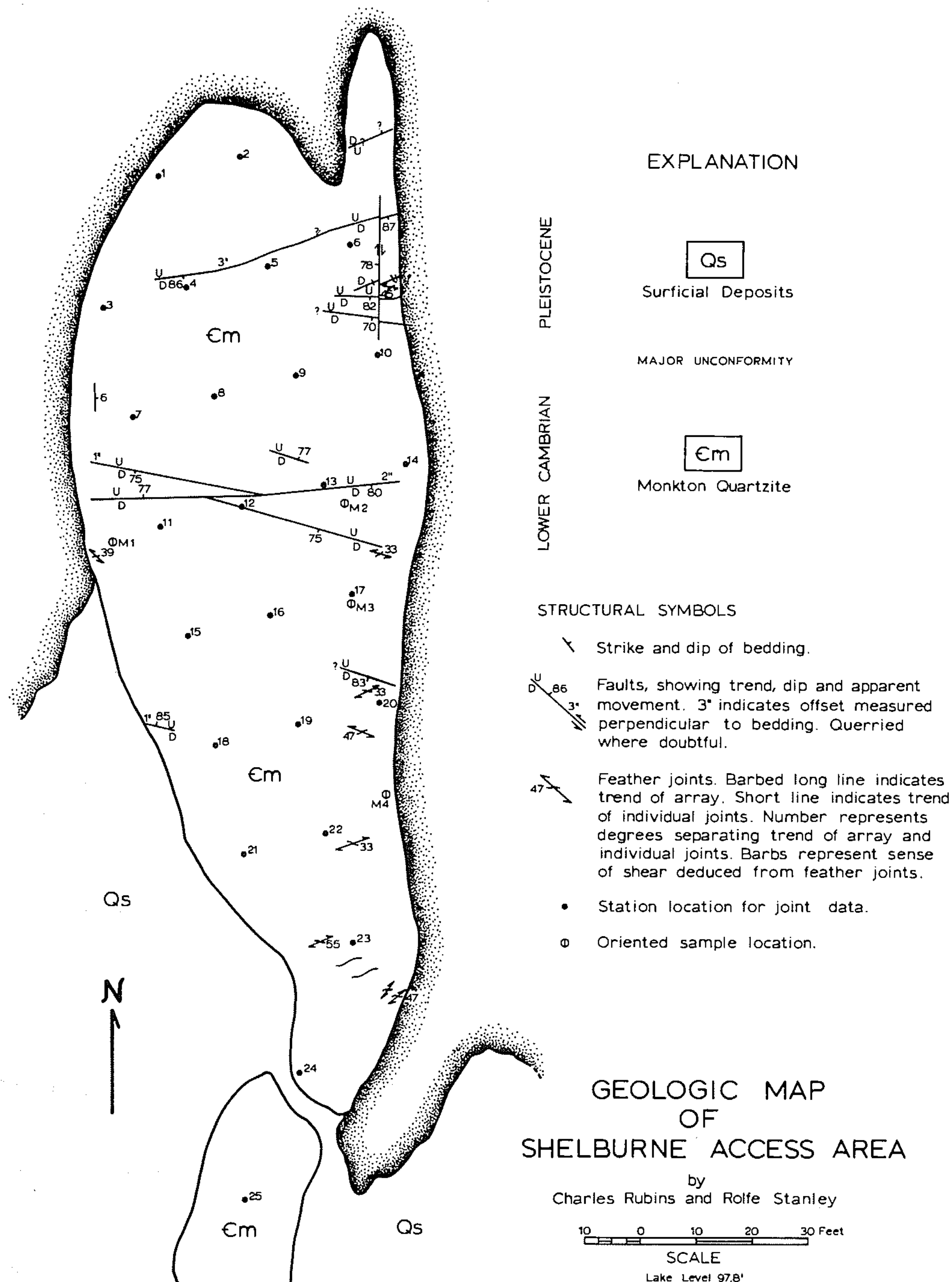


Figure 4 Structures in the Monkton Quartzite at the southern end of Shelburne Bay, Vermont.

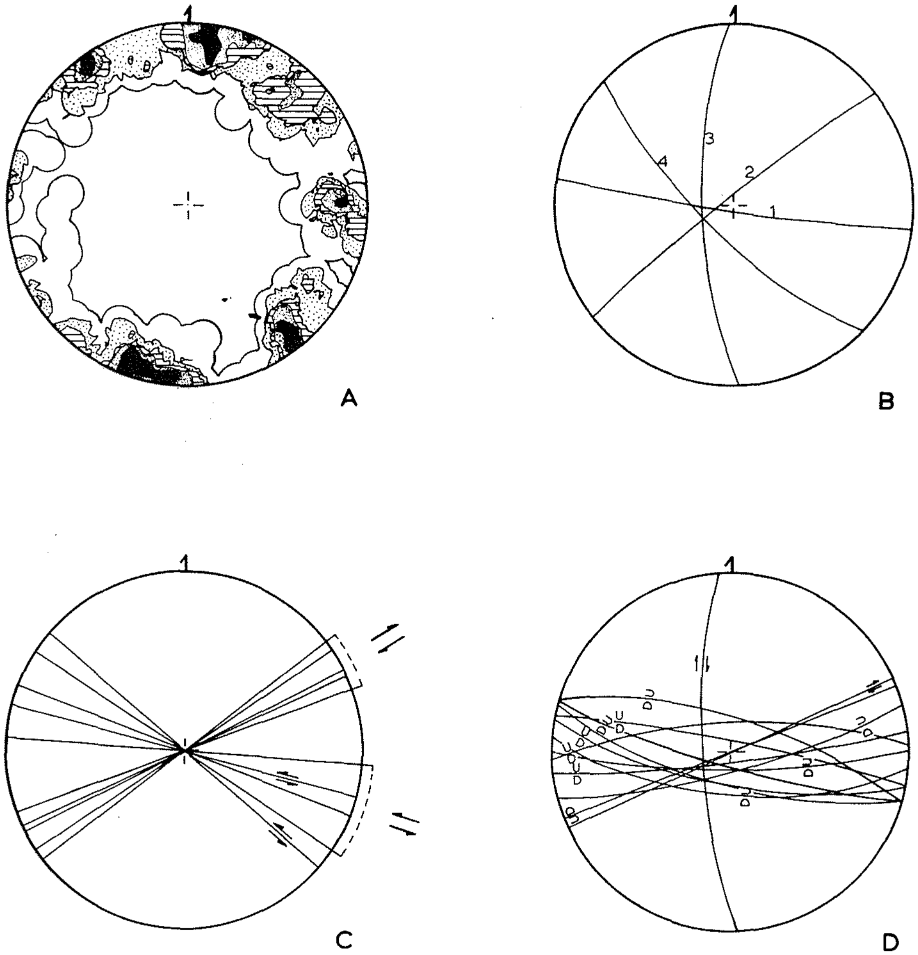


Figure 5 Lower hemisphere equal area projections showing fractures present in the Monkton Quartzite at the Shelburne Access area (fig. 4), Diagram A shows 248 poles to joints. Contour intervals are 0.4, 1.2, 2.0, 2.8, 3.6 respectively per 1 percent area. Diagram B shows planes corresponding to the 3.6 percent maxima of diagram A. Diagram C shows the trend and sense of shear of feather joints. Diagram D shows the orientation and apparent movement of 13 faults. Modified from an unpublished research report of Charles Rubins.

Diagrams 1 through 3 (Fig. 3) were determined at localities in the Iberville Formation beneath or just west of the Champlain thrust. With the exception of diagram 1, the slip directions indicate movement to the northwest (see diagram 7, Fig. 3 for synopsis). This direction is approximately parallel to fault mullions at Long Rock Point and slickensides on calcite-veneered surfaces at Shelburne Point (diagram 3) and, therefore, suggests that the deformation plane containing σ_1 and σ_3 for this phase of movement did trend northwesterly. A unique explanation for the easterly slip direction in diagram 1 is not known at present, but several possibilities will be discussed.

Stop 2. Shelburne Access Area This locality is one of several places in the Monkton Quartzite that displays a sufficient number of joints, faults and feather joints for dynamic analysis. The location and orientation of faults and feather joints are shown on the geologic map (Fig. 4). At each numbered station the orientation, relative abundance, and surface features of approximately 10 joints were recorded. Diagram A of figure 5 shows the poles to 248 joints and diagram B shows four planes corresponding to the maxima in diagram A. The trend of each of the 10 feather joint arrays and their sense of shear are shown in diagram C.

Diagram D shows 13 faults with letters or arrows indicating the apparent movement. All faults trend approximately eastward (hereafter called cross faults) except for one that is striking to the north. As indicated on the map the dip slip displacement of each fault is only several inches whereas the strike slip displacement on one of these faults is 9 feet (fault 13,[2] Fig. 4). Feather joint arrays adjacent to several cross faults further attest to strike slip displacement. This relationship is further emphasized by comparing diagram C and D (Fig. 5). Although Welby (1961, p. 204) assumes that all cross faults are normal faults, the above data definitely supports a wrench fault interpretation. Petrofabric analysis of quartz deformation lamellae in three samples (M1, M3, M4) further confirms this conclusion (Fig. 6).

Dynamic Analysis Discussion of the stress configurations for each structure is based upon techniques and principles summarized by Friedman (1964). In stress analysis only the direction and relative magnitudes of the principal stresses can be evaluated. Furthermore, it is assumed that the manner in which these structures formed in nature is similar to the way analogous structures are formed in the laboratory.

Joints: The planes in diagram B (Fig. 5) corresponding to the maxima of diagram A can be interpreted in several ways. The acute angle between joints 1 and 3 is 80 degrees and the acute angle between joints 2 and 4 is 83 degrees.

[2]Faults are identified by the station number nearest them on the geologic map (Fig. 4).

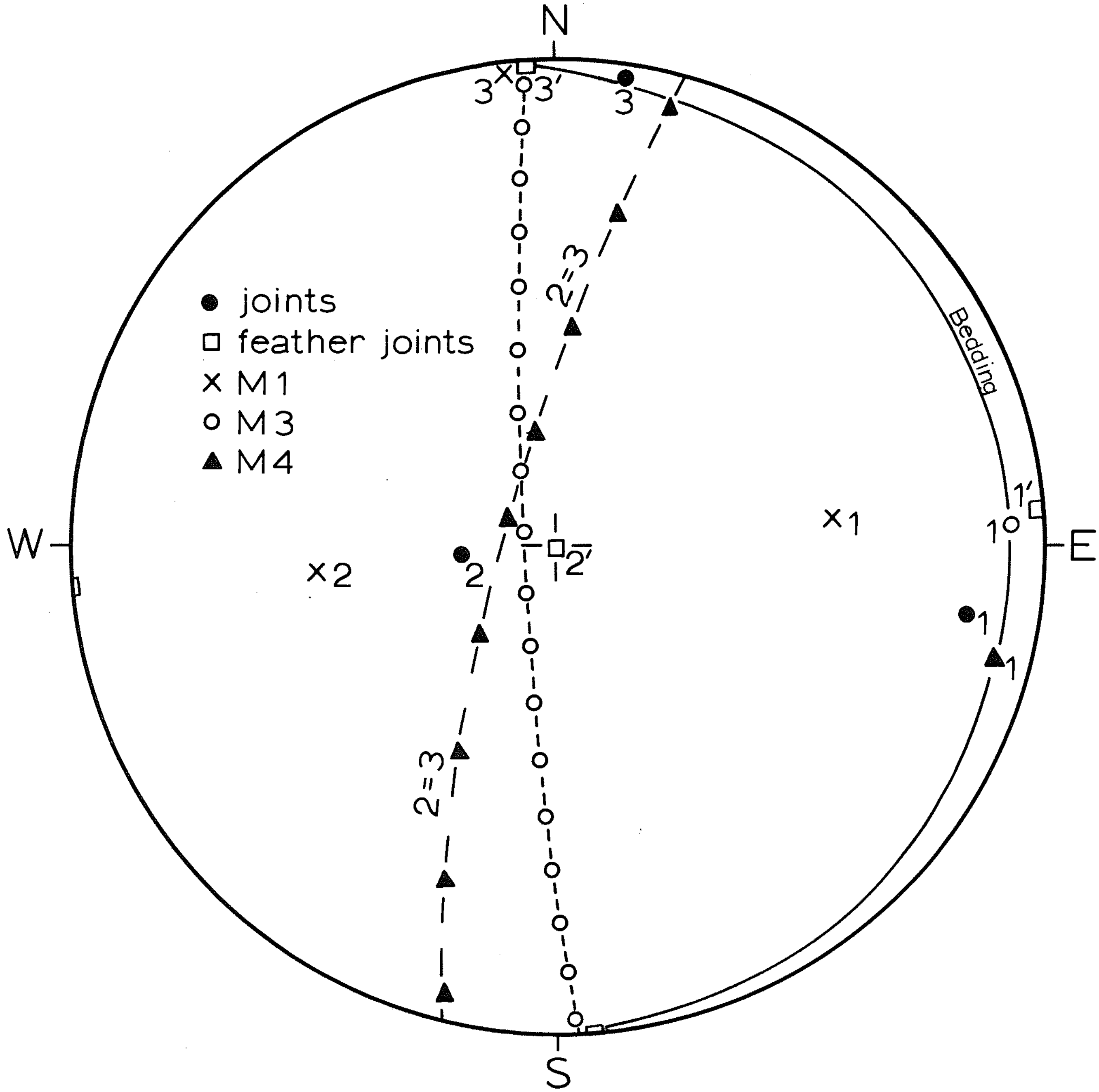


Fig. 6: Synoptic diagram of principal stress directions (σ_1,σ_2,σ_3) deduced from joints (n = 248), feather joints (n = 11), and deformation lamellae in quartzite samples M 1 (n = 204), M 3 (n = 119), M 4 (n = 62). Numbers 1, 2, 3 refer to σ_1 (maximum compressive stress), σ_2,σ_3 (minimum compressive stress) respectively.

Hypothesis 1: Joints 2 and 4 formed in the shear position. Joints 1 and 3 formed as extension and release joints respectively.

Hypothesis 2: Joints 1 and 3 formed in the shear position. Joints 4 and 2 formed as extension and release joints respectively. In both hypotheses σ_2 is defined by the intersection of all joints. In hypothesis 1, σ_1 would parallel joint 1 and trend to the east whereas in hypothesis 2 σ_1 would parallel joint 4 and trend to the southeast. Hypothesis 1 is preferred because joint 1 is commonly filled with calcite. Furthermore, this configuration is similar to the stress configurations deduced from the other structures (Fig. 6).

Feather joints: The 10 feather joint arrays in diagram C (Fig. 5) with their respective senses of shear indicate that σ_1 is oriented east-west. σ_3 is oriented north-south and σ_2 is oriented vertically. Because the dip of the feather joint array could not be measured, the inclination of the principal stress axes could not be pinpointed. Although the principal stress directions are not as accurately known for feather joints as they are for other structures, their symmetrical relationship to the cross faults support a wrench fault interpretation and thus suggest that those faults trending north of west are left lateral whereas those trending north of east are right lateral.

Wrench Faults: The apparent vertical movement on all faults trending north of west is up to the north. This movement can be realized on left lateral wrench faults if σ_1 dips more steeply to the east than the bedding. Those faults trending north of east that are downthrown to the north would then be right lateral wrench faults.

The apparent vertical movement of faults 4 and 13 trending north of east is up to the north. These faults are right lateral wrench faults if the movement on 4 is assumed to be the same as the displacement of fault 12 by 13. The horizontal and vertical components of the actual movement at these faults could be caused by σ_1 trending westward and dipping eastward more gently than the bedding.

North-South Fault: This fault cuts several of the wrench faults and hence is younger in age. It could result from the following two stress configurations:

Hypothesis 1: If the fault is a normal fault downthrown to the west, then σ_1 would be vertical or steeply inclined to the north so as to produce the right lateral movement. σ_3 would trend eastward or dip gently westward. σ_2 would trend northward.

Hypothesis 2: If the fault is a wrench fault, then σ_1 would trend northeasterly. σ_3 would trend northwesterly and σ_2 would be approximately vertical.

Quartz Deformation Lamellae: Approximately 380 deformation lamellae from 3 oriented samples have been measured from this outcrop. For each sample 100 (50 for M4) quartz grains were measured from each of three mutual perpendicular thin sections. For each grain the orientations of the c axis and deformation lamellae (if present) were measured. The results were analyzed using methods described by Carter and Friedman (1965), and Scott, et al., (1965). The deduced orientation for σ_1, σ_2, and σ_3 are shown in figure 6. In M3 and M4, σ_1 lies in the bedding and σ_2 appears to be equal to σ_3 in magnitude. In M1 σ_1 is inclined 40^o to the east, σ_2 dips 50^o to the west and σ_3 trends northward and is horizontal. Although these orientations are not parallel in all samples, they are consistent with the stress positions deduced from the megascopic structures and support the conclusions on the cross faults and joints.

Structural and Stress History Based on the above information it is possible to develop a structural history for this outcrop. This sequence is divided into the following three phases.

Phase 1: During this time all the wrench faults formed except fault 13, 4 and 10 (north-south fault), σ_1 was inclined more steeply to the east than the bedding so that faults trending north of west are upthrown to the north and faults trending north of east are downthrown to the north. σ_3 was oriented northward and σ_2 was inclined steeply to the west.

Phase 2: During this time faults 13 and 4 developed with σ_1 either horizontal or inclined more gently eastward than the bedding. This orientation permitted the right lateral faults (13 and 4) to be upthrown to the north. It should be emphasized at this point that the change in orientation of σ_1 relative to the bedding in the Monkton Quartzite can equally be attributed to a rotation of the bedding within a stress system of constant principal stress orientation.

Phase 3: During this time the north-trending fault developed and can be interpreted as a normal, wrench, or reverse fault.

The feather joints and deformation lamellae are thought to have formed during phase 1 and 2, although definite evidence for their timing is lacking at present. The formation of the joints shown in diagram A and B in figure 5 could well have spanned the first two phases.

Relationship to the Champlain Thrust Wrench faults are commonly associated with thrust faults. Both can be related to the same σ_1 direction and only require a switch of σ_2 and σ_3 in the stress configuration during thrusting to develop wrench or tear faults. In the same manner the small wrench faults in the Monkton Quartzite are thought to bear the same relationship to the Champlain thrust and as such suggests that other cross faults shown on the Geologic Map of Vermont (Doll, et al., 1961), may also be wrench faults.

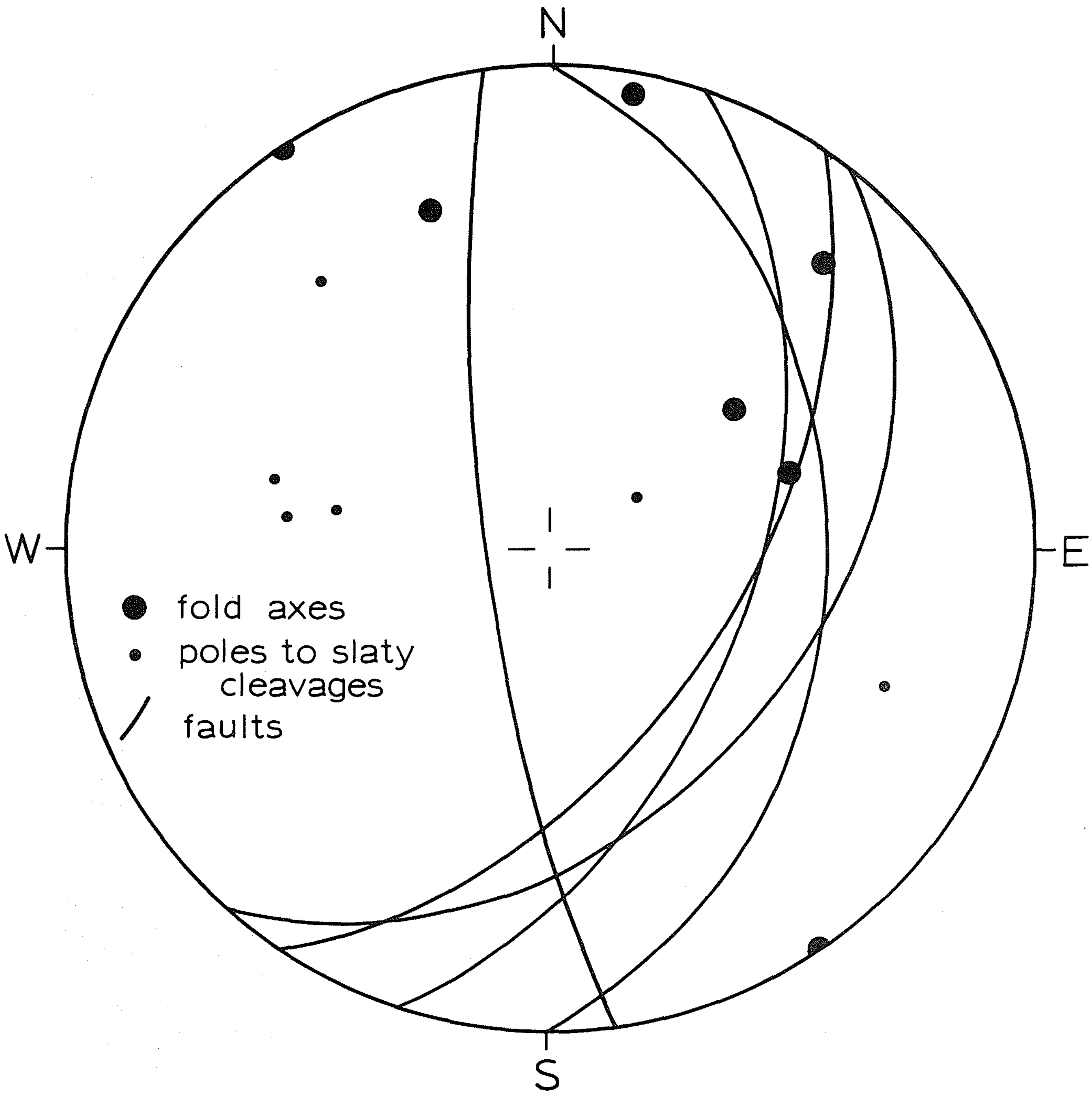


Figure 7 Lower hemisphere equal area projection showing the orientation of fold axes, slaty cleavage and faults in the Iberville Formation on the west side of Jones Hill.

The different stress orientations deduced at Long Rock Point (stop 1) and the Shelburne access area suggest several explanations. It is possible that they formed under the same stress system but have assumed their present orientation by subsequent deformation. Alternatively, they could have formed at different times under stress systems of different orientations. Future work will resolve this problem.

Stop 3. This locality is a fine exposure of the Winooski Dolomite in fault contact with the upper portion of the Monkton Quartzite. The bedding in both these units dips gently eastward. The most conspicuous of several northeastward-trending faults is downthrown to the north placing the Winooski Dolomite in fault contact with the upper portion of the Monkton Quartzite south of the fault.

Stop 4. Shelburne Falls just southeast of the village of Shelburne The Danby Formation dips gently eastward at this locality and is cut by several sets of joints. Ripple marks of varying sizes, cross beds and fossil burrows are well displayed in the quartzite and sandy dolomite on the west bank of the river. Current directions based on current ripples in the upper layers along the southern part of the exposure indicate flow to the southeast (S 10 E to S 40 E) whereas current directions based on cross bedding in the lower layers exposed near the falls to the north indicate flow to the north and west.

Stop 5. Jones Hill This stop is located just west of the Champlain thrust and shows several large folds and minor thrusts exposed in the Iberville Formation. A lamprophyre dike, striking N 78 W and dipping 75 degrees to the northeast, is exposed in the southern part of the outcrop. The laccolith of syenite on Barber Hill is located approximately a mile southeast of here near the village of Charlotte.

The orientation of fold axes, slaty cleavage and faults is shown on figure 7; most of the faults are thrusts dipping eastward or southeastward at gentle to moderate angles. As shown on plate 1 the trace of the Champlain thrust is sinuous and offset at several places in the area. Welby (1961, Plate 1a) shows normal faults at each of these places but the orientation of bedding, particularly between Jones Hill and Pease Mountain, suggests that the eastward reentrants can be explained equally well by anticlines plunging gently to the east. Minor folds with this orientation are present on the north side of Jones Hill.

Stop 6. This stop will consist of a short traverse through the fields north of the road. The Clarendon Springs Dolomite, the Shelburne Formation, and the Cutting Dolomite will be crossed. If time permits the traverse will be extended northward into the Bascom Formation. At least two generations of folds are present in this area. Axes of the earlier generation plunge to the southeast, east and northeast and possess a well developed axial surface cleavage which commonly dips eastward except when it is deformed by folds of the younger generation. The slip line orientations of diagrams 4 and 5 (Fig. 3) were determined from drag folds of the earlier fold generation at localities 4 and 5 on figure 1. These slip lines are almost parallel to those to the west of the Champlain thrust discussed under stop 1 and, therefore, strongly suggest that they may have formed contemporaneously as a result of the same stress system. The younger generation of folds plunges northward and folds the cleavage of the earlier fold generation.

Stop 7. Hinesburg thrust northwest of Mechanicsville This locality is one of the finest exposures of the Hinesburg thrust. The Bascom Formation forms the lower plate and the argillaceous facies of the Cheshire Quartzite forms the upper plate. Minor folds can be found in both these units. Inasmuch as only 4 folds could be found in the Bascom Formation little significance can be attached to southwestward movement direction shown in diagram 6 (Fig. 3).

Stop 8. The Bascom Formation crops out on several small hills just west of Brownell Mountain. To the east the Bascom grades into the Brownell Mountain Phyllite, a member of the Bascom Formation only recognized in the Hinesburg synclinorium. This stop may be cancelled if insufficient time is available.

End of Trip. Return to Burlington.

REFERENCES CITED

Berry, W. B. N., 1960, Graptolite faunas of the Marathon region, West Texas: Univ. Texas Pub. 6005, 179 p.

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

__________, 1960, Stratigraphic and geotectonic relationships in northern Vermont and southern Quebec: Geol. Soc. America Bull., v. 71, p. 531-576.

__________, 1967, Geosynclinal setting of the Appalachian Mountains in southeastern Quebec and northwestern New England, in: Clark, T. H., ed., Appalachian Tectonics: Royal Soc. Canada Spec. Pub. 10, p. 57-68.

__________, 1968, The lateral transition from the miogeosynclinal to the eugeosynclinal zone in northwestern New England and adjacent Quebec, p. 151-161, in: Zen, E-an, et al., 1968, Editors, Studies of Appalachian geology, Northern and Maritime: New York Interscience Publishers, John Wiley and Sons, Inc., 475 p.

Carter, N. L., and Friedman, Melvin, 1965, Dynamic analysis of deformed quartz and calcite from the Dry Creek Ridge Anticline, Montana: Am. Jour. Sci., v. 263, p. 747-785.

Clark, T. H., 1934, Structure and stratigraphy of southern Quebec: Geol. Soc. America Bull., v. 45, p. 1-20.

Doll, C. G., Cady, W. M., Thompson, J. B., Jr., and Billings, M. P., Compilers and Editors, 1961, Centennial geologic map of Vermont: Montpelier, Vermont, Vermont Geol. Survey, scale 1:250,000.

Dimon, J. R., 1962, Petrography of the laccolith at Barber Hill, Vermont: M.S. thesis, University of Vermont, Burlington, Vermont.

Fisher, D. W., 1969, Bedrock geology of the Plattsburgh and Rouses Point quadrangles, New York and Vermont: Vermont Geol. Survey Special Bull. No. 1.

Friedman, Melvin, 1964, Petrofabric techniques for the determination of principal stress directions in rocks, in: State of Stress in the Earth's Crust: American Elsevier Publishing Co., Inc., p. 451-552.

Hansen, Edward, 1967, Methods of deducing slip-line orientations from the geometry of folds, in: Carnegie Inst. Wash., Yearbook 65, p. 387-405.

Hansen, Edward, Scott, W. H., and Stanley, R. S., 1967, Reconnaissance of slip-line orientations in parts of three mountain chains, in: Carnegie Inst. Wash., Yearbook 65, p. 406-410.

Harper, C. T., 1968, Isotopic ages from the Appalachians and their tectonic significance. Can. J. Earth Sci., 5, pp. 49-59.

Hawley, David, 1957, Ordovician shales and submarine slide breccias of northern Champlain Valley in Vermont: Geol. Soc. America Bull., v. 68, p. 55-94.

Hitchcock, Edward, Hitchcock, Edward, Jr., Hager, A. D., and Hitchock, Charles, 1861, Report on the geology of Vermont, vol. 1, 558 pages; vol. 2, p. 559-988, Claremont, Vermont.

Howard, K. A., 1968, Flow direction in triclinic folded rocks: Am. Jour. Sci., v. 266, p. 758-765.

Keith, Arthur, 1923b, Outlines of Appalachian structure, Geol. Soc. Am., Bull., v. 34, p. 309-380.

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

Keith, Arthur, 1933, Outline of the structure and stratigraphy of northwestern Vermont, in: 16th Internat. Gol. Cong., Guidebook 1 (Eastern New York and Western New England), p. 48-61.

Logan, W. E., 1863, Report on the geology of Canada: Geol. Survey Canada, Rept. Prog. 1863, 983 pages.

Migliore, J. J., Jr., 1959, Petrography of some Champlain Valley dikes: MS thesis, University of Vermont, Burlington, Vermont.

Pavlides, Louis, Boucot, A. J., and Skidmore, W. B., 1968, Stratigraphic evidence for the Taconic orogeny in the northern Appalachians, p. 61-82, in: Zen, E-an, et al., 1968, Editors, Studies of Appalachian geology, Northern and Maritime: New York Interscience Publishers, John Wiley and Sons, Inc., 475 p.

Perkins, G. H., 1910, Geology of the Burlington quadrangle: Vt. State Geol., 7th Rept., p. 249-256.

Richard, M. J., 1965, Taconic orogeny in the Western Appalachians: experimental application of microtextural studies to isotope dating: Geol. Soc. Amer. Bull., v. 76, p. 523-536.

Rodgers, John, 1968, The eastern edge of the North American continent during the Cambrian and Early Ordovician, p. 141-149, in: Zen, E-an, et al., 1968, Editors, Studies of Appalachian geology, Northern and Maritime: New York Interscience Publishers, John Wiley and Sons, Inc., 475 p.

Scott, W. H., Hansen, E. C., and Twiss, R. J., 1965, Stress analysis of quartz deformation lamellae in a minor fold: Am. Jour. Sci., v. 263, p. 729-746.

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

Stone, S. W., and Dennis, J. G., 1964, The geology of the Milton quadrangle, Vermont: Vermont Geol. Survey Bull. 27, 79 p.

Thompson, J. B., Jr., and Norten, S. A., 1968, Paleozoic regional metamorphism in New England and adjacent areas, p. 319-327, in: Zen, E-an, et al., 1968, Editors, Studies of Appalachian geology, Northern and Maritime: New York Interscience Publishers, John Wiley and Sons, Inc., 475 p.

Thompson, J. B., Jr., Robinson, Peter, Clifford, T. N., and Trask, N. J., Jr., 1968, Nappes and gneiss domes in west-central New England, p. 203-218, in: Zen, E-an, et al., 1968, Editors, Studies of Appalachian geology, Northern and Maritime: New York Interscience Publishers, John Wiley and Sons, Inc., 475 p.

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

Zartman, R. E., Bruch, M. R., Heyl, A. V., and Thomas, H. H., 1967, K-Ar and Rb-Sr ages of some alkalic intrusive rocks from central and eastern United States: Am. Jour. Sci., v. 265, p. 848-870.

Zen, E-an, 1967, Time and space relationships of the Taconic allochthon and autochthon: Geol. Soc. America Spec. Paper 97, 107 p.

_________, 1968, Nature of the Ordovician orogeny in the Taconic area, p. 129-139, _in_: Zen, E-an, _et al_., 1968, Editors, Studies of Appalachian geology, Northern and Maritime: New York Interscience Publishers, John Wiley and Sons, Inc., 475 p.

Trip D

THE LATE PLEISTOCENE OF THE CHAMPLAIN VALLEY, VERMONT

by

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INTRODUCTION

The existing knowledge of Pleistocene events in the Champlain Valley can be discussed in terms of glacial stratigraphy, areal distribution of glacial deposits, and water planes of Lakes Vermont and "New York", (this paper) and the Champlain Sea. The purpose of this field trip is to examine new evidence of glacial and post-glacial events as recently deduced from field study of the Champlain Valley between Burlington and Middlebury, Vermont.

Included in the list of references are recent papers which pertain to the Pleistocene and Recent history of the Champlain Valley. The topographic quadrangles, in order of their first appearance on the field trip, are as follows: Burlington; Fort Ethan Allen; Essex Junction; Mount Philo; Hinesburg; Bristol; Monkton. Acknowledgement is made to Chester A. Howard, Jr. and William R. Parrott, Jr. for data on Stop 16, to Howard for information about surficial deposits near the mountains north of the Winooski River (Fig. 1), to Robert Switzer for field assistance in 1967 and 1968, and to Allen S. Hunt for a review of the manuscript. The work upon which this paper is based was supported in part by funds provided by the U. S. Department of Interior as authorized under the Water Resources Research Act of 1964, Public Law 88-379.

GLACIAL DEPOSITS

Four kinds of glacial deposits can be distinguished. Two varieties of till are found, including a lower, compact, gray-colored till, and a brown, but otherwise similar upper till. At one locality (Stop 16) near Shelburne, Stewart (1961a) has interpreted primarily fabric differences between gray and brown till units as indicative of multiple glaciation. Thus far no evidence of a relatively warm climate separating glacial episodes has been found in Vermont.

In several places is found a highly variable, poorly sorted and washed material that resembles both till and gravel. The deposit is informally referred to in figure 1 as "mantle material", indicating the

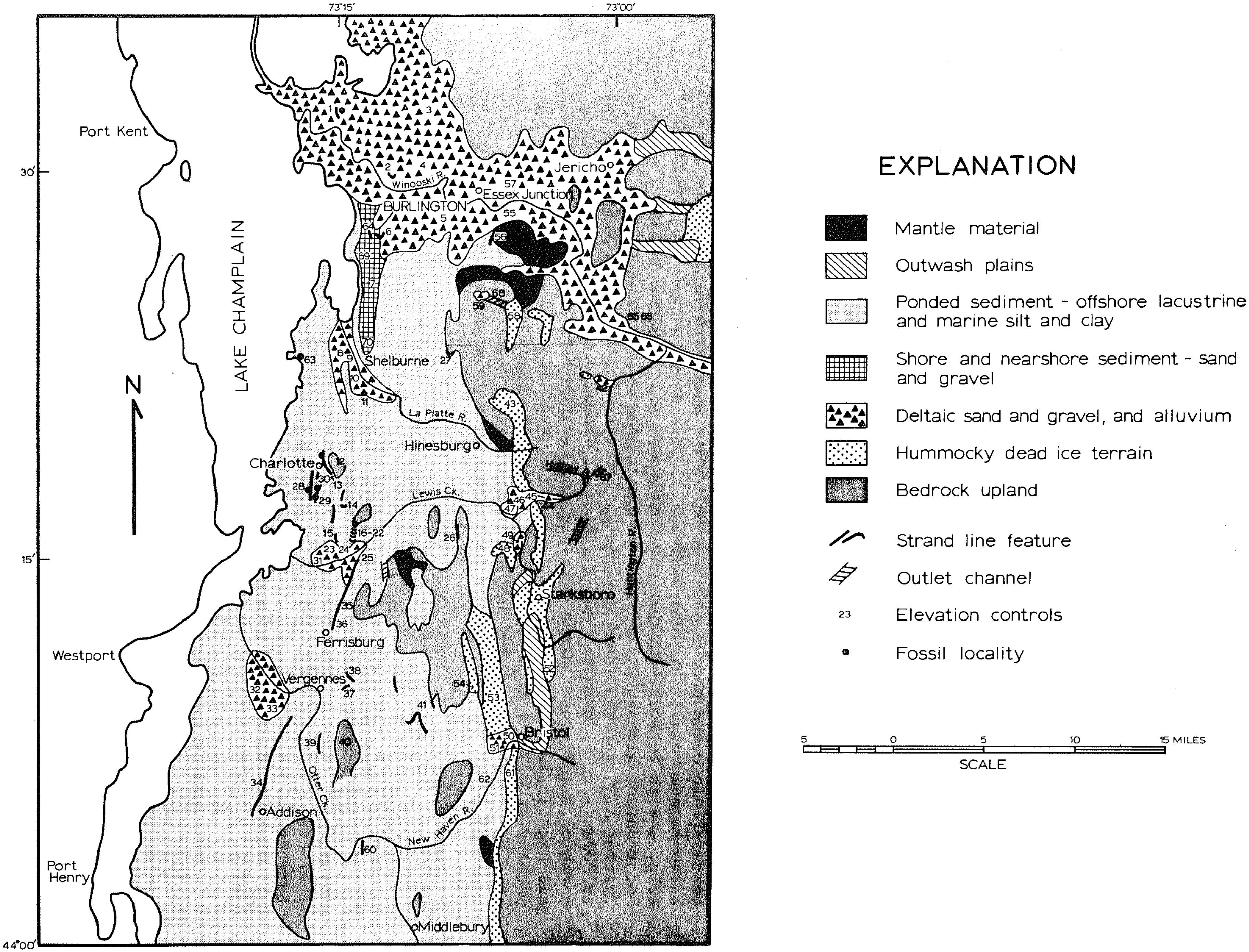


Figure 1 Map of the surficial deposits of the Champlain Valley between Middlebury and vicinity of Burlington.

tendency for the deposit to form a non-descript veneer on the terrain. Mantle material can be definitely related to glaciation by a few scattered exposures of interbedded masses of till; presumably the material in a superglacial drift. Mantle material appears to be most extensive on the south sides of large river valleys.

A fourth glacial deposit is distinguished from mantle material by a distinctive, constructional topography here informally termed "hummocky dead ice terrain". The pattern of this unit on Figure 1 suggests one or possibly more ice margin locations. The scattered patches of hummocky dead ice terrain may delimit an ice margin along the foothills of the Green Mountains. It is tempting to correlate such a margin with morainic features in southern Quebec (Gadd, 1964; McDonald, 1968), and on the north flank of the Adirondack Mountains (Denny, 1966; MacClintock and Terasmae, 1960), thus completing a picture of ice lobation in the Champlain Valley. Although this may be more or less valid, a slightly different, alternative explanation involves successively younger ice marginal features northward in the Champlain Valley. At the core of this problem is the nature of ice retreat in the area. For example, Stewart (1961b) considers regional versus marginal zone retreat of stagnant ice. MacDonald (1968) favors retreat of an active ice front in southern Quebec. Also involved is the degree of geological resolution possible for glacial events.

