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Field Trip Guidebook Number 3

Cretaceous Intrusions in the Northern Taconic Mountains Region, Vermont
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FIELD GUIDE TO CRETACEOUS INTRUSIONS IN THE NORTHERN TACONIC MOUNTAINS REGION, VERMONT

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Introduction

Northern New England and adjacent Canada were host to widespread, mid-plate, non-orogenic alkalic igneous intrusions during Early Cretaceous times. In several stages, from 130 to 100 million years ago, groups of plutons, dike swarms, and individual igneous complexes were intruded and now form an igneous province that stretches 400 km from the central Adirondack Highlands of New York eastward through southern Maine, and 350 km from the Monteregian Hills of southern Québec southeastward through New Hampshire (Fig. 1). We have been calling this the New England-Québec ("NEQ") igneous province (McHone and Butler, 1984). Still more of these intrusions are found in the continental shelf (Puffer, 1989), and they continue across the western North Atlantic as the New England seamount chain.

If an "eogeologist" could do field work in the Vermont of 110 million years ago, some members of the NEQ province would be mapped as volcanic mountains in a tectonic setting perhaps not too different from that of East Africa today. The evolution of our present topography of valley basins and mountain ranges was far from finished and "terrible lizards" walked upon stratigraphic formations that were several kilometers above the present surface. Those formations may well have included Mesozoic clastic and volcanic rocks, especially within the structural basins that flank Vermont (McHone, 1982). One proven area of volcanism in eastern Vermont is the great monadnock of Ascutney Mountain, which has blocks of volcanic rocks preserved within plutonic rocks near its peak. Other intrusions, such as the Cuttingsville complex and Barber Hill stock of western Vermont, show no evidence for associated extrusive rocks. We have no sure idea of whether dikes of hypabyssal magmas, such as those visited on this trip, ever reached the surface anywhere in the region.

Differential erosion, controlled by faulting, uplift, down-drop, and weathering, has removed the stratigraphic evidence of Vermont's Mesozoic volcanic and sedimentary rocks, yet that same erosion now provides views of the plutonic chambers that once lay beneath Cretaceous volcanoes. Erosion has also exposed thousands of dikes across the NEQ province, some of which we describe in this paper and others of which are described in works listed in the bibliography. High-angle faults, mostly normal, are also exposed and may be relicts of Mesozoic tectonism in the region (McHone, 1987).

NEQ dikes display a bimodal range of mafic and felsic types in overlapping sets and swarms, each with somewhat distinctive ages and physical characteristics. Given that they cooled in contact with rocks only a few kilometers below the surface, the dikes are naturally intermediate in crystal character between phaneritic plutons and aphanitic volcanic rocks. Good eyesight and a hand lens are required to make out the minerals and textures of the dikes, but with care and experience, most can be classified in the field. Unlike the great quartz tholeiite intrusions of southern and eastern New England, most NEQ dikes are too small to have produced flood basalts or large volcanic edifices. Yet, as shown by xenoliths of spinel peridotites and other mantle rocks (McHone and Williams, 1985), these magmas, in dikes with just a few meters of exposed width, ascended from mantle depths.

This field excursion presents work that characterizes the little-known northern Taconics (NT) subprovince of the New England-Québec igneous province. We have described only a few of the NT intrusions; the group is worthy of a thesis or other research efforts (let us know if you are interested). As in other regions, relatively few of the geologists who produced the quadrangle bedrock maps for the area paid much attention to these intrusions. Some older papers contain very useful information, such as those by Marsters (1889), Dale (1899), Eggleston (1918), and Fowler (1950). More recent work centers on the pluton at Cuttingsville (e.g. Laurent and Pierson, 1973; Robinson, 1990; Wood, 1984; and Eby, 1992).

Dikes at Stops 1, 2, 3, and 5 were first shown to J.G.M. by E. Stanley Corneille, who shared these geological interests while doing graduate work at the University of Vermont in the early 1970's. Other field visits were made with J. Robert Butler during and after J.G.M.'s Ph.D. work at the University of North Carolina at Chapel Hill (1974–1978) and some field work was also conducted with Chiasma Consultants, Inc. for the National

Uranium Resource Evaluation (1978–1980). We made a field tour in September of 1992 for this log.

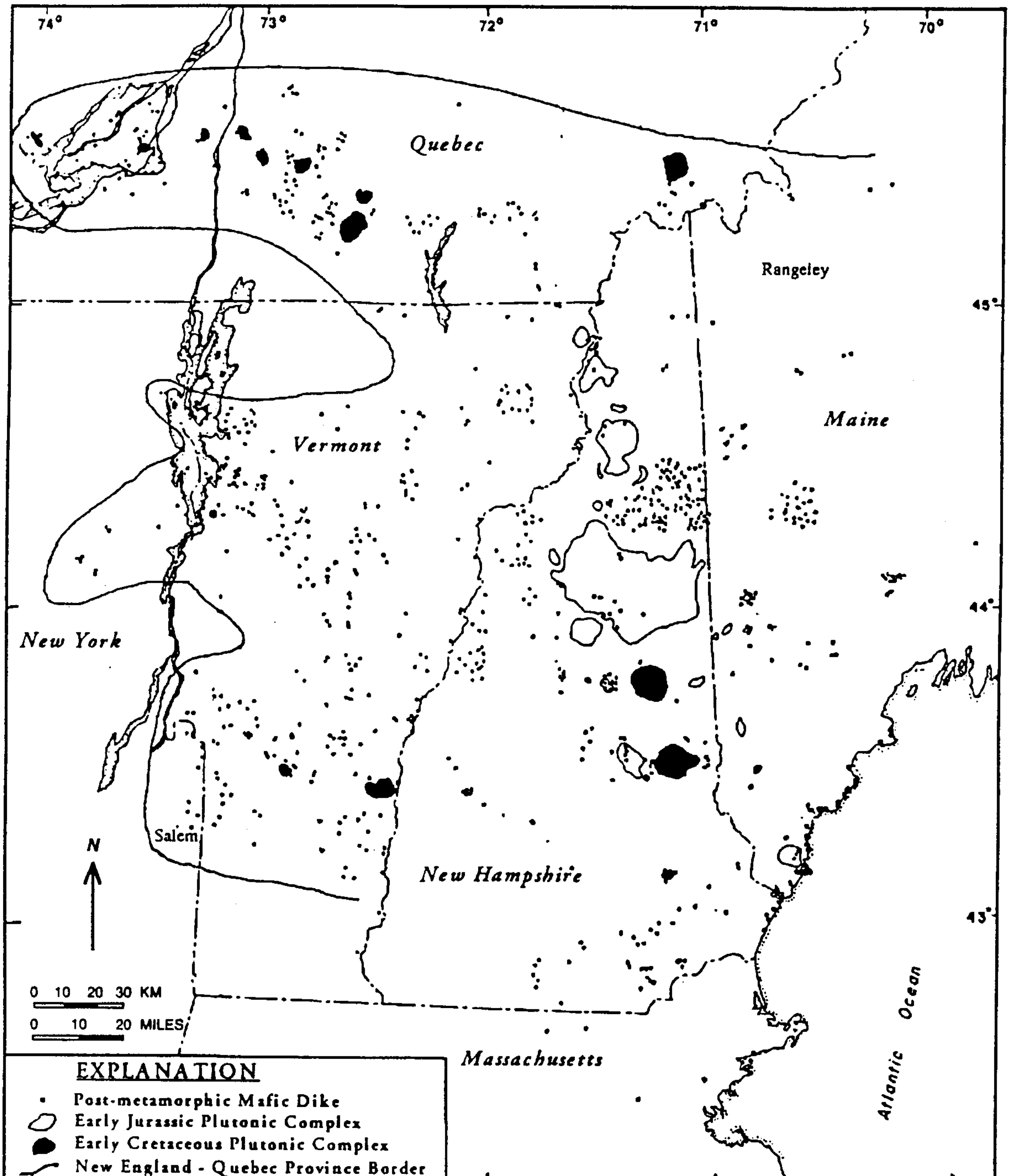

Figure 1. Geographic extent of the New England-Québec igneous province showing the distribution of unmetamorphosed (post-Acadian) dikes and larger intrusions.

The Northern Taconic NEQ Subprovince

The western border of the NEQ province outlines three lobes, or subprovinces, that extend westward from northern New England (Fig. 1). The northernmost lobe is the Monteregian Hills subprovince of southern Québec which is well known for its carbonatites and ultramafic stocks as well as for alkali lamprophyre dikes. Most radiometric dates are near 110 (±5) Ma.

Table 1. Radiometric Dates, Northern Taconic Igneous rocks

Site	Description	Date (Ma)	Reference
Stop 3 (PO-1)	Spessartite, Rte. 4 road cut Poultney quadrangle lat. 43°32'05"N long. 73°10'35"W	113 ±4	(1)
Stop 6 (WR-3)	Hbl spessartite, Rt.4 road cut West Rutland quadrangle lat. 43°30'49"N long. 73°03'20"W	108 ±4	(2)
Stop 9 (RT-7)	Andesitic breccia, Shrewsbury Rutland quadrangle lat. 43°30'57"N long. 72°53'56"W	101 ±4	(1)
Stop 10	Biotite syenite, Cuttingsville	102 ±2	(3)
	essexite	98.8 ±2	(3)
	essexite	103 ±4	(4)
	quartz syenite	108 ±1	(5)
	quartz syenite	100 ±3	(6)

Note: All dates are by K-Ar analysis of whole-rock and mineral samples, except for Cuttingsville date by Ref. 5, which is by Rb/Sr isochron. Dates have been revised, where appropriate, to newer IUGS decay constants.

References: (1) This paper; dates courtesy of H. Kreuger, Geochron Labs; (2) Zen, 1972; (3) Armstrong and Stump, 1971; (4) Stone & Webster unpub. date, Ref. Kanteng (1976); (5) Eby, 1992; (6) G.N. Eby, pers. comm., 1992.

Igneous rocks are unknown in the northernmost Lake Champlain Valley but, north and south of Burlington, Vermont there are several hundred lamprophyre and trachyte dikes exposed along shorelines, road cuts, streams, and hillsides (Fig. 1). Lamprophyre dikes of this Champlain lobe or subprovince are distributed westward into the central Adirondack Highlands of New York and eastward into north-central Vermont (McHone and Corneille, 1980). Champlain Valley dikes are identical to Monteregian dikes, including carbonate rich types, but associated plutonic complexes are fewer and smaller in the Lake

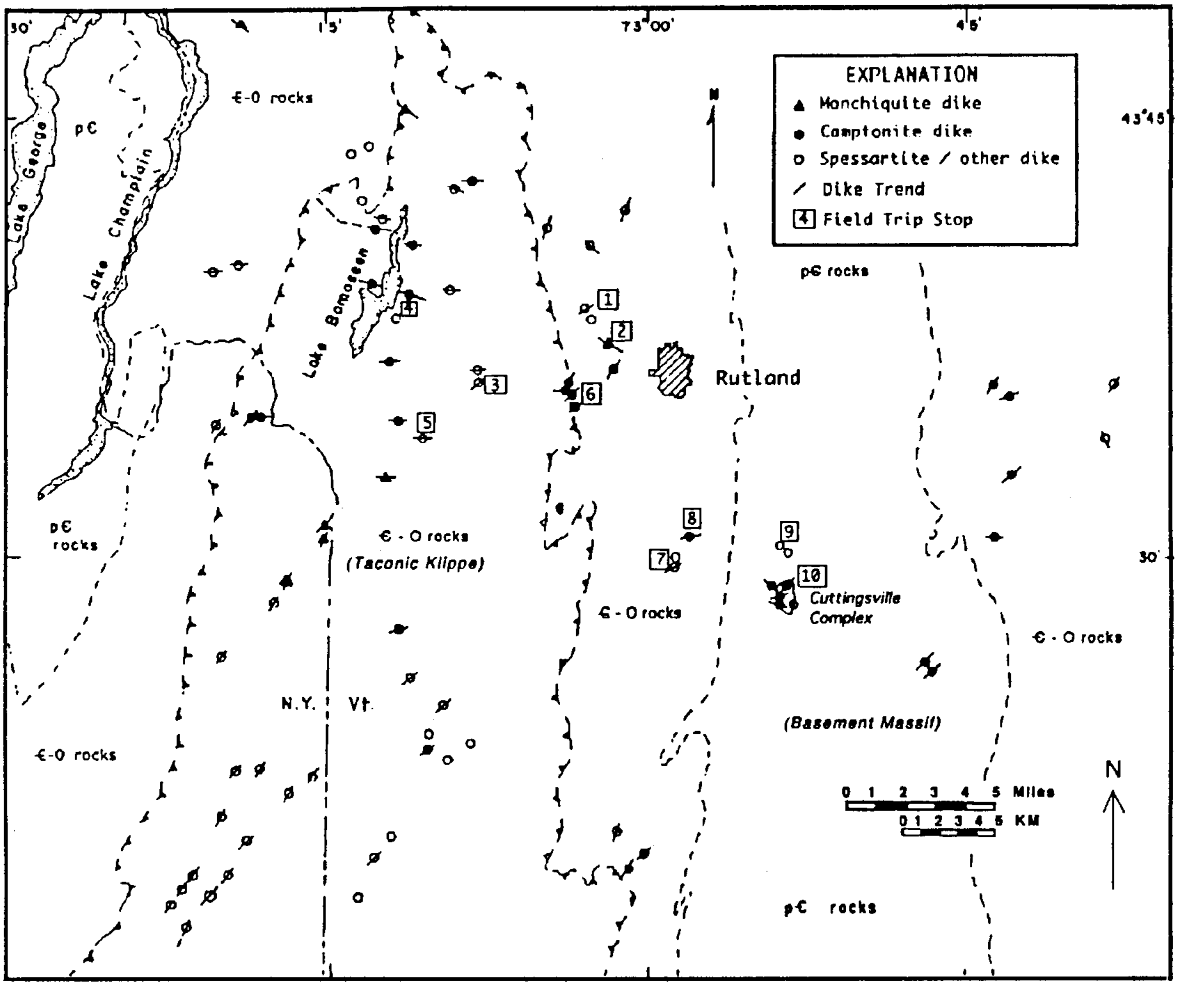

Figure 2. Dike locations in the northern Taconics region, Vermont and New York.