WATER PLANES

Retreat of the continental ice margin in the Champlain Valley was accompanied by the development of proglacial lake stages of Lake Vermont (Chapman, 1937), "Lake New York" (this paper), and lastly the Champlain Sea (Karrow, 1961). Numerous shoreline features have been identified in this study (Figs. 1 and 2; Appendix). Figure 2 is a cross-sectional plot of the elevation of shoreline and other features against distance along a line oriented N20W, approximately perpendicular to previously determined isobases in the area (Chapman, 1937; Farrand and Gajda, 1962). The positions of enumerated features on Figure 1 were extrapolated along a S70W direction to the cross-section, thus making it possible to compare items of similar uplift and to depict the true rather than apparent tilt of water planes.

The elevation ranges of individual features on Figure 2 reflect the contour intervals of the topographic maps used to determine elevations; more accurate determinations are desirable. Beaches and spits provide relatively precise markers of former water levels but in many cases they are difficult to identify and unquestionable beaches and spits are not numerous. Deltas are relatively inaccurate strandline indicators, but on the other hand they are abundant. Delta elevations given on Figure 2 refer to the delta surface, not the foreset - topset bed contact which, although a better indicator, is rarely observed in this area.

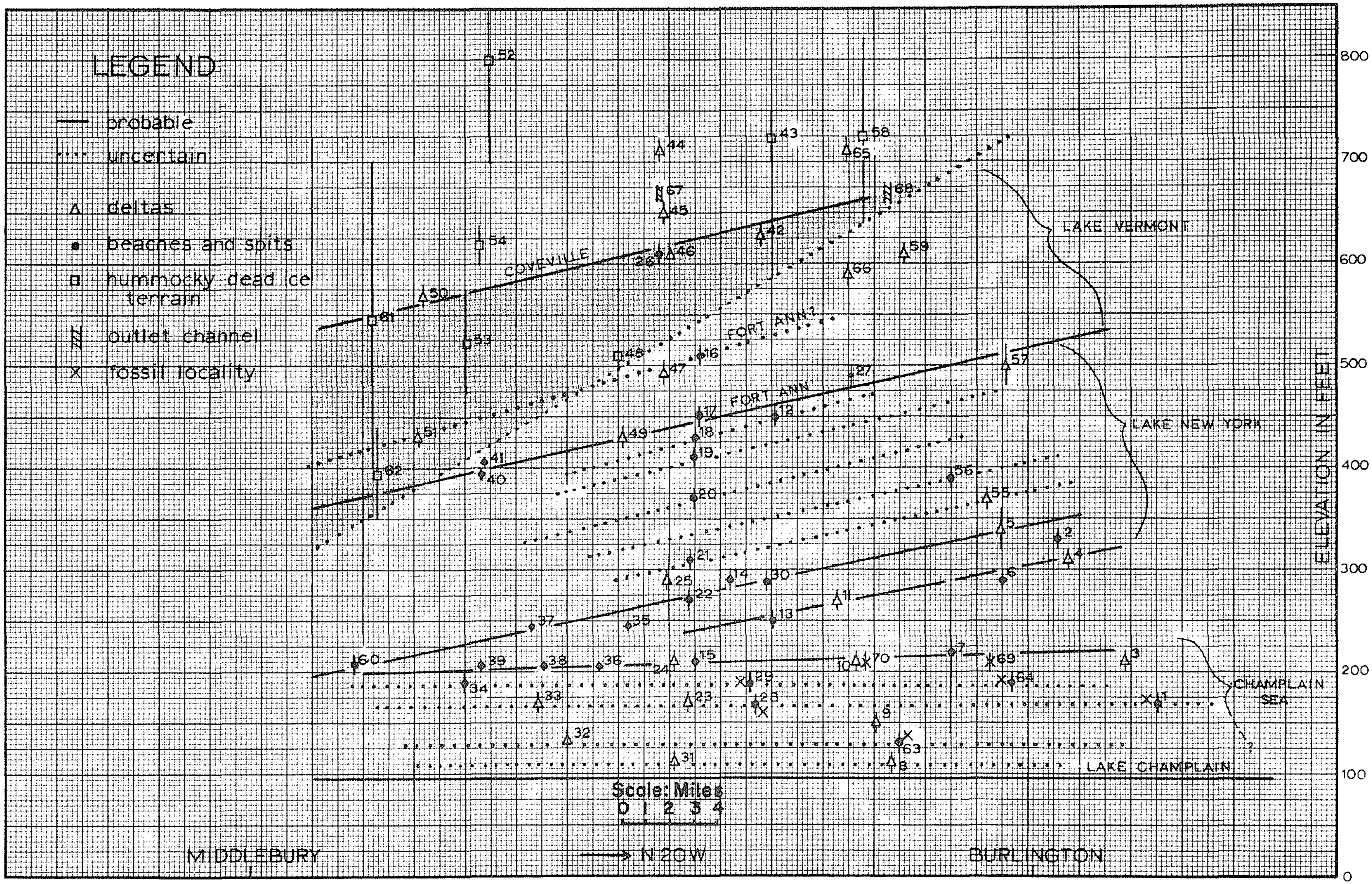


Figure 2 Cross-section of shoreline and other features in the Champlain Valley. Shaded area denotes hummocky dead ice terrain with lacustrine sediment veneer. Vertical bars indicate either the elevation range or the accuracy of the enumerated shoreline or other features.

The correlations of the shoreline features, shown by various lines on Figure 2, are made assuming that the orientations of water planes in the Valley are somewhat as previously described by others.[1] Although higher, older shoreline features do exist, there is not enough evidence to determine whether or not they are related to major, early stages of Lake Vermont (such as Woodworth's, 1905, "Quaker Springs Stage") or to local, small, and separate lakes. The features here identified as Coveville could be related to local lakes, but the upper limit of lacustrine cover on hummocky, dead ice terrain fits nicely with a distinct, widespread level (Fig. 2).

The tilted (5.8 feet per mile) Coveville plane was purposely drawn to intersect a Coveville delta at Brandon (Chapman, 1937), south of the area mapped, and to coincide with the previously recognized outlet channel near Coveville, New York (Woodworth, 1905). The northward extent of the Coveville plane is problematical and a variety of lines of evidence must be considered. Item 1. The 560-580 foot Bristol delta (No. 50, Figs. 1 and 2), which lies on the Coveville plane, can be correlated upstream via an outwash plain and longitudinal projection into hummocky dead ice terrain at Starksboro. Thus, the position of an ice margin is located at Starksboro during Coveville time. Item 2. Another Coveville delta exists at Hollow Brook (No. 46), north of Starksboro. In order for Hollow Brook to be confluent with Coveville waters an ice margin north of Hollow Brook is likely, an interpretation which conflicts with Item 1. Items 1 and 2 can be reconciled if: a) northward ice retreat occurred during Coveville time; or b) the configuration of the continental ice margin was highly irregular and complex; or c) the Starksboro hummocky dead ice terrain was constructed by a mass of ice separate from the main body of continental ice; or d) at least one of the deltas is misidentified as Coveville. Item 3. In any case, because the Hollow Brook deltas probably were formed by ponded water which escaped from the Winooski Valley via the Huntington River and Hollow Brook Valleys, an ice margin blocking the Winooski Valley (east of Williston) is indicated. Item 4. Other evidence of ice in the lower Winooski Valley during Coveville time includes: absence of a lacustrine sediment veneer on hummocky dead ice terrain below the Coveville level near Williston (No. 58); presence of a low-level delta (No. 59) formed by escape of local lake water from Winooski Valley via the Oak Hill outlet channel (No. 68); absence of a Coveville-level delta in the Winooski Valley. Item 5. However, Connally (1967) has identified a Coveville delta in the Lamoille Valley (Fig. 1). Also, a problematical delta occurs at the Coveville level in the Huntington Valley. Obviously the problem of the location of the ice margin during Coveville time is complex and cannot be resolved at the present time.

[1]Other interesting hypotheses include: control of Fort Ann level by the present Champlain - Hudson divide (147 feet) near Fort Edward, New York; merger of Lake Albany and Coveville water to form a single level with a hinge line, or temporary blockage of the Fort Edward divide by ice or drift, forcing Coveville drainage through South Bay or Lake George lowlands.

Below the Coveville stage are a few shoreline features which may or may not be correlative as a single plane. However, if connected as shown on Figure 2, the southward projection of the tilted (6.7 feet per mile) plane coincides with the outlet for the Fort Ann stage of Lake Vermont (Chapman, 1937). Thus, the level may be tentatively identified as Fort Ann. However, the slightly greater tilt of this level compared to the Coveville plane suggests an erroneous identification of either Fort Ann or Coveville planes, or both.

Numerous slightly lower shoreline features lie on a well-defined plane (5.3 feet per mile tilt) which also extends to the Fort Ann outlet. The abundance and well-developed character of shoreline features at this level suggest that a major episode of Lake Vermont is represented. No evidence is available to delimit the ice margin during this time but apparently it was north of the area of study.

Below the major Fort Ann level are numerous shoreline features, many of which are too scattered to clearly define any planes. However, several water planes are relatively well documented. "Lake New York" levels (Fig. 2) are so tilted (5.0 feet per mile) that their southward projection intersects Lake Champlain near its southern extremity. In these cases, therefore, southward drainage of Champlain Valley waters into the Hudson Valley is precluded. An alternative, that the southward-projecting planes pass through a hinge line and hence to the Fort Ann outlet, is rejected because of the absence of detectable hinges on Lake Vermont strandlines. It is suggested that after the Fort Ann stage of Lake Vermont, Champlain Valley drainage shifted northward to the St. Lawrence lowland, creating a new system of proglacial lakes informally referred to here as "Lake New York". Progressive ice retreat probably opened successively lower drainage routes on the northwest slope of the Appalachian uplands. The similarity between the tilts of the lowest Fort Ann stage and Lake New York suggests that no significant differential isostatic uplift occurred during this time. Physiographic considerations require that Lake New York drainage must have developed immediately after the lowest Fort Ann water plane on Figure 2, or in other words, that no Lake Vermont level below that shown on Figure 2 could have existed. The northern extent of the Lake New York planes is unknown.

Influx of saline water of the Champlain Sea into the Champlain Valley is indicated by the occurrence of certain mollusks (Wagner, 1967). The highest (and therefore oldest?) of such fossils occur in nearshore sediment at localities 69 and 70 (Fig. 2). Associated beach deposits and deltas delimit a well-defined, tilted (0.63 feet per mile) water plane (Fig. 2). A radiocarbon date of 11,230 $\pm$ 170 years B. P. (I - 3647) has been obtained from shells in a Champlain Sea beach (No. 29, Fig. 2) near the highest marine limit.

The absence of water planes with intermediate tilts between Lake New York and the Champlain Sea is curious. One possible explanation is that no sufficiently long stabilization of water level occurred for shoreline development just prior to the Champlain Sea episode; significant isostatic movements would have had to occur during this time. Another explanation is that ice retreat permitted drainage of the Champlain Valley to a low level, possibly below modern Lake Champlain, during which time isostatic movements occurred. The latter hypothesis is supported by MacClintock's (1958) similar interpetation based on oxidized lacustrine sediment beneath unoxidized marine clays.

The nature and timing of the transition from sea water back to fresh water is unknown. Marine mollusks occur at elevations almost as low as the present level of Lake Champlain but due to the relatively large depth tolerances of the species, the lowermost level of the Champlain Sea is unknown. A minimum estimate of 9,500 years B.P. for the lower age limit of the Champlain Sea was made by Terasmae (1959) from the St. Lawrence lowland.

FIELD TRIP STOPS[1]

Stop 1. Champlain Sea beach (No. 1, Figs. 1 and 2); 3.5 miles northeast of Burlington; exposures in gravel pits on both sides of barn show excellent beach structures and marine mollusks.

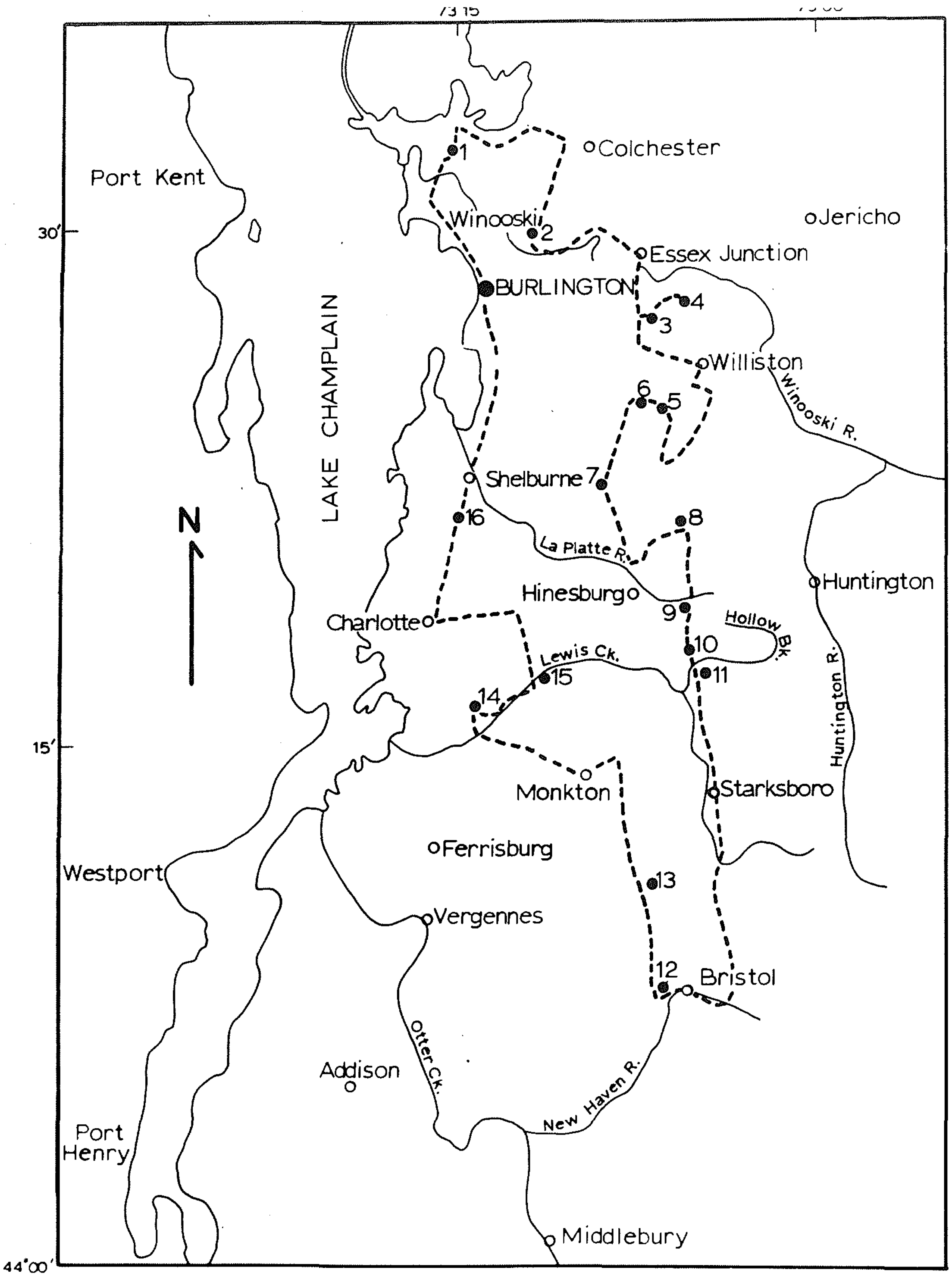
Stop 2. Lake New York beach-spit and delta (No. 2, Figs. 1 and 2); end of North Street at northern margin of Winooski; gravel pits are in beach material; slightly lower bench to west is composed of deltaic sand which correlates with No. 4, Figures 1 and 2.

Stop 3. Lake New York beach; No. 56; 1.7 miles southeast of Essex Junction; one of the best developed beach landforms known in the area; terrain immediately east of beach is lake clay and west of beach is a lower level Lake New York delta. Note dunic(?) landforms between stops 3 and 4.

Stop 4. Fort Ann delta; correlative with No. 57; 1.7 miles southeast of Essex Junction; gravel pit shows channel structures characteristic of deltas in this area.

Stop 5. Oak Hill outlet channel; No. 68; 4.8 miles southeast of Essex Junction; gravel pit in channel bottom shows very coarse, poorly sorted gravel; outlet drained a Winooski Valley Lake in post-Coveville (?) time

[1]Figure 3 depicts the field trip route.



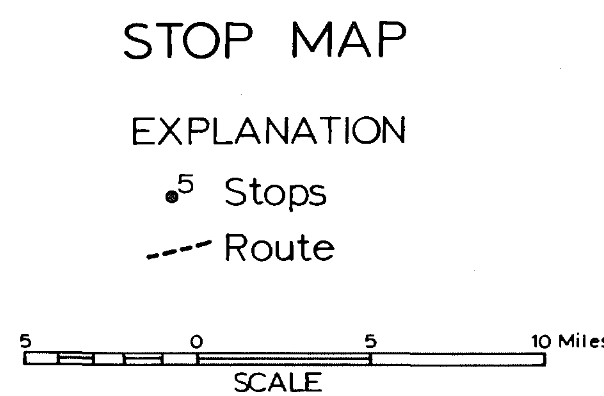


Figure 3 Map of field trip route and locations of stops.

and therefore proves (?) that ice blocked the Winooski Valley during Coveville time.

Stop 6. Ice-contact delta; No. 59; 1.2 miles northeast of stop 5; numerous gravel pits in vicinity show progressively finer material from stop 5 to stop 6 where cross-bedded fine gravel, sand, and silt are exposed in pit; ice-contact nature of delta is suggested by poorly sorted gravel in nearby vegetated area to southwest; delta was apparently formed by Oak Hill channel water drainage into local lake impounded by continental ice.

Stop 7. Fort Ann bench; No. 27; 3 miles northwest of Hinesburg at Junction of Routes 116 and 2A; road cut exposes coarse, poorly sorted gravel underlying (?) boulder-strewn (lag?) bench on hillside. A problem exists here in that the topography and the lag (?) suggest an erosional environment whereas the road cut indicates a depositional environment. Possibly mantle material (see text) has been modified by wave action.

Stop 8. Hummocky dead ice terrain; No. 43; 1.8 miles northeast of Hinesburg; stop is at gravel pit showing poorly sorted character of deposit; ice contact structures are visible in places in the vicinity from time to time.

Stop 9. Ice-contact delta?; 1.5 miles southeast of Hinesburg; this is a brief photography stop. Large scale cross-bedding is in kame terrace, delta, or ice-contact delta(?). The last named possibility is favored here.

Stop 10. Kame terrace; 2 miles southeast of Hinesburg along Route 116; gravel pits in vicinity show characteristic ice-contact features.

Stop 11. Pre-Coveville (Nos. 44 and 45), Coveville (No. 46), and upper Fort Ann (No. 47) deltas; 3 miles southeast of Hinesburg along Route 116; deltas were constructed by water from a local Winooski Valley lake which drained through an outlet channel (No. 67) at the divide between Hollow Brook and Huntington River; gravel pits show coarse gravel with large scale foreset bedding and less obvious topset bedding in places.

Stop 12. Coveville (No. 50) and upper Fort Ann (No. 51) deltas; the village of Bristol is located on the Coveville delta and the upper Fort Ann delta just west of Bristol; gravel pits show features similar to those at stop 11.

Stop 13. Lacustrine sediment veneered hummocky dead ice terrain; No. 53, 3 miles north of Bristol; gravel pit exposes ice-contact material overlain by and interbedded (?) with lacustrine sediment; features suggest glacial and lacustrine sedimentation in close proximity.

Stop 14. Shoreline features of Lakes Fort Ann and New York; Nos. 16-22; southwest side of Mount Philo; a spectacular succession of six wave-cut benches and one beach-spit.

Stop 15. Two(?) till locality; 3.2 miles northeast of North Ferrisburg; north bank of Lewis Creek. Stewart (1961a, Stop 6) reports two till units with intervening lake sediment. The present investigation showed complex vertical and horizontal variations. At the east end of the exposure occurs (bottom to top): gray till (25 feet); till or gravel (30 feet); slumped lake clays. At the west end: gray till (25 feet); lake sediment (25 feet). At the middle of the exposure as many as five till units with intervening lake sediment have been counted. Lateral relationships are difficult to determine due to the steepness of the exposure but facies changes are likely.

Stop 16. Two(?) till locality; 1.3 miles south of Shelburne Village along Route 7; stream bank exposures of a lower gray till and upper brown till; Stewart's (1961a, Stop 4) type locality for the Shelburne (gray) and Burlington (brown) tills. At the time of this writing, fabric and clay mineralogy analyses are being made. This information will hopefully be available for field trip presentation.

REFERENCES CITED

Chapman, D. H., 1937, Late-glacial and postglacial history of the Champlain Valley: American Jour. Sci., v. 34, p. 89-124.

Connally, G. G., 1967, Glacial geology of Mount Mansfield Quadrangle, Vermont: Program Abstracts, Second Annual Meeting, Geological Society of America, Northeastern Section, p. 20.

Denny, C. S., 1966, Surficial geology of the Plattsburgh area: The Empire State Geogram, v. 4, no. 3, p. 6-10.

____________, 1967, Surficial geologic map of the Dannemora Quadrangle, and part of the Plattsburgh Quadrangle, New York: United States Geol. Sur. Map GQ-635.

Farrand, W. R., and Gajda, R. T., 1962, Isobases on the Wisconsin marine limit in Canada: Geog. Bull., no. 17, p. 5-22.

Gadd, N. R., 1964, Moraines in the Appalachian region of Quebec: Geol. Soc. America Bull., v. 75, p. 1249-1254.

____________, 1967, Pattern of glacial recession in southeastern Quebec: Program Abstracts, Second Annual Meeting, Geological Society of America, Northeastern Section, p. 28.

Karrow, P. F., 1961, The Champlain Sea and its sediments, in: Leggett, R. F., editor, Soils in Canada, geological, pedological, and engineering studies: Royal Soc. Canada Spec. Pub., no. 3, p. 97-180.

McDonald, B. C., 1967a, Pleistocene events and chronology in the Appalachian region of Southeastern Quebec, Canada: Ph.D. Dissertation, Yale University, 161 p.

______________, 1967b, Wisconsin stratigraphy and ice movement directions in southeastern Quebec, Canada: Program Abstracts, Second Annual Meeting, Geological Society of America, Northeastern Section, p. 41-42.

______________, 1968, Deglaciation and differential post-glacial rebound in the Appalachian region of southeastern Quebec: Jour. Geol., v. 76, p. 664-677.

MacClintock, P., 1958, Glacial geology of the St. Lawrence seaway and power projects: New York State Mus. and Sci. Ser., 26 p.

______________, and Dreimanis, A., 1964, Reorientation of till fabric by overriding glacier in the St. Lawrence valley: American Jour. Sci., v. 262, p. 133-142.

______________, and Terasmae, J., 1960, Glacial history of Covey Hill: Jour. of Geol., v. 68, p. 232-241.

Schafer, J. P., and Hartshorn, J. H., 1965, The Quaternary of New England, in: Wright, H. E., and Frey, D. G., editors, The Quaternary of the United States, VII INQUA Congress Review Volume, 922 p.

Stewart, D. P., 1961a, The glacial geology of northern Vermont: in: Doll, C. G., editor, Guidebook New England Intercollegiate Geological Conference, Fifty-third Annual Meeting.

______________, 1961b, The glacial geology of Vermont: Vermont Geol. Sur. Bull. 19, 124 p.

______________, and MacClintock, 1964, The Wisconsin stratigraphy of northern Vermont: American Jour. Sci., v. 262, p. 1089-1097.

Terasmae, J., 1959, Notes on the Champlain Sea Episode in the St. Lawrence lowlands, Quebec: Science, v. 130, p. 334-336.

____________, 1965, Surficial Geology of the Cornwall and St. Lawrence seaway project areas, Ontario: Geol. Sur. Canada Bull. 127, 54 p.

Wagner, F. J. E., 1967, Published references to Champlain Sea faunas 1837-1966 and list of fossils: Geol. Sur. Canada Paper 67-16, 82 p.

Woodworth, J. B., 1905, Ancient water levels of the Champlain and Hudson Valleys: New York State Museum Bull. 348, 265 p.

APPENDIX

No.	Feature	Quadrangle	Miscellaneous	Elevation (ft.)
1.	beach	Ft. Ethan Allen	mollusks	160-180
2.	beach-spit	Ft. Ethan Allen	----	320-340
3.	delta	Ft. Ethan Allen	Winooski River	200-220
4.	delta	Ft. Ethan Allen	Winooski River	300-320
5.	delta	Burlington	Winooski River	320-360
6.	beach	Burlington	UVM Poultry Research	280-300
7.	beach	Burlington	UVM agricultural Research	140-300
8.	delta	Burlington	LaPlatte River	100-120
9.	delta	Burlington	LaPlatte River	140-160
10.	delta	Mt. Philo	LaPlatte River	200-220
11.	delta	Mt. Philo	LaPlatte River	260-280
12.	beach?	Mt. Philo	possibly kame terrace	440-260
13.	beach	Mt. Philo	----	240-260
14.	beach	Mt. Philo	----	280-300
15.	beach	Mt. Philo	----	200-220
16.	beach-spit	Mt. Philo	----	500-520
17.	beach	Mt. Philo	----	440-460
18.	beach	Mt. Philo	----	420-440
19.	beach	Mt. Philo	----	400-420
20.	beach	Mt. Philo	----	360-380
21.	beach	Mt. Philo	----	300-320

No.	Feature	Quadrangle	Miscellaneous	Elevation (ft.)
22.	beach	Mt. Philo	----	260-280
23.	delta	Mt. Philo	Lewis Creek	160-180
24.	delta	Mt. Philo	Lewis Creek	200-220
25.	delta	Mt. Philo	Lewis Creek	280-300
26.	beach?	Mt. Philo	possibly kame terrace	600-620
27.	beach?	Mt. Philo	possibly kame terrace	480-500
28.	beach	Willsboro	whale locality	160-180
29.	beach	Willsboro	mollusks	180-200
30.	beach?	Willsboro	----	280-300
31.	delta	Port Henry	Lewis Creek	100-120
32.	delta	Port Henry	Otter Creek	120-140
33.	delta	Port Henry	Otter Creek	160-180
34.	beach	Port Henry	----	180-200
35.	beach	Monkton	----	240-250
36.	beach	Monkton	----	200-210
37.	beach	Monkton	----	240-250
38.	beach	Monkton	----	200-210
39.	beach	Monkton	----	200-210
40.	beach	Monkton	----	390-400
41.	beach-spit	Monkton	----	400-410
42.	delta	Huntington	Huntington River tributary	620-640
43.	hummocky terrain	Hinesburg	----	700-750

No.	Feature	Quadrangle	Miscellaneous	Elevation (ft.)
44.	delta	Hinesburg	Hollow Brook	700-720
45.	delta	Hinesburg	Hollow Brook	640-660
46.	delta	Hinesburg	Hollow Brook	600-620
47.	delta	Hinesburg	Hollow Brook	460-500
48.	hummocky terrain	Hinesburg	lake sediment veneer	500-520
49.	delta	Hinesburg	Lewis Creek	420-440
50.	delta	Bristol	New Haven River	560-580
51.	delta	Bristol	New Haven River	420-440
52.	hummocky terrain	Bristol	----	700-900
53.	hummocky terrain	Bristol	lake sediment veneer up to 560-580 feet	480-580
54.	hummocky terrain	Bristol	----	600-640
55.	delta	Essex Junction	Winooski River	360-380
56.	beach	Essex Junction	----	380-400
57.	delta	Essex Junction	Winooski River	480-520
58.	hummocky terrain	Essex Junction	----	640-820
59.	delta	Essex Junction	Oak Hill channel	600-620
60.	beach	Middlebury	----	200-220
61.	hummocky terrain	South Mountain	----	400-700
62.	hummocky terrain	South Mountain	lake sediment veneer	360-440

No.	Feature	Quadrangle	Miscellaneous	Elevation (ft.)
63.	beach?	Willsboro	mollusks	120-140
64.	beach	Burlington	mollusks	180-200
65.	delta?	Richmond	Winooski River	700-720
66.	delta?	Richmond	Winooski River	580-600
67.	outlet channel	Hinesburg	Hollow Brook	660-680
68.	outlet channel	Essex Junction	Oak Hill	660-680
69.	nearshore sediment	Burlington	mollusks	200-220
70.	nearshore sediment	Burlington	mollusks	200-220

Trip E

STRATIGRAPHY OF THE CHAZY GROUP (MIDDLE ORDOVICIAN) IN THE NORTHERN CHAMPLAIN VALLEY

by

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INTRODUCTION

The Chazy Limestone (the oldest Middle Ordovician Group of the Champlain Valley) was first named by Emmons (1842) from exposures 15 miles north of Plattsburgh at Chazy, New York. Here and elsewhere in the northern Champlain Valley (Fig. 1) the unit outcrops on a variety of normal fault blocks. Given the low dips and heavy cover, Chazy stratigraphy is most easily understood from various shore outcrops around Lake Champlain. Valcour Island, southeast of Plattsburgh, offers perhaps the best section of the Chazy, and has been intensively studied (Raymond, 1905; Hudson, 1931; Oxley and Kay, 1959; Fisher, 1968; Shaw, 1968). The Isle La Motte, Vermont, exposures to be covered in this trip and trip F are those studied by many of the same authors and, in addition, display the lower contact of the Chazy with the underlying Ordovician dolostones of Canadian age.

In the northern Champlain Valley (Valcour Island and north to the International Boundary), the Chazy Limestone (now Group) consists of about 800 feet of quartz sandstones, calcarenites, dolomitic calcilutites and biohermal masses (Fig. 2). Three formations, Day Point, Crown Point, and Valcour, in ascending order, were proposed by Cushing (1905) and have persisted to the present, albeit with some controversy (Fisher, 1968; Shaw, 1968). Oxley and Kay (1959) further subdivided the Day Point and Valcour into members, those of the Day Point (Head, Scott, Wait, Fleury) coming from southern Isle La Motte in the area to be visited. Shaw and Fisher experienced difficulty in using the Valcour subdivisions outside of their type areas at South Hero, Vermont.

DESCRIPTION AND INTERPRETATION OF CHAZY GROUP LITHOLOGIES

Day Point Formation

With the exception of the biohermal masses on Isle La Motte (Trip F), the Day Point consists of a basal, cross-bedded quartz sandstone, followed by alternating units of shale, more sandstone, calcarenite, and topped with

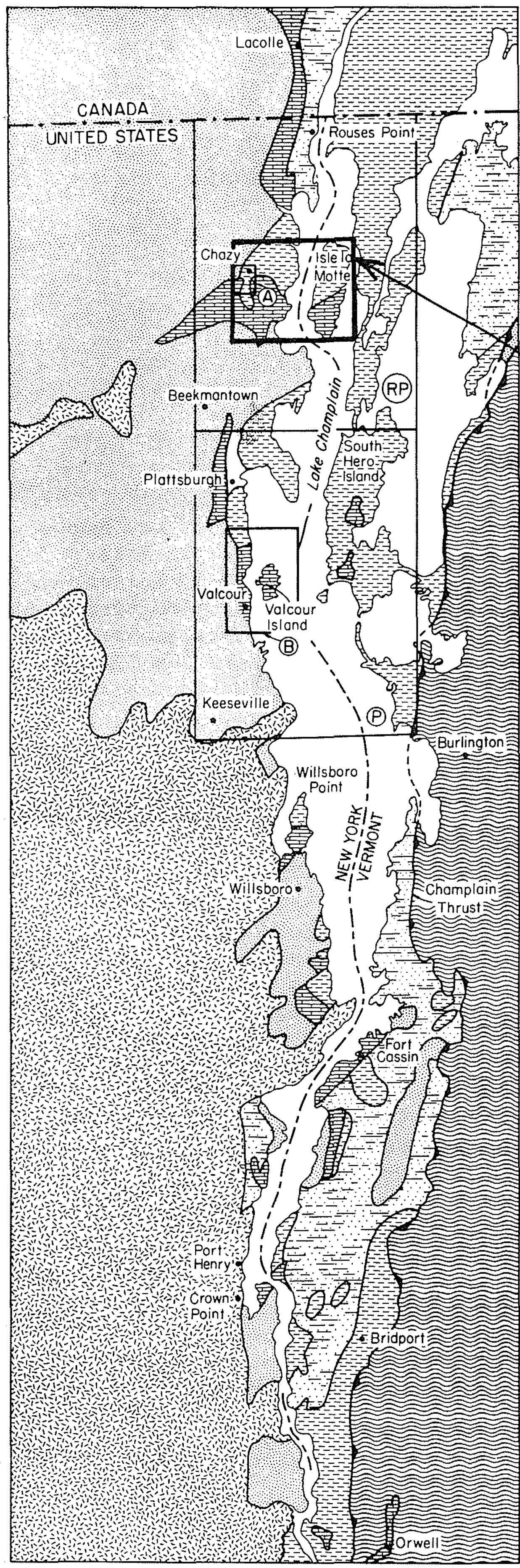

Figure 1. INDEX MAP

of
Champlain Valley
and portions of
New York, Vermont and Quebec
(area of figure 5)

EXPLANATION

- Pleistocene (sands, clays)
- Post-Chazy Ordovician sedimentary rocks (mainly shales)
- Chazy Group (limestones)
- Pre-Chazy Cambrian-Ordovician sedimentary rocks (sandstones, dolostones)
- Cambrian-Ordovician metamorphic strata
- Precambrian rocks (gneiss, metanorthosite, charnockite, marble, quartzite)
- Ⓐ Chazy area
- Ⓑ Valcour area
- Ⓟ Plattsburgh quadrangle
- (RP) Rouses Point quadrangle

Geology modified from D.W. Fisher et al., (1962)

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SCALE IN MILES

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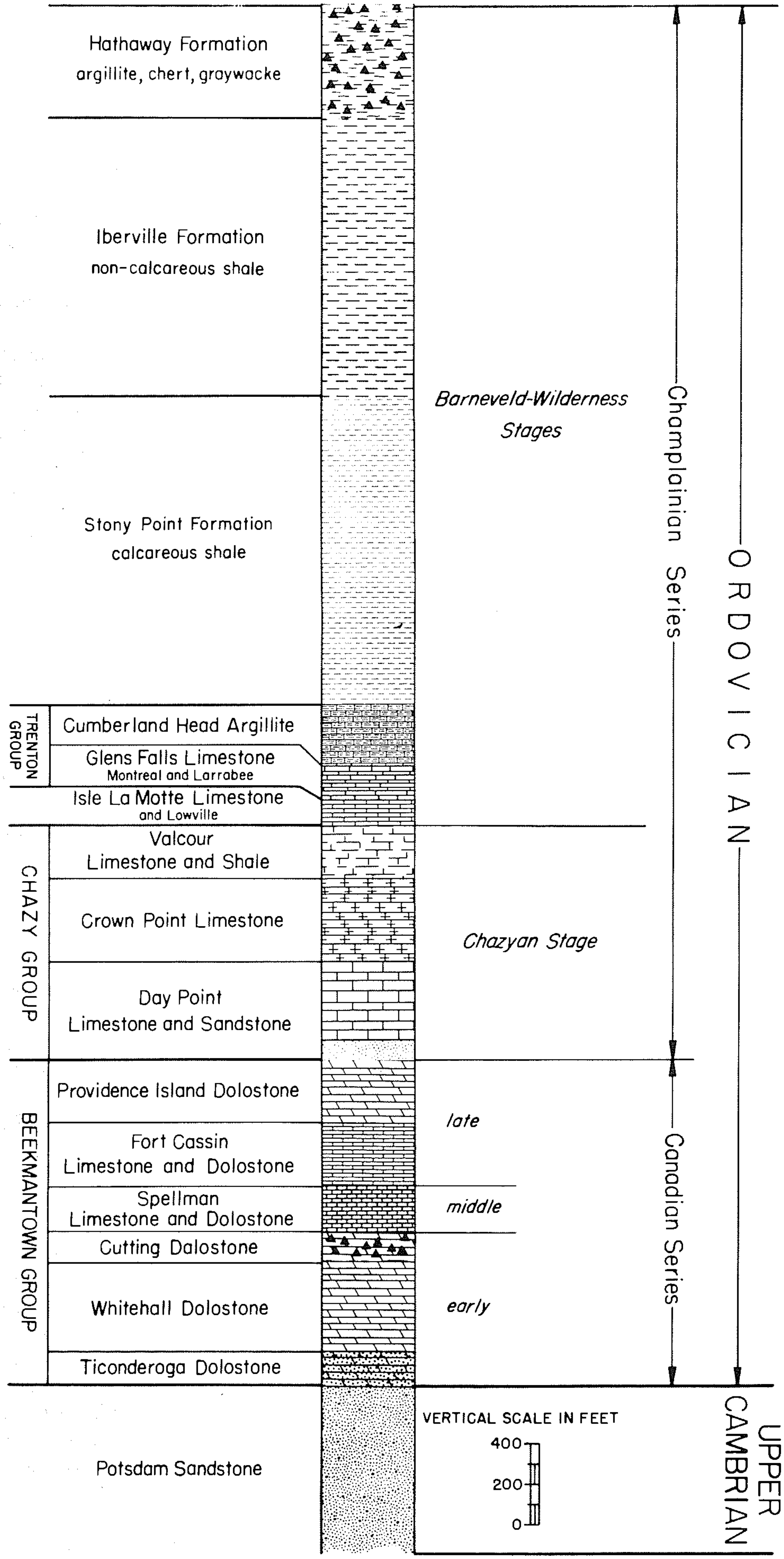


Figure 2. Generalized Stratigraphic Column—Champlain Valley

a relatively thick (35 feet) calcarenite unit (the Fleury Member). The lower sandstone, with its cross-bedding, presence of Lingula as nearly the only fossil, and overlying the supratidal Lower Ordovician, is clearly transgressive and probably of very shallow water origin. This is further borne out by the presence of an oolite band in some of the sections around Chazy, New York. The source of the sand is unknown and no petrographic studies on this formation or on most of the other Chazy units have been undertaken. Derivation from Cambrian sandstones exposed on lowly emergent land to the west appears feasible. The calcarenites are primarily echinodermal in origin, although bryozoans and trilobites also occur abundantly, particularly in the Fleury (Ross, 1963, 1964; Shaw, 1968). Again, shallow water seems indicated, although probably subtidal judging from the abundant faunas (compare Laporte, 1968).

At Valcour Island and on the adjacent shore at Day Point, the Upper Fleury calcarenites are interbedded with dark, muddy limestones, some of which contain the varied silicified trilobite fauna described by Shaw (1968).

Crown Point Formation

The Crown Point Formation begins where muddy limestones become the dominant lithology. A striking feature of this formation is the abundance of thin (maximum 1/2 inch thick) dolomite stringers. Thin section analysis of many of these irregular stringers indicates that they are composed of argillaceous material, calcite grains, and scattered dolomite rhombs (Barnett, pers. comm., 1969). Judging from the abundant faunas (gastropods, trilobites, ostracodes, brachiopods) and their preservation (some trilobites and ostracodes articulated), this lithology represents somewhat deeper and less agitated water. This leaves the origin of the dolomite to be explained inasmuch as recent discussions of dolomite have focused on a supratidal origin. Possibly this dolomite is secondary. Similar lithologies are known in the Ordovician of the southern Appalachians and Nevada and present a good petrologic problem. The 200-300 feet of the Crown Point Formation has never been subdivided into members, attesting to its homogeneity. Twenty-five miles south of Valcour Island, at Crown Point, New York, nearly the whole section (250 feet) is comprised of Crown Point Formation lithology (Fig. 3).

Valcour Formation

The Valcour Formation is characterized by a return to calcarenites, interspersed with limestones of Crown Point aspect. In addition, much of the Valcour as well as the underlying Crown Point display well-developed bioherms consisting of stromatoporoids, bryozoans, calcareous algae, and corals with an accompanying fauna of trilobites, brachiopods, cephalopods, and echinoderms. Spectacular examples of these will be covered on Trip F. The channels in these reefs, the packing of these channels with trilobite and nautiloid fragments, and the accompanying carbonate sands again argue for relatively shallow water, with the more typical muddy limestones occupying slightly deeper basins between.

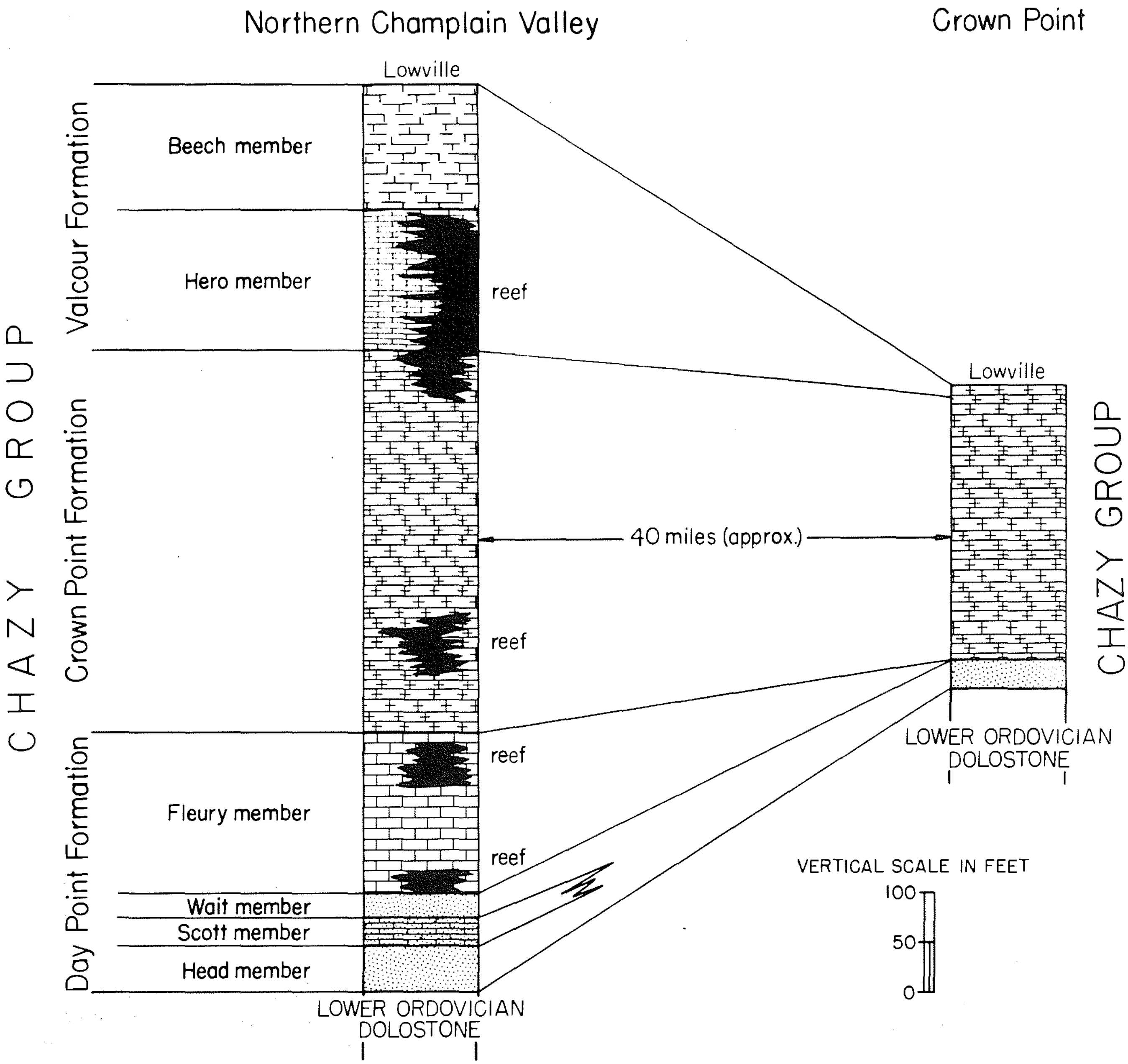


Figure 3. Lithologic Correlation

The Valcour is overlain by rock units usually assigned to the Black River Group, although outcrop or exact paleontological continuity with the type Black River of central New York and Ontario cannot be demonstrated. Paleontologically, the Isle La Motte (Fig. 2) is quite similar to the Chaumont, so that the 'Black River' designation for these units can be defended. The best exposure of the top of the Chazy will be seen at stop 5 south of Chazy.

PALEOGEOGRAPHIC SETTING OF THE CHAZY GROUP

As mentioned above, the lower Chazy Group evidently represents a transgressive sequence over Lower Ordovician dolostones. The relative thinness, lack of abundant clastics, shallow water features, lack of volcanics and predominance of 'shelly' rather than graptolitic faunas all argue for a setting on the platform or at best at the very edge of the miogeosyncline. Most paleogeographic reconstructions of Chazyan time (Kay, 1947) exhibit this relationship. Although the Chazy Group thins and disappears southward and westward into New York State, lithologically and faunally similar units persist northward to the Montreal area and eastward into the Mingan Islands of eastern Quebec (Hofmann, 1963; Twenhofel, 1938). Westward thrusting of perhaps as much as 100 km along Logan's Line has removed much of the miogeosyncline from view to the east, leaving us with either unfossiliferous or graptolite-bearing rocks which defy exact comparison to the Chazy Group. Speculation as to the exact geography of the Appalachian geosyncline in this area during Chazyan time thus is hazardous.

FAUNAS OF THE CHAZY GROUP

Raymond (1906) identified three faunal zones in the northern Champlain Valley Chazy Group, corresponding roughly to the three formations proposed by Cushing (1905). These were not really assemblage zones in the modern sense but relied heavily on two brachiopods and the large gastropod *Maclurites magnus* LeSueur (Pl. 1, Fig. 3). Raymond and later workers (Welby, 1961; Erwin, 1957; Oxley and Kay, 1959) also thought that the trilobite *Glaphurus pustulatus* (Walcott) first appeared at the base of the Valcour Formation. All of the above instances now appear to be examples of local abundance and/or facies control, although they are of some stratigraphic use locally in the Champlain area. Ross (1963, 1964), Cooper (1956), and Shaw (1968), using bryozoans, brachiopods and trilobites, respectively, were unable to make meaningful faunal subdivisions of the Chazy Group. Nevertheless, the Group as a whole is distinctive, marking as it does the first appearance of stromatoporoids, primitive tetracorals, bryozoans (?), and primitive pelecypods. In addition, twenty-four genera of trilobites appear first in the North American Ordovician here in the Chazy Group. By contrast, graptolites and several long-ranging groups of trilobites such as robergioids and agnostids are absent from the Group, probably as a result of facies control or restricted oceanic circulation.

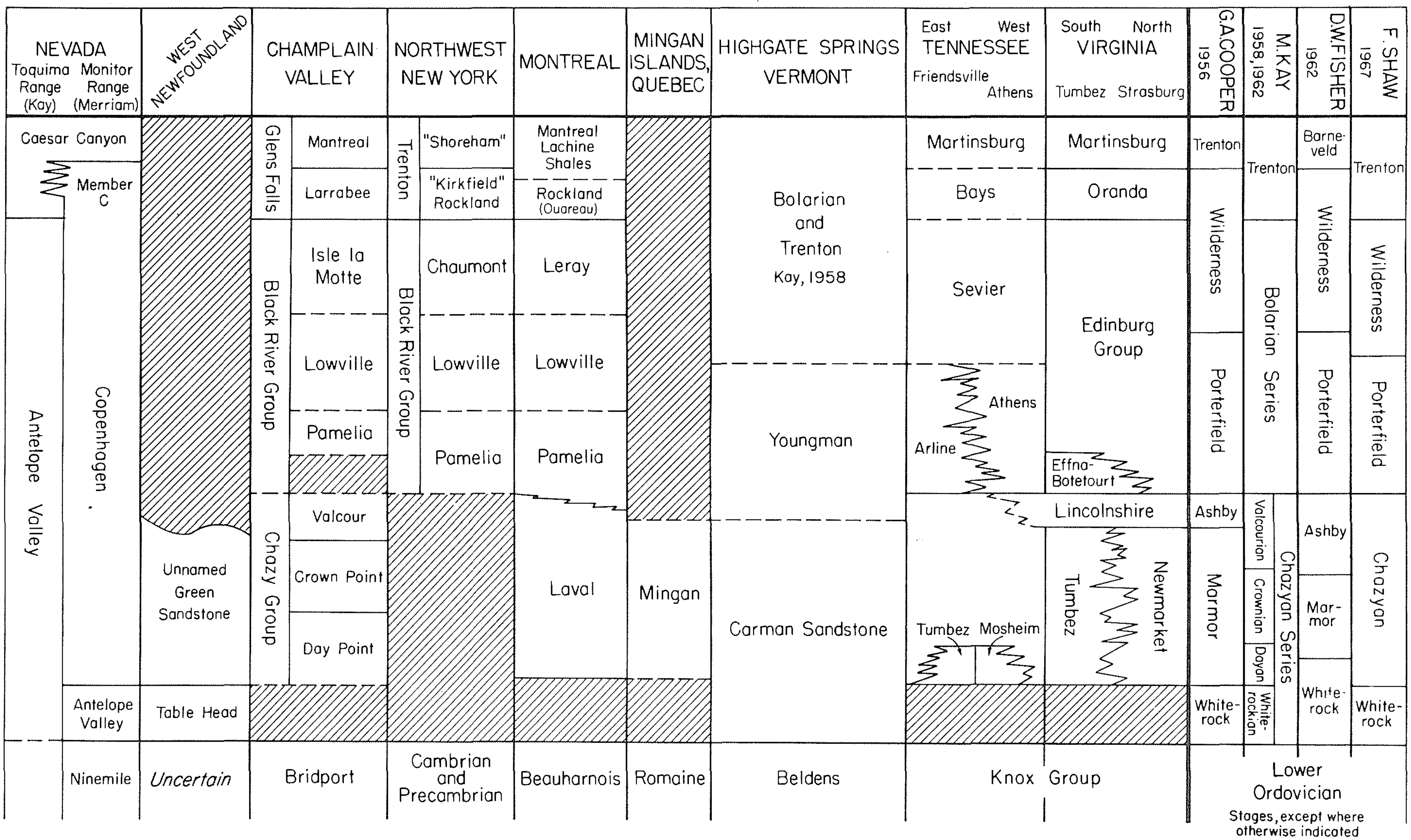


Figure 4 Correlation Chart

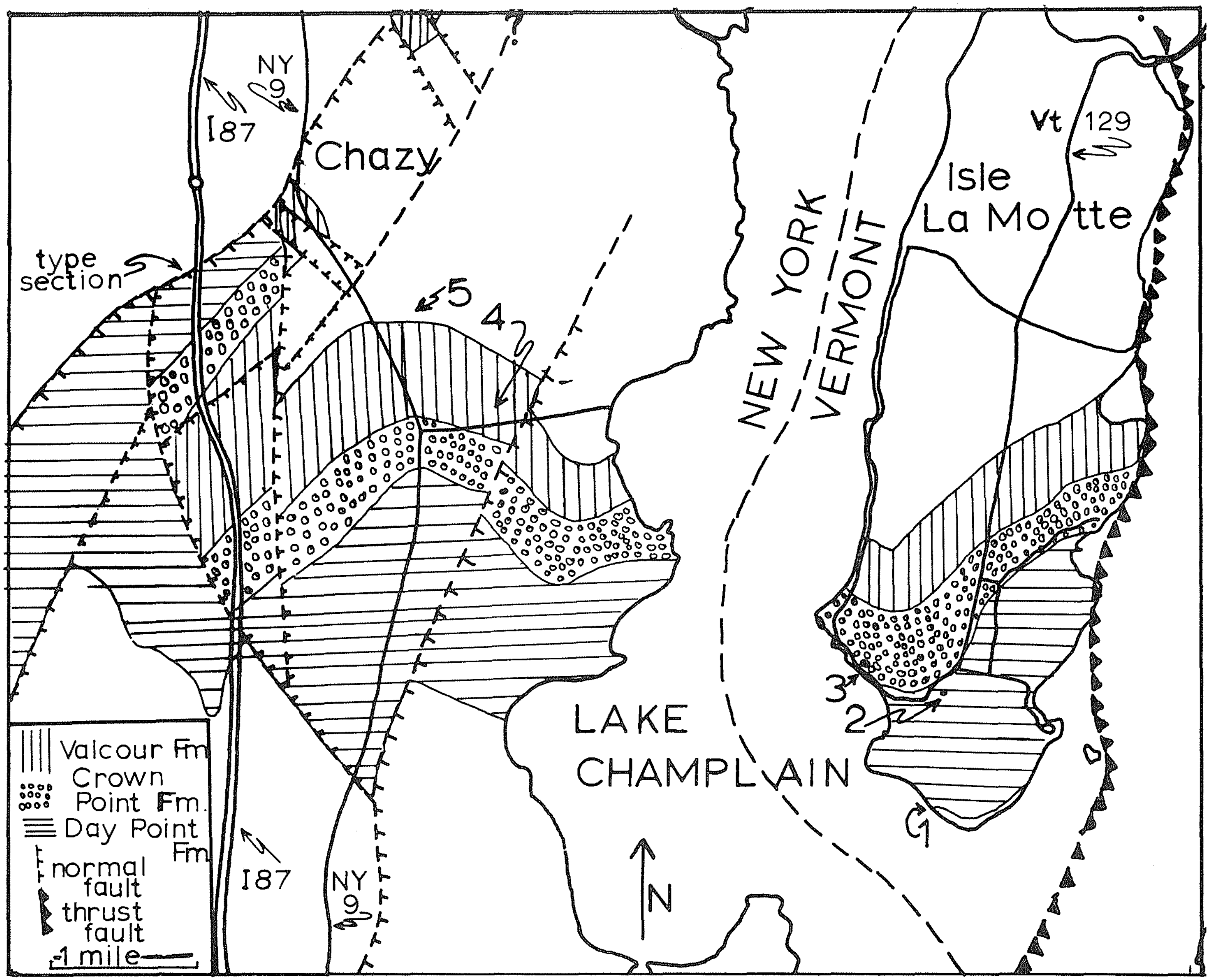
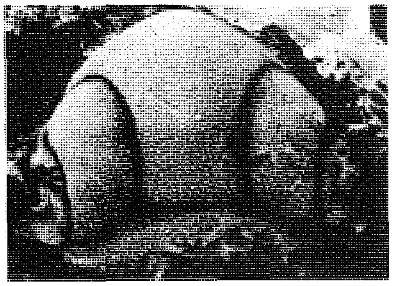


Figure 5 Geologic sketch map of the Chazy-Isle La Motte area, New York and Vermont. Numbers 1-5 are Field Trip Stops. (modified from Fisher, 1968, Plate 1)



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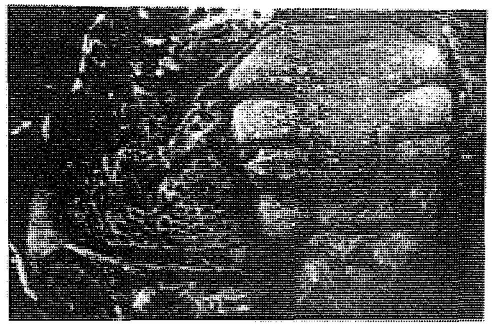

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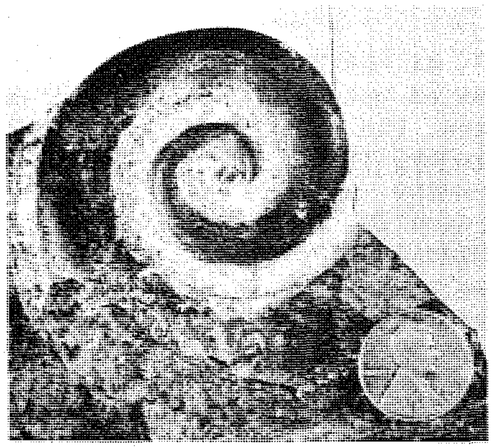

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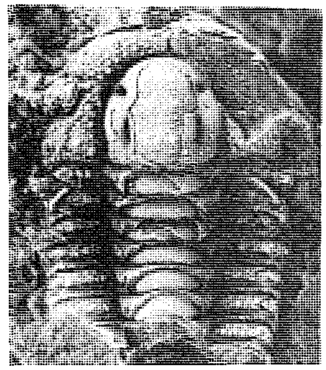

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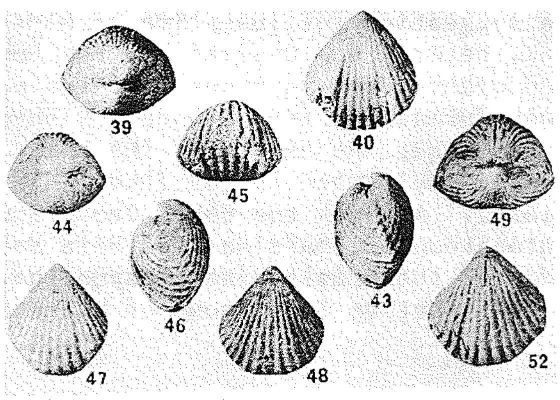

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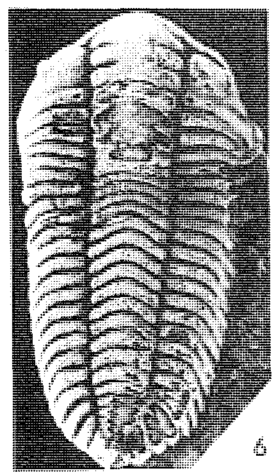

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Plate 1

1. Amphilichas minganensis: cranidium, dorsal view x2, from fine lime mud infilling reef framework at Sheldon Lane, New York (Stop 4).

2. Paraceraurus ruedemanni: cranidium, dorsal view x1, same lithology and locality.

3. Maclurites magnus: shell largely recrystallized. Crown Point Fm. Intersection of NY348 and I87, SW of Chazy Village.

4. Glaphurus pustulatus: partial cranidium and thorax, dorsal view x4. Same lithology and locality as 1 and 2.

5. Rostricellula plena: Valcour Fm., Chazy, New York. (From Cooper, 1956) x1.

6. Pliomerops canadensis: complete specimen, lower Valcour Fm., east side of Valcour Island, New York. x1.

Figures 1, 2, 4, 6 also appear in Shaw (1968)

In sum, the Chazy Group records a diverse marine fauna of cratonic aspect, including the very early representatives of a number of successful Paleozoic taxa. Exclusion of other taxa expected to be present, as well as facies dependence of organisms within the various facies of the Chazy Group (Shaw, 1968) generate some problems in correlating the Group to other North American Ordovician sequences.

THE CHAZYAN AS A STAGE

Since the days of Emmons (1842), it has been recognized that rocks of Chazy lithology and faunal content do not occur southwest of the Adirondacks in the Mohawk Valley. There, the transgressive Black River Group (Schopf, 1966; Young, 1943; Winder, 1960) overlies Lower Ordovician units with a pronounced unconformity (Fisher, 1954).[1] Ulrich, Grabau, Raymond and others thus proposed the Chazyan Stage, deemed the lowest stage of the North American Middle Ordovician. The geologic isolation of the Chazy Group makes close comparison to other standard Ordovician rock units difficult. However, since stages must ultimately be recognized by their faunas, the modern studies by Cooper (1956), Flower (1958), Ross (1963, 1964), Shaw (1968), Kraft (1962), Pitcher (1964), and Schopf (conodonts, proposed) leave the Group well known. Cooper (1956), expressing dissatisfaction with the traditional arrangement of Ordovician stages, proposed several new stage names for the Middle Ordovician based on sections in the southern Appalachians and Nevada. Of particular interest was the suggestion for inserting an older Middle Ordovician stage, the Whiterock, between the Chazyan and the Lower Ordovician. Kay (1960, 1962) and others have vigorously contested this, claiming that supposed Whiterock formations from Nevada, Oklahoma and Newfoundland are really Chazyan in age. At any rate, the Chazy faunas do differ from the nearest supposed Whiterock faunas (the Table Head Formation of western Newfoundland). As both the Chazy Group and Table Head Formation appear to be autochthonous, marginal cratonic sequences along strike on the western edge of the Appalachian geosyncline, the present writer favors the argument that at least some of Whiterock is older (Shaw, 1968).

FIELD TRIP STOPS

PLEASE NOTE: STOPS 1 AND 2 ARE ON PRIVATE PROPERTY WHICH WE HAVE SPECIAL PERMISSION TO ENTER. DO NOT SMOKE IN THE FIELDS, KNOCK OVER FENCES, ETC. OTHERS MAY WANT TO RETURN TO THIS CLASSIC LOCALITY AFTER YOU.

Stop 1. The Head, Isle La Motte, Vermont, 3 miles SSW of Isla La Motte Village. Lakeshore outcrops of the Providence Island Dolostone (Lower Or-

[1] Schopf and Raring (pers. Comm., 1969) appear to have found some close relationships between Black River and Chazy Group conodonts. The break in the Mohawk Valley section may thus not be as large as previous workers have supposed.

dovician) and the Day Point Formation (Chazy Group). Dip of both units several degrees to the north. Contact well-exposed and at least locally unconformable. Locality discussed by Shaw (1968), Erwin (1957). Section measured and described by Oxley and Kay (1959).

Approximately 40 feet of Providence Island Dolostone is exposed, being very fine-grained, massive, thinly laminated, and unfossiliferous. Mudcracks and a few ripple marks complete the picture of a unit deposited in very shallow water. Following Laporte (1967), the environment of formation was probably supratidal, closely paralleling modern day environments of dolomite formation described from Florida and the Bahamas. No detailed petrologic work has been done on this unit. In the absence of fossils, the age of this unit is not known. The underlying Fort Cassin Limestone (not exposed here) is known to be Late Canadian.

The Chazy Group begins here with about 20 feet of quartz sand and siltstones together with minor amounts of greenish shale (Head Member of Oxley and Kay, 1959). Ripple marks and cross-bedding are common. The fossils consist primarily of 'fucoids' (probably recording a variety of trails, worm tubes and the like) and Lingula. The succeeding Chazy unit (Scott Member of Oxley and Kay) consists of about 40 feet of echinodermal lime sand, cross-bedded in some places. Brachiopods (Orthambonites?) and indeterminable trilobite scraps are the chief recognizable fossils. The overlying 15 feet of quartz sandstone (Wait Member of Oxley and Kay) appears very similar to the initial sandstone.

This second sandstone is followed by a thick (115 feet) lime sand (Fleury Member of Oxley and Kay) which occupies most of Scott Point and The Head south of the road. Much of the unit is composed of echinoderm fragments, although little is known about the actual morphology of the creatures involved. Both the fragmental nature of the fossils and the frequently observed cross-bedding argue for considerable agitation of the ocean bottom.

Stop 2. Same vehicle location as stop 1. Upper Fleury Member of Day Point Formation and overlying Crown Point Formation. 200 yards south of the right angle bend in the road is locality R 25 (Shaw, 1968) which yielded 12 genera of trilobites, including Sphaeroxochus and Ceraurinella, from a particularly coarse pocket in the upper Fleury lime sands. Gastropods (Raphistoma) and brachiopods (Orthambonites?) are also present. This same stratigraphic level elsewhere, particularly 1 mile to the NE, displays spectacular bryozoan bioherms and the very early tabulate coral Lichenaria (Pitcher, 1964). These will be viewed on Trip F.

About 50 yards north of the road at this same stop, the silty, Maclurites-bearing limestones of the Crown Point appear. The actual contact with the Day Point is not visible but the lithologic change is evident. The Crown Point here contains several modest bioherms which have not been studied in detail. The earliest known stromatoporoids (Pitcher, 1964) are known to be important reef builders nearby in this unit and doubtless are dominant here as well (see Trip F).

Stop 3. Fisk Quarry, 2.5 miles SSW of Isle La Motte Village. Middle Crown Point Formation, consisting of fine-grained, dark, silty limestone with buff-colored dolomitic partings. This is 'typical', non-reef Crown Point lithology. However, in the quarry wall and some of the cut blocks, small reeflets can be seen. These are assumed to be largely stromatoporoids and calcareous algae, although they have not been studied as intensively as the reefs at the same horizon to the east (Trip F). Evidently, these reef masses could grow at some depth in relatively silty waters. The mechanism of their establishment thus does not appear to be tectonic. Maclurites (large gastropod, rare) and a few trilobites and brachiopods may possibly be collected from the limestone, although they are not abundant.

Stop 4. Sheldon Lane, 2 miles SE of Chazy Village, New York. Upper Crown Point and Valcour Formations; dip north several degrees. Locality R3 of Shaw (1968) and "Road to Little Monty Bay" locality of Raymond and others.

South of the road, a very low-dipping section of Crown Point and Day Point extends for almost 2 miles across fields. The transition between the two units is similar to that on Isle La Motte. Bioherms appear not to be present in the Day Point in this area. The Crown Point, however, is only about 100-150 feet thick here, being composed largely of biohermal masses. These can be seen just south of the road. Just north of the road, in and around the abandoned quarry, are exposed bioherms usually classified as basal Valcour. They are characterized by somewhat more silt than the reefs south of the road. Similar reefs on Isle La Motte have been classed by Pitcher (1964) as having a higher percentage of bryozoan and algal components than the Crown Point reefs. Billingsaria (tabulate coral) may also be important here.

Impressive cephalopod and trilobite faunas have come from the Valcour reefs here, including Glaphurus (trilobite, Pl. 1, Fig. 4) and large asaphids, illaenids and ceraurids (trilobites). Shaw (1968) has discussed the restriction of some of these forms to the reef environment. Most of the fossils appear in fine lime mud which apparently was trapped in channels and pockets in the reef framework. Lime sands of pelmatozoan origin drape the reefs.

Stop 5. Abandoned quarry of International Lime and Stone Company, 1 mile SE of Chazy Village, New York. This is the uppermost part of the same section covered at stop 4 and shows well the contact of the Valcour with the overlying Black River Group (see also Fisher, 1968, Fig. 25). The upper Valcour, here, probably equals the Pamelia (lowest Black River), and is a shaly dolostone bearing Rostricellula (brachiopod, Pl. 1, Fig. 4). The contact with the overlying Lowville Limestone, a massive, gray limestone is gradational. The early horn coral Streptelasma is present in this unit, although possibly not here. The Isle La Motte appears correlative to the Chaumont west of the Adirondacks.

REFERENCES CITED

Cooper, G. A., 1956, Chazyan and related brachiopods: Smithsonian Misc. Pubs. v. 127 (2 parts), pp. 1023.

Cushing, H. P., 1905, Geology of the northern Adirondack region: N.Y. State Mus. Bull. 95, pp. 271-453.

Emmons, Ebenezer, 1842, Geology of New York, Part 2 - Survey of the 2nd Geological District: Albany, p. 437.

Erwin, R. B., 1957, The geology of the limestones of Isle La Motte and South Hero Island, Vermont: Vermont Geol. Surv. Bull. 9, pp. 94.

Fisher, D. W., 1954, Lower Ordovician (Canadian) stratigraphy of the Mohawk Valley: Geol. Soc. Amer. Bull. v. 65, pp. 71-96.

____________, 1968, Geology of the Plattsburgh and Rouses Point, New York-Vermont Quadrangles: New York State Museum and Science Service, Map and Chart Series, #10, pp. 51, pl. 2.

Flower, R. H., 1958, Some Chazyan and Mohawkian Endoceratida: Jour. Paleont. v. 32, pp. 433-458.

Hudson, G. H., 1931, Fault systems of the northern Champlain Valley, New York: N.Y. State Mus. Bull. 286, pp. 5-80.

Kay, Marshall, 1960, Classification of the Ordovician system in North America: Rep. 21st Int. Geol. Cong. pt. 7, pp. 28-33.

____________, 1962, Classification of Ordovician Chazyan shelly and graptolite sequences from central Nevada: Geol. Soc. Amer. Bull. v. 73, pp. 1421-1430.

Kraft, J. C., 1962, Morphologic and systematic relationships of some Middle Ordovician Ostracoda: Geol. Soc. Amer. Mem. 86, pp. 104.

Laporte, Leo, 1968, Ancient environments, Prentice-Hall, pp. 116.

Pitcher, Max, 1964, Evolution of Chazyan (Ordovician) reefs of eastern United States and Canada: Bull. Can. Pet. Geol. v. 12, pp. 632-669.

Raymond, P. E., 1905, The fauna of the Chazy limestone: Am. Jour. Sci. ser. 4, v. 20, pp. 353-382.

Raymond, P. E., 1906, The Chazy formation and its fauna: Ann. Carnegie Mus. v. 3, no. 4, pp. 498-596.

Ross, J. P., 1963a, Chazyan (Ordovician) leptotrypellid and atactotoechid Bryozoa: Palaeont. v. 5, pp. 727-739.

__________, 1963b, Ordovician Cryptostome Bryozoa; standard Chazyan Series, New York and Vermont: Geol. Soc. Amer. Bull. v. 74, pp. 577-608.

__________, 1963c, The bryozoan trepostome Batostoma in Chazyan (Ordovician) Strata: Jour. Paleont. v. 37, pp. 857-866.

__________, 1964, Champlainian Cryptostome Bryozoa from New York State: Jour. Paleont. v.38, pp. 1-32.

Schopf, T. J. M., 1966, Conodonts of the Trenton Group in New York, southern Ontario, and Quebec: N.Y. State Museum and Science Service, Bull. 405, pp. 105.

Shaw, F. C., 1968, Early Middle Ordovician Chazy Trilobites of New York: N. Y. State Museum and Science Service, Memoir 17, pp. 163.

Welby, C. W., 1961, Bedrock geology of the central Champlain Valley, Vermont: Vermont Geol. Surv. Bull. 14, p. 296.

Winder, C. G., 1960, Palaeoecological interpretation of Middle Ordovician stratigraphy in southern Ontario, Canada: Int. Geol. Cong. Rep. 21st Session, pt. 7, pp. 18-27.

Young, F. P., 1943, Black River stratigraphy and faunas: American Jour. Sci., v. 241, 141-166, 209-240.

THE PALEOECOLOGY OF CHAZYAN (LOWER MIDDLE ORDOVICIAN) "REEFS" OR "MOUNDS"

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REEFS AS BIOLOGIC COMMUNITIES

Geologists have commonly concerned themselves with reefs as a problem in the building and maintenance of a framework in the face of the destructive effects of wave-energy. From the point of view of the reef organisms the paramount aspect of the reef environment may well be the opportunities it provides for interactions between organisms. To use an analogy which may be appropriate on many levels, the former approach is like viewing a city as a problem in architecture, while the latter is like viewing a city in terms of social interactions. What draws people to cities is that they provide a maximum availability of functional relationships, or to use the equivalent biological term, of ecologic niches. The high population density of a city, as well as of a reef, is both the necessary condition for such functional diversity as well as an ultimate result of it. A measure of the uniqueness of a reef as a focus of energy transformation is shown by the fact that its long-term average rate of primary productivity by photosynthesis is the highest of any living aquatic ecosystem, except for short-term peaks, largely in polluted fresh-waters.

Table: Average primary productivity of aquatic ecosystems calculated from gas exchange (Odum, 1959)

Ecosystem	Production (g dry organic matter/m^2/day)
Infertile open ocean, Pacific	0.2
Shallow, inshore waters, Long Island Sound	3.2
Estuary, Laguna Madre, Texas	4.4
Fresh-water lakes (various)	0.3-9.0
Silver Springs, Florida	17.5
Coral reefs, average 3 Pacific reefs	18.2

To paleontologists fossil reefs have a two-fold importance. For one, they are a sample of the marine life of the time at its most complex and diverse. For another, they are, so far as the sessile benthos goes, a true natural community, and enable one to avoid one of the perennial questions of paleoecology: "Is this community truly in place and representative of its environment?" The answer can only be a resounding "Yes".

As with cities, one of the important problems of a reef is the pollution of its environment by its own metabolism. In particular, the depletion of oxygen and the production of nitrogenous wastes, are the most acute problems. The solution adopted by the inhabitants of modern coral reefs is the development of a symbiosis with certain phytomastigophora, termed zooxanthellae, which live in the tissues of reef organisms and have been shown to absorb nitrogenous materials. In living coral reefs zooxanthellae are present in scleractinians, sponges, and even the giant clam Tridacna. It is likely that the Scleractinia had this adaptation well back into the Mesozoic, for hermatypic Scleractinia and their reefs go back at least to the Jurassic, and they form the last part of an ecologic succession, or sere, in late Triassic reefs (Sieber, 1937). Whether this was true of Paleozoic reef organisms is not known. Certainly, calcareous algae were present in Paleozoic reefs and have continued to those of the present day. These free-living algae also contribute to oxygen replenishment and nitrogenous waste absorption. They must have been as essential to ancient reefs in this metabolic function as well as in their well-known frame-building properties.

TROPHIC RELATIONS IN THE CHAZY REEFS

The principal reef-building organisms of the Chazy reefs are stromatoporoids (Cystostroma, Pseudostylodictyon or Stromatocerium), lithistid sponges (Zittelella, Anthaspidella), tabulate corals (Lamottia or Lichenaria, Billingsaria, Eofletcheria), bryozoa (Batostoma, Cheiloporella, Atactotoechus), and calcareous algae (Solenopora, Sphaerocodium or Rothpletzella, Girvanella). In addition, a trilobite (Glaphurus) and numerous pelmatozoan fragments are found in such intimate physical association with the reef as to be likely inhabitants of its surface. The large gastropod, Maclurites, and many genera of large nautiloids, are very abundant in the pelmatozoan-brachiopod calcarenites between the reefs, and in channels cut into the reefs, but there is no direct evidence that they actually lived on the reef while it was active. However, their presence in the near vicinity suggests that they participated in the overall food chain.

Unlike modern reefs, in which macrophagous, carnivorous coelenterates are the dominant element, and feed upon an abundant fauna of small nekton, these middle Ordovician reefs are dominated by suspension feeders. This is especially true when one considers the recent reinterpretation of stromatoporoids as sponges (and therefore suspension-feeders) (Hartman and Goreau, 1966, in press, and personal communication) belonging to a little-known group that today participate in living coral reefs. The only possible non-suspension feeders in the Chazy reefs are the tabulate corals, presumably carnivorous, and the trilobite, which is possibly a detritus-feeder (if not a suspension-feeder).

The off-reef Maclurites is an archaeogastropod, and presumably grazed on algae. It is the largest of the primary consumers of the Chazy beds. The nautiloids may have fed on the Maclurites. If so, they ate the soft parts without breaking the shells, for most of the large shells are whole. The possibility that some of the nautiloids "grazed" on the sessile benthonic invertebrates should not be discounted, for these middle Ordovician cephalopods are not far removed in evolution from the late Cambrian ellesmeroceratids, whose short, relatively non-buoyant shells indicate a vagile benthonic adaptation. Although the Chazyan nautiloids had buoyant shells and were probably good swimmers, they may have retained an interest in bottom feeding. Apart from the cephalopods we have no evidence for other large carnivores.

The Chazy reefs are thus a community in which the benthos was fed primarily from suspended matter or plankton. Even the corals had very small polyps and could not have eaten anything very much larger than a few millimeters across. This is surely a reflection of the paucity of larger nekton or vagile benthos on which to feed. Only the snails and cephalopods provide a larger fauna, and may have formed a side loop to the general food chain, the cephalopods feeding primarily on the snails.

Reefs earlier than the Chazy consist only of algae, or else include certain primitive possible sponges, such as Archaeocyatha and Calathium, which were not likely to be anything other than suspension feeders. In the Silurian, corals become much more important, and one begins to see carnivorous macrophagy becoming a more important element in the trophic relationships of reef faunas. Silurian reefs are still dominated by tabulate rather than rugose corals among the carnivores, and thus consumed mainly small vagile animals. The suspension feeding element (bryozoa, stromatoporoids) is also still strong in Silurian reefs. It is not until Devonian times that the large rugose corals become a dominant element in reefs, probably not without connection with the fact that this was the first time that fish and other large nekton appear in abundance.

DEVELOPMENT OF REEF FAUNAS

Within the Chazy group the reef faunas show a progressive increase in diversity with time (Pitcher, 1964). The earliest reefs, in the lower Day Point Formation (Scott Member), are built of bryozoans only, and chiefly of one species, or at most two. These early benthic concentrations are, like their predecessors, of suspension feeders only. In the middle of the Day Point (Fleury Member) the Lamottia bistrome, introduces the oldest-known coral in the world, which is also the oldest-known sessile carnivore with a skeleton. (The only older sessile carnivore is a possible anemone from the middle Cambrian Burgess Shale, and the only older vagile carnivores are the early Ordovician starfish, and the late Cambrian and early Ordovician nautiloids). It is possibly at this moment in the history of the earth that it first became profitable for a carnivore to sit and wait for its food to come to it. This coral appears to have lived in a different environment from the bryozoa, although probably nearby. The corals are often fragmented and the fragments overgrown by bryozoa. Pitcher (1964, p. 648) considers the corals to have been transported into the area of outcrop, and there to have acquired their coatings of bryozoa.

In the latest Day Point, immediately above the Lamottia bed, bryozoan mounds again develop, not much different from before except that occasional individuals of the lithistid sponge Zittelella foreshadow the richer fauna of the Crown Point.