Champlain region than in Québec. Radiometric dates indicate ages near 135 Ma for monchiquites, 125 Ma for trachytes and syenites, and 115 Ma for camptonites (McHone, 1987), although this very neat correlation of rock type and age needs better confirmation.

In the southern or upper Champlain Valley (the lake flows northward), there is a "virtual" gap in igneous rocks, with only a few stray dikes known at Vergennes, Middlebury, Westport (New York), and Orwell (Fig. 1). The third lobe (herein labeled "NT" for Northern Taconics) has around 70 known dike localities (some of which are probably exposures of the same dike) distributed along the northern Taconic region between Proctor and Dorset, westward a few kilometers into eastern New York, and eastward across the Vermont Valley into the Green Mountains southeast of Rutland (Fig. 2). Many of these intrusions have petrologic characteristics that are distinct from the northern NEQ dikes, but there are also some very similar examples. Except for a few trachytes near Rutland, all of the dikes so far studied are lamprophyres. Dates are mostly 100–110 Ma (Table 1).

While the dikes in the NT subprovince have only been subject to reconnaissance study, the Cuttingsville plutonic complex, southwest of Rutland, has received attention since the 1970's from mineral companies as well as by research geologists (e.g. Laurent and Pierson, 1973; Robinson, 1990; Wood, 1984; and Eby, 1992). We consider the Cuttingsville intrusions to be part of the Northern Taconic subprovince on the basis of age and petrology, although the density of presently known dikes diminishes towards Cuttingsville (Fig. 2). Eastward from Cuttingsville dikes remain fairly common (2 to 6 dikes have been sampled per 15' quadrangle) where they merge with the regional camptonite swarms of eastern Vermont, New Hampshire, and Maine (McHone, 1984).

Trends and Structures

Each of the three western NEQ lobes has different orientation maxima for dikes (Fig. 3). The Monteregian subgroup has a WNW–ESE maxima for dikes, examples for which are found the entire distance from Montreal to northwestern Maine (McHone, 1978a). Dikes of the Champlain Valley subprovince have a very distinct E–W preference (McHone and Corneille, 1980). When plotted together, dikes in the northern Taconics lobe display a NE–SW maxima (Fig. 3, 'Rose C'). Most of the dikes in the southern and eastern portions of the subprovince show this NE–SW trend, but there are also several dikes in the northern portions that trend between E–W and SE–NW (Fig. 2).

We believe that most dike magmas intrude along directions of simple extension, widening fractures against the direction of minimum compression within the upper crust (McHone, 1988). But how can dikes of (presumably) the same generation have different trends across such a small area? Possibilities include: (1) Dikes radiate from a plutonic center or chamber analogous to the exposed Spanish Peaks (Colorado) volcanic center. Such a hidden pluton would have to exist somewhere near Rutland to be roughly at the intersection of local dike trends. (2) Dikes have originated at different times in the same areas, responding to a changing stress field as they formed. (3) Some dikes have filled "shear" fractures in addition to extensional fractures during the same event. Finally, (4)

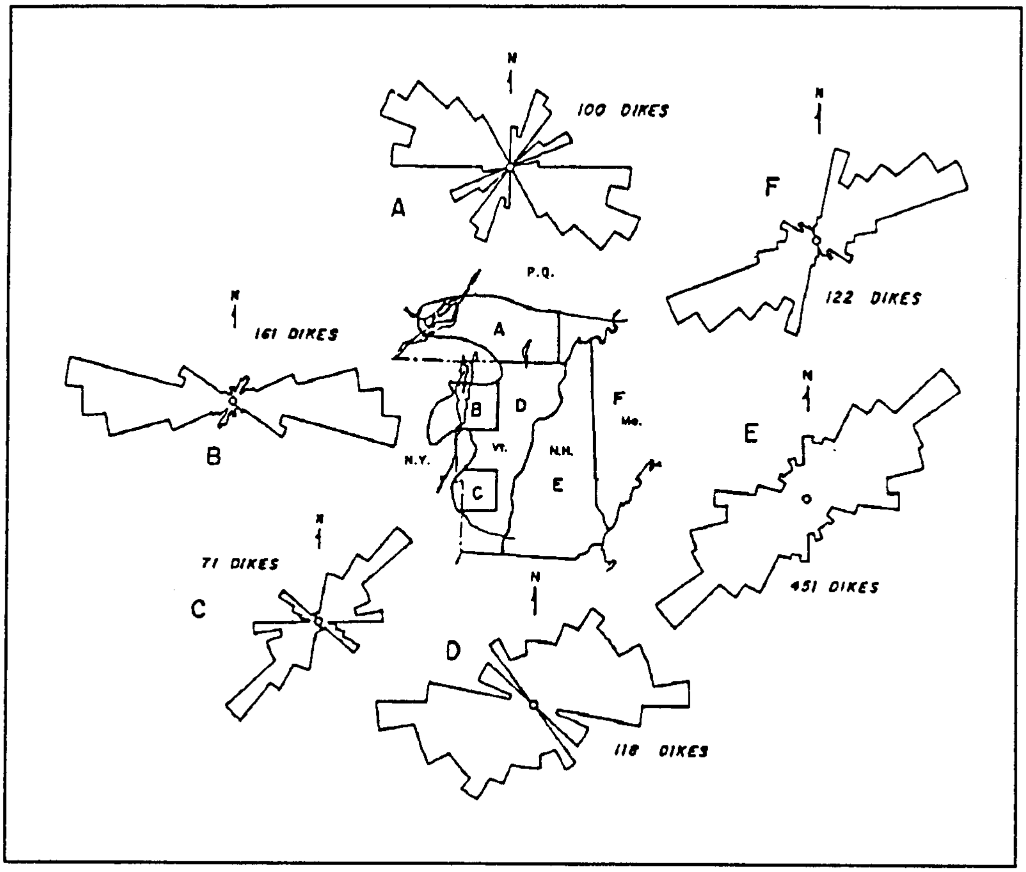

Figure 3. Dike trends (rose diagrams) in northern New England. Rose diagram labels refer to areas shown in the central figure, with rose C representing northern Taconic dikes.

we are intrigued by the notion of a major lithospheric break (see below) that acted as a tectonic boundary to stress (and strain) fields in the region.

McHone and Shake (1992) suggested that the shift of the Cretaceous NEQ dikes from E–W and ESE–WNW trends in northern areas to a NE–SW trend toward southeastern New England (Fig. 3), is controlled by a lithospheric cross-structure that is partly expressed by major topographic lineaments (Shake and McHone, 1987). The proposed structure underlies a lineament that extends across the region in a northeasterly direction from the vicinity of Salem, New York at least to Rangeley, Maine (Fig. 1). The Salem-Rangeley zone is contiguous with the Alabama-New York lineament of King and Zietz (1978), which they proposed to be due to a major high-angle structure. This proposed structure was a boundary that reoriented stress fields in New England which in turn controlled the orientation of dikes in the NEQ province. We have, however, no direct field evidence for such a major "basement break."

Northern Taconic Igneous Rock Types

The NT dike types include monchiquite (nephelinite), camptonite (basanite), bostonite (trachyte), and spessartite (andesite), all of which can be examined during this trip. During this trip, observe the different colors for mafic and felsic dikes types. The quartz syenite member of the Cuttingsville complex is visited on the final stop, depending on field conditions. All of these igneous rocks are presumed to be related through an event that included fractional melting, differentiation, and crystallization, but it is unlikely that they were at one time all co-magmatic.

Monchiquite is a very mafic, granular, analcite-bearing, olivine-bearing, augite-rich alkali basalt, often with appreciable calcite (in spheroidal bodies), phlogopitic mica, and kaersutitic hornblende. Feldspar (Ca-plagioclase) is poorly developed or lacking. Monchiquite is commonly dark gray in color.

Camptonite can look much like monchiquite, except that olivine is rare or absent, kaersutite is common to abundant, and plagioclase is more abundant than analcite. Phenocrysts are only mafic (augite and/or kersutite), rather than felsic. Camptonite dikes usually have a brownish to medium gray range of colors.

Spessartite dikes lack olivine and analcite, but plagioclase (intermediate Ca) is well developed and present both as phenocrysts as well as intergrown with augite in the groundmass. Phenocrysts (or megacrysts) of kaersutite serve to distinguish spessartite from tholeiitic dolerite (diabase) dikes that are common in other parts of New England. Spessartite often shows a distinctly greenish or purplish cast as well as gray colors.

Bostonite is a name that in a strict sense applies only to felsic (anorthoclase-rich) dikes that have a "felty" clumped-grain texture, which actually is not always present. **Trachyte,** although used for volcanic rocks as well, is a better general term. Minor minerals include oxidized biotite, quartz, and clay products. Some examples show well-formed alkali feldspar and/or quartz phenocrysts. Trachyte dikes may be iron-stained, but they are generally light brown to cream-colored on fresh surfaces. The quartz syenite of Cuttingsville can be chemically like trachyte (Table 2), but at Stop 10 the syenite has been enriched by sulfides.

TABLE 2. Chemical Analyses of Intrusive Rocks and Minerals, Northern Taconics

OXIDE (Wt. %)	WR-4B	PO-1	MPMB	QTZ-SYEN	PO-1k	PO-1a	PO-1p	PO-1i
SiO_2	45.09	50.85	69.00	65.86	41.32	50.77	55.32	6.61
TiO_2	2.89	1.71	0.24	1.03	4.84	1.08	0.15	55.62
Al_2O_3	13.58	16.66	16.50	17.19	11.60	4.43	28.08	2.56
FeO*	10.97	6.71	2.25	2.88	12.03	5.45	0.70	18.15
MnO	0.19	0.22	0.20	0.17	0.28	0.17	0.01	0.84
MgO	7.02	3.86	0.06	0.06	12.09	15.83	0.06	2.58
CaO	10.38	7.32	0.50	0.84	11.86	21.35	10.48	1.58
Na_2O	2.94	5.15	4.30	6.78	2.82	0.80	4.66	0.51
K_2O	1.25	2.63	4.00	5.24	1.09	n.a.	0.38	0.18
P_2O_5	0.71	0.35	0.10	0.03	n.a.	n.a.	n.a.	n.a.
H_2O^+	1.10	1.49	3.00	n.a.	n.a.	n.a.	n.a.	n.a.
CO_2	3.28	2.85	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
H_2O^-	0.30	0.39	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Total	99.70	100.19	100.15	100.08	97.75	99.88	99.84	88.63

TRACE ELEMENTS (ppm)	WR-4B	PO-1	MPMB	QTZ-SYEN
Rb	25.9	58.3	43.0	111
Sr	805	843	205	29.0
Y	28.7	27.5	n.a.	n.a.
Zr	279	261	532	469
V	234	144	n.a.	5.0
Cr	267	179	n.a.	50.9
Ni	142	51.0	n.a.	12.8
Ba	609	1120	n.a.	181

Note: FeO* = total Fe as FeO
n.a. = not analyzed
WR-4B = camptonite near Stop 6, ref. McHone, 1978b
PO-1 = spessartite at Stop 3, ref. McHone, 1978b
MPMB = trachyte dike at Charlotte, ref. McHone and Corneille, 1980
QTZ-SYEN = average of 5 Cuttingsville quartz syenite analyses, ref. Wood, 1984
PO-1k = kaersutite phenocryst from dike PO-1, ref. McHone, 1978b
PO-1a = augite phenocryst from dike PO-1, ref. McHone, 1978b
PO-1p = plagioclase phenocryst from dike PO-1, ref. McHone, 1978b
PO-1i = ilmenite phenocryst from dike PO-1, ref. McHone, 1978b

FIELD TRIP DESCRIPTION

General Information

The region traversed on this field trip is generally rural and famous for its scenery (Fig. 4). Motels and other amenities are most abundant in Rutland, but that city also has the most unpleasant traffic flow of the area. An excellent campground (in season) is Bomoseen State Park located on the west side of Lake Bomoseen a few miles north of Route

4. Some of the ten sites described in this field guide may be difficult to visit during bad weather (snow, ice) or high stream flow...judge for yourself from the stop descriptions.