In all the Day Point reefs algae are seemingly missing and one can scarcely speak of these assemblages as functionally integrated communities. In the Crown Point Formation the reef faunas are more diverse, and include algae, stromatoporoids, lithistid sponges, and corals (Billingsaria, not Lamottia) along with the bryozoan species that built the earlier mounds. The algae were not immediately available to the reef animals for food, but probably performed an anti-pollutant function. They may have been eaten by soft-bodied meiobenthos which were subsequently consumed by the corals, but the principal flow of organic matter must have been from the phytoplankton directly, or through zooplankton, to the reef animals, and from there, through the intervention of bacteria, back to the benthic algae and to the phytoplankton as dissolved molecules, or recycled through the sponges in the form of whole bacteria, which may be a principal food of sponges. (Rasmont, in Florkin & Scheer, 1968, Madri, 1967). That the Crown Point environment was in general one of a high trophic level is demonstrated by the abundance of the large snail Maclurites magnus, which is virtually a guide fossil to the formation, as well as of the large nautiloids that may have fed on it. The abundance of algae outside the reef environment (dead Maclurites and nautiloid shells are frequently encrusted by them) undoubtedly provided a firm base for the overall food chain as well as a food supply for the Maclurites.

In the Valcour Formation the faunal complexity is maintained and some new species are introduced in the non-reef environments.

The Crown Point reefs are the earliest reefs that can be considered functionally-integrated communities. It is worth noting that the animals that populate them (stromatoporoids, lithistid sponges, tabulate corals, and bryozoa) are very nearly the earliest-known representatives of their respective taxonomic groups.

The abundance of sponges in the Crown Point and early Valcour reefs deserves consideration, for it can be related to the general evolution of hermatypic organisms. Sponges are not common reef-building animals. During the periods when tabulate and rugose corals were abundant, and during the periods when the scleractinians were abundant, including today, sponges were a very minor element in the construction of reefs. It is only before the corals first become abundant (before the Silurian), and also during the interval between the decline of the Paleozoic corals and the rise of the scleractinians, (Permian and Triassic), that sponges were important reef builders. This statement leaves out the stromatoporoid sponges, which managed to coexist with the Paleozoic corals through the Silurian and Devonian, though often in different reefs, and presumably in different environments. The bryozoans show a similar inverse relationship to the corals, but seem to have been sturdier competitors than the sponges. Bryozoa are still present in Silurian reefs alongside corals and stromatoporoids, though they tend to be replaced by the latter in ecologic successions (Lowenstam, 1957). In Devonian times bryozoans are rarely present in reefs, but reappear to some extent in the Permian (Zechstein of Germany) when corals declined.

In the Cambrian the calcareous Archaeocyathids participate in reefs, and in the early Ordovician, Calathium (a receptaculitid) does. The sponges of the Chazy beds are of interest in that they are lithistid demosponges with a siliceous skeleton (now completely calcified). The middle Ordovician (Chazyan and Black River) is the only time during the Paleozoic that siliceous sponges were significant frame builders. This time coincides with the first radiation of the lithistid demosponges. It may be that the higher rate at which stromatoporoids, corals and bryozoa could secrete calcium carbonate skeletons was the reason for the disappearance of lithistids from the later reefs. When the corals declined at the end of the Paleozoic it was the calcareous Sphinctozoan sponges that replaced them as important reef-builders, in Permian and Triassic times, along with the ever-present calcareous algae.

It should be noted that most lithistid sponges, although their skeletons are rigid, do not by themselves bind sediment or build up massive structures. In the Crown Point reefs sediment-binding was probably carried on only by stromatoporoids, laminar bryozoa, corals, and calcareous algae. Nevertheless, the lithistids cover, on the average, from 22% to 50% of the surface of the reefs in which they are most abundant (Pitcher, 1964, pp. 662, 675). They thus contributed significantly to the bulk of the reef mass. They also served to trap sediment. That this by itself can be a potent factor in mound formation is indicated by the late Jurassic sponge "reefs" of Germany (Roll, 1934) in which siliceous sponges built mounds apparently solely by trapping sediment and without the significant presence of binding organisms. These Jurassic mounds are the only known examples of siliceous sponge reef-like structures in post-Paleozoic times.

It is possible that some of the laminar Anthaspidella was actually of encrusting habit and may have helped to bind other skeletal material, but its role would have been minor compared to that of the more abundant binding organisms.

ECOLOGIC SUCCESSION

Ecologic succession in the Day Point reefs can hardly be said to exist, since the reefs consist only of one species of bryozoan. The encrusting of the coral Lamottia by the bryozoan Batostoma is probably not true succession if the corals are not in place. It is worth noting, however, that the Lamottia bed is immediately succeeded by Batostoma reefs which were built on the coral debris (Pitcher, 1964, p. 650) as shown by cores.

In the more complex Crown Point reefs no clear succession is evident though there are suggestions of it. The stromatoporoid Pseudostylidocton frequently forms small reeflets by itself, resting on pelmatozoan calcarenite. It also often forms the basal parts of larger reefs, together with subordinate ramose bryozoa. Subsequently there succeeds a more diverse fauna of the lithistod sponges Zittelella and Anthaspidella, the coral Billingsaria, the bryozoan Batostoma and a flora of Sphaerocodium and Solenopora. Some reefs on Valcour Island end with this community. Others on Isle La Motte often have a capping of Pseudostylodictyon alone. In this mature reef community the lithistid sponges and the stromatoporoids occupy by far the largest surface area. Billingsaria, bryozoans and the algae are distinctly subordinate. The stromatoporoids can be considered to form a pioneer community which initiates reef development. It apparently provides a favorable substrate for the lithistid sponges and for the encrusting corals, bryozoans and algae. The lithistid sponges (Zittelella, Anthaspidella) can be quite common in the calcarenites away from the reefs, and therefore do not need the stromatoporoids as a base. Their participation in the reef is facultative rather than obligatory. The development of this rudimentary succession may be a matter of building up into somewhat shallower water, as is suggested by the change from ramose to laminar algae. It may also be a matter of the development of a firmer substrate than is provided by the surrounding shell sand. Biotic factors, such as the availability of food, probably also enter into the picture. Laminar algae, favored in their growth by a hard substrate, may attract herbivorous meiobenthos, which may in turn provide abundant food for Billingsaria, and indirectly, more bacteria for the sponges.

In the Chazyan mounds of Quebec, a better-defined ecologic succession has been ascertained (see accompanying paper of Toomey and Finks). Here pioneer communities of the encrusting bryozoan Batostoma are succeeded by a mixed bryozoan-coral community (Batostoma, Chazydictya, Billingsaria, Eofletcheria). Finally the corals (Billingsaria or Eofletcheria) become dominant over the bryozoans at the top of the mound, perhaps a foretaste of things to come.

COMPETITION

Bryozoans tend to show a somewhat inverse relationship of abundance with reference to stromatoporoids and sponges (Pitcher, 1964, fig. 44) suggesting competition, as might be expected from the fact that they are all suspension feeders. At the top of the Crown Point, bryozoan reefs occur side by side with stromatoporoid-lithistid reefs. They tend to dominate the Valcour reefs again, almost as they did in the earlier Day Point. The variability of the proportions of reef organisms in the Crown Point from one reef to the next, also suggests that there was near-equality in competition between many of these organisms. At least one reef in the pasture on Isle La Motte is composed of 50% Billingsaria throughout (Pitcher, 1964, p. 666). Other reefs in the same pasture contain, on the surface, anyway, about 50% lithistid sponges (Zittelella, Anthaspidella). The corals and the sponges did not compete for food but they probably competed for substrate space. Occurrences of Billingsaria and Zittelella together on the flanks of reefs in this same pasture indicate that they had the same environmental tolerances.

VERTICAL ZONATION

In the early Day Point byrozoan mounds, the mounds are built of laminar Batostoma or Cheiolporella, while the interreef areas contain abundant branching Atactotoechus. This may be considered a rudimentary sort of depth zonation, with the branching bryozoa occupying the deeper quieter water, and the laminar bryozoa the rougher shallower zones. However, the total relief at any one time was scarcely more than a foot or two (see Pitcher, 1964, fig. 10) and the differences in wave energy could not have been very great. Nevertheless, the presence of branching bryozoa, along with stromatoporoids, in the basal parts of Crown Point reefs, and their replacement by laminar bryozoa higher up (Pitcher, 1964, fig. 8) suggests that there may be something to this form distribution in relation to depth. Certainly in living sponges, corals and bryozoa there is a similar confinement of branching forms to the less rough water areas.

The surface distribution of organisms on a Crown Point mound was studied by Pitcher (1964, fig. 26) from the low flanks up to its crest. This should reflect bathymetric differences. He found that the stromatoporoids were most abundant at the crest, the bryozoa most abundant somewhat lower down, and the corals and lithistid sponges most abundant still lower on the flanks with the sponges remaining abundant further down than the corals. This again would correspond to a well-known pattern of morphological distribution., with the conical or cup-shaped lithistids (Zittelella) being characteristic of quieter, deeper water, while the laminar bryozoans, corals and stromatoporoids are characteristic of rougher water. The total vertical relief involved is scarcely six feet, and except for the absence of stromatoporoids at the base and the absence of lithistids and bryozoa on the crest, all the forms occur over the whole reef. Thus the environmental differences cannot have been very great.

A more pronounced bathymetric differentiation may be shown by some of the Crown Point reefs on the southwest shore of Valcour Island, on the point of land north of the concrete boat dock. Here the flanking beds pass laterally into dark calcilutites with numerous hexactinellid sponge root-tufts and body fragments. These are much more delicate sponges and may have occupied a depressed area with genuinely quiet water peripheral to the reef.

ORIENTATION AND CURRENTS

Bryozoan mounds in the Day Point (Pitcher, 1964, fig. 19) and stromatoporoid reeflets in the Crown Point (on both Isle La Motte in the Goodsell Quarry, and on the mainland at Sheldon Lane) tend to have a roughly north-south orientation. This is parallel to the paleoshore, and the mounds may have grown either in belts of optimum depth or into the set of longshore currents. An indication that currents may be involved is shown by the fact that hexactinellid sponge root-tufts in non-reefy beds of the Crown Point at South Hero, Vermont (Pitcher, 1964, fig. 32) show the same preferred orientation on the bedding planes. Orthocone nautiloid shells are less clearly oriented in a preferred direction, but in the Crown Point of Isle La Motte and South Hero (Pitcher, 1964, fig. 32) they show a broad peak at about 45° to a N-S line, perhaps the result of some being rolled around to a position perpendicular to the current while others were swung parallel to it.

In the channels that cut the Crown Point reefs, nautiloid shells are most commonly oriented parallel to the axis of the channel, obviously parallel to currents sweeping through. Maclurites shells are also often piled together in pockets in these channels, probably as a result of current action. The channels, however, may not be strictly contemporary with the reefs they cut.

Channels

The Crown Point reefs are cut by numerous channels, mostly one to three feet wide and as much as two feet deep, filled with a black calcarenite that contrasts sharply with the light calcilutite of the reef rock. There is a considerable body of evidence that these channels may have been formed subaerially by solution, possibly by enlargement of tectonic joints, following consolidation and diagenesis of the reef rock. The entire sequence of events would have to have taken place entirely within Crown Point time, perhaps several times. The evidence is as follows:

1. The channels have sharp boundaries against the reef rock along smooth surfaces that cut through the middle of stromatoporoid colonies, lithistid sponges, and calcilutite matrix in a continuous sweep. The matrix must have been consolidated, and the lithistid sponges may have already been changed from silica to calcite, for they show no effects of differential hardness on the erosion surface.

2. The channels usually end in rounded culs-de-sac, or sometimes have an ovoid shape, suggesting either pothole-like abrasion or sinkhole-like solution. There are essentially no quartz clasts in the surrounding sediments, so that abrasion would seem to be unlikely, thus leaving solution as the alternative.

3. The channels tend to intersect at close to right angles and most frequently, though by no means universally, are oriented roughly north-south and east-west. This suggests that they may follow a tectonic joint pattern. Participants in the trip are invited to compare the form of the channels with that of solution-enlarged joints now being eroded in the same rock.

4. If the channels were surge channels present in the active reef, we would expect to find them bordered with at least some entire outlines of reef-building organisms, or where these were broken by contemporary wave-action, to find that the broken outlines, and the margins of the channel as a whole, would be irregular rather than smooth.

Because the calcarenite filling the channels contains Crown Point guide fossils identical to those beneath and to either side of the reef, and because such channeled reefs occur at more than one level within the Crown Point beds in the same area, we must assume that the entire process postulated took place repeatedly entirely within Crown Point time. If Crown Point time is assumed to be one-third of Chazy time, and that one-sixth of Ordovician time, and Ordovician time to be 60 million years long, then we have 3.3 million years for these processes to take place in. Admittedly this may be hard to swallow, and we have not had the opportunity to test the hypothesis adequately, but participants in the field trip may wish to think about these possibilities while examining the outcrops.

ITINERARY

The walking-tour will start at the north end of the picnic ground and trailer camp on the north side of Wait Bay in southeastern Isle La Motte. It may be reached by following the main north-south road down the center of Isle La Motte to its southern end, turning left (east) to the trailer park entrance, and then turning left (north) up the hill to the picnic ground. Please note that the entire trip is on private property, and that permission must be secured from the landowners for visits.

Cross the fence and walk north to the bare exposures of the *Lamottia* biostrome in the Fleury Member of the Day Point Formation. CAUTION! DO NOT STEP INTO SOLUTION-ENLARGED JOINTS. SOME ARE PARTLY CONCEALED BY VEGETATION. WALK ONLY ON BARE ROCK SURFACES. THE JOINTS ARE OVER A FOOT DEEP.

The hemispherical to discoidal heads of *Lamottia* are closely packed in a calcarenite matrix. Joints offer an opportunity to observe their orientation in section. More than half are overturned over much of the area. Many are broken. The proportion of broken ones increases to the north and east, where the biostrome passes into calcarenite with ever fewer and smaller fragments of *Lamottia*. In the central area of the exposure there are belts some 10 feet wide in which fragmentation, proportion of overturned specimens, and quantity of calcarenite matrix, are higher than elsewhere. These may represent surge channels. In the peripheral area to the northeast one may see much laminar *Batostoma chazyensis* surrounding the *Lamottia* fragments.

This is the type locality for the genus *Lamottia* Raymond, 1924. Although Raymond's description of this bed as the "world's oldest coral reef" may be disputed, it still seems to be unchallenged as the world's oldest occurrence of corals of any kind.

Walk northwestward upsection, so far as fence lines, cultivated fields, and vegetation permit. DO NOT DISTURB FENCES OR LEAVE GATES OPEN! NO SMOKING WHILE WALKING THROUGH THE FIELDS; THERE IS A DANGER OF FIRE. ALSO, PLEASE KEEP OFF CULTIVATED GROUND.

At about 2000 feet N 45° W of the Lamottia outcrop, we will find small mounds of Batostoma chazyensis at the top of the Day Point Formation. They tend to show a N-S alighment. the zoaria are mostly branching rather than laminar. Possibly this is a deeper water environment than that of the Lamottia biostrome.

Continue to walk northward to a small dirt road, then walk west along it to a T-junction with a larger dirt road. Turn left and follow it southwest to a house and barn on the right. We will enter a gate into the large pasture behind the house and barn. Mr. Ira LaBombard, the present owner of the property, has kindly given us permission to enter his pasture to study the reefs in the Crown Point and lower Valcour Formations. He has requested, as a condition of permission, that NO SPECIMENS WHATEVER be collected. PLEASE RESPECT THIS ORDER. We will have an opportunity later in the day to collect from these same beds at another locality. The fossils are so beautifully displayed here that relationships may be seen without disturbing the rock. They may be photographed very advantageously on the glacially polished surfaces.

The reefs exposed here are mainly in the Crown Point formation and are the ones intensively studied by Pitcher (1964). You may examine contemporaneous reefs by walking northeastward along strike. You may examine successively younger reefs by walking northwestward upsection (dip is about 10° NW).

The reefs are exposed as mounds of light rock. The calcarenite between the reefs, and filling the channels in the reefs, is nearly black. The reefs outcropping nearest the fence were mapped by Pitcher as his Assemblage A, consisting of the stromatoporoid Cystostroma and the alga Solenopora. Those beyond to the northwest, and covering most of the pasture up to a distinct linear rise in ground, belong to Pitcher's Assemblage B. These show interesting variations from reef to reef as well as changes in faunal distribution from flanks to tops of the mounds. The fauna consists of the stromatoporoid Pseudostylodictyon eatoni, the lithistid demosponges Zittelella varians and Anthaspidella sp., the tabulate coral Billingsaria parva, the bryozoan Batostoma chazyensis, and the calcareous algae Solenopora, Sphaerocodium and Girvanella.

The fossils may be identified readily on weathered surfaces as follows:

1. Pseudostylodictyon eatoni: Large whitish masses with fine, dark laminae forming concentric patterns about centers an inch or two apart. These concentric patterns represent the mamelons and their small size is characteristic of the species.

2. Zittelella varians: Circular, dark gray bodies two to three inches in diameter, with a central circular light area representing the matrix-filled cloaca, and radial light areas, or ovoid dots, a few millimeters wide, representing the canals. In longitudinal section, the sponge is conical, and oblique sections will show the expected intermediate shapes. Some specimens have an irregular outline in cross section.

3. Anthaspidella sp.: Similar to Zittelella in color and texture, but shaped like long sinuous bodies, an inch or so thick and several inches long, when seen in cross section. A surface view of the sheet-like sponge shows a somewhat irregular mass with-

out a cloaca. The complete sponge has a short stalk, the whole being shaped somewhat like a distorted cake-plate. The open, "spongy" texture may help when shape fails. Needless to say, the shape and geometric arrangement of the spicules in thin section is necessary for a secure identification. Not every shapeless mass is a sponge.

4. Billingsaria parva: Small, black, oval patches, a few inches across. The dark color is very distinctive. Close inspection with a hand lens will reveal the stellate outlines of the corallites with their characteristic septal ridges.

5. Bryozoa: These weather white, either as small, branching twiglets, or as laminated sheets. Identification requires thin sections, but the outlines of the zooecia are usually visible on the weathered surface and suffice to identify it as a bryozoan.

6. Solenopora: White concentric circles, often sparry. A few inches across. This is the most common form of Solenopora seen on the reef surfaces.

7. Girvanella: Small black ovoid bodies, less than an inch in length. These are oncolites, or algal-coated shell fragments.

8. Maclurites magnus: Large coiled shells a few to several inches across. No septa. The shell substance is white in cross section.

At the rise in ground is a one-foot stromatolitic layer with many orthocone cephalopods. Pitcher called this his Assemblage C and assumed it was laid down as a blanket during a relative drop in sea level. It forms a dip slope through which appear, apparently, the tops of Assemblage B mounds, as well as small mounds of Batostoma chazyensis alone which Pitcher called Assemblage D. At the west end of this cuesta-like feature, nearest the main road, a good cross section of an Assemblage B mound is exposed (see Plate 2).

Down the dip slope, above a ten-foot interval of grey calcarenites, are mounds in the lower part of the Valcour Formation. Theyare composed of Batostoma campensis, together with the alga Solenopora. Some Zittelella may be found. The bryozoa are clearly dominant.

Walk northeastward along strike for about a half-mile, observing Crown Point mounds as you go. You will eventually reach the Goodsell Quarry, operated by the Vermont Marble Company. The quarry is opened in the lower beds of the Crown Point which are relatively lacking in reefs except for small stromatoporoid-algae mounds. The quarry has been intermittently active, and the stone, which makes a beautiful black marble when polished, has been widely used as an interior trim. The rock weathers light gray, and has also been used locally as a dimension stone. It was used to build the old border fort, Fort Montgomery, visible from the Rouses Point bridge.

CAREFULLY avoiding falling into the water-filled quarry, one may observe vertical sections through stromatoporoid-algal mounds and their relationships with the surrounding calcarenite (see Plate 5). By tracing laminae from the mounds into the surrounding sediment, one can see that the mounds never stood more than a foot or two above the sea floor at any one time, though the total thickness is much greater because of the persistence of the mound population on the same spot. On the quarry benches, especially the glacially polished upper surface, one may see plan views of mounds and note their tendency to a N-S lineation. On these surfaces also, especially when wet down, one may see orthocone and other nautiloid shells, and Maclurites shells, overgrown by algal coatings.

This ends our examination of the Isle La Motte exposures. On our way back we will stop at the Sheldon Lane section, south of Chazy, where we may collect from the Crown Point and lower Valcour, in essentially the same facies.

The Sheldon Lane beds dip north. We will begin with cross-bedded upper Day Point (Fleury) calcarenite at the south end and walk through the entire Crown Point section including reef mounds with the same fauna as in LaBombard's pasture. North of the road is an old quarry in which the lower Valcour is well exposed.

Acknowledgements

We wish to thank Mrs. Malvina Bruley and Mr. Ira LaBombard of Isle La Motte, for permission to visit the classic exposures of reefs on their respective properties. Their cooperation has made this excursion possible.[1]

[1] I am indebted to Mr. Rodney V. Balasz for information concerning possible channels in the Lamottia biostrome, and to my paleontology class for mapping some of the channels in the various reefs in the fall of 1968 (RMF).

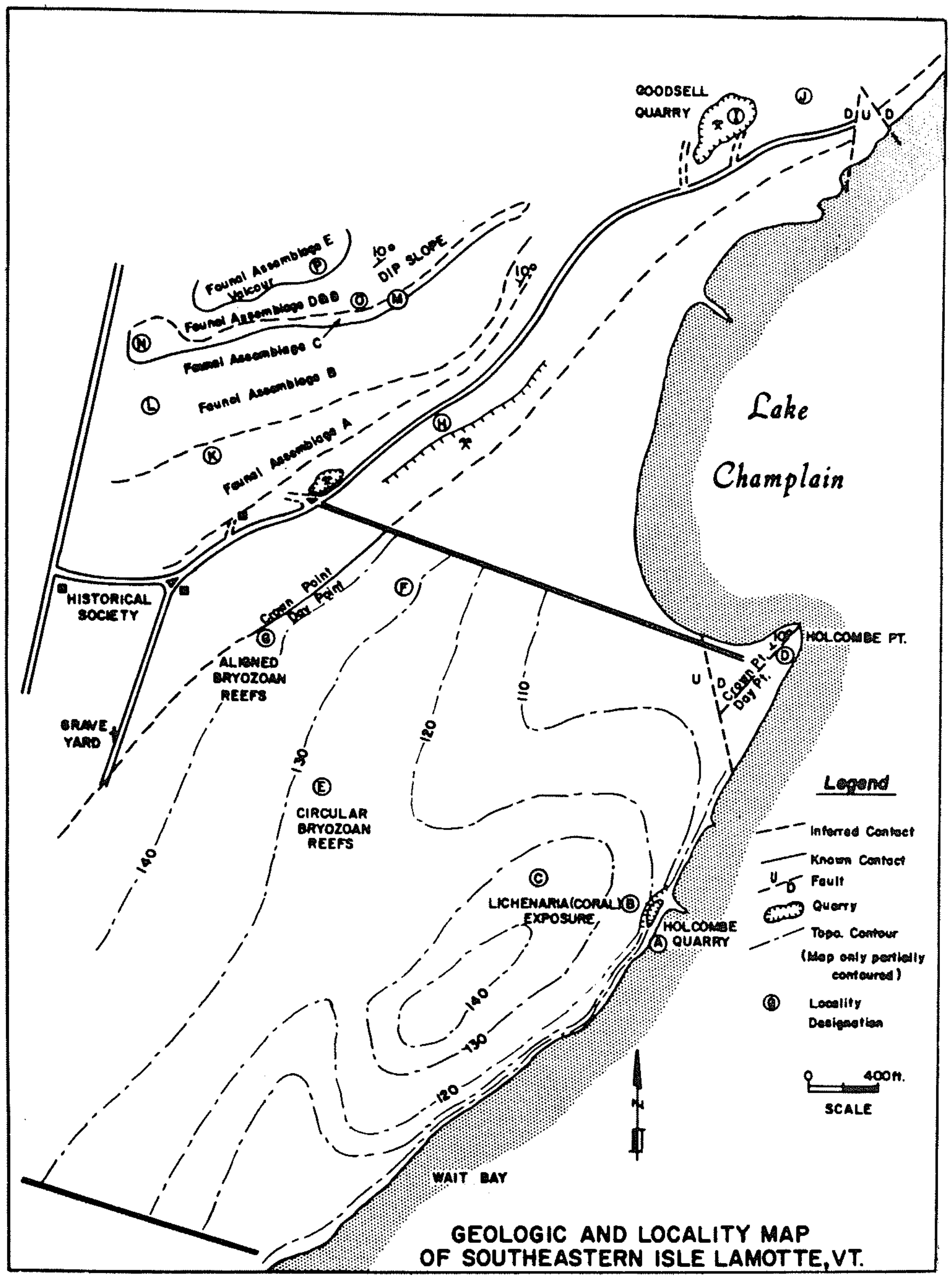

Figure 1. Geologic and locality map of Southeastern Isle La Motte, Vermont. After Pitcher, 1964.

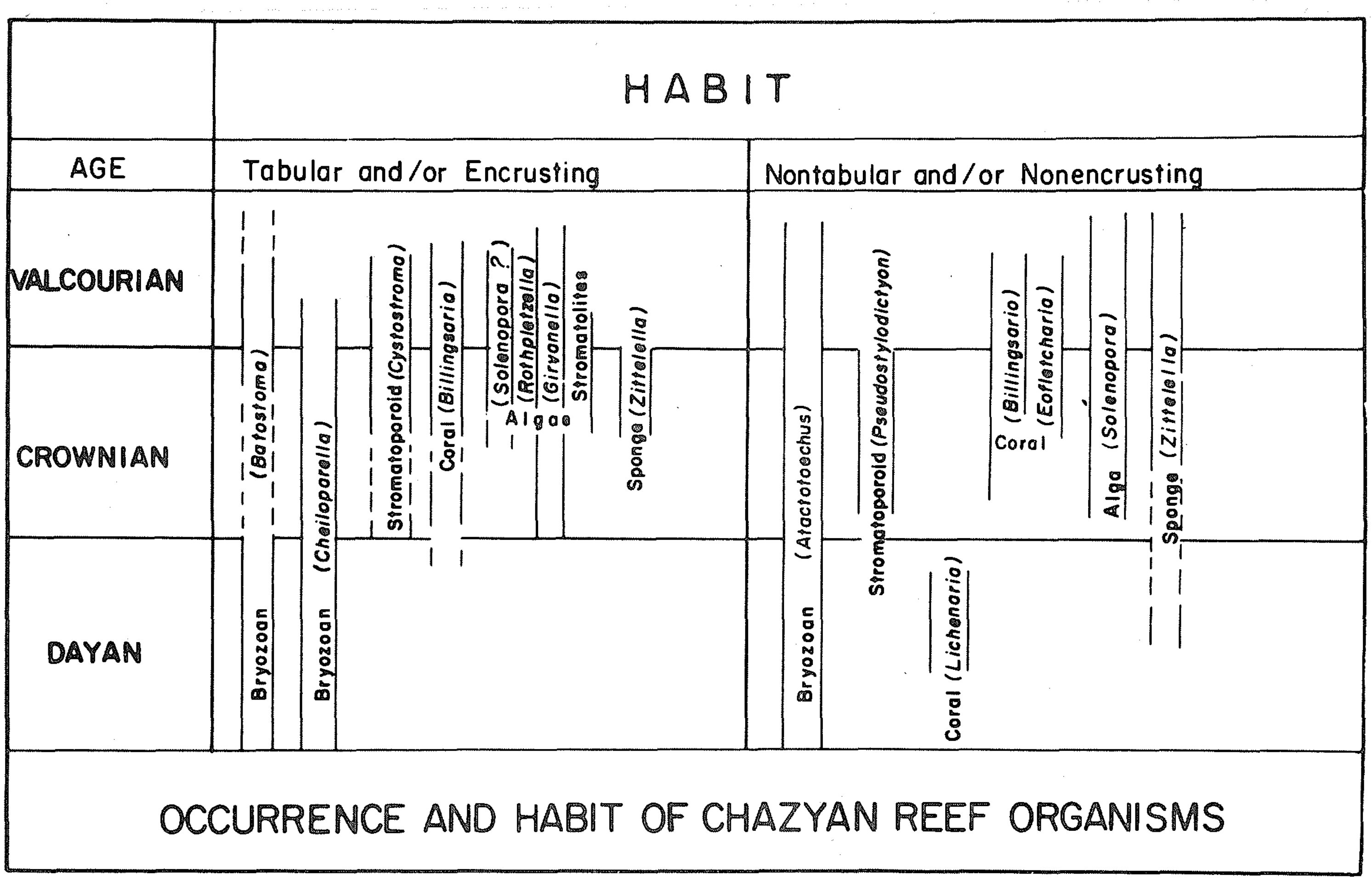

Figure 2. Occurrence and habit of Chazyan Reef organisms. After Pitcher, 1964.

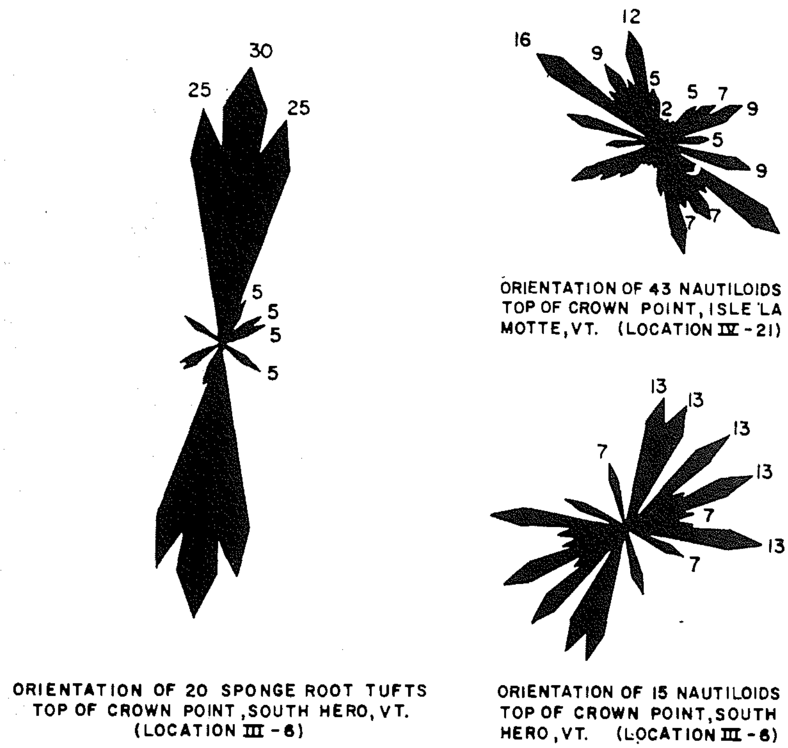

Figure 3. Orientation of Crown Point sponge root-tufts and nautiloids. After Pitcher, 1964.

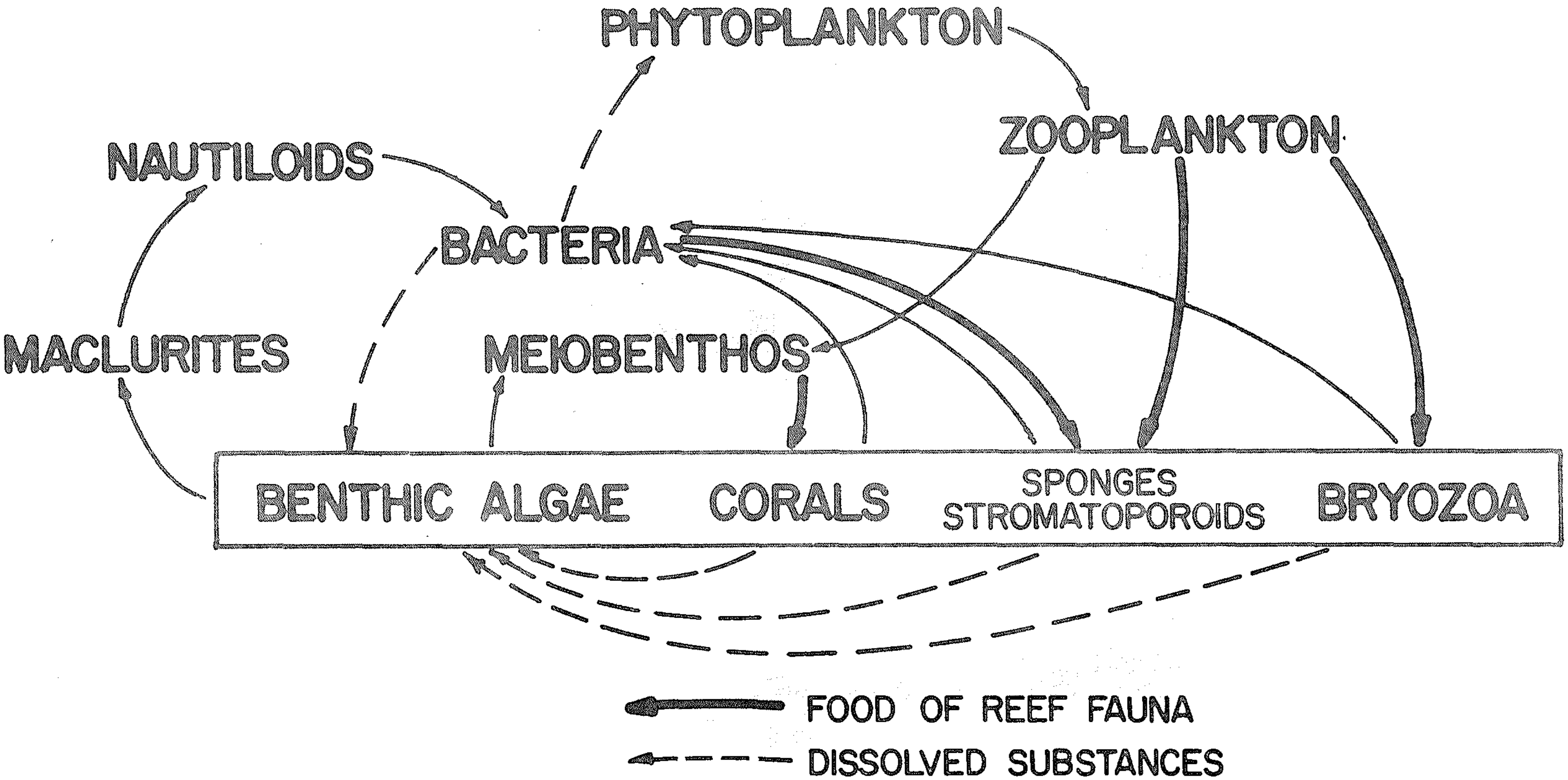

FIGURE 4. SIMPLIFIED FOOD-CHAIN OF CROWN POINT REEFS

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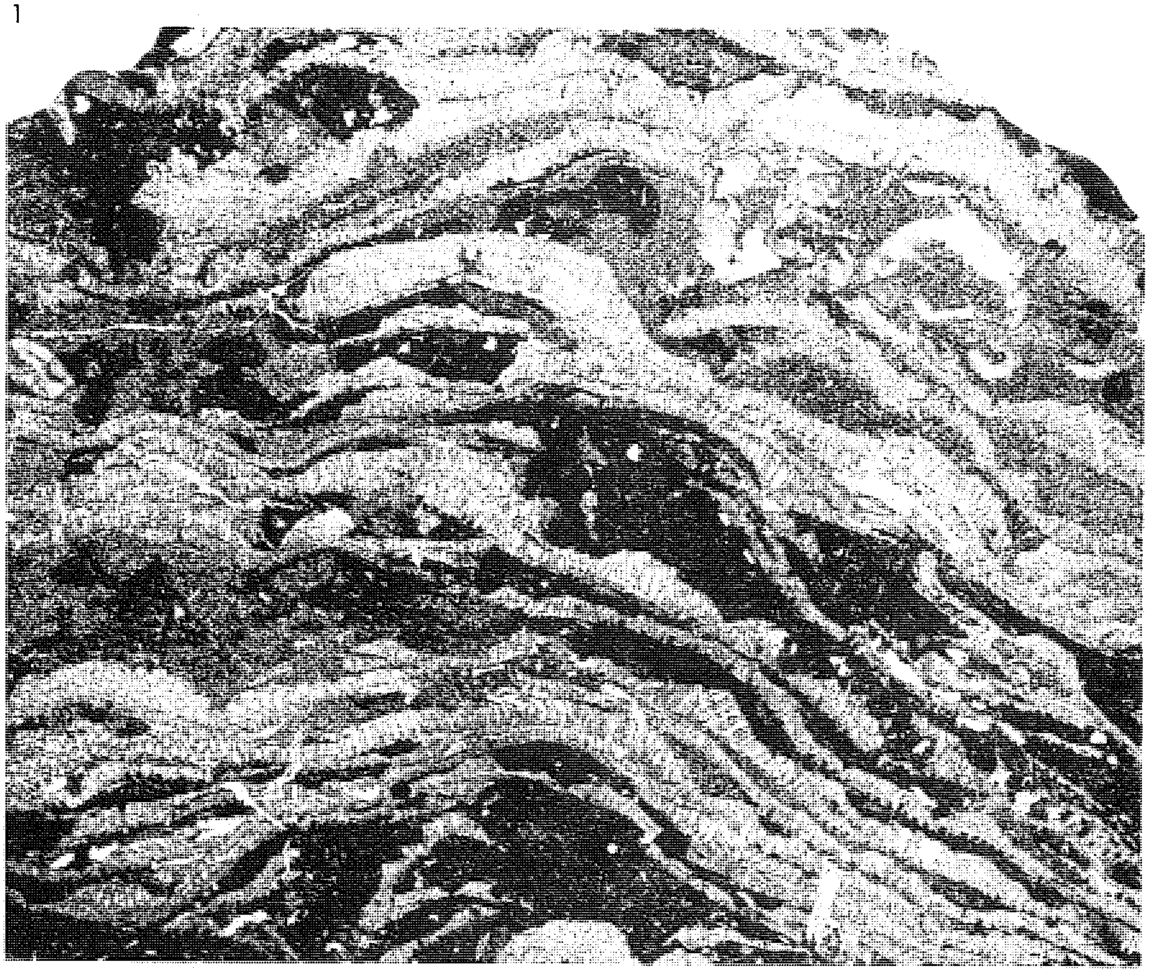

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PLATE 1

Figure 1 — Outcrop photograph of the surface of a Middle Ordovician (Chazyan) bryozoan mound exposed in the Day Point Formation (uppermost Fleury Member) on Isle La Motte, western Vermont. Note general lineation of the bryozoan colonies; length of hammer approximately 14 inches.