U.S.G.S. topographic maps are all available at 1:24,000 scale for the area. The route traverses six 7 1/2' quadrangle maps in the following order: Proctor (dated 1944, Rte. 3 no longer as shown), West Rutland (1972), Poultney (1972), Bomoseen (1944), Poultney again, West Rutland again, Rutland (1980), and Wallingford (1986). We have found The Vermont Atlas and Gazeteer (DeLorme Mapping Co.) to be generally useful and widely available. The U.S.G.S. Planimetric Maps for Ticonderoga (Fig. 4) and Rutland include the sites.

Meeting Place

Directions to start: From Burlington, drive south on Route 7 to Pittsford, then follow Route 3 southwest to Proctor. From southern/eastern/western approaches, turn north onto Rte. 3 off Business Route 4 between Rutland and West Rutland, then north to Proctor. The Vermont Marble Company is just across the bridge in the western part of the village. We have permission to assemble in their lot near the information kiosk.

Along the way, a lunch break can be made in the village of Castleton, which is crossed before and after Stop 5. Gasoline is available in Castleton, but not in many other places along the trip route. Restaurants and gas stations can also be found along Rte. 7 in Rutland before or after the trip. The total travel distance from start to finish is about 60 miles.

ROAD LOG AND SITE DESCRIPTIONS

0.0 miles: **START**

Meet at the Vermont Information kiosk in the parking lot of the Vermont Marble Company, Proctor, Vermont. The marble industry in Vermont is nearly 200 years old and this is the only remaining large-scale operation out of many former businesses. If you have time, take a tour of their showcase museum. Some bargains of polished stone seconds can be had at the adjoining sales yard and we can personally recommend their handsome stone table tops from the store inside.

Head south on Rte. 3 for 1.7 miles.

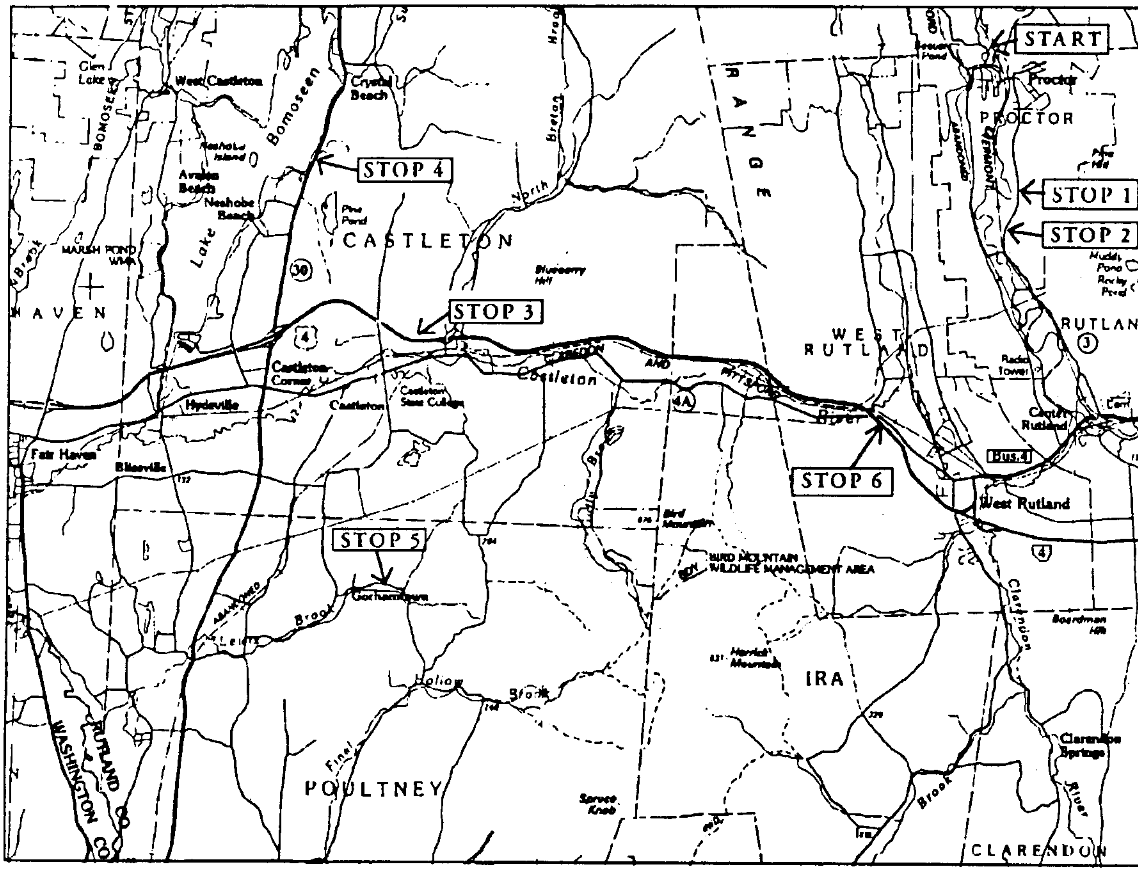

Figure 4. Roads and locations of Stops 1–6, northern Taconics region.

STOP 1: PROCTOR TRACHYTE DIKES

Miles between points: 1.7 Cumulative Mileage: 1.7

Road cut on eastern (left) side of the the highway. Watch for the traffic as you pull onto the wide shoulder on the eastern side, about 3/4 of the way down the cut.

There are two trachyte dikes towards either ends of the cut about 40 m apart; our dike numbers PR-1 to the north and PR-2 to the south. PR-2 is oriented Azimuth 264,55 (when looking along strike, to 264 or 6 degrees south of west, dip is 55 degrees to the right, or north), and is 42 cm wide (measured at chest height). PR-1 is AZ 236,71; 91 cm wide. These two dikes are fairly typical of "bostonite" dikes with their tan color, rhythmic weathering stains, and brittle fracture patterns. Their mineralogy is dominated by anorthoclase, with minor Na-plagioclase, quartz, oxi-biotite, and iron sulfides. The country rock is the Cambrian Dunham dolostone.

The dikes cross a thick, steeply east-dipping quartzite bed that shows much fracturing along the intrusion margins. A few small faults also occur here. Small syn-intrusional steps in the dike walls produced curving flow lines in the dike. Trachyte magma was much more viscous than the magmas that produced the more mafic dikes we will see. Generally the trachyte dikes appear in the vicinity of larger, syenitic plutons or differentiated plutonic complexes. In our discussion of Champlain Valley dikes (McHone and Corneille, 1980), we inferred that trachyte dikes are offshoots of such magma chambers while the widespread lamprophyres are not. Here we are about 25 km northwest of the Cuttingsville complex and so these trachytes are rather isolated...unless there is another pluton, still hidden.

During the National Uranium Resource Evaluation (NURE) of the late 1970's trachyte dikes were found to have gamma radiation levels that are 5 to 10 times higher than surrounding rocks. An analysis of the PR-1 dike shows 9 ppm U_3O_8 and 60 ppm eTh (McHone and Wagener, 1980). These values are high but within the range of 1 to 15 ppm U_3O_8, and up to 105 ppm eTh measured for syenitoids and granitoids of the Cuttingsville and Ascutney plutons. U-enriched trachytes have been mined in Europe.

Continue south on Rte. 3 for 0.4 miles.

STOP 2: PROCTOR MONCHIQUITE DIKE
Miles between points: 0.4 Cumulative Mileage: 2.1

This is another good dolostone road cut on the eastern side of Rte. 3. Dike WR-2 occurs toward the southern end of the outcrop, is oriented AZ 132,85 (a rare NW-SE strike), and is 148 cm wide. This dark gray monchiquite shows "typical" monchiquite features, such as a zone of small pink-white ocelli towards the center, some development of pebbly alteration-texture, and even schistose weathering zones. Note that obvious feldspar is lacking and the fine, granular texture of mafic minerals (augite and olivine). The blocky, cobblestone fracturing evident here is also common among lamprophyres. There is a monchiquite dike west of Orwell village, on strike with WR-2 (also with a NW orientation), but almost 30 km away.

Being basaltic rocks, lamprophyres are not very radioactive. However, two Vermont lamprophyre dikes to the north have 2.2–2.5 ppm U and 9.7–10.5 ppm Th (McHone, 1978b), which is several times higher than the average tholeiite or oceanic basalt. Lamprophyres are relatively rich in other "incompatible" elements as well, probably because they are small melts derived from an "enriched" mantle source.

Continue south on Rte. 3

Miles between points: 1.8 Cumulative Mileage: 3.9

Not a stop, but there is another lamprophyre dike (WR-1) hidden behind brush along this cut. It has a northeasterly strike like others to the south and east.

Miles between points: 0.7 Cumulative Mileage: 4.6

Intersection of Rte. 3 with Business Rte. 4. Turn right, head west on Business Rte 4.

Miles between points: 1.6 Cumulative Mileage: 6.2

Turn right into the interchange with Rte. 4, heading west (bear right where the road splits).

Miles between points: 1.3 Cumulative Mileage: 7.5

For the next mile, there is good exposure along the other lane. Stop 5 is near the northwestern end on the way back.

Miles between points: 6.1 Cumulative Mileage: 13.6

Continue past Exit 5. Outcrop hosting the Castleton dike is along the north side of Rte. 4, 0.8 miles beyond Exit 5.

STOP 3: CASTLETON DIKE

Miles between points: 0.8 Cumulative Mileage: 14.4

This handsome intrusion (PO-1), near the western end of the cut, has a purplish color that contrasts with the surrounding green Mettawee slate. This is a spessartite dike, quite unlike most of the the "alkali lamprophyre" dikes in the Champlain Valley to the north that are camptonite or monchiquite. Its orientation is AZ 076,81; 281 cm. Notice how the northern contact is stepped, in places, 10–20 cm outward from the dike and that, despite its size, thermal metamorphism is not apparent in the Mettawee slate along the margins.

The dike contains phenocrysts of plagioclase and large (up to 1 cm) rounded megacrysts of kaersutite (brown Ti-hornblende). There are also small xenoliths of dark quartzite and gneiss, presumably derived from the Grenvillian basement beneath the Taconic and

Champlain Valley lithologic sequences. The age determination (Table 1) and chemistry (Table 2) make this one of the best-characterized of the northern Taconics dikes. In keeping with its feldspar-rich nature, the chemistry of the rock shows higher concentrations of Si, Al, and Na relative to alkali lamprophyres (Table 2). The kaersutite in this dike (PO-1) is not different from kaersutite in the type-camptonite and in other lamprophyres (Table 2). Similar kaersutite is known from harzburgites and other mantle lithologies that are found as xenoliths in dikes at North Hartland, Vermont and Ayres Cliff, Québec (McHone, 1986). The date of 113 ±4 Ma is a bit older than other NT ages (Table 1).

Continue west on Rte. 4.

Miles between points: 1.5 Cumulative Mileage: 15.9

Exit 4; turn north onto Rte. 30. Follow Rte. 30 north for 2.7 miles. The Stop 4 outcrop is along the east side of the road.

STOP 4: BOMOSEEN DIKE

Miles between points: 2.7 Cumulative Mileage: 18.6

Park in the small lot across from the restaurant just north of the outcrop. Note: because of limited parking and no road shoulder, this stop can only be made safely with a small group, and few vehicles. Walk back (south) about 200 m, past the driveway of a fairly new, contemporary-style house. Exposures are along the east side of the road.

NARROW SHOULDER—WATCH FOR CARS

This dike (our BO-1) was mapped by Fowler (1950, Plate II), who shows it as extending in a WNW direction for about 4 km. The dike runs across the lake just to the north of the slate quarry, visible from here. When we first visited this dike in 1981 it was exposed on both sides of the road. We measured an orientation of AZ 293,89 along its southwestern contact and a width of about 12 meters. More recent construction has since masked much of its northern side, but the remarkably coarse texture of what is left will attest to the great mass of this intrusion.

Thin sections from this dike show a hornblende-bearing plagioclase-rich rock that we are calling spessartite. Groundmass augite is greatly altered to brown minerals. The rock is rather stained and weathered in hand sample, but still looks nothing like the narrower mafic dikes of the NEQ province. Because of its relatively slower cooling, the Bomoseen dike has plutonic textures, properly an alkali diorite, such as occurs at Ascutney or in the Monteregian Hills. There is another "great dike" parallel to BO-1 on the western side of the lake, but farther north, which we have not visited. Fowler (1950, p. 58) reports that dikes in the Castleton 15' quadrangle have no preferred orientation. We note above that dikes farther south and east are predominantly NE–SW trending, while there are several ESE–WNW trends among dikes in this area.

Continue north on Rte. 30 to turn around. Watch blind corner!

Miles between points: 0.5 Cumulative Mileage: 19.1
Crystal Beach; turn around in the parking area to the left; go south on Rte. 30.

Miles between points: 3.3 Cumulative Mileage: 22.4
Cross under Rte. 4.

Miles between points: 0.5 Cumulative Mileage: 22.9
Cross Rte. 4A. Continue south on Rte. 30, unless a lunch stop or break is needed. Castleton village is nearby to the east on Rte. 4A.

Miles between points: 1.1 Cumulative Mileage: 24.0
Turn left (E) at the first crossroad, off Rte. 30.

Miles between points: 0.6 Cumulative Mileage: 24.6
Go straight through the first crossroad and then pass under the old D&H RR bridge, now a foot/bike path.

Miles between points: 0.2 Cumulative Mileage: 24.8
Pond Hill Farm: At the 'T' intersection, just beyond the bridge, turn right (S) onto a gravel road.

Miles between points: 1.3 Cumulative Mileage: 26.1
At the first intersection past the 'T', turn left (E), past a restored farmhouse. Follow this road to a parking area above Lewis Brook.

STOP 5: LEWIS BROOK FLUME DIKE

Miles between points: 0.8 Cumulative Mileage: 26.9

Park at the small pull off along the south side of the road. Note: Visiting this stop is contingent on low stream flow and no ice. As at Stop 4, only a small group can be accommodated. The site requires an athletic scramble down and back up a steep stream bank. Walk towards the west (downstream) to find a place to scramble down. Be careful along this bank.

The Lewis Brook flume is controlled by erosion along a very large, spessartite dike that is oriented AZ 092,90 and is 480 cm wide (our dike PO-2, Fig. 5). This gorge is a little larger than 'The Flume' at Franconia Notch, New Hampshire, but much smaller than Quechee Gorge of Vermont, both of which are also formed by stream erosion along dikes. Such flumes develop both because the mafic dike rock weathers (chemically) faster than the country rock and because fractures along the dike path are more abundant than in the country rock. Often a stream will be "captured" by the dike for a portion of its length, usually starting with a waterfall into the flume and continuing to a less-steep point where the stream can escape. The Lewis Brook dike is well exposed for several hundred meters,

Figure 5. View downstream (west) within the Lewis Brook flume. The dike fills the canyon floor and forms the streambed.

depending partly upon boulders moved by each spring flood. Maps show a very linear stream segment of more than 1 km along and below the flume, but dike exposure is poor downstream because of road fill.

Fowler's (1950) map of the Castleton 15' quadrangle shows several dikes nearby, but misses this one. We found it only because of a sketch in a 19th century reference that describes the wonders of this site (reference since lost to us). It is very rare for a dike to have such a great horizontal exposure; most are only known in vertical segments at road cuts or waterlines. This site has a peaceful if eerie ambience, like being in a large cave.

At the eastern end of the flume, the dike is faulted about 4.5 m in a left lateral sense, so that it disappears not far into the stream bank along the northern side. The fault is partially exposed, and clearly truncates the dike along AZ 140. Lewis Brook follows this fault from upstream, so that the stream turns at this point into the east-west flume not far downstream from a series of falls. The fault must have had lateral or oblique movement, because pure dip-slip would not produce much offset on this near-vertical dike. The fault is not mapped by Fowler (1950) or Zen (1964). The country rock is Mettawee slate.

Retrace path back to Rte. 30 and then north.

Miles between points: 4.0 **Cumulative Mileage: 30.9**
Cross Rte. 4A.

Miles between points: 0.5 **Cumulative Mileage: 31.4**
Turn right onto Rte. 4 East.

Miles between points: 0.2 **Cumulative Mileage: 31.6**
Pass the Castleton uranium occurrence in the slates on your left, a small vein with 150 ppm U_3O_8. Slates and black shales in this region are two to three times more radioactive than are most other rocks, but U concentrations are small.

Continue east along Rte. 4. Stop 6 is a long outcrop along the south side of Rte. 4.

STOP 6: ZEN'S DIKE

Miles between points: 7.5 Cumulative Mileage: 39.1

Pull well off the pavement on the right shoulder of Rte. 4, about a hundred meters past the start of the road cut (Fig. 4). Dike WR-3 (faulted) should be close by (Fig. 6).

This dike is another spessartite with altered augite, but still showing abundant brown hornblende (kaersutite) and plagioclase; it approaches camptonite in its petrography. This dike is oriented AZ 015, 83; is 160 cm wide (our sample code is WR-3), and has been dated (108 ±4 Ma; Zen, 1972; Table 1). We are not far above the Bird Mountain thrust fault of Zen (1964), which here divides the allochthonous Mettawee slate from autochthonous Ira phyllite below.

This dike attracted attention because of its displacement by a prominent normal fault, first described by E-an Zen (1972). The fault is oriented about AZ 019, 66 and has a displacement of about 80 cm. The fault zone has fairly thick gouge, especially near the dike. The fault has not been traced.

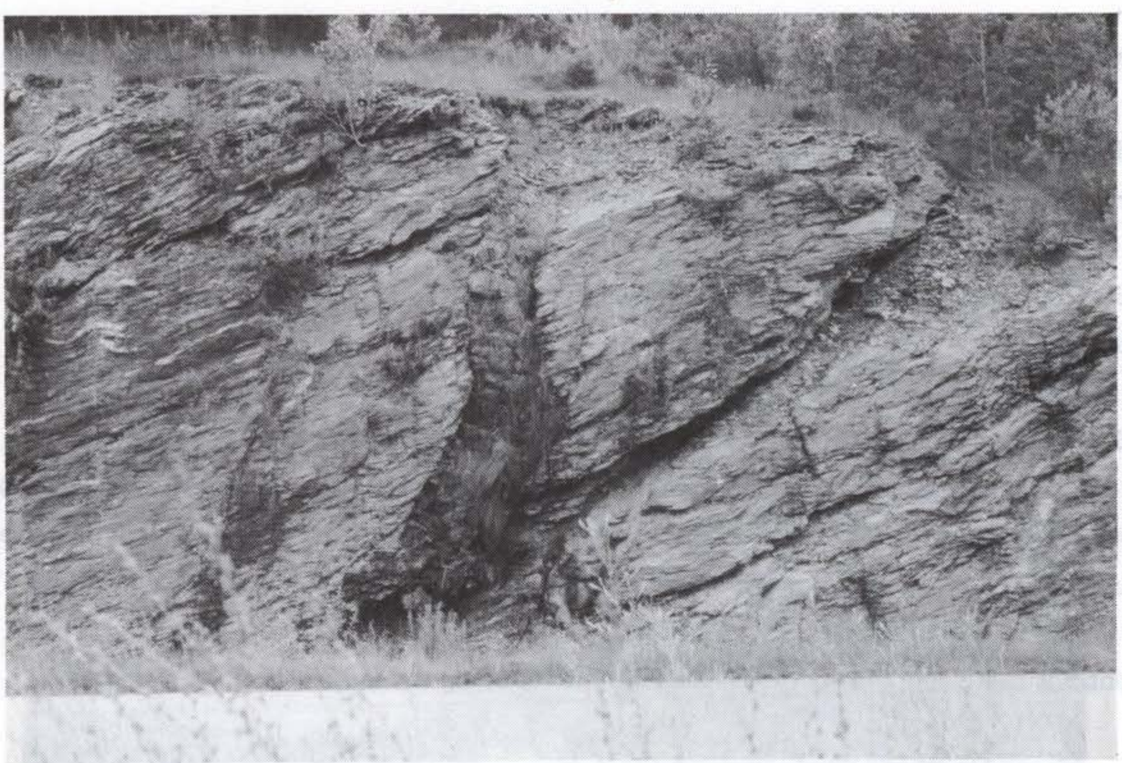

Figure 6. Faulted dike WR-3 exposed along the east bound lane of Rte. 4, West Rutland, Vermont. View is to the southwest.

This road cut extends 1.5 km (0.9 miles) to the southeast and shows at least four more dikes, all of which are similar to WR-3 in type and NE trend. The next dike (WR-4) is about 100 m to the south and was chemically analyzed (Table 2). As is evident in Table 2, camptonite has less silica and sodium than does spessartite, reflecting its Na-rich plagioclase content. There are few such continuous outcrops in the area, and the smaller road cuts happen to show only one or two dikes as a rule. But it appears that dikes often do occur together in small parallel groups, as shown by this cut and as observed in several other areas within the NEQ province.

Continue east on Rte. 4 to Rte. 7.

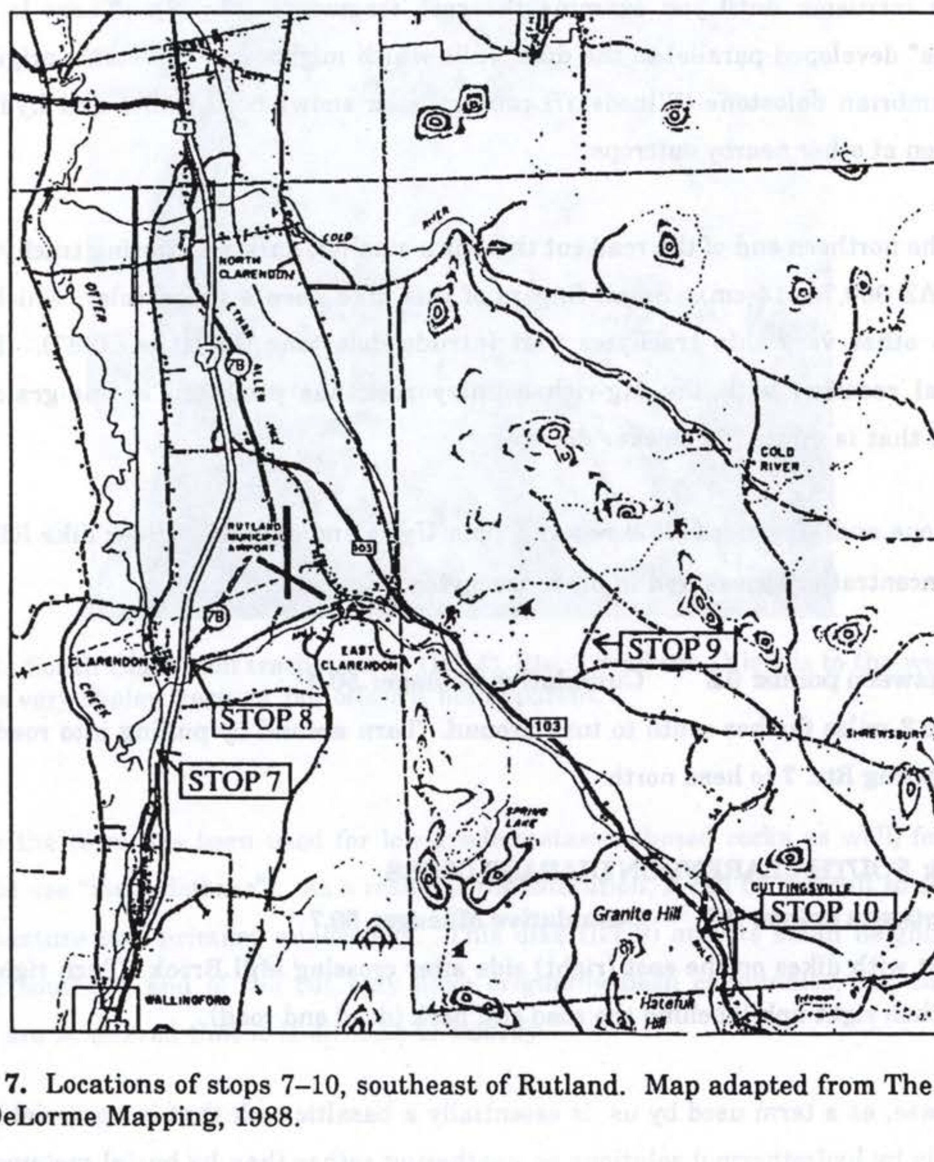

Figure 7. Locations of stops 7–10, southeast of Rutland. Map adapted from The Vermont Atlas, DeLorme Mapping, 1988.