Figure 2 — Thin section photomicrograph (X3) of characteristic bryozoan mound rock that forms conspicuous mounds in the uppermost Day Point Formation, Fleury Member, on Isle La Motte, western Vermont. The mound rock is primarily composed of consecutive sheets or layers of the colonial trepostome bryozoan Batostoma chazyensis Ross, separated by lime mud layers containing relatively abundant, although quite small, dolomite rhombs (small grey flecks).

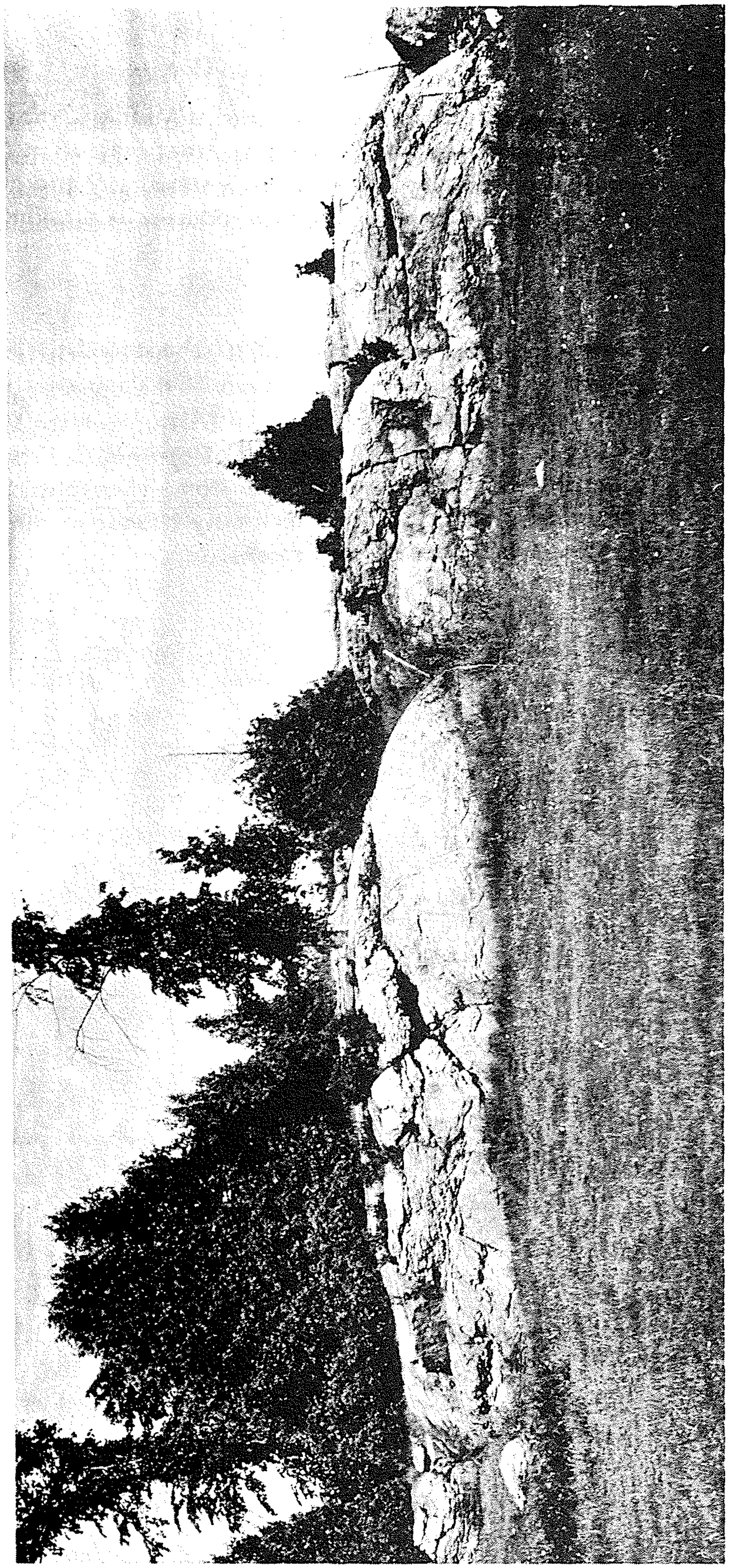

PLATE 2

Outcrop photograph of a series of typically small-sized Crown-pointian (Middle Ordovician-Chazyan) mounds exposed in La-Bombard's Pasture, Isle La Motte, western Vermont. Rounded mound structures are composed of lime mud containing abundant algae, sponges, stromatoporoids, and trepostome bryozoans. The beds filling-in the surface irregularities and capping the mounds are dominantly composed of relatively coarse-textured pelmatozoan debris (see Plates 3 and 4). The length of the sledge hammer located on the right-hand side of the prominant mound is approximately 3 feet.

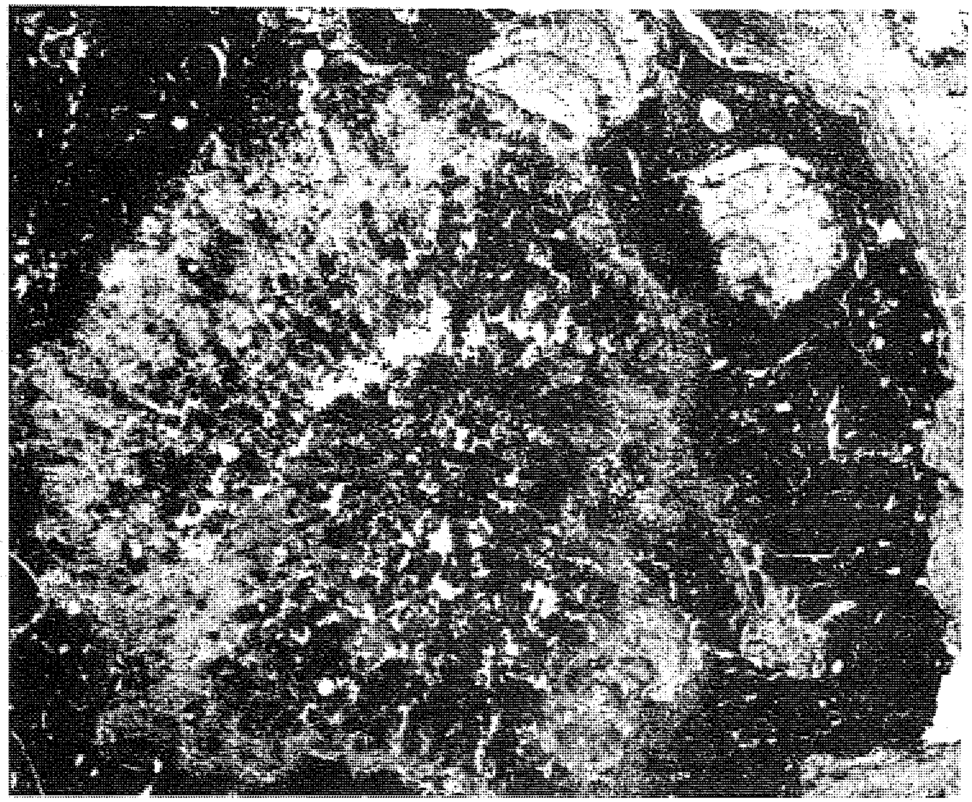

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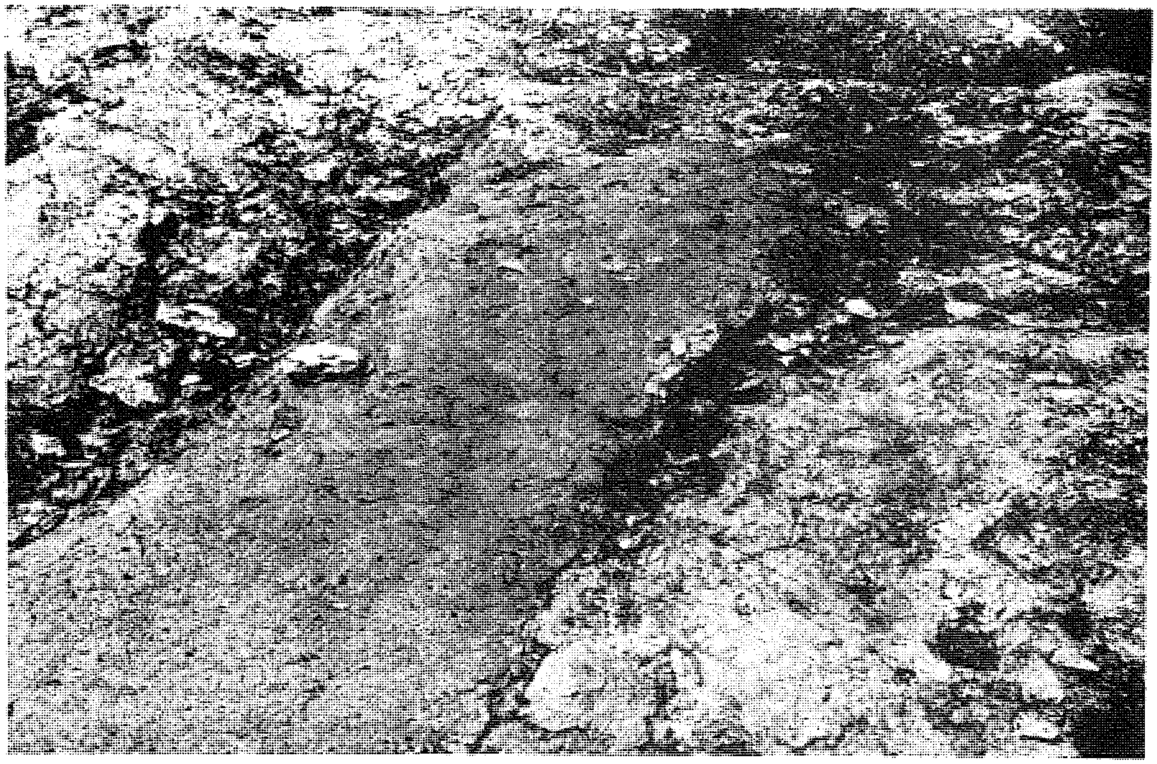
PLATE 3

Figure 1 Thin section photomicrograph (X3) of a transverse cut through the sponge Zittelella in what is typically Crownpointian mound rock, LaBombard's Pasture, Isle La Motte, western Vermont. Note overall muddy character of the rock, and the appearance of an encrusting bryozoan? on the outer surface of the sponge.

Figure 2 Thin section photomicrograph (X14) of Crownpointian mound rock with relatively abundant encrusting (bead-like segments) algae of the genus Sphaerocodium, LaBombard's Pasture, Isle La Motte, western Vermont. Again, note the dominantly muddy character of the rock. Scattered small grey flecks are floating dolomite rhombs.



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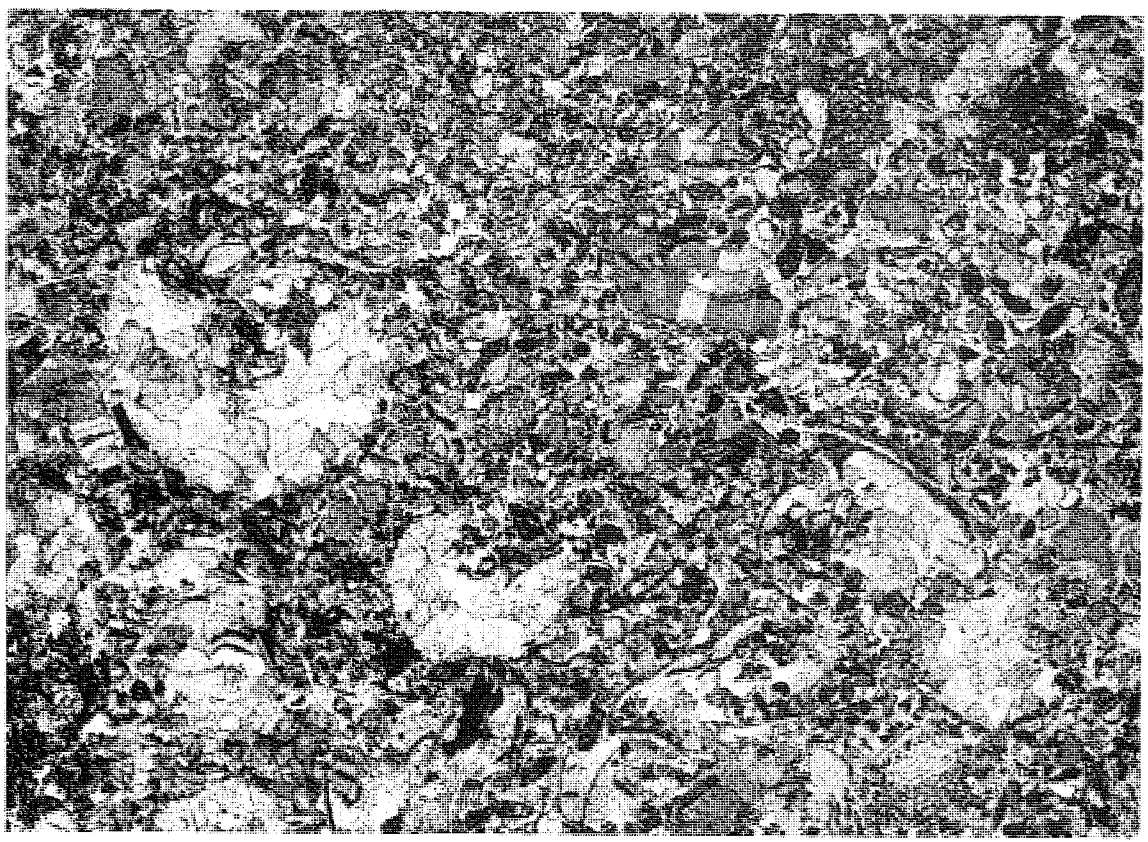

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PLATE 4

Figure 1 — Outcrop photograph of a channel cutting mound rock in the Middle Ordovician (Chazyan) Crown Point Formation, west of the Goodsell Quarry, Isle La Motte, western Vermont. The width of the channel is approximately 18 inches. Note lighter-colored mound rock on either side of darker-colored channel rock.

Figure 2. — Thin section photomicrograph (X4) of channel rock from the above locality. Rock is primarily a pelmatozoan calcarenite, although intraclasts (small rounded dark grains), bryozoan and brachiopod fragments are also present. Cavities or original void spaces filled with secondary granular sparry calcite are also common within the channel rock.

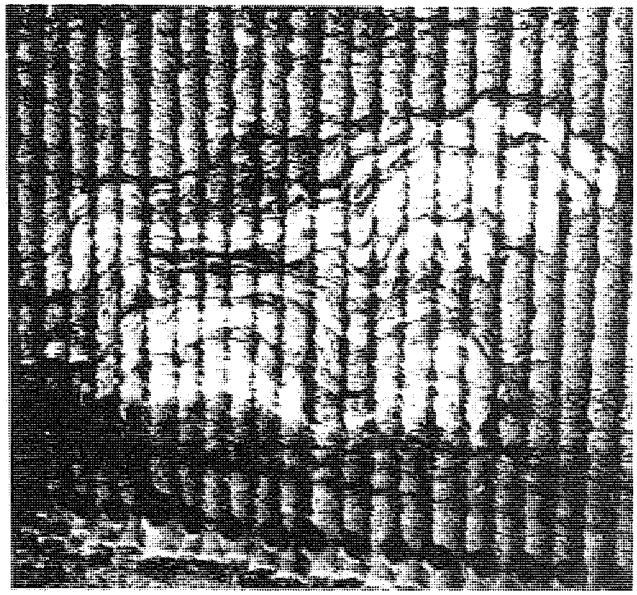

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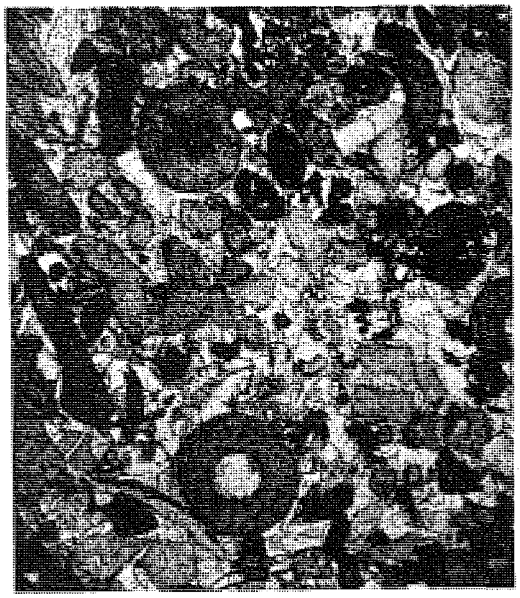
PLATE 5

Figure 1 — Stromatoporoid (Pseudostylodictyon ?) mound exposed on the south wall of Goodsell Quarry (June, 1962); Middle Ordovician (Chazyan) Crown Point Formation, Isle La Motte, western Vermont. Stromatoporoid mound is approximately 4 feet in width 2 1/2 feet in height.

Figure 2 — Thin section photomicrograph (X4) of a vertical section of the stromatoporoid Pseudostylodictyon ? chazianum (Seely) from the lower part of the Crown Point Formation, LaBombard's Pasture, Isle La Motte, western Vermont. Specimen shows characteristic thin laminae separated by pronounced layers of lime mud.

Figure 3 — Thin section photomicrograph (X4) of a horizontal section of the stromatoporoid Pseudostylodictyon ? eatoni (Seely) showing mamelons of various sizes, from the lower part of the Crown Point Formation, LaBombard's Pasture, Isle La Motte, western Vermont.



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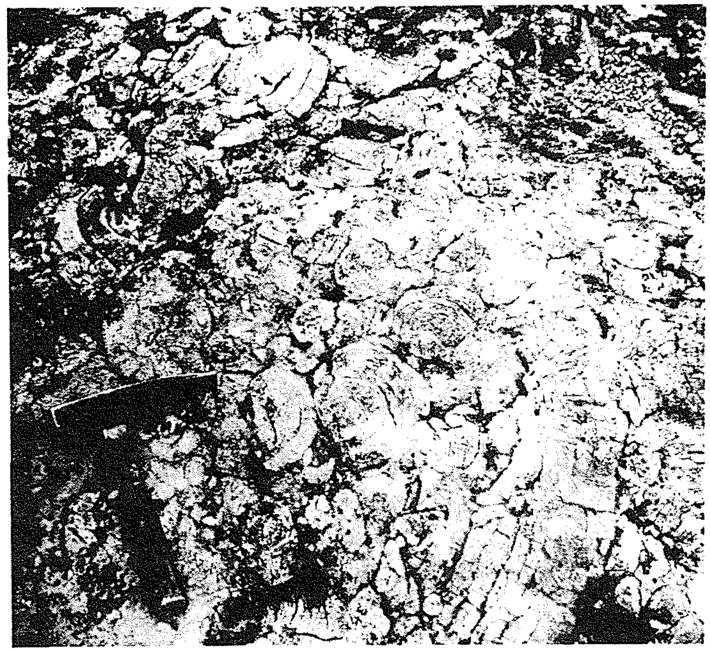

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Figure 1 Thin-section photomicrograph (X8) of a pelmatozoan grainstone from a channel associated with the Lamottia buildup ("Lamottia reef" of Raymond), Day Point Formation (Middle Chazyan) middle Fleury Member, Isle La Motte, western Vermont. Note abundance of pelmatozoan ossicles (probably cystoid and/or blastoid), and the dominant sparry calcite matrix; many of the pelmatozoan ossicles have calcite overgrowths. The small black grains are intraclasts and/or Girvanella pellets (diagnostic structures not seen at this magnification).

Figure 2 Outcrop photograph of the Lamottia accumulation ("Lamottia reef") located near the center of the buildup. Note jumbled mass of coral "heads" which appear to be heaped together and overturned, probably due to wave and/or current sorting. Length of hammer is 11 inches; Day Point Formation, middle Fleury Member, southeastern Isle La Motte, western Vermont.

Figure 3 Thin-section photomicrograph (X4) of the muddy rock matrix between the massive Lamottia "heads". Rock can be classified as a skeletal wackestone. Note large fragment of the tabulate coral Lamottia heroensis Raymond set within a muddy matrix with included intraclasts and abundant skeletal debris. Thin-section taken from rock near the center of Raymond's "oldest coral reef" within the Day Point Formation, Middle Fleury Member, southeastern Isle La Motte, western Vermont.

REFERENCES

Hartman, W. D., and Goreau, T. F., 1966. Ceratoporella, a living sponge with stromatoporoid affinities, Amer. Zoologist, v. 6, p. 262.

Lowenstam, H. A., 1957, Niagaran reefs in the Great Lakes area, IN Ladd, H. S., ed., Treatise on marine ecology and paleoecology, volume 2, Paleoecology, Geol. Soc. Amer. Mem. 67, p. 215-248.

Madri, P. P., Claus, G., Kunen, S. M., and Moss, E. E., 1967. Preliminary studies on the Escherichia coli uptake of the redbeard sponge (Microciona prolifera Verill), Life Sciences, v. 6, p. 889-894.

Odum, E. P., 1959. Fundamentals of Ecology, 2nd ed., Wiley, New York

Pitcher, M., 1964, Evolution of Chazyan (Ordovician) reefs of eastern United States and Canada, Bull. Canadian Petrol. Geol., v. 12, p. 632-691.

Rasmont, R., 1968. Nutrition and digestion. IN Florkin, M., and Scheer, B. T., Chemical Zoology, volume II, Porifera, Coelenterata and Platyhelminthes, p. 43-51.

Raymond, P. E., 1924. The oldest coral reef, Vermont State Geologist, 14th Rept., p. 72-76.

Roll, A., 1934, Form, Bau und Entstehung der Schwammstotzen im süddeutschen Malm, Palaeontol. Zeitschr., v. 16, p. 197-246.

Sieber, R., 1937, Neue Untersuchungen über die Stratigraphie und Oekologie der Alpinen Triasfaunen. I. Die Fauna der nordalpinen Rhätriffkalke, Neues Jahrb., Beilage Band, v. 78, pt. B, p. 123-187.

MIDDLE ORDOVICIAN (CHAZYAN) MOUNDS*, SOUTHERN QUEBEC, CANADA: A SUMMARY REPORT

by

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INTRODUCTION

Data relative to the biotic composition, petrology, and gross morphology of four small Middle Ordovician mounds exposed in southern Quebec, Canada, are summarized. It is noted: (1) that the Quebec Chazyan mounds are probably uppermost Crownpointian-to lowermost Valcourian in age, (2) the known outcrop distribution of these mounds is very patchy, (3) the mounds are very small scale structures, and (4) the biotic composition appears to be relatively simple. Brief comparison is made with the Chazyan mounds exposed in the Lake Champlain region of the northeastern United States, and marked differences in biotic composition, gross morphology, and occurrence are stated. It is thought that the mounds probably grew and thrived in relatively shallow marine waters, within a depth range of perhaps a few tens of feet.

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*The term mound is herein defined as an organic carbonate buildup, commonly of relatively small size, devoid of obvious bedding features, and containing a biota different from the usually bedded surrounding sediments. Used in this sense, the Chazyan mounds are thought to have been centers of organic activity whose fossil skeletons are assumed to be in growth position, and whose growth directly influenced surrounding sedimentational and biotic patterns because of their relative relief in relationship to the surrounding sea floor. The term "reef", used in a present-day sense, i.e., a wave resistant structure composed principally of hermatypic corals welded together by lime secreting algae, is rejected for the Chazyan mounds discussed here since the biotic potential necessary to build and perpetuate a massive dominantly wave resistant structure did not exist this early in geologic time.

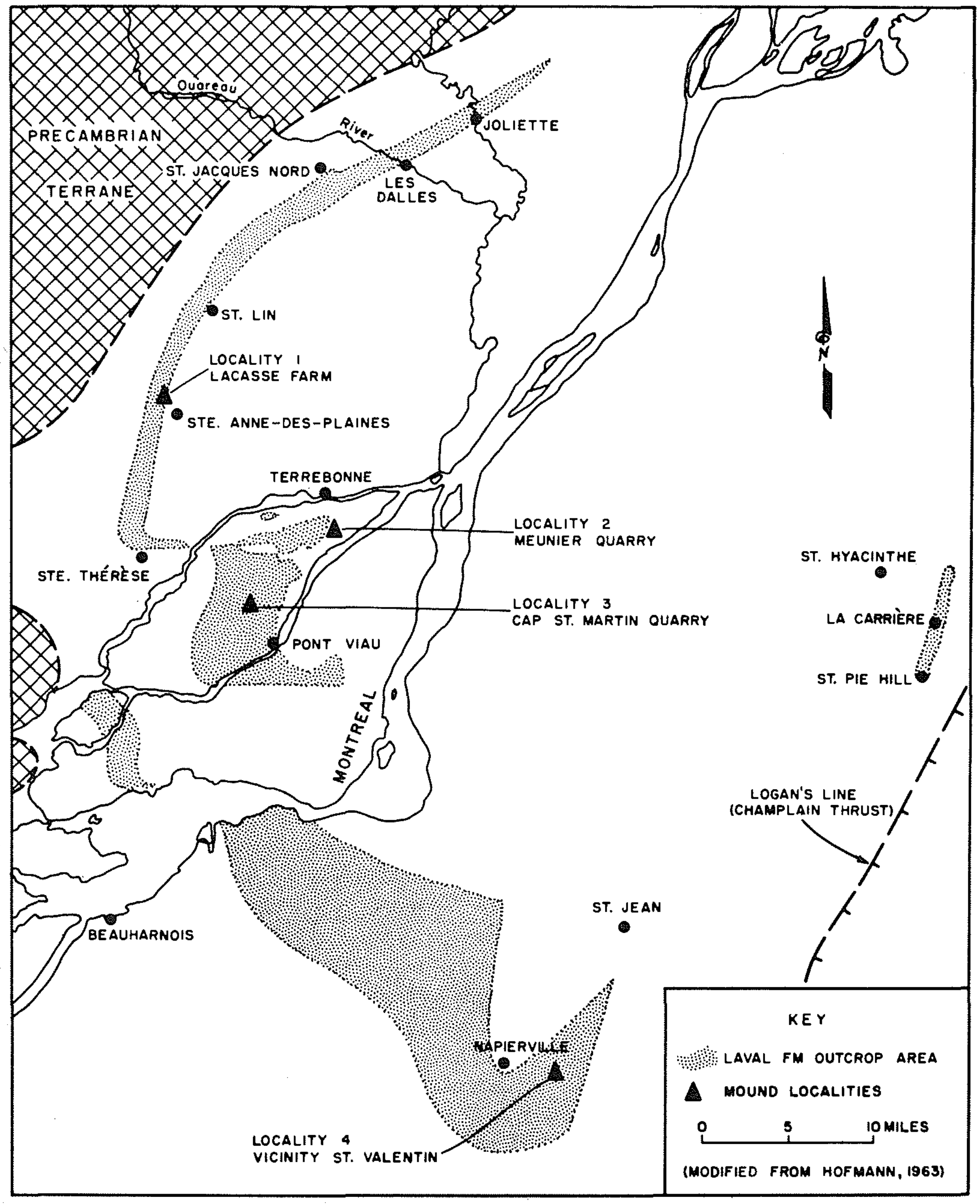

Fig. 1 Regional locality map of southern Quebec, showing the locations of known Chazyan mound outcrops.

STRATIGRAPHY OF THE QUEBEC CHAZYAN

Carbonate mounds are exposed in the Middle Ordovician (Chazyan) Laval Formation of southern Quebec, Canada, see Fig. 1. According to the data presented by Hofmann (1963, p. 273) this formation attains a thickness of 970 feet approximately 20 miles northeast of Three Rivers, Canada. Towards the Montreal region the formational thickness ranges from 205 to 390 feet, whereas south of St. Jean the thickness ranges from 450 to 670 feet. The thickness isopach contour map given by Hofmann (p. 274) shows a somewhat northerly trend.

Hofmann (1963, p. 271-273) informally subdivided the Laval Formation into three distinctive lithologic units: (1) the lower Laval consisting of quartz-sandy beds, (2) the middle Laval of shaly limestones and calcarenites, and (3) the upper Laval beds of predominantly argillaceous calcisiltite, which gradually becomes more dolomitic towards the top. Within this lithologic subdivision two main biostratigraphic units are recognized: (1) an upper division (upper Laval beds) in which the brachiopods Rostricellula plena (Hall) and Linguella? sp. are very abundant, and (2) a lower, pre-Rostricellula plena division (= middle and lower Laval) characterized by the cystoid Bolboporites americanus Billings. The carbonate mounds are confined to a stratigraphic position at the top of the Bolboporites Zone (Hofmann, p. 280). These structures have only been recognized in outcrop and their distribution appears to be very spotty (Fig. 1).

Stratigraphically, the Rostricellula plena Zone can be correlated with the Valcour Formation of the Lake Champlain region of northeastern New York and western Vermont. The Chazyan section below the Rostricellula plena Zone of the Laval Formation is characterized by several fossils, amongst which are the forms Bolboporites americanus and the brachiopod Sphenotreta acutirostris (Hall), species originally found by Raymond (1906), and thought by him to be restricted to the lower Chazy and lower parts of the middle Chazy. On this basis Hofmann (p. 290) considers that the middle and lower Laval Formation may be correlated with the Day Point-Crown Point Formations, undivided. Although Pitcher (1964, p. 658) in his study of the evolution of the Chazyan mounds in the Lake Champlain region thought that the mounds which he examined at Terrebonne (Locality 2) were definitely Valcourian (upper Laval) in age due to the presence of the trepostome bryozoan Batostoma campensis Ross, found with the brachiopod Rostricellula plena, a reliable Valcourian indicator. However, the mound exposure near St. Valentin (Locality 4) carries abundant Batostoma chazyensis Ross, a form that Pitcher believes is restricted to the Crown Point and Day Point Formations (middle and lower Laval), see Pitcher, 1964, p. 674. Accordingly, Pitcher (p. 639) correlates the St. Valentin mound exposure with the upper Crown Point Formation of the Lake Champlain region.

MOUND FACIES

The distribution of four known Chazyan mound outcrops in southern Quebec, Canada, is shown in Fig. 1. All of the mounds are relatively small structures, not exceeding 7 feet in height, and usually less than 50 feet in width. All have somewhat irregular form (see Fig. 2) and are not simple dome-shaped features. The mounds are totally devoid of bedding structures, although bedded sediments of different lithologies underlie, overlie, and abut and grade into the mounds. Some of the off-mound lithologies dip away from the mound at 10 to 25 degrees, which rapidly decreases to zero degrees a relatively few feet away from the mound.

Petrologically, the mounds can be classified as trepostome bryozoan (*Batostoma*) and/or tabulate coral (*Eofletcheria* and/or *Billingsaria*) boundstones (Dunham, 1962). The more muddy portions of the mound would be classified as skeletal wackestones indicating a general absence of a binding biota. The mound rock matrix is dominantly muddy, although scattered floating dolomite rhombs and a few mud intraclasts can usually be recognized. Accessory biotic components common to the mound rock are the red alga *Solenopora*, sponge spicules, pelmatozoan debris (cystoid and blastoid), trepostome bryozoans (*Atactotoechus* and *Chazydictya*), brachiopods (*Hebertella* and/or *Rostricellula*), gastropods, trilobites and ostracodes. The beds that overlie, underlie, abut, and grade into the mound are usually relatively mud-poor coarse pelmatozoan packstones/grainstones with some bryozoan fragments and mud intraclasts. The channel rock, well developed at Locality 4 (vicinity of St. Valentin) is identical to the latter (see Figure 4). Most of the non-mound rock skeletal grains, principally pelmatozoan debris, contain pronounced calcite overgrowths. A summary of pertinent mound attributes for each outcrop locality is given in Table 1.

Paleoecologically the mound rock can be dominated by either one of two prevailing biotic components. Either the mound is composed predominantly of tabulate corals (*Eofletcheria* and/or *Billingsaria*), as at Localities 1 and 4, or the mound biota may be primarily composed of sheet-like encrusting trepostome bryozoans (*Batostoma*), as at Localities 2 and 3 (see Table 1).

When the mound biota is dominated by tabulate corals, one or two growth forms are usually characteristic. *Eoflectheria incerta* (Billings), the dominant species at Locality 1, commonly forms a series of colonies growing one upon another. The colonies are less than 1 foot in length and up to 6 inches in height. These colonies thin towards their edges and appear to be circular in plan view. On the other hand, *Billingsaria parva* (Billings) the dominant species at Locality 4, usually forms flat-lying colonies no more than 12 inches in length and only two inches in height. At the other two localities, Localities 2 and 3, where tabulate

LOCALITIES	GENERAL DIMENSIONS	ROCK TYPES	PRIMARY MOUND BUILDERS	MOUND DETRITUS CONTRIBUTORS	DESCRIBED BY
1. Lacasse Farm	Height: 5 ft. Width: 30-50 ft. Length: about 700 ft.	Mound mainly a tabulate coral boundstone with pelletoidal & intraclastic wackestone; Overlying and underlying beds are coarse pelmatozoan packstones	Eoflectcheria incerta* Billingsaria parva Batostoma sp.	Hebertella vulgaris Rostricellula plena Indet. gastropods	MacGregor (1954 p. 44-46); Clark (1960, Fig. 1D)
2. Meunier Quarry	Height: 7 ft. Width: 8 ft.	Mound mainly a bryozoan boundstone; Overlying and underlying beds are coarse pelmatozoan packstones	Batostoma campensis* Eoflectcheria incerta	Malocystites murchisoni Rostricellula plena Hebertella sp. Solenopora sp. Sponges (spicules)	MacGregor (1954 p. 23); Hofmann (1963, p. 281); Pitcher (1964, p. 373)
3. Cap St. Martin Quarry	Height: 1 ft. Width: 2½ ft.	Mound mainly a bryozoan boundstone; Overlying and underlying beds are coarse pelmatozoan packstones	Batostoma campensis* Billingsaria parva	Blastoidocrinus carchariadens Hebertella vulgaris Rostricellula plena Solenopora sp.	MacGregor (1954, p. 43)
4. Vicinity St. Valentin	Height: 6 ft. Width: 30 ft. Length: 120 ft.	Mound mainly a tabulate coral boundstone; Overlying and underlying beds are bryozoan and pelmatozoan grainstones with intraclasts; Channels are pelmatozoan packstones	Billingsaria parva* Batostoma chazyensis Chazydictya sp.	Rostricellula plena Indet. gastropods Indet. ostracodes Solenopora sp. Sponges (spicules); Channels contain many: Atactotoechus? sp. Chazydictya sp. Batostoma sp.	Hofmann (1963, p. 280-281); Pitcher (1964, p. 667); and original observations

*Dominant biotic element present as the primary mound-builder

TABLE 1 Comparison of Some Chazyan Mound Attributes, Southern Quebec, Canada corals are only of secondary importance they may be found as either small bun-shaped colonies up to 12 inches long and 9 inches high, or as flat-lying colonies up to 9 inches long and only 2 inches high. The encrusting sheet-like bryozoan Batostoma overgrows many of the tabulate corals and probably functions as a binding and cementing agent that readily stabilized the primary mound builders--the tabulate corals.

When the dominant mound organism is bryozoans, as at Localities 2 and 3, the encrusting trepostome bryozoan Batostoma campensis Ross, appears as consecutive sheets stacked one above the other in conspicuous layers less than one-half inch in thickness. Layers of muddy sediment occur between the bryozoan colonies and usually contain varying amounts of floating dolomite rhombs. In appearance, this series of bryozoan sheets make up cabbage-head masses piled one on top of another. The result of this stacking of bryozoan colonies is the formation of a distinctive, somewhat massive, unbedded, irregular mass of boundstones. Small colonies of the tabulate corals Eoflectcheria and Billingsaria may occur within the mound proper or on the tops of the mound.

Figure 5 schematically shows the biotic attributes of a composite Middle Ordovician (Chazyan) mound community. Principally, this figure shows which organisms are present within the mound community, what their contribution to the perpetuation of the mound structure might be, and their vagility and specific feeding types. As such, this diagram presents a functional grouping of the participating biotic elements within the mound community. In the Chazyan mounds of southern Quebec, the role of the primary mound builders is a relatively simple one, since all that appears to be necessary is a colonial organism possessing reasonable constructional ability. In this instance, the tabulate corals Billingsaria parva (Billings) and Eofletcheria incerta (Billings) satisfy this requirement. To further insure perpetuation of the constructional aspect of the mound, sheet-like trepostome bryozoans, especially species of the form Batostoma, encrust and bind the coral heads together, and in doing so, thus reinforce and strengthen the structural unit. In some cases encrusting trepostomatous bryozoans may be the sole dominant biotic element forming some of the mounds, especially the very small mounds. It thus appears that when bryozoans are the sole primary builders there is a definite limitation as to their size building potentialities. It is to be noted that all of these primary mound building organisms are cemented sessile benthos, and all are suspension feeders.

The detritus contributors live on, within, and around the mound and are an integral segment of the mound community. Of especial note are the abundant cystoids and blastoids which probably formed "meadows" around, about, and also lived within the mound itself. Their importance in a biological sense cannot be underestimated, since they undoubtedly affected sedimentational and biological patterns around the mound. Most apparently, acting as a baffle element in trapping finer sediment which normally might by-pass the area, but more importantly, their very presence created additional ecological niches within which a host of other organisms could flourish, and in essence be able to transform the immediate mound area into a center of organic activity. Other detritus contributors include the red alga Solenopora, sponges, bryozoans--specifically trepostomatous types, pedunculate brachiopods, gastropods, trilobites, and ostracodes. Obviously, there were probably other organisms, in particular those without preservable hard parts. However, it is difficult to estimate what percentage of the total biota would define this category. Perhaps, if reliably known it probably would not be an appreciable percentage, especially in Middle Ordovician time when a good many of the potential ecologic niches on a mound were probably not fully exploited even by the indigenous biota. It should be emphasized that these are the organisms that contribute the bulk amount of mound detritus. In vagility they range from cemented and rooted forms to nektonic creatures, and feeding types represented bridge the gamut from suspensions feeders to carnivores and autotrophs.

In summary, interpolation of the data presented by MacGregor (1954), Hofmann (1964), Pitcher (1964), and original observation, a generalized sequence of the mound building events, as exemplified by the four Chazyan mounds of southern Quebec, may be recorded as follows:

(1) an initial stage consisting of a limesand (packstone/grainstone) substrate usually containing abundant rhynchonellid brachiopods (Rostricellula), some cystoids (Malocystites), and cryptostome bryozoans (Chazydictya);

(2) the inclusion of small scattered patches of encrusting trepostome bryozoans (Batostoma);

(3) discontinuous thin layers of the tabulate coral Billingsaria growing over and encrusting the earlier bryozoans and the lime sand substrate, associated with relatively small colonies of the tabulate coral Eofletcheria filling-in available interspaces;

(4) sheet-like encrustations of the bryozoans Batostoma and Chazydictya encrusting the tabulate corals and binding some of the coral heads together;

(5) piling-up, one on top of another, of tabulate coral colonies (usually only one form being dominant in a particular mound), and the binding together of the corals by the encrusting brozoans thus strengthening and welding together the mound structure.