Miles between points: 5.6 Cumulative Mileage: 44.7

Turn right onto Rte. 7, heading south.

Miles between points: 2.7 Cumulative Mileage: 47.4

Pass turnoff to Rte. 103.

STOP 7: SOUTH CLARENDON TRACHYTE DIKES

Miles between points: 2.4 Cumulative Mileage: 49.8

Road cut on west (right) side of Rte. 7. Pull off near southern end of cut (Fig. 7).

This very fractured trachyte dike (RT-4; AZ300,86; 230 cm) is hardly recognizable as an igneous intrusion until you examine the rock fragments (Fig. 8). There is a "shaley cleavage" developed parallel to the dike walls which might have a tectonic origin (Fig. 8). The Cambrian dolostone (Winooski?) country rock shows both faults and hydrothermal alteration at other nearby outcrops.

On the northern end of the road cut there is a smaller, dark-weathering trachyte exposed (RT-1; AZ 060,75; 14 cm). Small fingers of this dike have a green color, which we have seen in other very thin trachytes that intrude dolostone (McHone, 1987). Perhaps a chemical reaction with the Mg-rich country rock has produced a fine-grained green mineral that is diluted in thicker dikes.

McHone and Wagener (1982) report 6 ppm U_3O_8 and 44 ppm eTh for dike RT-4, in line with concentrations measured in other trachytes.

Miles between points: 0.2 Cumulative Mileage: 50.0

Travel 0.2 miles farther south to turn around. Turn around by pulling into road on right, then crossing Rte. 7 to head north.

STOP 8: SOUTH CLARENDON DIABASE DIKES

Miles between points: 0.7 Cumulative Mileage: 50.7

Road cut with dikes on the east (right) side after crossing Mill Brook. Turn right onto old Rte. 7, then right uphill behind the road cut, park (dead end road).

Diabase, as a term used by us, is essentially a basaltic rock that is appreciably altered, generally by hydrothermal solutions or weathering rather than by burial metamorphism

Figure 8. South Clarendon trachyte dike (RT-4), Rte. 7 road cut. View is to the west. The dike has a very shaley fracture, but offset is not apparent.

(although the term has been used for low-grade metamorphosed rocks as well, for which one should use "meta-diabase"). As a result of the alteration, it can be difficult to see much original texture and primary mineralogy. This dike (RT-2) and its small neighbor (RT-3) on the southern end of the cut may have originally been camptonite, but the mafic minerals are so altered that it is difficult to classify.

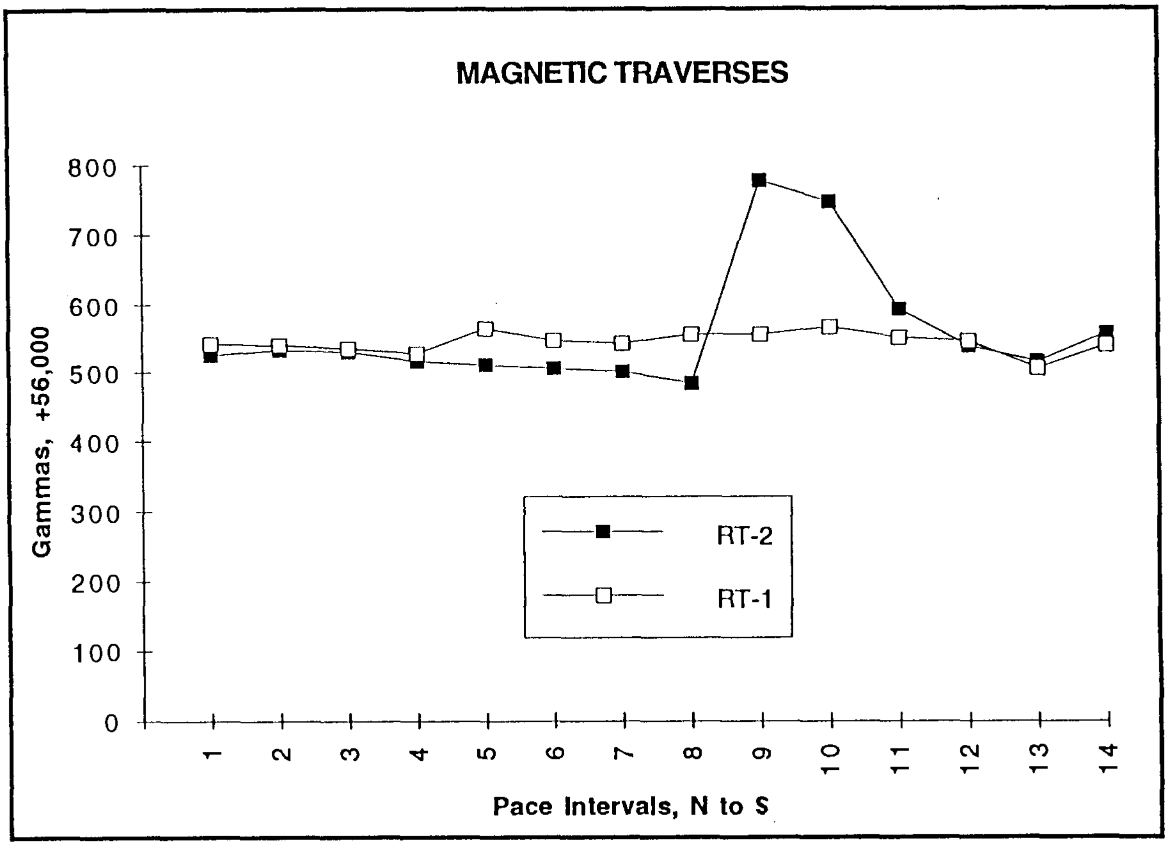

Figure 9. Magnetic anomaly across diabase dike RT-2, traversed along the road ditch. A similar traverse across trachyte dike RT-4 reveals no anomaly.

This E–W striking dike (AZ 084,86; 124 cm wide) has a small but distinct positive magnetic anomaly, which we measured with a portable proton precession magnetometer (Fig. 9). We traced its magnetic expression through the field to the east for several hundred meters, as far as the Mill River. The dike is exposed in the river gorge on this magnetic line, and could possibly be traced much farther. We are interested in discovering whether datable dikes are offset by faults along the Green Mountain front. A similar magnetic traverse of the trachyte dike at Stop 7 (RT-4) revealed no anomaly (Fig. 9).

Turn right (ESE), away from Rte.7 to follow gravel road along the Mill River, to East Clarendon.

Miles between points: 1.6 **Cumulative Mileage: 52.3**

Turn right at stop sign.

Miles between points: 0.5 Cumulative Mileage: 52.8

Turn right (southeast) onto Rte. 103.

Miles between points: 1.1 Cumulative Mileage: 53.9

Turn left onto Maplecrest Farm Road. Go uphill, eastward.

Miles between points: 0.7 Cumulative Mileage: 54.6

Bear left at first 'Y' intersection.

Miles between points: 0.1 Cumulative Mileage: 54.7

Bear left at intersection.

STOP 9: SHREWSBURY INTRUSIVE BRECCIA

Miles between points: 0.1 Cumulative Mileage: 54.8

Stop along road near edge of woods to your right (east). Outcrops are low rock mounds in the woods on the east side of the road.

This site is on private property of Mr. Arthur Pierce, whose residence is at the last intersection. Please do not damage the fence or other property.

The map by Brace (1953) shows this breccia and a few others to the southeast in vague ovals because exposure is poor. The true form of the intrusion may consist of several "pipes." It was certainly a violent intrusion; the outcrops are full of clasts of local metamorphic rocks of the Grenvillian Mt. Holly complex. This site provided samples for Paul Doss, who cataloged many of the lithologies within the breccia (Doss, 1986). Doss looked especially for sedimentary clasts of the Champlain Valley sequence, which would prove an overthrust relationship of the western Green Mountains, but none were identified.

The dike matrix is fairly fresh in a few places between xenoliths, and has a very volcanic, andesitic look in thin section. The date of 101 ±4 Ma (Table 1) is reasonable and indicates little contamination by K or Ar from the country rocks.

Turn around, head back to Rte. 103.

Miles between points: 1.0 **Cumulative Mileage: 55.85**

Turn left (southeast) onto Rte. 103.

STOP 10: CUTTINGSVILLE QUARTZ SYENITE

Miles between points: 3.5 **Cumulative Mileage: 59.3**

Enter village of Cuttingsville and pull into the Ford dealer lot on the right. Park in the back, away from dealer stock. We will walk south about 150 meters along the Mill River if conditions permit (low water is helpful).

Good exposures of biotite quartz syenite member of the Cuttingsville igneous complex are present here. From this point, the Mill River cuts southward across the complex for approximately 2 km exposing other members of the complex and contacts with essexite, non-quartz syenite, and other petrographic varieties. The juxtaposition of nepheline syenite with quartz syenite is an interesting problem in several plutons of New England. Probably, the quartz-bearing magmas were formed by interaction with crustal rocks, while the Si-poor magmas are closer to differentiates of mantle magmas. Many quartz-bearing trachyte dikes in the Champlain valley are very similar chemically to the quartz syenite here at Cuttingsville (more so than samples MPMB and QTZ-SY, Table 2). Although no grabbroic analog of lamprophyre is exposed at Cuttingsville, there is a strong magnetic anomaly over the stock that indicates a mass of gabbro beneath the felsic phanerites.

McHone and Wagener (1982) report 15.5 ppm U_3O_8 and 105 ppm eTh for syenite in this area. This is higher than most "primary" values, even for alkalic rocks. These relatively high concentrations may result from the hydrothermal enrichment of ore minerals that shows here. The sulfide enrichment of the syenite is evident in the stream bank and a large mass of iron-copper sulfides ("copperas") was once mined on the hillside to the northeast. A tunnel is still present on the hillside above the river, no doubt a prospect for metals. Good crystals of pyrite and quartz are easy to find along this stop, at least during low water. Even more interesting may be the native gold that Robinson (1990) has described from this area.

End of trip: Return to Burlington via Rte. 7, or if you have lots of time, we suggest continuing southeast on Rte. 103 to intersect Rte. 100 and other scenic roadways.

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A SUCCESSION OF DEPOSITIONAL ENVIRONMENTS IN THE MID-ORDOVICIAN SECTION AT CROWN POINT, NEW YORK

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Introduction

The Crown Point section, exposed at the Crown Point State Historic Site, is a wonderful place for using fossils, sedimentary textures and structures, and lithologies to interpret changing environments of deposition. Formations exposed in the 120 meter (400 feet) thick section include the Crown Point, Valcour, Orwell, and Glens Falls limestones, deposited between about 458–444 million years ago on the eastern margin of North America. The rocks record the onset of the continent-arc collision known as the Taconic Orogeny. The lower half of the section records deposition on a slowly subsiding, passive continental margin. The upper half records a swift transition to deeper water environments as the continental margin entered the subduction zone. At Crown Point these rocks comprise a homoclinal section dipping about 8 degrees to the west-northwest.

The Middlebury College geology department uses the Crown Point section as a field exercise for both first- and second-year geology students. Their field trip as well as yours consists of a walking tour beginning about 500 meters southeast of Fort Crown Point, heading towards and through the Fort, and then continuing west for about 200 meters along the Lake Champlain shoreline. We will visit most of the lettered stations shown on the index and air photo maps (Figs. 1 and 2). The lettered stations are also shown on the detailed columnar section (Fig. 3) and the student columnar section (Fig. 4). For historical purposes Figure 3 also shows Raymond's (1902) sections B and C; he listed a large number of fossils from those sections. Appendix A (reprinted from Baldwin and Mehrtens, 1985) diagrams the fossils occurring in this section.

Absolutely no collecting or rock hammers are allowed in the Crown Point State Historic Site. Its pristine condition is maintained because the many geologists who have visited over the years have honored this rule. It is useful to circle fossils with chalk for the aid of others.