It should be noted that many of the smaller Chazyan mounds, usually those that range in size from somewhere under three feet in length and two feet in height, are composed almost entirely of encrusting bryozoans and hence, do not show this evolutionary sequence as noted above.

It is thought that the mounds grew and thrived in relatively shallow waters, within a depth range of perhaps a few tens of feet. The mound facies (boundstones/wackestones) appears to have been deposited under quiet water conditions, and it is probable that most of the fine mud present within the mounds was accumulated and deposited mainly by the baffling action of rather commonly occurring pelmatozoans. The off-mound packstones appear to represent more agitated water environments, whereas the coarse grainstones, usually adjacent to the mounds, suggest relatively high energy open-circulation conditions. There is no substantial evidence to support Hofmann's contention (1963, p. 281) that the mounds were subaerially exposed and eroded. On the contrary, most evidence seems to indicate permanent submergence for these small, geologically short-lived mounds.

COMPARISON WITH THE CHAZYAN MOUNDS OF THE LAKE CHAMPLAIN REGION

Mound structures occur quite commonly in the Middle Ordovician (Chazyan) rocks of the Lake Champlain region of northeastern New York, and western Vermont and have been reported by Oxley (1951), Erwin (1957), Oxley and Kay (1959), and Pitcher (1964). In this area, the Chazyan is subdivided into three formations: Day Point, Crown Point, and Valcour (in ascending order). The maximum thicknesses of the Day Point and Crown Point Formations are in the general vicinity of Valcour Island, New York, where each formation exceeds 300 feet; the greatest thickness of the Valcour Formation occurs on South Hero Island, Vermont, where it is approximately 200 feet (Oxley and Kay, 1959, p. 840).

The mounds are usually oval in shape with widely variable dimensions. In general, they range in size from small masses 1 to 3 feet in length and 2 feet in height, to an average of approximately 25 feet in height and up to 300 feet in length. The mounds of the Day Point Formation are the smallest in size, whereas the Crown Point mounds are the largest and best developed; the dimensions of the Valcour mounds lie somewhat between these two extremes.

Pitcher (1964) has demonstrated that the Chazyan mounds show distinct changes in organic composition through time. The early Chazyan Day Point mounds were primarily constructed by trepostome and cyclostome bryozoans, which built relatively small linearly aligned mound structures. The mound core of carbonate mud and skeletal debris (boundstone) differs from the cross-bedded, mud-free skeletal packstones/grainstones surrounding the mounds.

The middle and upper Chazyan (Crown Point and Valcour) mounds can be subdivided into five principal biotic component variants. These can be considered as primary mound builders with distinct and variable gradations existing amongst them. The primary mound builders are: (1) algae (*Girvanella*, *Sphaerocodium*, *Solenpora*, and possibly *Nuia*), (2) lithistid sponges ("*Zittelella*" and others), (3) tabulate corals (*Billingsaria*), (4) bryozoans (*Batostoma* and others), and (5) stromatoporoids (*Cystostroma* and *Pseudostylodictyon*). The mound core matrix is muddy and can be classified as a boundstone. Surrounding and overlying the mounds are well sorted calcarenites that can be classified as pelmatozoan grainstones or packstones, in which *Girvanella* pellets and *Solenopora* colonies are conspicuous elements. Channels have been cut through many of the mound structures, and these form a rather prominent anastomosing pattern. On Isle La Motte, where channels are particularly well developed in the Crown Point mounds, they stand out quite distinctively from the surrounding massive gray-colored mound structures. The channels are usually brownish in color, and this is believed due to the higher dolomite content of the calcarenites. The various mound biotas are believed to have existed contemporaneously, and appear to have been closely associated throughout middle and upper Chazyan time. Significantly, Pitcher (1964, p. 659) observed that there is distinct vertical biotic differentiation within any single mound. It appears that stromatoporoids are usually dominant through scattered biotic elements at the base and tops of a mound, but they are usually very sparse within the middle portion. In the Crown Point mounds, exposed in Patnodes Pasture on Isle La Motte, stromatoporoids are particularly common on the upper surfaces, and as such may represent the dominant element of a climax community. This seems to be substantiated by shallow cores taken by Pitcher, which show that the stromatoporoids do not extend for any depth below the surface.

Compared to the mounds of southern Quebec, those of the Lake Champlain region are much more abundant, and they are stratigraphically present throughout most of the Chazyan interval. Most importantly, the Lake Champlain mounds are biotically more complex and diversified, and exhibit a progressive sequence in the evolutionary development of mound building organisms not recorded from any other locale. Petrologically, the presence of mound rock types (boundstones) and offmound calcarenites (packstones/grainstones) are identical to that observed in the four Chazyan mounds of southern Quebec. However, channel cuts through the mounds, a feature perhaps closely analogous to modern-day reef surge channels, are a much more extensive and commonly occurring feature on the Lake Champlain mounds, especially those in the Crown Point Formation.

ACKNOWLEDGEMENTS

We are particularly grateful to Dr. Max Pitcher of Continental Oil Company for personally guiding us to and helping to collect the St. Valentin mound exposure, and for loaning the senior writer his collection of thin-sections that were prepared from the Terrebonne mounds. Dr. June Ross of Western Washington State College gave invaluable aid with identifications of the bryozoans.

MacGregor (1954) in an unpublished thesis had originally described three of the four mound localities (Localities 1-3) and presented pertinent paleoecologic observations as to the overall biotic composition of these Chazyan mounds. MacGregor's observations pertaining to the mound exposures near Ste. Anne-des-Plaines (Locality 1) have subsequently been summarized by Clark (1960, p. 26-27, Fig. 1D). Data from all of the above localities, including those of Pitcher (1964), along with original observations made at Localities 2 and 4 have been consolidated and interpolated within this summary.

A. MOUNDS OF SOUTHERN QUEBEC, CANADA

1. SIZE: Relatively small features, <7 feet in height and 50 feet in width

2. STRATIGRAPHIC POSITION: Restricted to within a very narrow stratigraphic interval of the Chazyan; mounds present at only four outcrop localities; unknown in the subsurface, would probably be too small to detect; Hofmann (1963, p. 291) shows the mounds restricted to the uppermost part of the middle Laval Formation (Crownpointian) and extending into the lower portion of the upper Laval Formation (Valcourian); Pitcher (1964, p. 638-639) believes that the mounds in the vicinity of St. Valentin correlate with the upper Crown Point (middle Middle Chazyan) whereas those near Terrebone (vicinity Meunier & Cap St. Martin Quarries) are Valcourian (upper Middle Chazyan)

3. PETROLOGY: Mounds may be classified mainly as trepostome bryozoan and/or tabulate coral boundstones associated with skeletal wackestone intervals; overlying and underlying beds are coarse pelmatozoan packstones/grainstones with some bryozoans; tidal or surge channels are pelmatozoan/bryozoan packstones with intraclasts; channels only well developed at the St. Valentin locality

4. BIOTIC COMPOSITION: Relatively simple; may be dominated by either tabulate corals (Eofletcheria and/or Billingsaria) with trepostome bryozoans, or trepostome bryozoans (Batostoma) with tabulate corals (Eofletcheria and/or Billingsaria); accessory biotic components include algae (Solenopora), sponges (spicules), blastoids, cystoids, trepostome bryozoans (Atactotoechus & Chazydictya), brachiopods (Hebertella and/or Rostricellula), gastropods, ostracodes, and trilobites

B. MOUNDS OF THE LAKE CHAMPLAIN REGION, NEW YORK & VERMONT

1. SIZE: Much variability in size; from very small to relatively large

2. STRATIGRAPHIC POSITION: Abundantly scattered throughout the Chazyan (Day Point, Crown Point, & Valcour); best developed and most numerous in the Crown Point interval on Isle La Motte, western Vermont; unknown in the subsurface

3. PETROLOGY: Day Point mounds are primarily bryozoan boundstones associated with pelmatozoan and bryozoan packstones/grainstones; Crown Point mounds are algal-sponge-stromatoporoid boundstones with associated pelmatozoan packstone channels; Valcour mounds are mainly algal-trepostome bryozoan boundstones with associated bryozoan-brachiopod packstones; channels best developed in the Crownpointian

4. BIOTIC COMPOSITION: Very complex; the Day Point mounds are mainly composed ot trepostome (Batostoma) and cyclostome (Cheiloporella) bryozoans; the Crown Point & Valcour mounds can be subdivided into five principal biotic component variants with distinct and variable gradations existing amongst them. The primary mound builders are: (1) algae (Girvanella, Sphaerocodium, Solenopora, and possibly Nuia), (2) sponges ("Zittelella" & others), (3) tabulate corals (Billingsaria), (4) bryozoans (Batostoma & others), & (5) stromatoporoids (Cystostroma and Pseudostylodictyon); accessory components include cystoids, blastoids, brachiopods, trilobites, ostracodes

TABLE 2 Comparison Of The Chazyan Mounds In Southern Quebec With Those Of The Lake Champlain Region

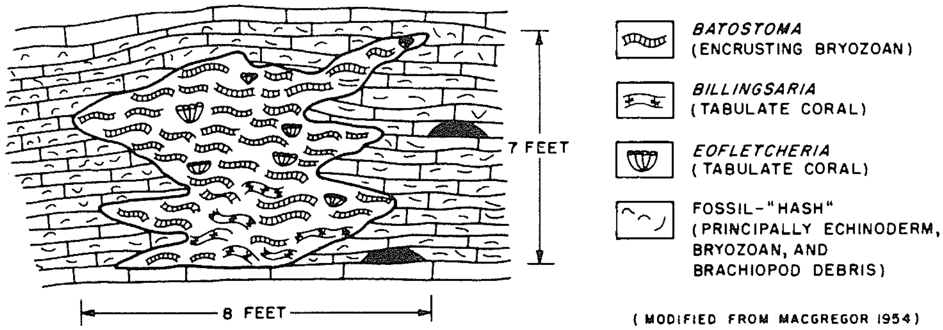

Fig. 2 Generalized diagram of a mound exposed in a quarry near Terrebonne (Locality 2).

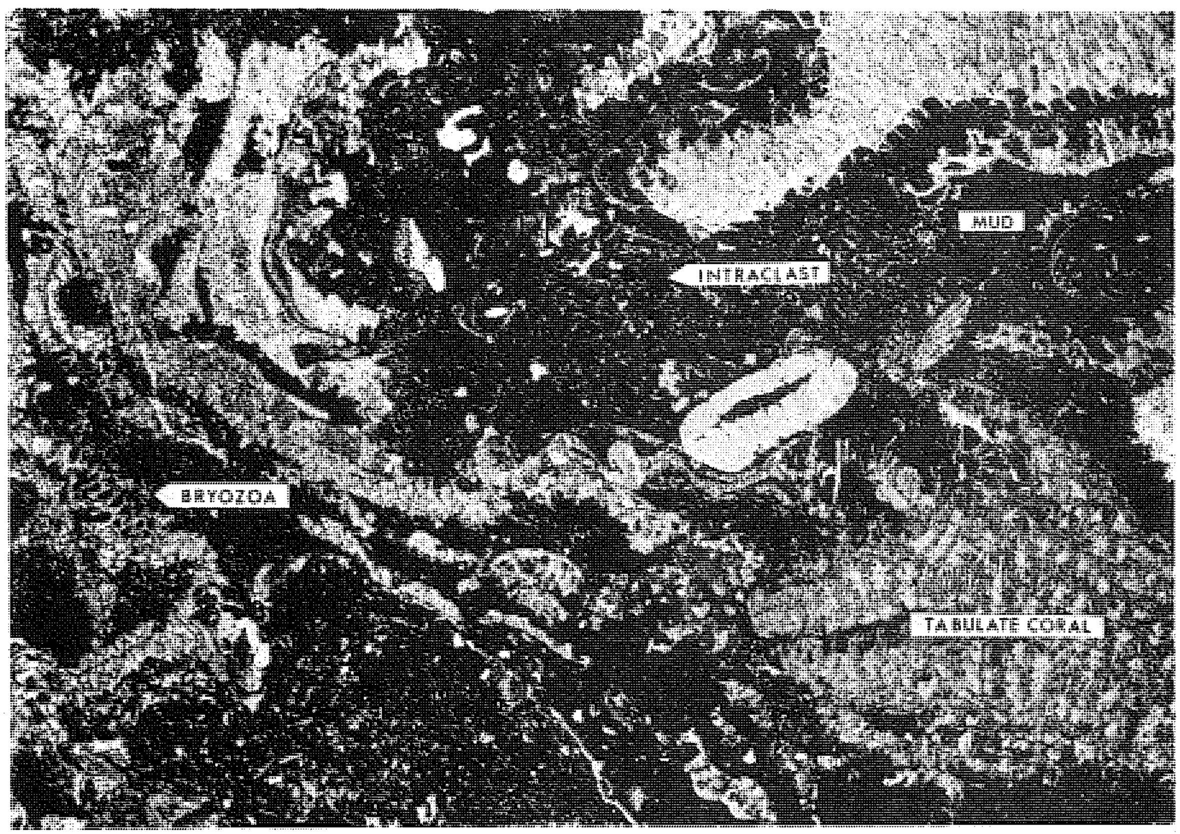

Fig. 3 Thin-section photomicrograph (X4) of mound rock (boundstone) Laval Formation, at Locality 4 (vicinity of St. Valentin), southern Quebec, Canada. Note the relatively abundant remains of the tabulate coral Billingsaria parva (Billings), encrusting bryozoans, and varied shell debris. Matrix is primarily a fine-grained mud containing scattered dolomite rhombs (small white flecks) and a few mud intraclasts.

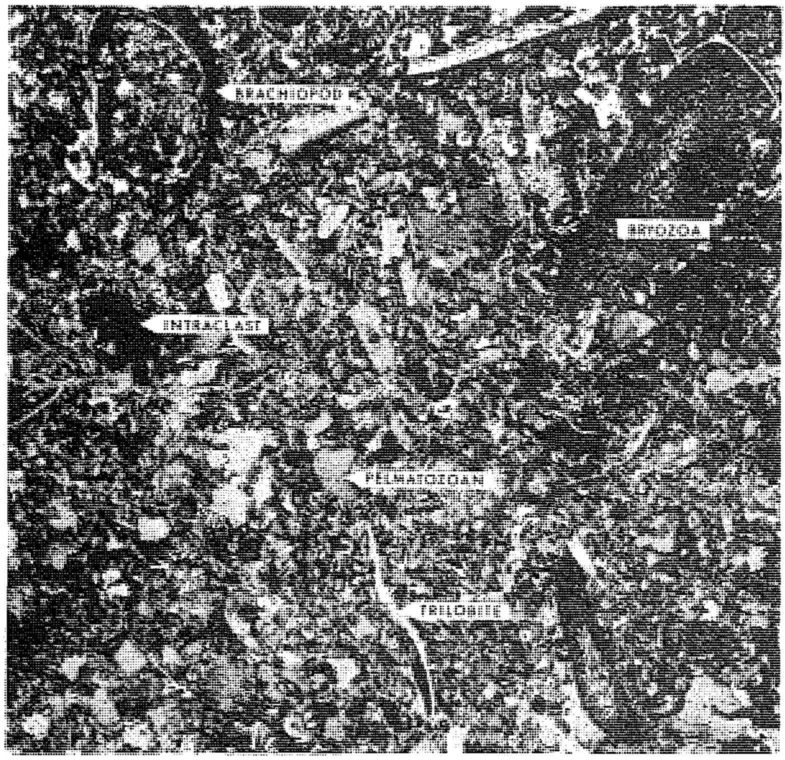

Fig. 4 Thin-section photomicrograph (X4) of channel rock (packstone) Laval Formation, at Locality 4 (vicinity of St. Valentin), southern Quebec, Canada. Note abundance of pelmatozoan debris (many grains with calcite overgrowths), along with conspicuous trepostome bryozoans (cf. Chazydictya sp.), brachiopod and trilobite fragments. Mud intraclasts commonly occur within the channel rock.

CHAZYAN MOUND COMMUNITY (SOUTHERN QUEBEC, CANADA)

CONTRIBUTION	ORGANISM	VAGILITY	FEEDING TYPE
PRIMARY BUILDERS	1. TABULATE CORALS a. *BILLINGSARIA PARVA* (BILLINGS) b. *EOFLETCHERIA INCERTA* (BILLINGS)	CEMENTED	SUSPENSION FEEDERS
	2. TREPOSTOME BRYOZOA a. *BATOSTOMA* SPP.	CEMENTED; ENCRUSTING	SUSPENSION FEEDERS
DETRITUS CONTRIBUTORS	1. PELMATOZOANS a. *MALOCYSTITES MURCHISONI* BILLINGS b. *BLASTOIDOCRINUS CARCHARIAEDENS* BILLINGS	ROOTED	SUSPENSION FEEDERS
	2. TREPOSTOME BRYOZOA a. *CHAZYDICTYA* SP. b. *ATACTOTOECHUS* ? SP.	CEMENTED	SUSPENSION FEEDERS
(DECREASING ABUNDANCE)	3. BRACHIOPODS a. *ROSTRICELLULA RAYMONDI* COOPER b. *R. PLENA* (HALL) c. *HEBERTELLA VULGARIS* RAYMOND	ROOTED	SUSPENSION FEEDERS
	4. OSTRACODES	VAGRANT BENTHONIC; NEKTONIC	DEPOSIT FEEDERS AND / OR SCAVENGERS
	5. TRILOBITES	VAGRANT BENTHONIC; NEKTONIC	DEPOSIT FEEDERS, SCAVENGERS, CARNIVORES
	6. RED ALGAE a. *SOLENOPORA* SP.	CEMENTED	AUTOTROPHIC
	7. SPONGES	ROOTED	SUSPENSION FEEDERS
	8. GASTROPODS	VAGRANT BENTHONIC	DEPOSIT FEEDERS, SCAVENGERS

Fig. 5 Biotic attributes of a composite Middle Ordovician (Chazyan) mound community, southern Quebec, Canada.

REFERENCES

Clark, T. H., 1960, Palaeozoic Rocks, IN: Osborne, F. F., Clark, T. H., "New Glasgow-St. Lin Area", Geol. Rept. 91, Quebec Dept. Mines, pp. 20-29

Dunham, R. J., 1962, "Classification of Carbonate Rocks According to Depositional Texture", IN: Classification of Carbonate Rocks, A Symposium, Ham, W. E., Ed., Amer. Assoc. Petrol. Geol., Mem. 1, pp. 108-121.

Erwin, R. B., 1957, "The Geology of the Limestone of Isle La Motte and South Hero Island, Vermont", Vermont Geol. Survey Bull. 9, 94 pp.

Hofmann, H. J., 1963, "Ordovician Chazyan Group in Southern Quebec", Amer. Assoc. Petrol. Geol., Bull., Vol. 47, No. 2, pp. 270-301.

MacGregor, A. R., 1954, "Chazyan Corals and Reefs", unpub. M. Sci. Thesis, McGill Univ., 83 pp.

Oxley, Philip, 1951, "Chazyan Reef Facies Relationships in the Northern Champlain Valley", Bull. Denison Univ., Jour. Sci. Lab., Vol. 42, Art. 10, pp. 92-111.

____________ and Kay, Marshall, 1959, "Ordovician Chazyan Series of Champlain Valley, New York and Vermont and its Reefs", Bull. Amer. Assoc. Petrol. Geol., Vol. 43, No. 4, pp. 817-853.

Pitcher, Max, 1964, "Evolution of Chazyan (Ordovician) Reefs of Eastern United States and Canada", Bull. Canadian Petrol. Geol., Vol. 12, No. 3, pp. 632-691.

Raymond, P. E., 1906, "The Chazyan Formation and its Fauna", Ann. Carnegie Mus., Vol. 3, pp. 498-598.

Ross, J. P., 1963, "The Bryozoan Trepostome Batostoma in Chazyan (Ordovician) Strata", Jour. Paleontology, Vol. 37, No. 3, pp. 857-867.

__________ 1963, "Ordovician Cryptostome Bryozoa, Standard Chazyan Series, New York and Vermont", Bull. Geol. Soc. America, Vol. 74, No. 5, pp. 577-608.

Toomey, D. F., 1967, "Additional Occurrences and Extension of Stratigraphic Range of the Problematical Micro-organism Nuia", Jour. Paleontology, Vol. 41, No. 6, pp. 1457-1460.

Trip G

SURFICIAL GEOLOGY AND GEOMORPHOLOGY OF WHITEFACE MOUNTAIN AND KEENE VALLEY

by

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The Pleistocene history of the Adirondack Mountains has largely been interpreted from geomorphic evidence. Kemp, (1905), Alling (1911, 16, 19, 21), Ogilvie (1902), Fairchild (1913, 19, 32) and others wrote of cirques, tarns, aretes, horns and lateral moraines as evidence of local glaciation. However, they pictured this local glaciation development as a minor phase of the rapid deglaciation of the mountainous region, leaving an ice free island surrounded by the continental ice mass. This author believes that the local glaciers were very active during the time of continental deglaciation with valley glaciers extending as much as 10 miles down valley from their cirques.

A new theoretical model is proposed to explain this climatic situation of deglaciation in the St. Lawrence Lowland to the north and active glaciation in the Highlands to the south.

The model is based on the relationship of large pro-glacial lakes formed at the margin of the continental glacier and the effects of local storms developing over the lakes and moving eastward into the Adirondack Mountains of New York. Glacial conditions were maintained locally by the high snow fall from these "lake effect storms". Ablation of the local snow would be retarded by low temperatures related to the cooling at higher elevations, effect of the continental glacier at the northern edge of the Adirondacks and the cloud cover that would develop over the High Peaks area due to orographic uplift of the eastward-moving moist air.

These local climatic conditions would cease to exist when the continental glacier retreated sufficiently to open the St. Lawrence Valley, lowering the pro-glacial lakes and decreasing their size considerably. Therefore, the existence of mountain glacial conditions was dependent on the existence of pro-glacial lakes and the time of mountain glaciation is directly related to the history of those lakes. Local accumulation of snow started after the development of the first large pro-glacial lake west of the Adirondack Mountains and ended when the lake system drained below its present level.

All the evidence for local glaciation can not be seen on this trip. Some of the key exposures are 5 to 10 miles in from the roads and can be reached only by foot.

The trip is set up to show the best examples of the geomorphic and stratigraphic evidence that can be reached from buses and within the time limit allowed.

Topographic maps useful on this trip:

Lake Placid	Paradox Lake
Mount Marcy	Schroon Lake
Elizabethtown	Santanoni
Ausable	Newcomb

Stop 1 - Whiteface Mountain Ski Centre, Wilmington, N.Y.
Lake Placid Quadrangle

Proceed from the bus to the chair lift and ride to the end of first lift.

"Coon Pit" stop. After getting off the chair lift walk up stream to the waterfall. There is a deep glacial grove enlarging a joint. Which direction was the ice flowing to cut this grove?

Walk over to the second chair lift and proceed to the top of the mountain.

Stop 2 - Top of Whiteface Mountain

After leaving the chair lift walk to the top of the mountain. Gather at the elevation marker.

From this point you can observe the only indication I have observed of continental glaciation over riding the mountain tops in the high peaks region. This is the "roches moutonnees" form of the crest on the south side of the peak.

Observe the "horn" configuration of the peak area. The "cirque form" of the valleys leading away from the peak and the very narrow "arete" ridge to the west of the peak. Could these features develop from Continental Glaciation?

Walk down the north trail (not the tourist trail) to Wilmington Turn House and have lunch.

Stop 3 - Atmospheric Sciences Research Centre, Whiteface Mountain
Field Station, Wilmington, N.Y.

Leave the bus and walk up the jeep road to the Whitebrook valley trail (unmarked). We will be going about ½ mile up the valley. Anyone that wishes to, may stay at the centre and look around the lodge.

Whitebrook Valley Moraine. This ridge was described by Alling (1919) as a lateral moraine of a local glacier occupying this "cirque". Notice the high percentage of Potsdam sandstone pebbles lying on the surface.

Proceed back to the bus.

Stop 4 - County Gravel Pit on 9 N, 1 mile south of Keene, N.Y.
Mount Marcy Quadrangle

10' till, grey oxidize moderately stoney, very sandy at base, large folds of underlying sand, silt and clay carried upward into till.

50 to 100' fine sand, silt and clay upper 5' strongly contorted by load folding.

Return to Plattsburgh.

Trip H

META-ANORTHOSITE OF THE JAY-WHITEFACE NAPPE, AUSABLE FORKS-LAKE PLACID QUADRANGLES, NORTHEASTERN ADIRONDACKS, NEW YORK

by

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GEOLOGIC SETTING AND OUTLINE OF PRECAMBRIAN DEVELOPMENT OF ADIRONDACK META-ANORTHOSITE

The Adirondack Mountains comprise an outlier of the Canadian Shield. They are geographically and geologically subdivided into the Highlands, which are principally underlain by granitic gneisses and anorthosite and the Lowlands, composed mainly of metasedimentary rocks of the Grenville Series (Fig. 1). Table 1 is an outline of two opposing theories on the origin of the anorthosite and related Precambrian rocks of the Adirondack region (from de Waard and Walton, 1967). A principle difference between the two theories is the time of Grenville sedimentation and subsequent metamorphism. Buddington (1939, 1952) considers Grenville sedimentation as the earliest recognizable Precambrian event which was followed by magmatic intrusion of the anorthosite series, and later by metamorphism. DeWaard et al. propose Grenville sedimentation on an older, partly anorthosite terrain.

NOMENCLATURE OF THE ADIRONDACK META-ANORTHOSITE

The Adirondack meta-anorthosite is subdivided into three broad, general groups: 1) The Marcy anorthosite, consisting of massive, porphyroclastic, generally homogeneous meta-anorthosite, which is typical of the Mt. Marcy massif, 2) the Whiteface anorthosite, consisting of more heterogeneous, gneissic, and typically gabbroic variety, which is characteristic of the Whiteface Mountain massif, and, 3) the subordinate Keene Gneiss, a hybrid type of andesine-augen-mesoperthite gneiss intermediate between anorthosite and charnockite. These three types are gradational with each other and are not restricted to the areas from which their names are derived. Plate I shows their general distribution in the Lake Placid and Ausable Forks Quadrangles.

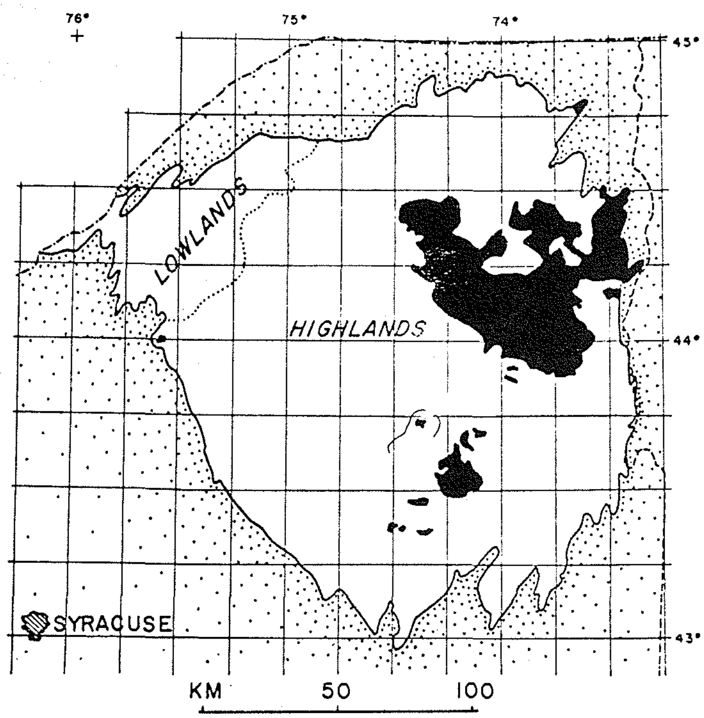


Fig. 1. Highlands - Lowlands division of the Adirondack Mountains, showing meta-anorthosite bodies in black. From deWaard, D. and W. D. Romey, 1963, Boundary relationships of the Snowy Mountain anorthosite in the Adirondack Mountains, Proc. Koninkl. Nederl. Akademie van Wetenschappen - Amsterdam, Ser. B, 66, No. 5

With increasing mafic mineral content, the Adirondack meta-anorthosites (and anorthosites in general) grade into meta-gabbros (clinopyroxene predominant) and meta-norites (orthopyroxene predominant). Intermediate varieties are thus classed as gabbroic or noritic meta-anorthosites, or anorthositic meta-gabbros or meta-norites. Charnockites, in which garnet and hypersthene are principle minerals, are also very common in the Adirondack region.

FIELD STUDY OF ADIRONDACK ANORTHOSITES

Anorthosites generally are very coarse grained. This is particularly true of the Marcy type, in which plagioclase crystals a foot or more in length are not uncommon. The normal procedure of field-and-thin section studies used with finer-grained rocks cannot be strictly applied to study of the Adirondack anorthosite, and some special methods have been devised. Representative modal analyses for instance, are obtained by point counting large areas on the outcrop with a transparent overlay net having evenly spaced points. Field differentiation of the meta-anorthosites without point counting is difficult, but a method used by several Adirondack workers is to differentiate on the basis of 1) percent mafic minerals (color index) and 2) degree of granulation. A third criterion which may be applied either on the outcrop or on the hand specimen in the laboratory is the K-feldspar content (Estimated from hand specimen stained with sodium cobaltinitrite)*.

*This staining technique was extensively used in the Santanoni quadrangle southwest of the Lake Placid quadrangle where Marcy type predominates. High K-feldspar contents were found only in intensely sheared zones, indicating probable metasomatism.

Buddington (1939, 1952)	Walton and De Waard (1963 a, b)
~ orogenic deformation and metamorphism	
intrusion of granites and alaskites	~ Grenville orogeny and metamorphism
~ orogenic deformation and metamorphism	
intrusion of hypersthene diabase	intrusion of olivine dolerite - gabbro
~ orogenic folding	
intrusion of quartz syenite and charnockite series	- deposition of supracrustal sequences
intrusion of olivine diabase or gabbro	
intrusion of anorthosite series, anorthosite and gabbroic anorthosite	erosion
Grenville series	~ pre-Grenville orogenic cycle includes:
	sedimentation, orogeny, metamorphism, and plutonism, forming an extensive granitic terrain, origin or emplacement of anorthosite-gabbro suite

Table 1. Geologic evolution of the Adirondacks, a correlation of two concepts. (de Waard, D. and M. Walton 1967, Precambrian geology of the Adirondack highlands, a reinterpretation, Geol. Rundschau 56 - 2, 596-629)

FIELD TRIP STOPS

NOTE: THE FOLLOWING NOTES ARE THOSE OF DR. PERCY CROSBY WHICH APPEARED IN THE GUIDEBOOK CONNECTED WITH THE GEORGE H. HUDSON SYMPOSIUM ON THE ORIGIN OF ANORTHOSITE, HELD IN PLATTSBURGH, OCTOBER 9, 1966.

INTRODUCTION

A sub-horizontal to gently northward-dipping sheet of meta-anorthosite and gabbroic meta-anorthosite 24 miles long and up to 12 miles wide lying north of the main High Peaks massif will be visited. It lies entirely within the Ausable Forks and Lake Placid quadrangles. The Jay sheet on the east forms most of the highest peaks of the Ausable Forks quadrangle and the Sentinel Range of the Lake Placid quadrangle. It is bordered on the northwest by the Whiteface sheet which occupies the Mt. Whiteface massif and much of the Stephenson and Wilmington Ranges to the north.

Massive porphyroclastic meta-anorthosite similar to that exposed in the main massif, often referred to in the literature as Marcy anorthosite, composes most of the Jay sheet whereas a typically gabbroic, gneissic border facies of the meta-anorthosite is particularly characteristic of the Whiteface sheet although it is widely distributed elsewhere near the boundaries of the Marcy anorthosite. It is a much more heterogeneous unit and locally contains substantial amounts of non-anorthositic material. It has been traditionally known as "Whiteface anorthosite." The distribution of the two types is shown on the Geologic Map, Plate 1, within the area we will visit, together with a hybrid rock, the Keene (andesine augen-mesoperthite) gneiss intermediate between the anorthosite and charnockite series.

Plate 2 is a northwest-southeast geological cross-section through the long dimension of the sheet which is connected near Lewis to the main anorthosite mass. It is interpreted by me as a meta-anorthosite nappe less than a mile thick rooted in the High Peaks meta-anorthosite. The Jay-Whiteface meta-anorthosite overlies a massive, inverted charnockitic section to the southwest. Similar, but less extensive, charnockites appear on the northeastern normal limb. Bordering the meta-anorthosite is a discontinuous metasedimentary shell which is best developed along the connecting neck to the southeast and also near the leading edge of the nappe within a major digitation east of Franklin Falls. For the sake of simplicity the charnockite- metasedimentary rocks have not been differentiated on Plates 1 and 2 where they are unlabelled.

ANORTHOSITE AND MESOPERTHITE GNEISS OF LAKE PLACID AND AU SABLE FORKS QUADRANGLES NEW YORK

GEOLOGY BY P. CROSBY

EXPLANATION

MASSIVE PORPHYROCLASTIC META-ANORTHOSITE (MARCY FACIES)

BORDER FACIES META-ANORTHOSITE AND GABBROIC META-ANORTHOSITE (WHITEFACE FACIES)
Locally includes charnockite and charnockite-anorthosite hybrid rocks in layers

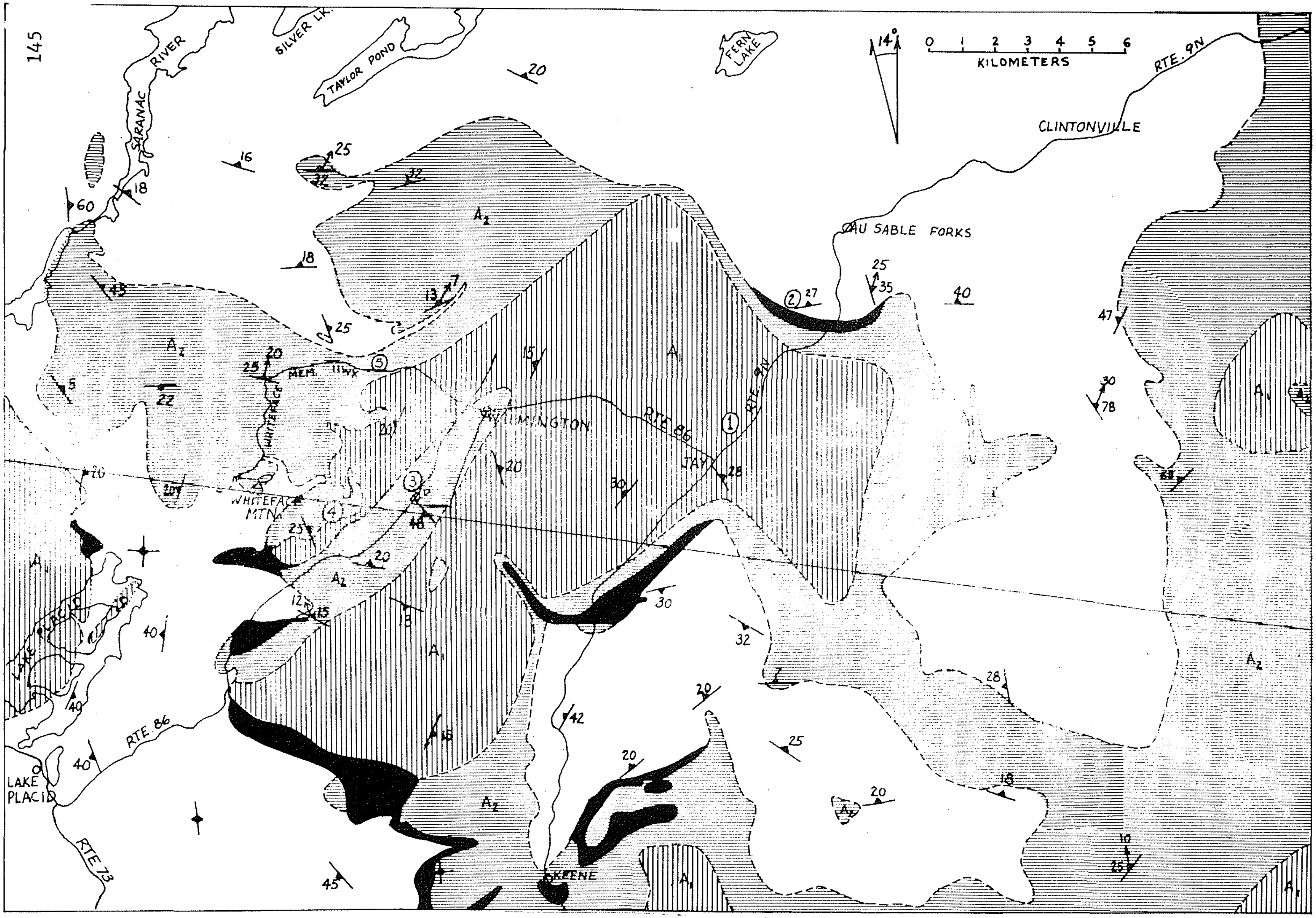
ANDESINE AUGEN MESOPERTHITE GNEISS (KEENE GNEISS)

Other lithologic units omitted for simplicity

Strike and dip of foliation with strike and plunge of lineation

Field trip stops

Plate I



SARANAC RIVER
SILVER LK.
TAYLOR POND
FERN LAKE
14°
0 1 2 3 4 5 6
KILOMETERS
RTE. 9N
CLINTONVILLE
AU SABLE FORKS
WILMINGTON
RTE 86
JAY
WHITEFACE MTN.
MEM. HWY.
LAKE PLACID
RTE 86
RTE 73
KEENE
A_1
A_2

Plate I

The nappe was presumably emplaced from the south with minimum apparent displacement 10 to 15 miles. Supporting this interpretation is a pervasive gently-plunging north-northeast mineral lineation which is particularly prominent in the strongly deformed gabbroic anorthosite gneisses on Whiteface Mt. near the base of the nappe. The nappe hypothesis is in general agreement with gravity data. Most of the Jay-Whiteface sheet lies outside the residual anomaly contour corresponding to a 3Km thickness for the anorthosite.