[1] Brewster Baldwin passed away on July 12, 1992.

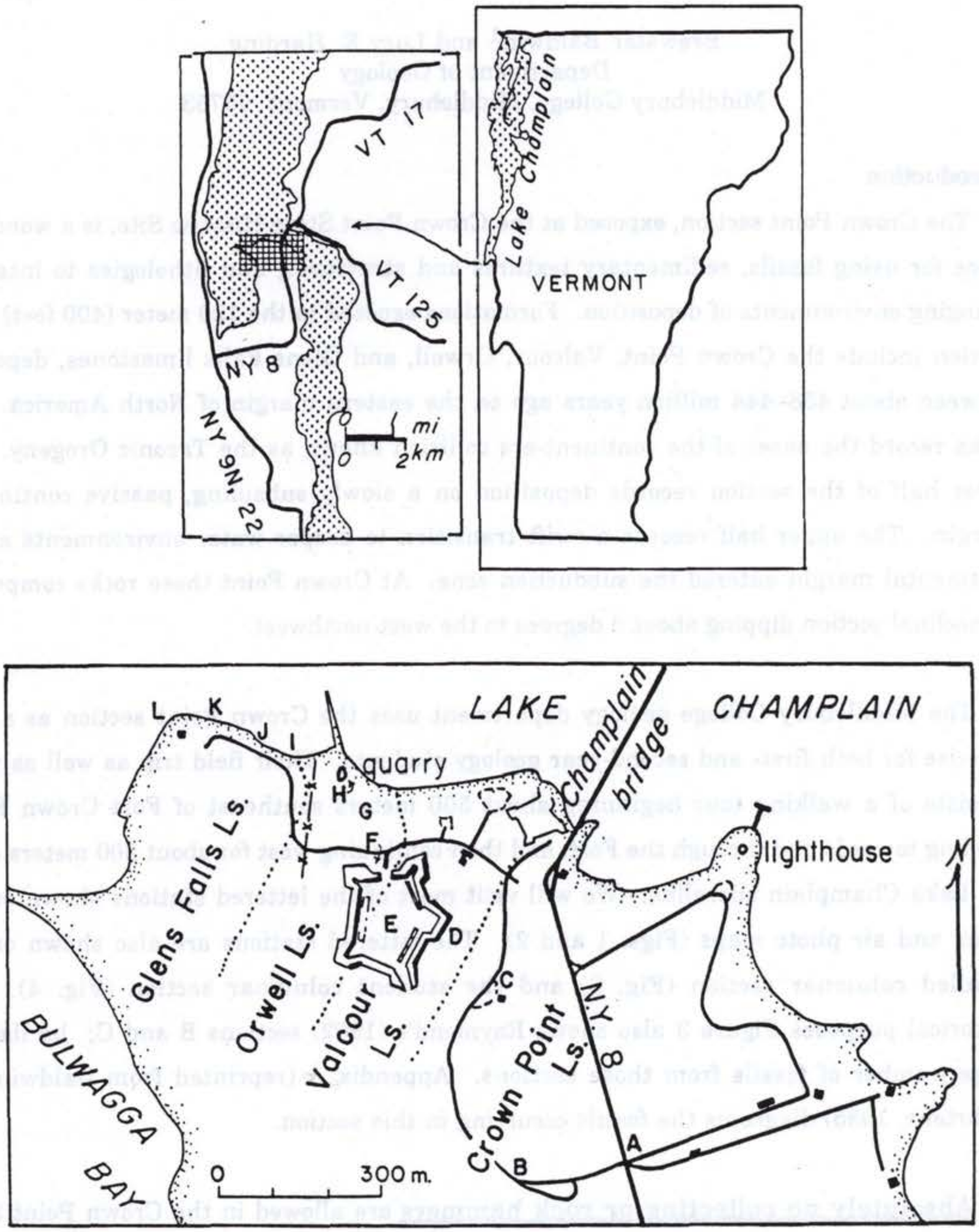

Figure 1. Maps showing location of the Crown Point Historic Site along the western shore of Lake Champlain. Detailed map shows roads, landmarks, and Stations A–L that are described in this field guide. Contacts between the Crown Point, Valcour, Orwell, and Glens Falls limestones are also shown. Compare with Figure 2. Maps after Baldwin (1980); reprinted with permission from *Northeastern Geology.*

Figure 2. (Opposite page) Aerial photograph of Crown Point showing field stations and historic ruins. Scale approximately 1:7,500.


LAKE CHAMPLAIN
N
0
500
1000
feet
pier
lighthouse
Fort St. Frederic
French 1734
light infantry redoubt
A
gate
B
C
D
E
F
G
H
I
J
K
pavilion
visitor center
park
quarry
private
Fort Crown Point
English 1759
CROWN POINT
N.Y.
Bulwaga
Bay

D3

For a more detailed tectonic analysis and petrographic descriptions of this section consult Baldwin (1980, 1982, 1987), and Baldwin and Mehrtens (1985). You may want to follow along with Figures 2 and 4 and Appendix A during the field trip. These figures and the appendix, along with some general instructions, are given to our students at the beginning of their Crown Point adventure! Students are asked to measure a strike and dip at A and then to calculate the thickness of the covered interval from A to B and from A to D using the air photo (Fig. 2). They also sketch a columnar section at A and describe the rocks at A and B. For Stations D and above they use Figure 4 as a guide.

The Crown Point limestone — Stations A, B, and C

A six-meter interval of the Crown Point limestone, belonging to the Chazy group, is exposed at Station A (Figs. 2 and 3). Uneven dolomite laminae (brown- or cream-weathering) are interlayered with lime-mud (gray-weathering) containing abundant rounded grains of calcite and fossil fragments. Bedding is horizontal and locally extensively burrowed. Fossils include abundant *Girvanella* (algae), nautiloids, *Maclurites* (gastropods), along with trilobite fragments and several brachiopods. These strata are interpreted to have formed in a low-energy tidal to subtidal environment. The dolomitic laminae may represent algal mats of the supratidal zone.

Much of the interval between Stations A and B is covered (Fig. 3). The two meter thick exposure of Crown Point limestone at Station B consists of bidirectionally cross-bedded lime-sand with rare grains of quartz. These characteristics suggest oscillating currents and a higher energy environment than at Station A, possibly a beach zone or the lower sand flat region of the tidal zone.

The limestone and dolostone beds near the picnic pavilion make up Station C. These beds, along with strata beneath the rampart between here and Station D, have not been studied but contain *Maclurites* and some trilobite fragments.

The Valcour limestone — Stations D and E

Station D is just outside the easternmost corner of the British fort, up a concrete sidewalk from the Crown Point Visitor Center. Here the Valcour limestone, which forms the upper part of the Chazy group, consists of laminated to thinly-bedded limestones and dolostones cut by channels and scour surfaces. Many strata show low-angle, bidirectional cross-bedding. The exposed face at Station D is about 3 meters high and the outcrop continues in the moat along strike to the south-southwest for about 100 meters to the southern entrance to the fort (Figs. 1 and 2). Several scour surfaces can be followed for some distance along the moat to the south. The lower beds at Station D consist of limestone with fossil fragments and dolomitized burrows. Interbedded limestone and dolostone (dolomite rhombs are visible with a hand lens) form the upper beds at Station D. Near the top of the face are dolostone beds with floating limestone clasts. In thin section the dolostone layers contain pellets. Did the dolomite only replace layers which originally had the greatest porosity? Fossils include *Maclurites* and mounds of gastropod shells. The lack of fossil diversity, cross-bedding, scour surfaces, channels, and dolostone suggest a tidal environment of deposition.

Inside the fort, the flat exposures in front of the southernmost barrack make up Station E (Figs. 1 and 2). The rock is similar to exposures at B, a bidirectionally cross-bedded lime-sand with scattered quartz grains. There are small plano-spiral gastropods and limestone clasts in some of the highest beds. As at B, the sand-sized grains, bimodally oriented cross-beds, and lack of faunal diversity suggest a relatively higher energy environment, possibly a beach, shoal, or lower sand flat in the tidal zone. Above E to the northwest is a low ledge of brown dolostone and next is a 0.6 meter thick ledge of medium- to coarse-grained, almost pebbly quartzite. A three meter thick covered interval separates the Valcour from the overlying Orwell limestone, exposed just outside the northern entrance to the fort. The dolostone and pebbly quartzite suggest that a shallow water, high-energy environment continues through Station E.

The Orwell limestone — Stations F, G, H, and I

The Orwell limestone differs from the Crown Point and Valcour limestones in that it is a thick-bedded, massive, light-gray limestone with areas of secondary black chert (present both as nodules and entire beds). The limestone is predominantly a lime-mud with only a few beds of lime-mud clasts and rare evidence of cross-bedding. Dolostone layers or quartzite are not present in the Orwell, although some burrows have been dolomitized. The massive bedding suggests extensive burrowing. Fossils are diverse and plentiful.

The Lowville beds form the lowermost part of the Orwell limestone (Fig. 3). They are exposed at Station F in a 5 meter high face, just outside the northern fort entrance, and also in the building stone in the southern barrack. This interval consists of very light gray-weathering, massive lime-mud and contains the vertical tubes of the trace fossil *Phytopsis*. The fine grain size suggests a very low-energy environment, possibly a lagoon adjacent to the Valcour tidal zone. Near the top of the face are more typical gray limestones

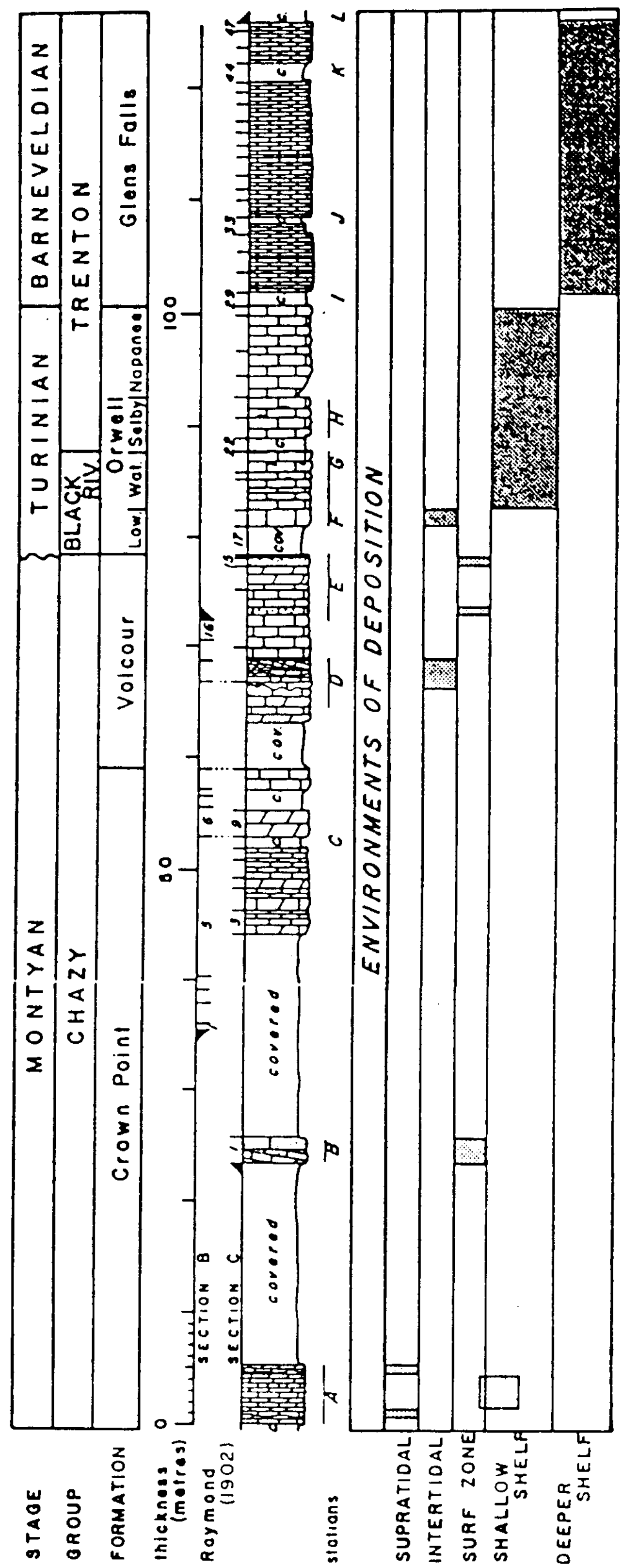

Figure 3. Columnar section for Crown Point. Station letters A–L correspond to the localities shown in Figures 1 and 2. Stratigraphic terminology is from Fisher (1977). The Orwell is divided into the Lowville (Low.), Watertown (Wat.), Selby, and Napanee units. Column from Baldwin (1980); reprinted with permission from *Northeastern Geology*.

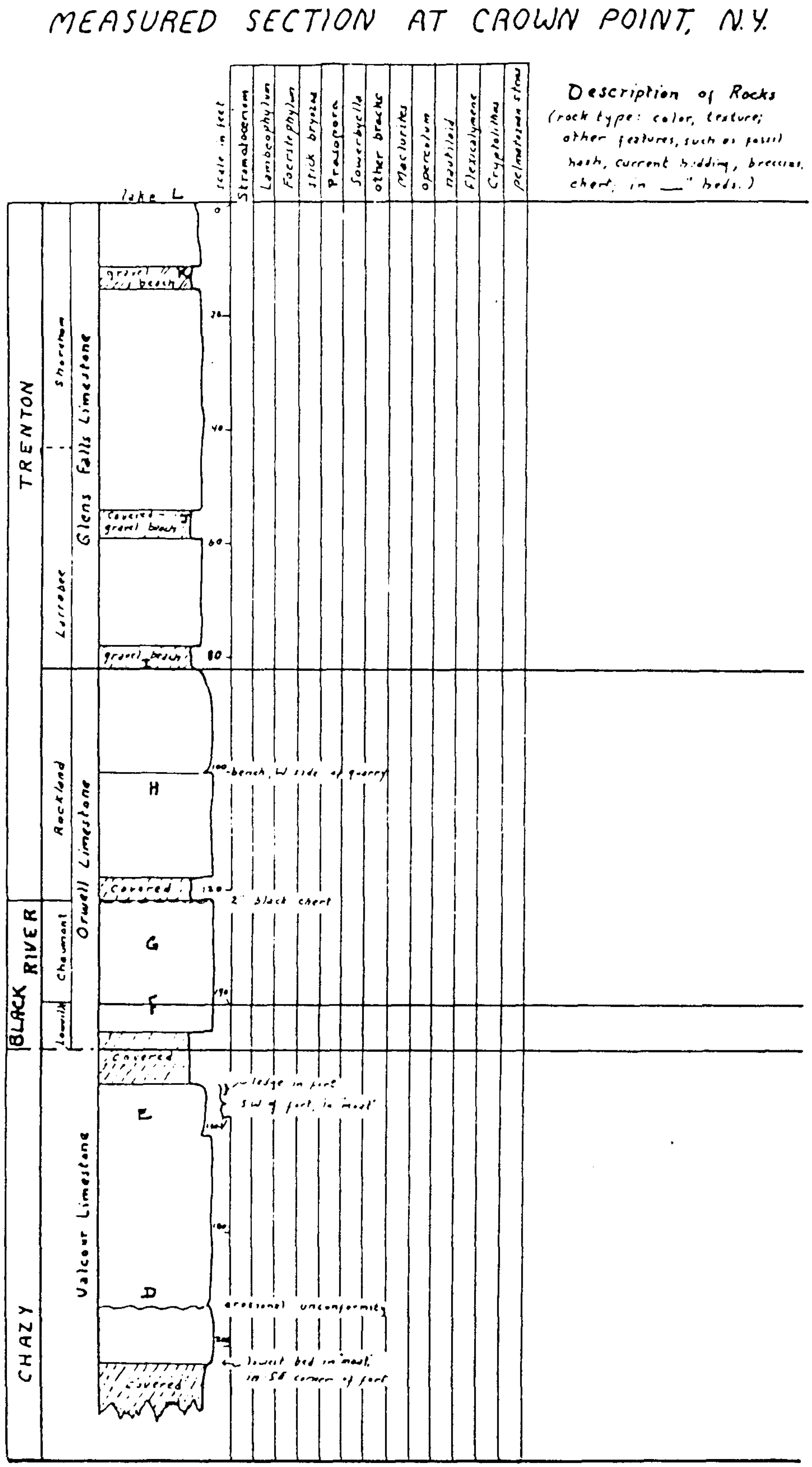

Figure 4. Measured section at Crown Point, N.Y., student's copy. Letters refer to stations on Figures 1 and 2.

of the Orwell and a layer containing black chert nodules. Above the Lowville begins the section of abundant fossils.

Walk north across the road and onto an extensive Orwell ledge towards Station G and Lake Champlain. Along the way, on various levels of the ledge, are exposed clusters of the solitary coral *Lambeophyllum*, nautiloids, the stromatoporoid *Stromatocerium*, gastropods including *Maclurites* , *Lophospira*, and the high-spired *Hormotoma*, the colonial coral *Foerstephyllum* , the trilobite *Isotelus*, brachiopods, some stick bryozoan fragments and a few pelmatozoan stems. Appendix A will help in identification. At the lake shore is a 2 cm thick black chert bed containing beautifully replaced costate brachiopods. This chert bed marks the top of the Black River group (Fig. 3). Note the glacial striations in the shoreline rocks.

Station H is an old quarry located along the lake shore, just west of the chert bed. Large quarry blocks contain many stick bryozoans, the "derby hat" *Prasopora*, pelmatozoan stems, brachiopods, and an occasional *Isotelus*. Just above the western quarry wall is a ledge containing cross-bedded limestone. Walk north from the quarry back to the lake shore just south of the spit. Note the opercula (lids) of *Maclurites* along with *Lambeophyllum*, *Foerstephyllum*, *Stromatocerium*, and the clam *Ambonychia* in the upper beds of the Orwell limestone.

The fine grain size and rare cross-bedding in the Orwell suggest a low-energy depositional environment. The corals and grazing snails suggest deposition within the photic zone under conditions of normal salinity. The depositional environment was probably shallow but subtidal, possibly in a sheltered bay or lagoon with good circulation.

The Glens Falls limestone —Stations I, J, K, and L

Station I is the gravel beach covering the contact between the Orwell limestone and the overlying Glens Falls limestone. Stations J and K are also gravel beaches and K and L are on private property. **Please do not go onto this private property.** After examining the Glens Falls limestone return to the visitor center by walking east along the shoreline or on the road that is just inland.

The gray Glens Falls limestone is immediately distinguishable from the Orwell by its thin-bedded, flaggy outcrop habit. Beds are 5 to 10 cm thick and many are graded or consist of two parts. The lower part contains coarse-grained fossil fragments; the upper part is lime-mud. Near Station K are some thin shaley laminae above the lime-mud. This shale is interpreted to be the precursor of the Trenton shale which overlies the Glens Falls in northern Vermont and Québec. Fossils include diverse brachiopods *(Rafinesquina, Dinorthis, Dalmanella, Sowerbyella)*, fragments of trilobites *(Flexicalymene and Cryptolithus)*, stick bryozoans *(Eridotrypa and Stictopora)*, *Prasopora*, and pelmatozoan stems. *Stromatocerium, Maclurites*, and the corals are absent.

The graded beds and the absence of corals and snails suggest deposition below wave base and the photic zone. The environment of deposition is interpreted to be deep water on an open continental shelf.

Depositional environments and tectonic interpretation

The following discussion comes largely from Baldwin (1987). Fossils, sedimentary textures and structures, and lithologies are used here to reconstruct three sequential depositional environments for the Crown Point section. The Crown Point and Valcour limestones (along with the older Cambrian and Ordovician sediments of west-central Vermont) appear to have been deposited at about sea level on the passive margin of proto-North America (Fig. 5). The Orwell was deposited in slightly deeper water but still within the photic zone, possibly in a quiet lagoon or bay. The Glens Falls limestone was deposited in still deeper water, below the photic zone and wave base on an open continental shelf. Thus the Crown Point section records a deepening of the ocean with time. To the north this deepening of the ocean was recorded by deposition of 1.5 km of marine shale of Trenton age, indicating more rapid subsidence of the shelf. This episode of subsidence reflected in the Crown Point section and the Trenton shale, in which the ocean deepened more rapidly than sediments could accumulate, marks the onset of the Taconic Orogeny.

Figure 5 uses the regional stratigraphy of Cady (1945) to illustrate the subsidence history. The time scale is from Harland and others (1964) and Churkin and others (1977). The thickness at the time of deposition (T_d) of the proto-American platform section and overlying deeper water shale was calculated from the present thickness of strata in west-central Vermont (Baldwin, 1980). This was done by decompacting the present thickness using compaction data of Baldwin and Butler (1985).

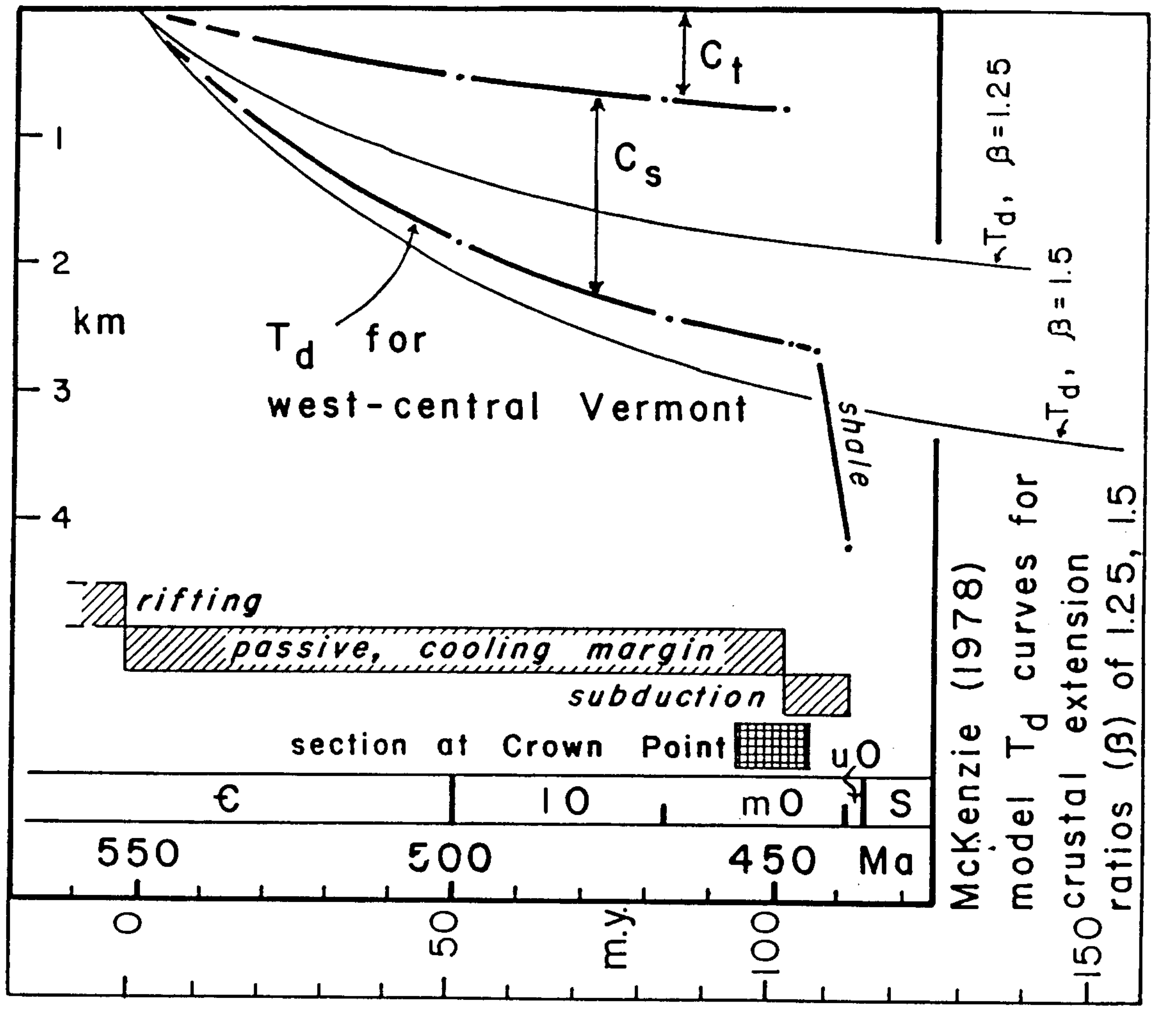

Figure 5. Cumulative thickness (T_d) of the west-central Vermont section, compared with model curves of McKenzie (1978): T_d = deposited thickness; C_s = crustal subsidence due to sedimentary loading; C_t= crustal subsidence due to tectonism; ß = crustal extension ratio. From Baldwin (1987).