Stop 1. Locality A, Jay, Ausable Forks Quadrangle

Take woods road beginning opposite Tirolerland restaurant, 0.8 miles N. of Jay on Route 9N. Follow woods road to point where it begins to veer sharply south, 0.25 miles from Route 9N. Climb 100 yards south to exposed slabs on hilltop which has been stripped and cleared by Rock of Ages Corp. Professor Brewster Baldwin of Middlebury College prepared base map by plane-tabling on which geology and sample localities have been plotted (Fig. 2).

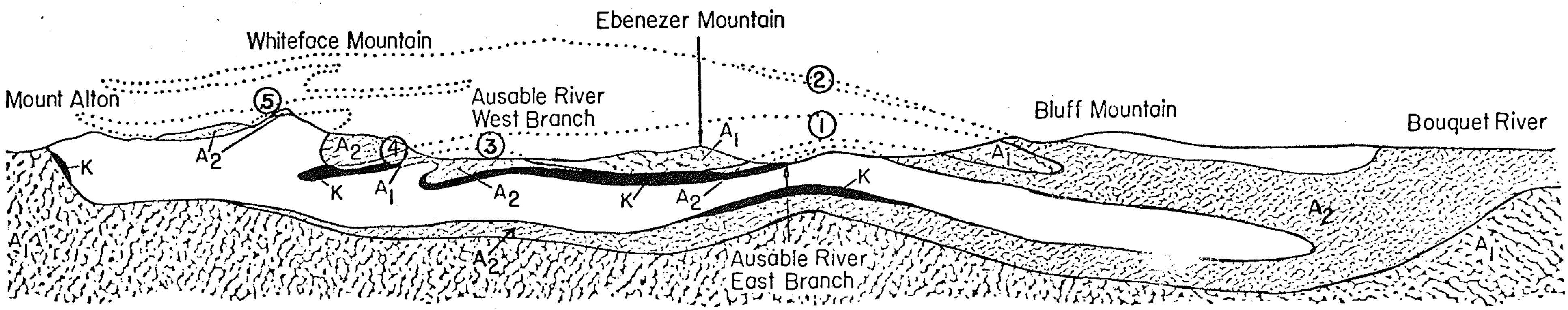
The exposure is representative of massive, generally porphyroclastic, core meta-anorthosite of the "Marcy-type," but shows the considerable local compositional and textural variation encountered in this map-unit. Three local facies have been distinguished here, viz:

Facies I: A fine-to-medium grained granoblastic equigranular gabbroic meta-anorthosite, locally passing into an anorthositic metagabbro, largely confined to the northwestern portion of the outcrop. It shows many analogies with the border facies anorthosite. (Whiteface-type") and perhaps should be so mapped. Clinopyroxene ($Wo_{40}En_{35}Fs_{25}$) takes place of hypersthene in this facies, and garnet is absent.

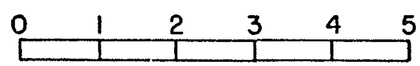
Facies II: This is the predominant lithic type in the exposure and consists of inequigranular meta-anorthosite with plagioclase megacrysts forming from about 25 to 70 per cent of the total rock volume, the proportion of megacrysts decreasing toward the southeast. Hypersthene is the major ferromagnesian constituent, but locally incipient rims of garnet and clinopyroxene are visible. Before concluding that the finer-grained plagioclase matrix between the megacrysts is of protoclastic origin or produced by post-crystallization deformation, examine the hypersthene grains intergrown with the matrix. Many will be seen to be optically continuous as shown by reflections from parallel cleavage surfaces and may indicate an ophitic intergrowth suggestive of a final, possibly eutectic, assemblage.

Plate 2 CROSS-SECTION of JAY WHITEFACE ANORTHOSITE NAPPE

Ausable Forks–Lake Placid New York Quadrangles



SCALE IN MILES



EXPLANATION

A1 — Massive porphyroclastic meta-anorthosite. ("Marcy" facies)

A2 — Border facies meta-anorthosite and gabbroic meta-anorthosite ("Whiteface" facies). Locally includes charnockitic and charnockite anorthosite hybrid rocks in layers

K — Andesine augen mesoperthite gneiss ("Keene" gneiss)

② Approximate location of field trip stops (along structural plunge)

TABLE 2
MODES, LOCALITY A, JAY

	FACIES I			FACIES II								
Mineral	1	3	4	2	5	6	8	9	10	11	12	13
Plagioclase (matrix)	81.7	75.0	61.4	50.4	51.2	51.4	31.6	27.4	24.1	30.4	36.4	68.0
Plagioclase (megacrysts)				44.9	46.2	45.2	59.3	68.8	70.7	64.4	57.5	27.1
Clinopyroxene	18.3*	25.0	38.6									
Hypersthene				4.7	2.6	3.4	9.1	3.8	5.2	5.2	6.0	4.9[1]
Garnet				tr.		tr.	tr.		tr.		tr.	tr.
Quartz												
Magnetite						tr.					0.1	
Pyrite		tr.										
Total Plag.	81.7	75.0	61.4	95.3	97.4	96.6	90.9	96.2	94.8	94.8	93.9	95.1
Color Index	18.3	25.0	38.6	4.7	2.6	3.4	9.1	3.8	5.2	5.2	6.1	4.9
Per cent An											56	
Matrix	72						44[2]					
Megacrysts							44[2]					

*Composition approx. $Wo_{40}En_{35}Fs_{25}$
1 also includes probably clinopyroxene
2 from excavated block 25 ft. SSE
(south southeast)

TABLE 2 (continued)
MODES, LOCALITY A, JAY

Mineral	FACIES II 14	15	16	17	18	19	FACIES III 7	20	Mean Facies I N=3	Mean Facies II N=15	Mean Facies III N=2	Mean Facies N=20
Plagioclase (matrix)	67.9	40.0	29.2	31.8	48.9	60.9	8.3	15.7	72.7	43.3	12.0	44.6
Plagioclase (megacrysts)	27.8	55.0	62.8	65.9	44.7	35.1	66.9	74.8		51.6	70.8	45.8
Clinopyroxene	tr.		tr.						27.3			4.1
Hypersthene	4.3	5.0	8.0	2.3	6.4	3.8	24.8	8.8		4.8	16.8	4.6
Garnet		tr.	tr.	tr.	tr.	tr.	tr.	0.7			0.4	0.1
Quartz						0.2						
Magnetite					tr.		tr.			other 0.3 (cpx, gar, mag, qtz)	tr.	other 0.8 (mag, Py, qtz)
Pyrite												
Total plag.	95.7	95.0	92.0	97.7	93.6	96.0	75.2	90.5	72.7	94.9	82.8	90.4
Color Index	4.3	5.0	8.0	2.3	6.4	3.8	24.8	9.5	27.3	5.1	17.2	9.6
Percent An	56											
Matrix			48						72	52		57
Megacrysts							50			46	50	47
All plag.										49.6		52.9

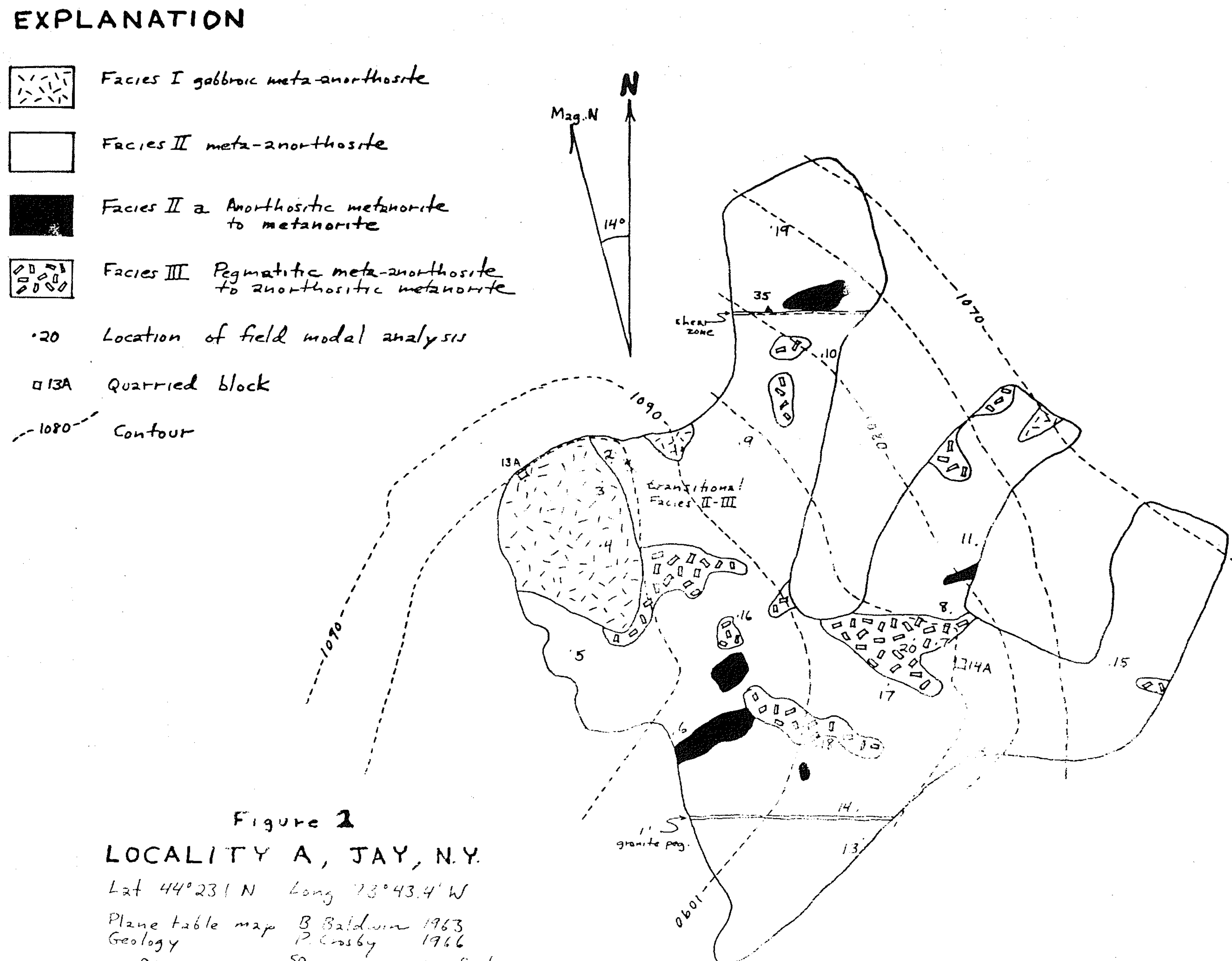


Figure 2

LOCALITY A, JAY, N.Y.

Lat 44°23.1' N Long 73°43.4' W

Plane table map B. Baldwin 1963
Geology P. Crosby 1966

0 50 100 feet
SCALE

Contour interval 5 feet

Facies III: Appearing as irregular patches up to 50 feet across in facies II meta-anorthosite is a pegmatitic phase, varying from meta-anorthosite to anorthositic metanorite. A coarse ophitic intergrowth of little-granulated plagioclase crystals and hypersthene, both attaining lengths up to one foot, is characteristic of facies. Garnet rims may be seen around some of the hypersthene.

In Table 2 are given modes measured on the outcrop (numbers painted on outcrop and located on Fig. 2) covering 6.5 feet^2 apiece (619 cm^2) with reticular spacing of one inch (2.54 cm.). Means for each of the facies and a mean for all modes, which have been apportioned among the various facies proportional to their aerial extent, have been calculated. The anorthite content of a limited number of plagioclase samples, both megacrysts and the finer grained matrix, has been determined by refractive indices. The compositional range is from calcic andesine (An_{44}) to sodic bytownite (An_{72}) with the average of all determinations An_{53}. Where both matrix and megacrysts were measured in a single sample, the composition was identical in each. It is likely, however, that many of the megacrysts show compositional zoning. This is indicated by one crystal having a chatoyant core and a non-chatoyant rim.

Stop 2. Omitted.

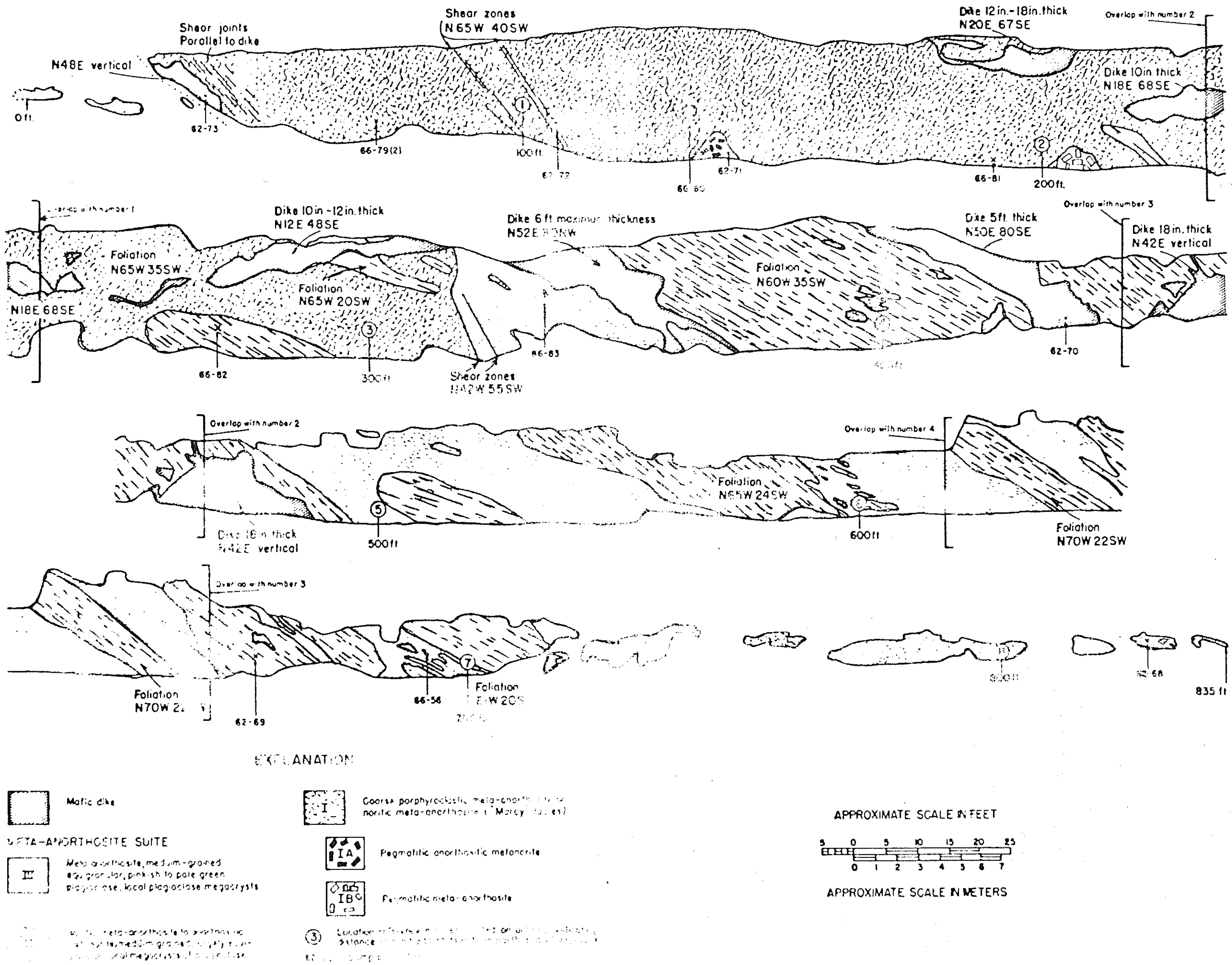
Stop 3. Roadcut South of the Flume, Route 86 Between Wilmington and Lake Placid

In this extensive exposure (for cross-section 835 feet long prepared from a photomosaic refer to Plate 3), the relationships between three distinctive anorthositic varieties are well displayed. Coarse-grained porphyroclastic meta-anorthosite to noritic meta-anorthosite (Facies I) with blue-black plagioclase megacrysts composing from 10-40% of the total rock flanked by a medium-grained, generally equigranular noritic meta-anorthosite to anorthositic meta-norite gneiss (Facies II) and a medium-grained, granoblastic anorthosite with mottled pale green and pink to lavender plagioclase (Facies III). For petrographic details, see Table 3. Despite the considerable compositional & textural variation, the anorthosite content of plagioclase remains remarkably constant (ca. An_{55}) with the principle difference being the Fe-Ti oxide inclusions of the dark plagioclase megacrysts and the absence of impurities and characteristic coloration of the (recrystallized?) plagioclase of Facies III.

Plate 3

ROADCUT SOUTH of THE FLUME, ROUTE 86, WILMINGTON — LAKE PLACID, NEW YORK

Trend of roadcut – N30E; looking S60E



Facies I meta-anorthosite closely resembles massive core meta-anorthosite of the Marcy-type although here it appears to be isolated from any extensive exposed body of such meta-anorthosite. Facies II is most closely related to border facies (Whiteface-type) meta-anorthosite and Facies III which forms most of the gorge (Flume) along the Ausable River just to the north is found interlayered with noritic meta-anorthosite at numerous localities in the Whiteface sheet to the west.

The meta-anorthosite facies are numbered in the order of their presumed relative ages as shown at this exposure. Block structure with inclusions of Facies II noritic meta-anorthosite gneiss in the much more leucocratic Facies III is especially well developed 600 feet south of the northern end of the cut. The position of Facies I meta-anorthosite in the sequence is much less certain, and it is placed first largely because of the inclusion within Facies II near the 400 feet marker. Large masses of Facies II, however, are surrounded by Facies I between the 200 and 300 feet markers. There is thus some evidence of reciprocal intrusion if geometric factors are ignored. Another example is patches of Facies III meta-anorthosite surrounded by Facies II at 400 feet. Some discussion of field criteria for age relationships will undoubtedly be warranted. In general elsewhere in the eastern Adirondacks as described in the literature, it is more common to find gabbroic meta-anorthosite intruding and enveloping blocks of a more leucocratic meta-anorthosite. At the next stop we shall examine further evidence bearing upon the anorthosite sequence.

There are small inclusions of diopsidic or garnet-biotite rich field spathic gneisses or granofelses in Facies III meta-anorthosite at 690 feet (specimen 66-56). More extensive layers of diopsidic meta-sediments in Facies II gabbroic meta-anorthosite are to be found 100 yards west of the bridge over the Flume just to the north.

The outcrop shows a number of fine-grained diabase dikes from 10 inches to 6 feet thick nearly parallel to the general trend of the cut (estimated mode, samples 62-70, Table 3).

Many shear surfaces of indeterminate displacement are present, the most prominent set perhaps being about N30E, parallel to the cut face and on strike with the pronounced linear through Wilmington Notch visible to the south, a locus of probably major faulting. Augen gneiss zones in the meta-anorthosite with considerable microclinization may be seen at 100 feet and at 320 feet (estimated mode, sample 62-72, Table 3). Garnet in the meta-anorthosite appears to be localized in these gneiss zones, a phenomenon observed at many other exposures. Many secondary minerals occur along the late shear surfaces, including calcite, quartz, epidote, and chlorite. Natrolite has also been identified.

TABLE 3, MODES VICINITY OF THE FLUME, AUSABLE R., W. BRANCH

	FACIES I				FACIES II		FACIES III			Micro clinized gneiss zone	Diabase dike
Mineral	66-79	66-80	66-81	62-71	62-69	62-79[3]	62-68	62-74	62-81[4]	62-72*	62-70*
Plagioclase[1]	Plagioclase determination only	Plagioclase determination only	Plagioclase determination only	Orthopyroxene determination only	77.6	85.9	92.4	96.6	93.4	30	54
Hypersthene					4.9	3.2					
Clinopyroxene					11.8	2.7		0.3	3.0	4	35
Hornblende					5.6	6.5		1.8	0.3		4
Chlorite						1.6	4.4	0.7	2.4	2	1
Garnet										<1	
Quartz										2(?)	
Microperthite										60	
Biotite											tr.
Calcite											1
Scapolite							3.1		0.5	tr.	
Sphene								0.1	0.1	<1	
Apatite						tr.		0.1		tr.	
Magnetite								0.4	0.4	<1	5
Ilmenite					0.1		tr.				
Pyrite[2]						tr.	tr.				
Color Index					22.4	14.1	4.5	3.2	6.0		
An plag. (megacrysts	56	56	53								
An plag. (matrix)	53	53	53		54	55	53	55	48	57	60
En hypersthene				65							

FOOTNOTES:

[1] includes Kaolinite alt'n

[2] includes hematite alt'n

[3] 300 ft. upstream from bridge along Ausable River

[4] Lower end of the Flume 0.3 mi. NE of Rte. 86 bridge

*estimated mode

Stop 4. Stream-worn slabs under lower ski lift, Whiteface Mt., elevation 1800 - 2000 feet between powerhouse and mid-lift station.

Drive into Whiteface Ski Center from Route 86 between Wilmington and Lake Placid. The lift will probably not be operating, so climb ski trail 600 feet vertically right (north) of stream along lift line to bare stream slabs above powerhouse.

Three distinct facies of the meta-anorthosite series are represented at this exposure, two of them related to those seen at the Flume. The earliest member of the series (Facies I) is not, and is preserved as several rounded inclusions mostly under running water in the stream bed as it veers north of the lift line towards the upper end of the exposure. It is a fine-to-medium-grained equigranular mafic anorthositic meta-gabbro enclosed by both Facies II and Facies III rocks. For thin-section mode see specimen 65-175 in Table 5. Notable is the calcic content of the plagioclase (An_{70}) compared to that of Facies II (An_{52-55}).

Facies II comprises most of the outcrop. For thin-section modes, see Table 5; for larger-area point counts with the screen employed at Stop 1, see Table 4 -- the numbers are painted on the outcrop. The average rock is a noritic-meta-anorthosite (color index 17.85) with mean per cent plagioclase megacrysts 24.25 and plagioclase matrix 57.80. The proportion of megacrysts varies from a minimum of 7.6% to a maximum of 42%. There is also considerable textural variation with a medium-grained rock showing subtle compositional banding and a suggestion of sub-ophitic texture in the matrix the preponderant type (modes, no. 1-10). Mineralogically similar, but coarser-grained and showing better developed ophitic texture is mode no. 11, whereas modes no. 13 and 14 are somewhat more leucocratic than the average but otherwise related texturally.

Facies III is an inequigranular medium-to-coarse-grained leucocratic meta-anorthosite apparently cutting Facies II noritic meta-anorthosite in irregular dikes with apophyses. It is interpreted as definitely younger than Facies II, and bears out the sequence deduced at the Flume. Similar dike-like bodies of Facies III in Facies II meta-anorthosite are present in slabs along the Boreen ski trail 100 feet higher to the northwest, and may be visited if time permits.

It is worthwhile inspecting some of the coarser-grained ophitic Facies II noritic meta-anorthosite in the lower narrow portion of the stream slabs for textural evidence of protoclasis or several stages of granulation followed by crystallization. Representative of the textures seen are large, partially granulated, plagioclase megacrysts

TABLE 4
MODES, LOWER WHITEFACE MTN. (SKI LIFT)

Mineral	FACIES II 1	2	3	4	5	6	7	8	9	10	11	13	14	Mean Facies II, N=13	Facies III 12
Plagioclase (matrix)	48.2	53.6	58.0	54.5	56.2	63.8	57.4	64.4	78.9	67.4	43.9	56.4	48.7	57.80	80.5
Plagioclase (megacrysts)	29.3	25.5	21.8	25.2	22.4	19.0	25.3	17.3	7.6	10.7	36.4	33.1	41.6	24.25	17.0
Hypersthene	20.8	19.2	19.0	18.8	18.0	15.6	16.4	16.7	11.9	17.5	16.4	9.8	8.2	16.02	1.5
Pyribole	1.7	1.7	1.2	1.5	3.4	1.6	0.9	1.6	1.6	4.4	3.3				1.0*
Garnet												0.7	1.5	1.93	
Magnetite															
Total plag.	77.5	79.1	79.8	79.7	78.6	82.8	82.7	81.7	86.5	78.1	80.3	89.5	90.3	82.05	97.5
Color Index	22.5	20.9	20.2	20.3	21.4	17.2	17.3	18.3	13.5	21.9	19.7	10.5	9.7	17.95	2.5

* includes magnetite

TABLE 5
THIN SECTION MODES
LOWER WHITEFACE MTN. (SKI LIFT)

Mineral	FACIES I	FACIES II	
	65-175	63-62	65-177
Plagioclase	66.5	79.7	88.1
Hypersthene	7.8	3.2	
Clinopyroxene		13.0	6.0
Hornblende	20.8	3.6	4.0
Biotite	4.4		0.1
Chlorite			1.4
Apatite		0.2	0.2
Magnetite		0.1	0.2
Ilmenite		0.1	
Pyrite	0.5	0.1	
Color Index	33.5	20.1	11.7
An Plag. (matrix)	70	52	55

with inserted ungranulated hypersthene crystals and shattered plagioclase augen veined by noritic meta-anorthosite. Relationships may be explained by varying paths of crystallization in Di-Ab-An ternary coupled with one or more stages of cataclasis. Locally there appear to be eutectic intergrowths of plagioclase and hypersthene between plagioclase megacrysts.

Observations bearing upon the metamorphic history of the meta-anorthosite at this locality are the following:

1. Garnet is generally absent except along shears. At the slabs in the Boreen ski trail a little higher, leucocratic meta-anorthosite "dikes" in gabbroic meta-anorthosite have garnet and clinopyroxene + hornblende rims to hypersthene grains whereas hypersthene does not have such reaction rims in the gabbroic meta-anorthosite.
2. Hypersthene generally is rimmed by clinopyroxene or hornblende or an intergrowth of the two in the noritic meta-anorthosite under the lift. Locally, a hypersthene core is rimmed by clinopyroxene and this in turn by hornblende. It is possible that the clinopyroxene rim represents exsolution from a primary orthopyroxene or inverted pigeonite host. No exsolution lamellae have, however, yet been detected in the hypersthene. If the clinopyroxene is a result of unmixing, then the only metamorphic mineral here is hornblende.

<u>Stop 5 (6, 7...)</u> Whiteface Memorial Highway and summit of Whiteface Mountain.

If time permits, we will ride up the Whiteface Memorial Highway, making one or more stops, and take one of the trails to the summit, preferably the northeast ridge trail from the "Wilmington turn" at the large cut below the main parking lot.

The meta-anorthosite rocks exposed in the Whiteface sheet extending from the upper slopes of Whiteface Mt. on the south to Catamount Mt. on the north and the Franklin Falls area on the west show many mineralogical, textural, and structural contrasts with the border facies of the meta-anorthosite we have thus far examined.

In general, they are much more recrystallized and relict igneous textures as observed under the ski lift at Stop 4 are rare. Foliation, and often lineation as well, is strongly developed and plagioclase megacrysts, mostly small and augen-shaped, seldom constitute more than a few per cent of the total rock. Banding is also common, and gabbroic meta-anorthosite alternates with leucocratic meta-anorthosite layers over short intervals. Garnet becomes a major phase and hypersthene a correspondingly less important constituent. There are layers of varying thickness from less than an inch to many tens of feet of mangeritic rocks, of Keene gneiss and of metasediments including coarse-grained gray quartzite and garnet-diopside granofels. Graphite is present locally near marble or calc-silicate contacts. I believe many of the physical and mineralogical characteristics of the Whiteface meta-anorthosite sheet can be explained by its tectonic position at the leading edge and base of the nappe described in the introduction to these notes.

Table 6 presents 20 modes of anorthosite rocks on Whiteface Mt. along the Memorial Highway and summit ridges in order of increasing elevation to emphasize any depth-controlled trends that may be present in the sheet. As we climb the northern slope of the Whiteface massif we are slowly passing upward through a nearly dipslope compositional layering parallel to a regional foliation.

At the Lake Placid turn (First switchback) there is a spectacular view of Lake Placid to the south. The contact between gabbroic meta-anorthosite, here somewhat more mafic than average, and mangeritic rocks is exposed just to the east of the turn. At the contact the mangerites are greenish but further away they become progressively pinker and quartz content increases. Boudins of anorthosite are enclosed in the mangerite-charnockite sequence -- is this an intrusive or tectonic relationship?

At the Wilmington turn, gabbroic meta-anorthosite structurally overlies green mangerite at the eastern end of the large cut. Meta-anorthosite blocks in the retaining wall excavated from the cut provide a good sample of the average rock in the cut. Garnet coronas are fairly common. On the upper, slightly weathered surfaces of the cut strong gently northeast-plunging lineation shown by aligned mafic minerals is shown to advantage. There are also thin quartzitic beds parallel to the foliation in the meta-anorthosite. One meta-anorthosite block not quite in place shows sharp folding of a quartzite layer.

Continuing up the trail along the northeast ridge of Whiteface Mt. to the summit, the open slabs will show the strongly gneissic, lineated, and banded character of the gabbroic meta-anorthosite. Thin plate-like quartzite layers parallel to foliation are common.

TABLE 6
MODES OF ANORTHOSITIC ROCKS ALONG WHITEFACE MEMORAL HIGHWAY
TO SUMMIT RIDGES OF WHITEFACE MT. IN ORDER OF INCREASING ELEVATION

Mineral	65-88	65-89	65-90	65-91	65-92	65-93	65-100	65-101	65-103	65-204
Plagioclase	97.9	52.7	76.0	90.2	58.5	74.2	86.1	86.2	79.1	16.9
Clinopyroxene		9.6	16.5	3.6	11.9	6.2	8.2	3.7	12.0	11.1
Hornblende		26.7	6.8	5.8	21.6		5.3	8.7	5.3	4.6
Hypersthene		5.8			7.9			0.8		
Garnet	0.1	2.9				7.6			0.3	56.0
Biotite	0.1	0.4	0.2						1.5	0.3
Chlorite	0.8							0.5	1.7	2.0
Scapolite										
Sphene	0.2	1.8	0.1			2.8	tr.	tr.	tr.	0.1
Apatite	0.1		0.2	tr.	0.1	1.1	0.4			
Epidote										2.4
Calcite	0.5									
Magnetite		0.1	0.2			8.1		tr.	tr.	6.6
Ilmenite	0.2			0.3				tr.		
Pyrite	0.1			tr.						
Color Index	1.3									
An plag.	68	70	60	70	68	56	52	43	54	66
ELEV.	2710'	2710'	2850'	2877'	2930'	2960'	3980'	3980'	4080'	4150'

TABLE 6 (continued)
MODES OF ANORTHOSITIC ROCKS ALONG WHITEFACE MEMORAL HIGHWAY
TO SUMMIT RIDGES OF WHITEFACE MT. IN ORDER OF INCREASING ELEVATION

Mineral	65-109	65-111	65-113	65-118	65-119	65-121	65-126	65-122	WS-30a	65-125
Plagioclase	83.9	41.3	84.1	57.2	79.9	60.1	89.0	29.6	18.1	88.0
Clinopyroxene	8.0	8.8	5.1	4.1	6.9	24.1	4.9	40.6	27.2	6.6
Hornblende	7.0	26.4	10.1	1.2	11.0	4.5	5.5	29.1	10.6	4.9
Hypersthene				1.7		6.4				
Garnet		17.2		34.3					37.1	
Biotite	0.3		0.3				0.3			
Chlorite		0.4			1.6					
Scapolite					tr.					
Sphene	0.4	0.3			0.2	0.3		0.4	0.1	
Apatite	0.1	0.9	0.1	0.3	tr.	0.3	0.2	0.2	1.0	0.1
Epidote										
Calcite										
Magnetite	0.3	4.7	0.1	1.2		4.3	0.1	0.1	5.2	0.3
Ilmenite										
Pyrite			0.2						0.7	0.1
Color Index										
An plag.	45	50	52	52	60	35	49	63	65	58
	4230'	4310'	4440'	4510'	4590'	4630'	4640'	4680'	4755'	Summit 4860'

From the summit, if you descend the south face 100 vertical feet you will cross the meta-anorthosite-mangerite contact once more. There is no intermediate andesine augen mesoperthite gneiss (Keene gneiss) at the boundary. Quartz content varies widely from 30-40 per cent to less than 5% in adjacent layers.

If the day is fine, it will be possible to point out regional relationships between anorthosite and related rocks in the High Peaks to the south and in the Jay-Whiteface sheet to the southeast and northwest.

Trip I

DEGLACIAL HISTORY OF THE LAKE CHAMPLAIN-LAKE GEORGE LOWLAND

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GENERAL GEOMORPHOLOGY

The Lake Champlain and Lake George Valleys are two separate physiographic units, situated in separate physiographic provinces (see Broughton, et al., 1966). The Champlain Valley lies in the St. Lawrence-Champlain Lowlands while the Lake George trough lies in the Adirondack Highlands. It is the continuous nature of the deglacial history of the two regions that links them together for this field trip. Indeed, the deglacial history of the regions begins at the north end of the Hudson-Mohawk Lowlands near Glens Falls, New York.

Physiography

The Champlain Valley lies between the Adirondack Mountains of New York on the west and the Green Mountains of Vermont on the east. Lake Champlain occupies most of the valley bottom and has a surface elevation of about 95 feet. The Great Chazy, Saranac, Ausable, and Bouquet Rivers enter the valley from the Adirondacks while the Mississquoi, Lamoille, and Winooski Rivers and Otter Creek enter from the Green Mountains. Figure 1 shows the general physiography and the field trip stops.

According to Newland and Vaughan (1942) Lake George occupies a graben in the eastern Adirondacks. The surface of Lake George is about 320 feet above sea level and it drains into Lake Champlain via Ticonderoga Creek at its northern end. In general, Lake George marks the divide between southerly flowing Hudson River drainage and northerly flowing St. Lawrence River drainage.

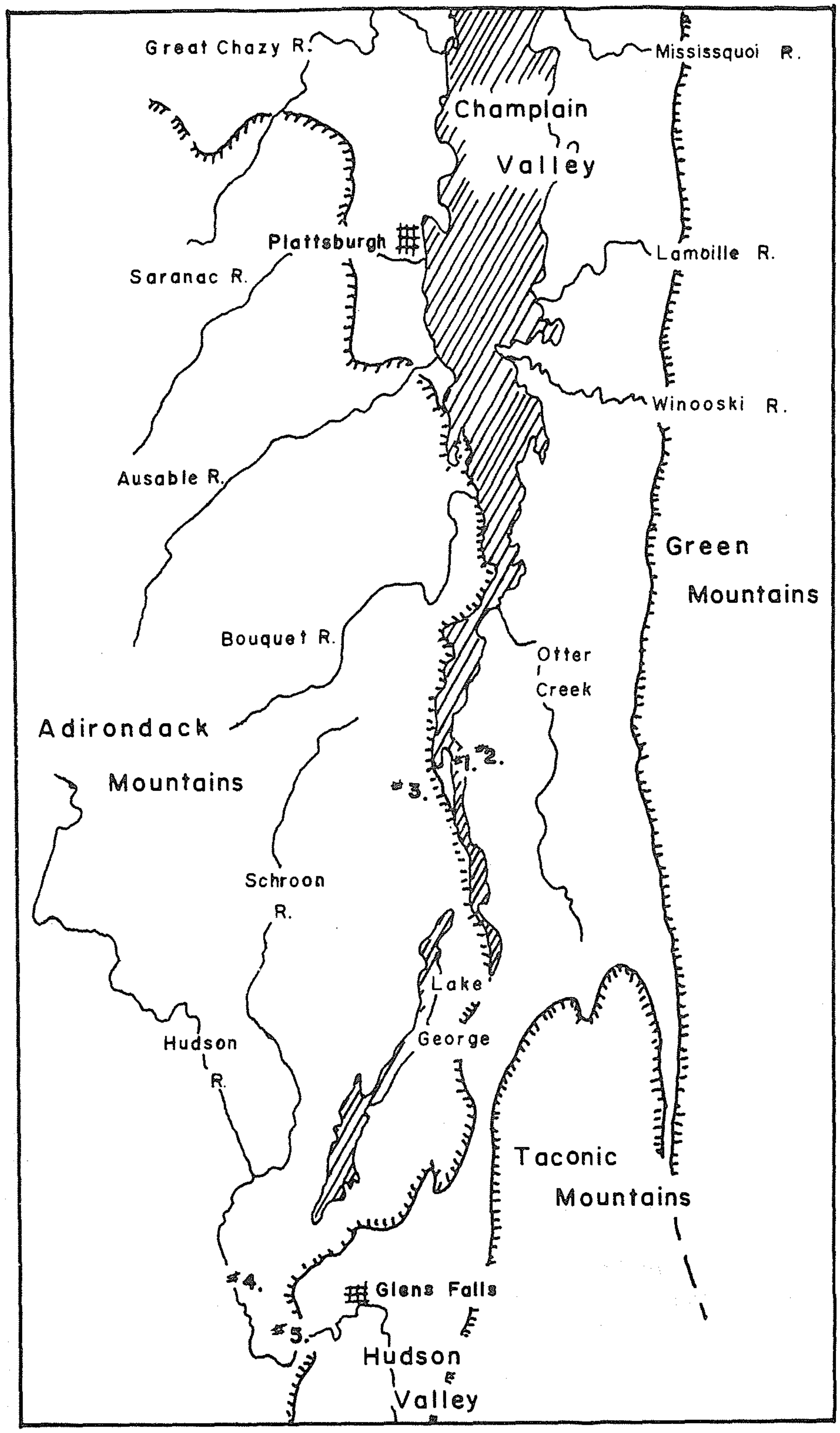

Figure 1. General physiography of the Lake Champlain-Lake George Lowland and field trip stops.

South of Lake George, the northern Hudson Valley lies between the Adirondack Mountains on the west and the Taconic Mountains on the east. This part of the valley is a lacustrine plain that varies in elevation from 180 to about 520 feet. The upper Hudson River and its chief tributary, the Schroon River, drain the Adirondack Highlands. The Hudson enters the Hudson Valley lowland via a gorge in the Luzerne Mountain range west of Glens Falls. The river descends approximately 175 feet over a series of waterfalls from Glens Falls to Fort Edward.

Bedrock

The Adirondack Highlands contain the highest mountains in New York State, in the High Peaks region of western Essex County. The High Peaks are underlain by Precambrian anorthosite that is highly resistant to erosion. Other igneous and metasedimentary rocks of lesser resistance are found in other areas of the Adirondacks. Newland and Vaughan describe the Lake George basin as a depressed fault block surrounded by Precambrian, granitic highlands. The Green Mountains, east of the Champlain Valley are also composed of crystalline, igneous and metamorphic rocks.

The Champlain lowland, and to a large extent the northern Hudson Valley is underlain by Cambrian and Ordovician sandstones, dolostones, and limestones. These rocks appear to dip gently away from the Adirondacks although they are structurally more complex on the east (Doll, et al., 1961). Although Ordovician shales predominate throughout much of the Hudson Valley they are subordinate to carbonates in the north.

GLACIAL DRIFT

The bedrock of the Lake Champlain-Lake George lowland is mantled by a variable thickness of glacial drift. These deposits consist of till, ice-contact stratified drift, outwash, and lacustrine and marine sediments.

Till

Two general types of till are found in and adjacent to the lowlands. Connally (1968a) has referred to the two types geographically as mountain till and valley till while Stewart (1961) has used the genetic designations ablation till and lodgement till. Although it is possible that these tills are products of separate glaciations, or separate processes, they are treated here as related magnafacies.

Mountain till is sandy to gravelly and exhibits limited size-sorting. It is variable in both texture and color however, usually it has a sandy-loam to loamy-sand matrix and is dark-yellowish-orange (10 YR 6/6) in color. This till is the dominant deposit in both the Adirondack and Green Mountains but is found in the northern Hudson Valley as well. On mountain tops and flanks the deposit is probably relatively thin but in intermontane valleys, thicknesses up to 100 feet may be encountered.

In general, mountain till is quite friable and is composed of quartz and feldspar grains derived from the crystalline rocks of the highlands. In the Lake George basin however, the till is quite tough and compact. Although no formal stop is planned during the trip a short stop will be made at an exposure found suitable at the time of the field trip.

Valley till is typical of the Champlain and Hudson Valleys (and the Taconic Mountains in addition). This type of till is very firm, usually very pebbley, and contains abundant carbonate in the valleys. The texture of the matrix is a loam to clay-loam in the Champlain Valley and clay-loam to silty-clay near Glens Falls. In the subsurface this till is usually light-olive-gray to medium-gray.

Three distinct valley tills have been described from both the Vermont (Connally, 1967a) and New York (Connally, 1968a) sides of Lake Champlain. One till is stoney, calcareous, and dark-gray (N 3) with a clay-loam matrix, or sometimes gray-black (N 2). A second valley till is stoney to bouldery, calcareous, and light-olive-gray (5 Y 5/2) with a loam to sandy-loam matrix. A complete weathering profile is necessary to distinguish this till from the dark-gray till because the dark-gray color oxidizes to light-olive-gray near the surface. The third valley till is dark- to moderate-yellowish-brown (10 YR 4/4) with a silt-loam matrix and very few till stones. This silt-rich till is found only in the Ticonderoga quadrangle, both in New York and Vermont, and probably represents the incorporation of lacustrine silty-clays during the Bridport readvance.

Multiple-Till Sections

Two sections will be visited during this trip. The first is the West Bridport section (Stop 1) located south of the abandoned ferry crossing on Lake Champlain, just west of West Bridport, Vermont. In 1964 the exposure showed:

0-2'	silty-clay containing ice-rafted(?) pebbles and boulders.
16'	laminated to thin-bedded lacustrine silt and sand.
5½'	dark-gray (N 3), clay-loam till divisable into three units: a lower 12-18", gray-black (N 2) till; a medial 12-18" oxidized gravel; and an upper, 4' dark-gray till.
3'	light-olive-gray (5 Y 5/2), calcareous, sandy-loam till.
	bedrock with striae oriented N 10°E.

The second section is the Luzerne Mt. section (Stop 5) located in the Hudson gorge west of Glens Falls, New York. In 1968 this section exhibited the following sequence:

4-10'	moderate-olive-gray (5 Y 4/2) till and colluvium overlain by spoil and vegetation.
5-12'	moderate-olive-gray till, very compact, very bouldery, with a sandy-loam matrix and many limestone and shale clasts.
4'	thinly-laminated to thin-bedded sand.
1-17'	gray-black (N 2), bouldery, silty-clay till containing clasts of dark-gray, contorted lacustrine sediment.

Another section close by and to the east shows:

6'	gray-black till, as above.
10'	dark-gray, laminated and rhythmically bedded lacustrine clay, silt, and fine sand, greatly contorted.

A third section a few hundred yards east shows:

4-20'	light-brown till(?), very sandy and choked with angular boulders, showing crude stratification.
2-3'	gray-black till as above.
20'	oxidized, pebbley sand.
20'	spoil.

These sections represent the type section for the Luzerne readvance, as demonstrated by the uppermost, moderate-olive-gray till.

Stratified Drift

As the glacier stagnated and melted upland areas soon emerged from beneath the ice cover leaving stagnant masses of ice in the lowlands. Water flowed into and over these stagnant masses and deposited material in contact with the melting ice. These ice-contact deposits are quite common in and adjacent to the Adirondack and Taconic Mountains. The deposits may vary in thickness from a few tens of feet to hundreds.

Outwash is waterlaid sediment deposited beyond the margin of the ice and it can frequently be traced back upstream to related ice-contact deposits. The outwash may be better sorted than the related ice-contact drift and may contain kettle holes where blocks of ice have been buried by, or deposited with, the outwash. Many sequences of outwash and ice-contact deposits have been described by Connally (1965) adjacent to the Green Mountains where they have been used to infer the retreat of an active ice margin.

Stop 3 will visit an outwash delta that was deposited into Lake Quaker Springs. This massive outwash body, containing kettle holes up to 120' deep, is located just west of the village of Streetroad, New York between Miller and Buck Mountains.

Lake and Marine Deposits

As the margin of the last glacier retreated northward, up first the Hudson and then the Champlain Valley, proglacial lakes were impounded. When the ice retreated north of the St. Lawrence Valley marine waters invaded and the Champlain Sea came into existence. Lacustrine and marine deposits can not be distinguished lithologically so all deposits are discussed as "lacustrine".

Beach ridges and hanging deltas represent former shorelines while thick, rhythmically bedded (varved?) clay deposits are the result of quiet water sedimentation. Reconstruction of former lakes has been discussed by Chadwick (1928), Chapman (1942), and LaFleur (1965).

GLACIATION

Very little is known about the glaciation of the Champlain lowland except what can be deciphered from striae. From striae and till fabric studies Stewart (1961) and Stewart and MacClintock (1964, 1967) have inferred three glaciations in Vermont. The latest (Burlington) glacier advanced from the northwest while the earlier (Shelburne) glacier advanced from the northeast. The earliest (Bennington) glacier also advanced from the northwest. However, in the Adirondacks only northeasterly striae appear to have been recorded. Both Denny (1966) and Connally (1967a) have suggested that the original advance was from the northeast, but that the most recent ice in the valley advanced in lobate form forming northeasterly striae west of the valley and northwesterly striae east of the valley. It is not clear whether patterns document separate glaciations or merely advancing and receding phases of the last Wisconsinan advance.

DEGLACIAL HISTORY

To trace the deglacial history of the Lake Champlain-Lake George lowland it is necessary to understand the complex interrelationships between moraines, readvances and proglacial lakes. Chapman (1942) has summarized proglacial lake history for Vermont and LaFleur (1965) for the Hudson Valley. This report attempts to relate these two studies to the readvances described in the Hudson (Connally, 1968a; 1968b) and Champlain (Connally, 1967a) Valleys.

Three lake levels in the Hudson Valley have been referred to collectively as "Lake Albany" and three lake levels in the Champlain Valley have been referred to collectively as "Lake Vermont". The highest level of "Lake Albany" is confined to the Hudson Valley while the lowest level of "Lake Vermont" is confined to the Champlain Valley and it is suggested here that these names be restricted to these levels. Inferior "Lake Albany" levels are equivalent to the superior "Lake Vermont" levels and it is further suggested that existing names Quaker Springs and Coveville be applied to these water bodies, both north and south of the Lake George graben. There does not seem to be any compelling evidence for Chadwick's Lake Bolton as a separate Lake George event.

Lake Albany

The most recent event south of the Lake George area was the readvance of the glacier near Rosendale, New York (Connally, 1968b). As far as can be shown at present the ice front retreated northward from Rosendale, defending a northward expanding proglacial lake - Lake Albany. LaFleur (1965) has shown that Lake Albany was contemporaneous with the receding ice margin as far north as Troy, New York. As defined here, Lake Albany reached its maximum extent when the ice margin reached Glens Falls or slightly northward.

Luzerne Readvance

The Luzerne Mountain section, west of Glens Falls, shows two ice advances. The uppermost event dammed a lake in the upper Hudson Valley that received proglacial outwash. The ice margin can be traced northward to the valleys followed by Route 9N that connect the Lake George basin with the upper Hudson River valley at Lake Luzerne. A morainal ridge blocks this valley east of Lake Luzerne. Lake Luzerne and associated lakes occupy kettle holes in the outwash on the distal side of the moraine. The Pine Log Camp bog (Stop 4) is situated in one of these kettles. As the Luzerne readvance apparently terminated in the Glens Falls region, and as the ice marginal deltas that date from this event are at Lake Albany levels, the readvance must have been contemporaneous with Lake Albany.

Lake Quaker Springs

As the ice retreated north of the Lake George basin the water level dropped to the lake referred to by Stewart (1961) as the Quaker Springs stage of Lake Vermont. LaFleur (1965) and Connally (1968a) have demonstrated that this level in the Champlain Valley is coextensive with the middle water level earlier referred to "Lake Albany". These two coelescing water bodies are here combined as Lake Quaker Springs. No evidence for Lake Quaker Springs levels has been found north of the Ticonderoga quadrangle. The ice-marginal outwash delta seen at Stop 3 marks the level of Lake Quaker Springs and confirms the presence of stagnant ice associated with this lake. A similar relationship is found south of Middlebury, Vermont.

Lake Coveville

As the ice margin retreated northward from Ticonderoga the lake level must have dropped quite rapidly to the level of Lake Coveville (Chapman's Coveville stage of Lake Vermont). Chapman traced this level as far north as Burlington and Connally (1967b) suggested its presence in the Lamoille River valley east of

Burlington. Since Lake Quaker Springs was restricted to the area south of Ticonderoga, the lacustrine sediments north of that quadrangle evidently relate to younger lakes.

Bridport Readvance

The Bridport readvance has been suggested from many lines of evidence most of which can be observed within the boundaries of the Town of Bridport, Vermont. Exposures of crumpled, lacustrine (Coveville) clays containing boulders are commonly exposed in excavations from the vicinity of Middlebury north to Burlington, Vermont. The uppermost unit at the West Bridport section is either a till whose matrix is composed of redeposited lacustrine silt with scattered boulders, or a lacustrine unit containing ice-rafted boulders. In either case, this unit is separated from the lower tills by boulder-free lacustrine sediments. One mile east of West Bridport (Stop 2) a gravel pit displays a firm lodgement till over stratified material with the stratified material incorporated along shear(?) planes in the base of the till.

Since the contorted clays occur from Bridport to Burlington, it is possible that the ice margin retreated all the way north to Burlington before readvancing to Bridport. However, the exact magnitude of the readvance is unknown and may be much less. It is suggested that the mountain front morainic system of Denny (1966) in the Plattsburgh region was formed during this readvance.

Lake Vermont

As the ice margin retreated north of the Lamoille River Valley the lake level lowered to the Fort Ann stage of Chapman or the restricted Lake Vermont of this report. It appears likely that the dam for this level was formed by the glacier as it deposited the Highland Front Moraine of Gadd (1964).

Champlain Sea

Following the recession of the ice margin north of the St. Lawrence River Valley, marine waters entered the Champlain basin and formed the Champlain Sea. Shells in the marine clays have been found as far south as Crown Point, New York. As the land rebounded the marine waters were cut off and modern Lake Champlain began.

POLLEN STRATIGRAPHY AND ENVIRONMENTS

Site Description and Methods

The Pine Log Camp bog has developed on the glacial outwash on the distal side of the moraine associated with the Luzerne readvance, as described above. The bog is located on the Lake Luzerne quadrangle; 43°21'N latitude, 73°50'W longitude. It is one of several bogs in this area which were probed for greatest depth and presumably longest stratigraphic record. The site itself is a kettle hole with over 30 feet of closure. The sedimentary record indicates that the kettle was formerly a lake, then a closed bog covered by a peat and rootlet mat.

The Lake Luzerne-Lake George region lies in the northern hardwood forest, as described by Küchler (1964), and also contains transitional elements of the Appalachian oak forest. Dominants in the local forest, a mixture of deciduous and evergreen trees; are oak, yellow birch, beech and hemlock along with other forest components; sugar maple, ash, white pine, black cherry, northern white-cedar, basswood and elm.

The bog was probed for greatest depth which was established at 8.0 m at the center of the basin. Coring was done with a Davis-type piston corer, with cores retrieved from the surface to the base in successive 25 cm segments. Since coring was done in the late fall (1967) and cores were extruded directly into tubes, local contamination was minimized. Core samples were processed by standard treatment schedules for pollen analysis, in which up to 300 pollen grains per slide were counted. Seven additional core segments were taken from basal sediments, 7.85-8.0 m, for radiocarbon age determination by Isotopes, Inc.

From the base, the bog consists of: 0.25 m, coarse clastics at base, grading to silt with wood fragments; 1.75 m, gray clay at base grading to black-brown gyttja; 2.25 m, compact fibrous peat with clay decreasing upwards; compact fibrous peat with wood fragments; 2.00 m, compact fibrous peat, water saturated; 1.00 m, brown fibrous peat, relatively dry, with plant detritus at the surface.

Pollen Stratigraphy and Results

The pollen stratigraphy for the Pine Log Camp bog and thus for the late- and postglacial environments in the northern Hudson Valley is summarized in Table 1. Correlations are also provided between the northern Hudson Valley and the southern Wallkill

POLLEN ZONES	NORTHERN HUDSON VALLEY	SOUTHERN WALLKILL VALLEY	SOUTHERN NEW ENGLAND	WESTERN LONG ISLAND	STAGES
C3a-C3b	Spruce Rise Pine, Birch, Hemlock	Oak, Hemlock	Spruce Rise Oak, Hemlock	Oak, Chestnut, Holly	POST GLACIAL
OAK C2	Beech, Pine, Hemlock, Oak, Birch	Oak, Hickory	Oak, Hickory	Oak, Hickory	
C1	Hemlock, Birch, Pine, Oak	Oak, Hemlock	Oak, Hemlock	Oak, Hemlock	
B2	Pine, Oak	Pine, Oak	Pine, Oak	Pine, Oak	
PINE B1	Pine, Birch	Pine	Pine	Pine	
A4	Spruce Maximum Alder, Pine (A3-4)	Spruce Returns	Spruce Returns	Spruce Returns	LATE - GLACIAL
SPRUCE A3		Pine, Spruce, Oak (12,850 ± 250)	Pine, Spruce, Oak	Pine, Spruce	
A1,2	Spruce, Pine, Grass (Spruce Park)	Pine, Spruce	Birch, Spruce	Pine, Spruce	
T3	Pine, Birch, Spruce, NAP, Park-Tundra (12,400 ± 200)	Pine, Spruce, Birch	Birch Park-Tundra	Pine, Spruce, NAP	
HERB T2	- - - - - - - - - -	Spruce, Pine, Fir	Spruce Park-Tundra	Spruce	
T1	Glaciated	Pine, Birch Shrub-Tundra	Tundra	Park-Tundra	
W		Glaciated	Glaciated	Park-Tundra Near Tundra	

TABLE 1: Correlation of pollen stratigraphy - eastern New York and southern Connecticut (after Connally and Sirkin, 1967; Deevey, 1958; Sirkin, 1967)

Valley (Connally and Sirkin, 1967), southern New England (Deevey, 1958) and western Long Island (Sirkin, 1967). As indicated, the pollen record began with the deposition of clastic sediments directly over glacial outwash in the base of the kettle. The lower 15 cm of the bog section contains some woody plant detritus and is radiocarbon date at *12,400 ± 200 years B.P. The pollen spectra of pine, birch, spruce and nonarboreal pollen (NAP), mainly grass, is interpreted as the initial vegetation in the vicinity of the bog. The presence of significant amounts of willow, alder and birch (mainly the small sized pollen grains often associated with dwarf or shrub species of these plants and of pine) along with the pollen of the rose family, sedge and several herbs, including the genus Shepherdia, is interpreted as representing a tundra vegetation, that of the Herb Pollen Zone. The NAP maximum of 36% of the pollen also occurs in this sequence and helps substantiate this interpretation.

The presence of conifer pollen indicates the rapidity of colonization of the deglacial surface by the boreal forest which was migrating northward with the glacial front at that time and invading and isolating the tundra or shrub-tundra vegetation. Pollen of deciduous hardwoods in the basal spectra indicate long distance transport of pollen from the hardwood forests to the south. The radiocarbon date, 12,400 ± 200 years B.P. is set as the minimum age for deglaciation of this site.

In the regional correlations of pollen stratigraphy from the end moraines to the northern Hudson Valley, the rate of deglaciation is suggested in part by the time transgressive nature of the Herb and Spruce Pollen Zones. For example, it has been postulated that deglaciation began on Long Island a minimum of 17,000 years B.P. (Sirkin, 1967), in southern New England about 15,000 years ago (Deevey, 1958) and in the Wallkill Valley about 15,000 years ago (Connally and Sirkin, 1967). Ages for the lower Spruce Pollen Zone average 14,000 years B.P. at Sandy Hook New Jersey (Sirkin, et al., 1968), range from 13,000 to 12,000 years B.P. in southern New England (Deevey, 1958), and are recorded at 12,850 ± 250 years B.P. in the southern Wallkill Valley (Connally and Sirkin, 1967). In the northern Hudson Valley the Herb Pollen Zone age of about 12,400 years might also be considered as the maximum possible age for

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the Spruce Zone. These ages provide a time scale for the time transgressive aspect of plant migration following deglaciation. Overall, a minimum of 4,600 years (17,000 - 12,400) is required for glacial recession from Long Island to the Luzerne position, and a minimum of 1,600 years (14,000 - 12,400) is required for the migration of the spruce forest from just south of the end moraines to the northern Hudson Valley region. It is believed that deglaciation was initially slow, with a fluctuation of the ice front along northern Long Island (Sirkin, 1968) and a span of at least 3,500 years between recession from the Harbor Hill position into southern New England (using pollen subzone T3 of Port Huron age, as in Deevey, 1958, for control). Thus, deglaciation must have accelerated after 13,500 B.P. in order to bring the ice front to the northern Hudson Valley by 12,600 B.P. (see Regional Correlations). The assumption made with respect to the northern Hudson age is that the basal age of the Pine Log Camp bog closely dates the age of the surficial glacial deposits.

The Spruce interval in the northern Hudson region as interpreted from the pollen evidence can be divided into at least two subzones, tentatively A 1,2 and A 3,4. Subzone A1,2 is interpreted as representing a spruce park or taiga, while subzone A3,4 is marked by the spruce increase and the late-glacial spruce and alder maximum. Significant peaks of pine, cedar and birch, and a fir peak, with the spruce and alder, indicate a spruce dominated, but mixed coniferous forest, possible with alder dominant at the bog site. Similar pollen data were obtained by Cox (1959) in his study of the Consaulus bog, to the southwest in the Amsterdam, New York, Quadrangle. In that study, the arboreal pollen record was interpreted as including subzones A1, A2 and A3. The tundra episode or Herb Zone is omitted from the record in the absence of a NAP record and radiocarbon ages. Although the NAP were apparently not counted, the basal arboreal spectra at Consaulus do not include the pollen of willow and birch, and contain only an insignificant amount of alder. It is presumed that the Herb zone was not encountered.

In southwestern Vermont, the Herb zone was also not encountered at Pownal bog (Whitehead and Bentley, 1963) where the pollen record does include the NAP. The Pownal record was terminated at Zone A, although there may have existed a longer section in that bog. The A zone at Pownal is characterized by the "spruce and fir maximum, low pine and deciduous trees", which correlates with the pollen spectra at Pine Log Camp and at Consaulus. Both Pine Log Camp and

Consaulus record two spruce and fir peaks, Pownal only the upper spruce and fir peak. The upper spruce peak incorporates the spruce maximum of 38% at Pine Log Camp, 60% at Pownal, and about 66% at Consaulus and at Pine Log Camp bog, the grass maximum of nearly 20%. The arboreal pollen vary from 76 to 86% of the pollen in the Spruce zone in this study.

The Pine Pollen Zone (the B zone) in the Pine Log Camp bog pollen record is subdivided into the B1 and B2 subzones. Subzone B1 is characterized by pine and birch dominance, mainly a pine peak and the birch maximum. Also in this subzone, spruce declines and the fir maximum is reached. The NAP are represented by grass, sedge and cattail. Subzone B2 is typified by the pine maximum of 58% at Pine Log Camp. This maximum also occurs in subzone B2 at Pownal and possibly at Consaulus, although the latter bog was not zoned in the pollen study. Significantly, oak reaches a peak in the present study and at Pownal.

The Oak Pollen Zone is represented in this study by three subzones, C1, C2, and C3, and the latter subzone may be split into the C3a and C3b subdivisions. The subzones of the Oak Zone differ from those in more southerly locations in the subordination of oak to hemlock, birch, pine, beech and lastly spruce. Subzone C1 contains rising hemlock and beech profiles, at the top of the C1 subzone. In addition, birch and oak increase and the chestnut maximum occurs here, while pine declines to its minimum amount.

In subzone C2, beech reaches its maximum of 22%, while hemlock drops to less than 20% during mid-C2 time, and pine gains. Hemlock increases in late C2 time with a corresponding falling off of pine, beech and ash. The oak maximum of 22% occurs in lower C3, that is C3a time, accompanied by an increase in beech and a gradual dropping off of hemlock. Subzone C3b shows a distinct change in forest composition with significant increases in pine and birch and by the spruce return which began in C3a time and reaches a peak in C3b time. Also significant at this time is the rapid influx of composite (cf. ragweed) pollen to a maximum of 14%. This phenomenon is typical of pollen records in the northeastern United States where it has been assigned either to pre-Colonial forest fires, i.e. those set by Indians as a device for clearing the land, or as a sign of European colonization and agriculture.

With respect to regional climatic and vegetational adjustments, it is postulated that the Spruce Zone represents vegetation responding to colder climates associated with deglaciation. The A4 subzone is generally correlated with the last advance of continental ice and the end of the late-glacial episode. The B zone has been associated with the climatic optimum, Hypsithermal, which probably extends well into the Oak Zone. Also, it appears that the lake at Pine Log Camp probably entered its bog stage during this warm-dry climate. The advent of the cooler, more moist sub-Atlantic climate is indicated in the Spruce rise of C3b time.

REGIONAL CORRELATIONS

Denny (1966) recognized areas of hummocky topography near the northern edge of the Adirondack Mountains and correlated this with discontinuous patches of moraine in the Saranac and Salmon River Valleys west of Lake Champlain. He proposed the informal name "mountain front morainic system" for these deposits, and interpreted this as a recessional position rather than a readvance.

East of the Champlain Valley, Gadd (1964) defined the Highland Front Moraine as a major morainic system that can be traced for 225 miles from Ganby to Rivier-de-Loup along the southeast border of the Eastern Quebec Uplands. Although the similarity of geographic position between the mountain front and Highland Front systems suggests that they are morphostratigraphic equivalents, Denny (1967) has projected the Lake Coveville shoreline on the mountain front system. Thus, the mountain front system is suggested as the morphostratigraphic equivalent of the Bridport readvance while the ice responsible for the Highland Front Moraine probably dammed Lake Vermont.

Recently McDonald (1968) has argued that the Highland Front Moraine must have been deposited about 12,600 years B.P. Lee (1963) established a maximum age of 12,720 ± 170 years B.P. while the initiation of the Champlain Sea at about 12,000 years ago appears to be a minimum age. McDonald uses a date of 12,570 ± 220 years B.P. for a bog on Mount St. Hilaire (Lasalle, 1966; Torasmae and Lasalle, 1968) as the crucial evidence in establishing the date of 12,600 for the Highland Front position. Although the authors accept McDonald's reasoning, the age is in apparent disagreement with their opinion that sedimentation at the base of the Pine Log Camp bog (12,400) closely postdates the Luzerne readvance. The Luzerne and Bridport readvances both predate the Highland Front system.

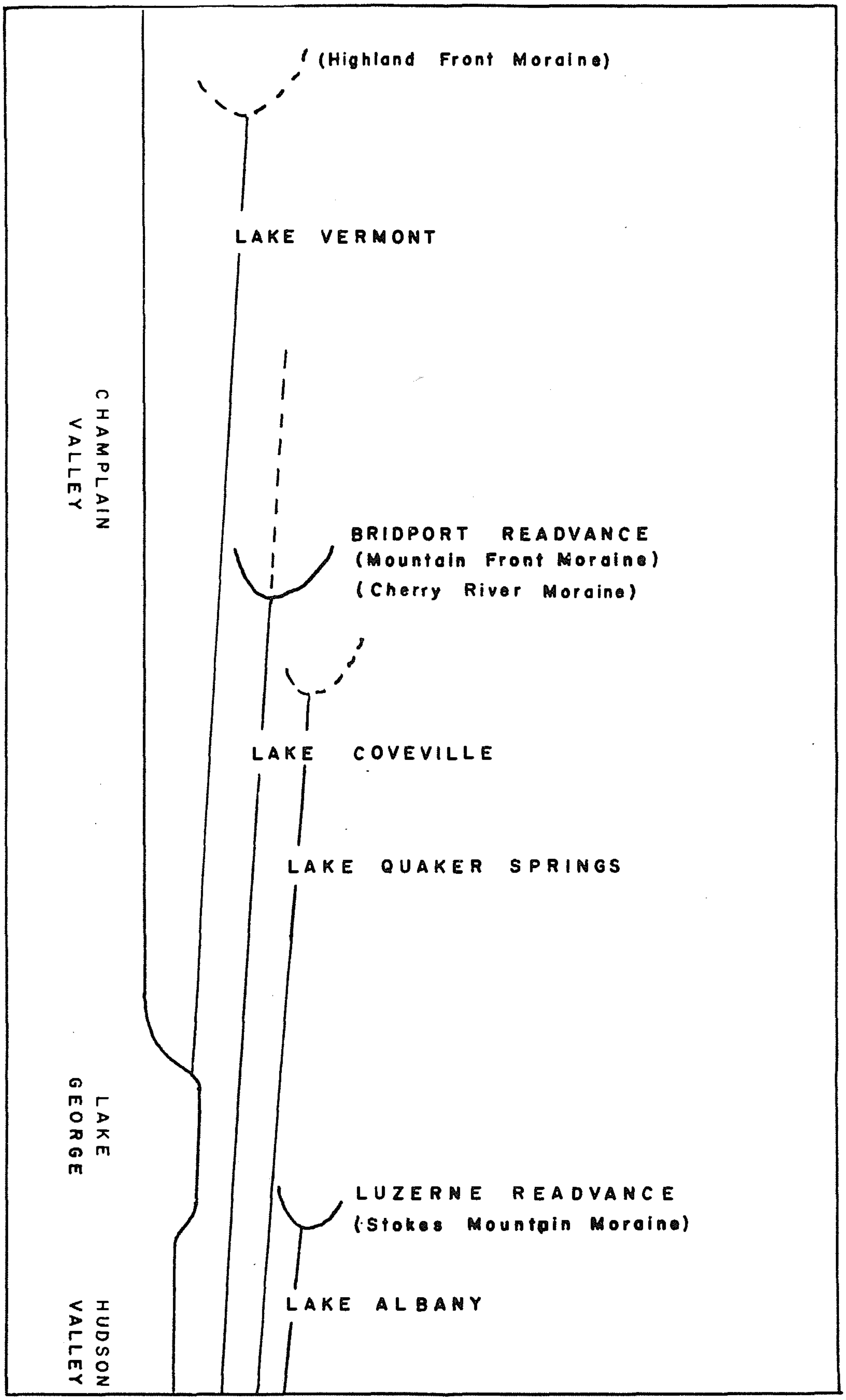

Figure 2. The relationship between lake levels and ice margins in the Lake Champlain-Lake George lowland and their possible upland equivalents.

If one subtracts the counting error from the St. Hilaire date and adds the counting error to the Pine Log camp date, the sequence of events outlined in this report is compatible with that reported by McDonald. If the St. Hilaire bog dates from 12,350 then McDonald could accept a date of 12,400 for the Highland Front Moraine. If the Pine Log Camp bog dates from approximately 12,600 then the Luzerne readvance precedes emplacement of the Highland Front system by 200 years. This leaves ample time for the Bridport readvance and consequent development of the mountain front system between these two events. Thus, the authors suggest that the St. Hilaire bog should be considered to date near its minimum and the Pine Log Camp near its maximum range, placing them in complete agreement with the stratigraphy of McDonald.

McDonald reports two moraines older than the Highland Front Moraine in the Megantic Hills. Both the Cherry River (younger) and Stokes Mountain Interlobate (older) moraines appear to border Glacial Lake Memphramagog. If the relationship between the various levels of this proglacial lake and the lakes of the Champlain Valley can be ascertained it will be possible to correlate the Luzerne and Bridport readvances with events in the Eastern Quebec Uplands. If the readvance to the Cherry River position is contemporaneous with drainage of Glacial Lake Memphramagog through the Lamoille River Valley to Lake Coveville, then it probably correlates with the Bridport readvance and the mountain front morainic system. This correlation is tentatively suggested here. The ice marginal positions corresponding to the Stokes Mountain Interlobate and other moraines of the Eastern Quebec Uplands are suggested as correlates of the Luzerne readvance. Fig. 2 summarizes the relationships.

REFERENCES SITED

Broughton, J. B., D. W. Fisher, Y. W. Isachsen, and L. V. Rickard, 1966, Geology of New York State, A Short Account; N.Y. State Museum, Ed. Leaflet 20, 47 p. and map.

Chadwick, G. H., 1928, Ice evacuation stages at Glens Falls, New York; Geol. Soc. America, Bull., v. 39, pp. 901-922.

Chapman, D. H., 1942, Late-glacial and postglacial history of the Champlain Valley; Vt. Geol. Survey, State Geol. 23rd Rpt., pp. 48-83.

Connally, G. G., 1965, Surficial geology of the Brandon-Ticonderoga quadrangles, Vermont; ms map and Rpt. to Vt. State Geologist, 41 p.

________________, 1967a, Glacial geology of the Brandon-Ticonderoga region, Vermont; Geol. Soc. America, Sp. Paper 115, pp. 255-256.

________________, 1967b, Glacial geology of the Mount Mansfield quadrangle, Vermont; Geol. Soc. America, Sp. Paper 115, p. 256.

________________, 1968a, Surficial resources of the Champlain Basin, New York; ms maps and Rpt. to N. Y. State Office of Pln. Coord., 111 p.

________________, 1968b, The Rosendale readvance in the lower Wallkill Valley, New York; Guidebook to Fieldtrips, NAGT, Eastern sec. meeting, New Paltz, N.Y. pp. 22-28.

Connally, G. G. and L. A. Sirkin, 1967, The Pleistocene geology of the Wallkill Valley; in R. Waines, ed., Guidebook NYSGA, 39th Ann. meeting, pp. A1-A16.

Cox, D. D., 1959, Some postglacial forests in central and eastern New York State as determined by the method of pollen analysis; N. Y. State Museum Bull. 377, 52 p.

Deevey, E. S., 1958, Radiocarbon-dated pollen sequences in eastern North America. Veröffentl. Geobotan. Inst. Rubel, Zurich, 34: 30-37.

Denny, C. S., 1966, Surficial geology of the Plattsburgh area; Empire State Geogram, v. 4, no. 3, pp. 6-10.

______________, 1967, Surficial geologic map of the Dannemora Quadrangle and part of the Plattsburgh Quadrangle, New York; U.S.G.S. Map GQ-635.

Doll, C. G., W. M. Cady, J. B. Thompson, Jr., and M. P. Billings, 1961, Geologic Map of Vermont, Vt. Geol. Survey.

Gadd, N. R., 1964, Moraines in the Appalachian region of Quebec; Geol. Soc. America Bull., v. 75, pp. 1249-1254.

Küchler, A. W., 1964, Potential Natural Vegetation, United States; AGS Spec. Pub. 36, New York, 150 p.

LaFleur, R. G., 1965, Glacial geology of the Troy, N. Y. quadrangle; N. Y. State Museum, Map & Chart Series No. 7, 22 p.

Lasalle, P., 1966, Late Quaternary vegetation and glacial history in the St. Lawrence Lowlands, Canada; Leidse Geol. Mededel. pt. 38, p. 91-128.

Lee, H. A., 1963, Pleistocene glacial-marine relations, Trois-Pistoles, Quebec; Geol. Soc. America, Sp. Paper 73, p. 195.

McDonald, B. C., 1968, Deglaciation and differential post-glacial rebound in the Appalachian region of south-eastern Quebec; Jour. Geol., v. 76, pp. 664-677.

Newland, D. H. and H. Vaughan, 1942, Guide to the geology of the Lake George region; N. Y. State Museum, Handbook 19, 234 p.

Sirkin, L. A., 1967, Late-Pleistocene pollen stratigraphy of western Long Island and eastern Staten Island, New York; in E. J. Cushing and H. E. Wright ed., Quaternary Paleoecology, Yale Press, New Haven, pp. 249-274.

______________, 1968, Geology, geomorphology, and late-glacial environments of western Long Island, New York; in R. M. Finks ed., Guidebook NYSGA 40th Ann. meeting, CUNY Queens, pp. 233-253.

Sirkin, L A., J. P. Owens, and J. P. Minard, 1968, Palynology of some buried Pleistocene peat samples from the New Jersey coastal plain. Geol. Soc. America, NE sec. meeting, Washington.

Stewart, D. P., 1961, Glacial geology of Vermont; Vermont Geol. Survey, Bull. 19, 124 p.

Stewart, D. P. and P. MacClintock, 1964, The Wisconsin stratigraphy of northern Vermont; Am. Jour. Sci., v. 262, pp. 1089-1097.

______________________________, 1967, Pleistocene stratigraphy in Vermont; Geol. Soc. America, Sp. Paper 115, p. 295.

Terasmae, J. and P. Lasalle, 1968, Notes on late-glacial palynology and geochronology at St. Hilaire, Quebec; Canadian Jour. Earth Sci., v. 5, pp. 249-257.

Whitehead, D. R., and D. R. Bently, 1963, A postglacial pollen diagram from southwest Vermont; Pollen Spores, 5: 115-127.

Trip J

SURFICIAL GEOLOGY OF THE INTERNATIONAL LEAD COMPANY MCINTYRE DEVELOPMENT AT TAHAWAS, N.Y.

by

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Take Northway south to Exit 29. Turn right onto North Hudson - Newcomb Road then right (south) on Route 9. Stop at sand pit on left side of road behind gas station.

Stop 1 - Schroon Lake Quadrangle

Deltaic sands, forset beds dipping west into valley wall. Delta built into Glacial Lake Warrensburg (Miller 1925).

Water seems to have carried material into the Valley from the Blue Ridge moraine which lies about 3 miles due west.

Stop 2 - Blue Ridge Moraine

This is a narrow moraine complex blocking off the east west valley. Outwash from this moraine can be traced into the deltaic deposits of the last stop. The north side has been modified by water flow from the north indicating ice occupying the valley to the W of this spot with the main N-S valleys free of ice.

Stop 3 - Outwash gravel and Kame deposits

Water was flowing out of Niagara Brook along the north side of the Blue Ridge Moraine into Glacial Lake Warrensburg.

Stop 4 - Newcomb Quadrangle

Striated bed rock face on Highway 128, 2 miles west of Newcomb. Just west of BM 1595 and Lodo Pond.

Which way was the ice flowing past this point?

Stop 5 - Tahawus, McIntyre Development, International Lead Co.; Newcomb 15' Quad.

Excavation of the present pit area was started in 1961. In clearing the overburden from the area formerly under Sanford Lake, a multiple till section was exposed with interglacial lake sediments between two tills. Disseminated wood fragments including material identified as Pinus strobus (David Bierhorst, Dept. of Botany, Cornell Univ.) were collected by E. Muller and the section was described (Muller 1966). The wood fragments were dated at an age greater the 40,000 years (W-1520). Muller describes two tills separated by the interglacial lake deposits.

This author observed what appeared to be a third till when visiting the pit in 1966. Detailed measurement in the pit verified the existence of three till units separated by stratified sediments.

Section at Sanford Pit (as exposed in 1966)

Top

Coarse sand and gravel, oxidized in zones - upper part mixed with excavation fill from Tahawas village	2 - 5'
Laminated sand, silt, some gravel lenses, upper parts oxidized, well developed ripple marks cut and fill structures throughout, bedding dips 8°N.	1 - 20'
Till, yellow brown oxidized moderately stoney, non-calcareous very few ore pebbles.	1 - 25'
Sandy gravel, laminated sand and silt. Numerous small folds overturned to the north. In some places this layer has been so disturbed it becomes till like in texture.	4'
Till, gray, moderately stoney, non-calcareous few Potsdam pebbles observed, no ore pebbles. Contact with overlying sediments marked by thin silt bands.	8 - 15'
Sand, yellow brown oxidized medium to coarse changes to sandy gravel a short distance to the west. Contact with underlying laminated clay not observed.	5 - 15'
Clay, brown with few pebbles and disseminated wood fragments including material identified as Pinus strobus (David Bierhorst, Dept of Botany, Cornell Univ.) Age greater than 40,000 years (W-1520).	3 - 12'

Gravel stratified	1'
Till yellow grey moderately stoney, non-calcareous, oxidized	5'
Till, stoney, numerous ore pebbles folded silt, sand inclusion, shear planes dipping south	1 - 30'

Return to Plattsburgh.

REFERENCES CITED

Alling, H.L., 1911, Geology of the Lake Clear region (parts of St. Regis and Saranac quadrangles): N.Y. State Museum Bull. 207, 208.

___________, 1916, Glacial Lakes and other glacial features of the central Adirondacks: Geol. Soc. Am. Bull. 27:645-672.

___________, 1919, Pleistocene geology of the Lake Placid quadrangle: N.Y. State Museum Bull. 211, 212: 71-95.

___________, 1921, Glacial geology of the Mt. Marcy quadrangle: N.Y. State Museum Bull. 229, 230: 62-84.

Fairchild, H.L., 1913, Pleistocene geology of New York state: Geol. Soc. America Bull. 24: 133-162.

___________, 1932, Closing stage of N.Y. glacial history: Geol. Soc. America Bull. 43: 603-626.

Kemp, J.F., 1905, The physiography of the Adirondacks: Science 21: 988-989 (abst.)

Miller, W.J., 1925, Remarkable Adirondack glacial lake: Geol. Soc. America Bull. 36: 513-520.

Muller, E.H., 1965, Adirondack Mountains, INQUA Field Conference, A. Guidebook, New England - New York State pp. 52. INQUA VIIth Congress.

Ogilvie, I.H., 1902, Glacial phenomena in the Adirondacks and Champlain Valley: Jour. Geol. 10: 397-412.

___________, 1902, The geology of the Paradox Lake quadrangle: N.Y. State Museum Bull. 96: 461-506.