Following principles of McKenzie (1978), space for T_d was provided by crustal subsidence due to tectonism (C_t) plus crustal subsidence due to loading (C_s). The magnitude of tectonic subsidence (C_t) depends on how much the continental crust was stretched and thinned during rifting. The rate of C_t was determined by thermal cooling.

One value each of T_d and C_s was calculated for the end of the Cambrian and for the Early Ordovician. An additional value of T_d and C_s was calculated within Mid-Ordovician time, corresponding roughly to the time of deposition of the Chazy Group (Crown Point and Valcour limestones). Values of T_d were also calculated for sediments accumulating later in Mid-Ordovician time corresponding to the end of deposition of the Glens Falls limestone and end of deposition of the Trenton shale. To do this, average densities were calculated for the accumulating section using density values and burial depths from Baldwin and Butler (1985). The sediment density averaged 2.3 g/cm^3. This value, compared with an assumed mantle density of 3.3 g/cm^3, indicates that C_s accounted for 70% of the total subsidence, and the remaining 30% was due to C_t.

The tectonic control of T_d is shown in Figure 5 which compares the west-central Vermont section with model curves of McKenzie (1978) for sediment thickness T_d. The McKenzie curves reflect thermal cooling accompanying crustal extension ratio (ß) of 1.25 and 1.5. The T_d curve for west-central Vermont suggests ß of about 1.4.

Thus, the Cambro-Ordovician section of the ancient continental platform, up through deposition of the Chazy group, fits McKenzie's data for a rifted and then thermally cooled, passive continental margin. Incidentally, Figure 5 also indicates that rifting ended about 550 Ma; this is in general agreement with more elaborate studies of Bond and others (1984). Strata younger than the Chazy group were deposited on a rapidly subsiding continental margin that was being forced to enter the subduction zone at the onset of the Taconic Orogeny.

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APPENDIX A

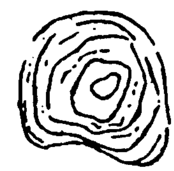
Fossils of the Crown Point, New York, section. From Baldwin and Mehrtens (1985); reprinted with permission from the Vermont Geological Society.

Algae

Girvanella (x 1) "cocktail-onion"; concentric layering, small, abundant.

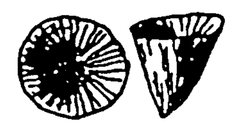
Stromatoporoid

Stromatocerium rugosum (x 1/10) irregular, sub-concentric masses to 30 cm across (looks as though a cow just went by)



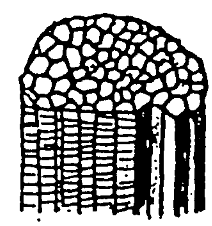
Corals (Coelenterates)

Lambeophyllum profundum (x 1) oldest genus of solitary coral in geologic record; shaped like ice-cream cone; presumably lived in shallow photic zone with algae.



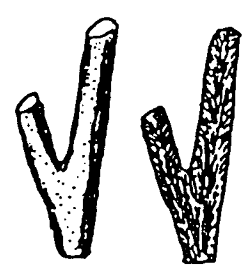
top side

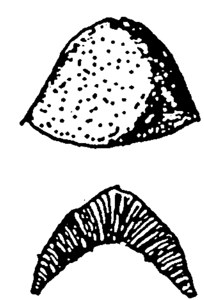
Foerstephyllum wissleri (x 1) honeycomb variety of colonial coral; presumably lived in shallow photic zone with algae; this species named after Professor Benjamin Wissler, Middlebury College. Sketch shows top (honeycomb) and side (columns; cut-away columns with interior tabulae).



Bryozoans (x 2, x 1)

Colonial coral-like animals with microscopic tubes for each individual; surface has tiny pin-hole apertures, visible with hand lens; interior has hair-like fibers. Bryozoans are filter feeders, needing clear water.





(shape and cross-section) "stick" Prasopora

Brachipods (2 shells) (x 1, 1/2)

Varieties are wide or narrow-hinged, smooth or marked with radiating lines or ribs; some show concentric growth lines. Diverse assemblage of brachiopods indicates open marine conditions.

a. 

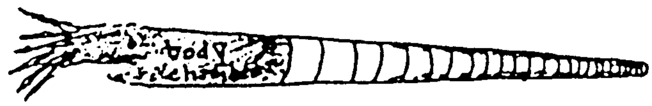
b. 

c. 

a. Sowerbyella; b. orthid; c. rhynchonellid

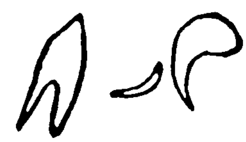
APPENDIX A (Continued)

Nautiloids (x 1/5, 1/10)
Belong to Mollusca, Cephalopoda. Related to squids; swimmers, predators. Chambers behind body chamber are preserved.

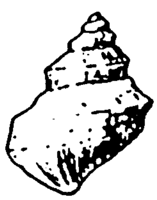


Gastropods (Mollusca; snails)
Maclurites is flat-coiled; probably grazed on algae in very shallow water. Its operculum is a thick-walled "hand"-shaped lid. Many snails have moderate to high spires (coils).

x 1/5
Maclurites

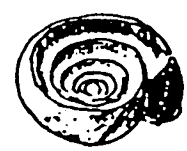


x 1/3
opercula



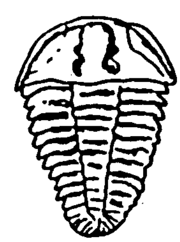
x 1
Lophospira

x 1
Hormotoma

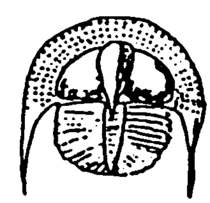


x 1/2
Lecanospira

Trilobites (x 1)
Belong to Arthropoda. Like locusts, they molted, so one animal could leave many exoskeleton fragments. The cephalon (head region) is most important part for identification.

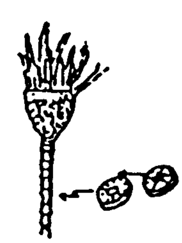


Flexicalymene



Cryptolithus

Pelmatozoans (x 1/2)
Belong to Echinodermata. Pelmatozoan stem (of crinoid, cystoid, etc.) comes apart in the sediment, leaving disks with round or star-shaped holes; cup with the living chamber seldom found here. Echinoderms live only in water of normal salinity.



THE ALTONA FLAT ROCK JACK PINE BARRENS: A LEGACY OF FIRE AND ICE

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Introduction

Altona Flat Rock is the largest (approximately 32 km^2) of a discontinuous, 5-kilometer wide belt of bare sandstone areas that extend approximately 30 km southeastward into the Champlain Valley from Covey Hill, near Hemmingford, Québec (Fig. 1). Created by catastrophic floods from the drainage of glacial Lake Iroquois and younger post-Iroquois proglacial lakes in the St. Lawrence Lowland more than 12,000 years ago (Denny, 1974; Clark and Karrow, 1984; Pair et al., 1988), the exposed sandstone today provides habitat for one of the largest jack pine *(Pinus banksiana)* barrens in the eastern United States. The relatively low-diversity jack pine community is maintained by fire, which has an important role in ecosystem regeneration in this nutrient-poor, drought-prone environment.

We will visit several sites in the southeastern portion of Altona Flat Rock on property owned by the William H. Miner Agricultural Research Institute. The area contains the remains of the "Million-Dollar Dam," part of a failed hydroelectric project begun by William Miner in 1910. This trip will address on-going efforts to understand the linkages between the hydrogeology and ecology of the jack pine barrens and will document the recent history of anthropogenic development in this unique ecosystem.

Geological Setting

Physiography

Altona Flat Rock, located in the northwestern Champlain Valley (Fig. 1), is entirely underlain by flat-lying Potsdam Sandstone (Cambrian). The lithology of the Potsdam ranges from orange-pink to pale red, very coarse to medium-grained, cross-laminated arkose with quartzitic green shale and conglomeratic interbeds to pinkish gray to very pale orange, well sorted, fine to medium-grained quartz sandstone (Fisher, 1968). The exposed rock surface slopes north and east from an elevation of more than 300 meters a.s.l. (above sea level) to below 200 meters a.s.l. where it passes beneath surficial deposits in the Champlain Lowland (Denny, 1974). The sloping surface is broken into a series of stairlike bedrock treads separated by risers that range from a few decimeters to tens of meters in height (Fig. 2). The tread surfaces have little local relief except near stream channels and risers. The eroded edges of truncated trough cross-beds, ripple marks, and solution pits are common minor surface features. Shoreline deposits from the highstand of glacial Lake Vermont (Fort Ann Stage) (Chapman, 1937; Denny, 1970, 1974) lap onto the northern and eastern margins of Altona Flat Rock.

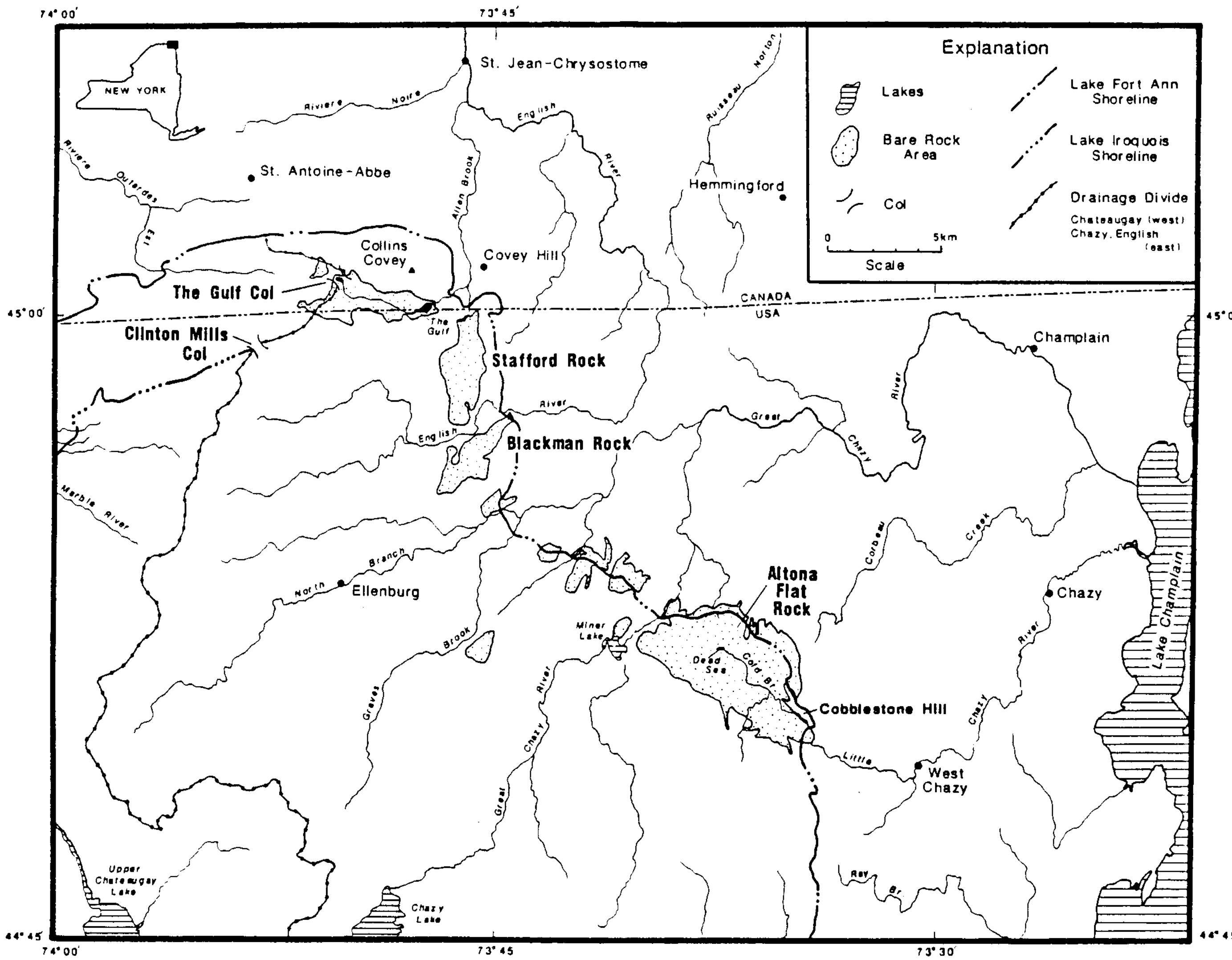

Figure 1. Location map showing the principal bare rock areas east of the divide between the Chateaugay (west) and Chazy and English (east) river watersheds in northeastern New York and adjacent parts of Canada (from Woodworth, 1905a; Denny, 1974; LaSalle, 1985).

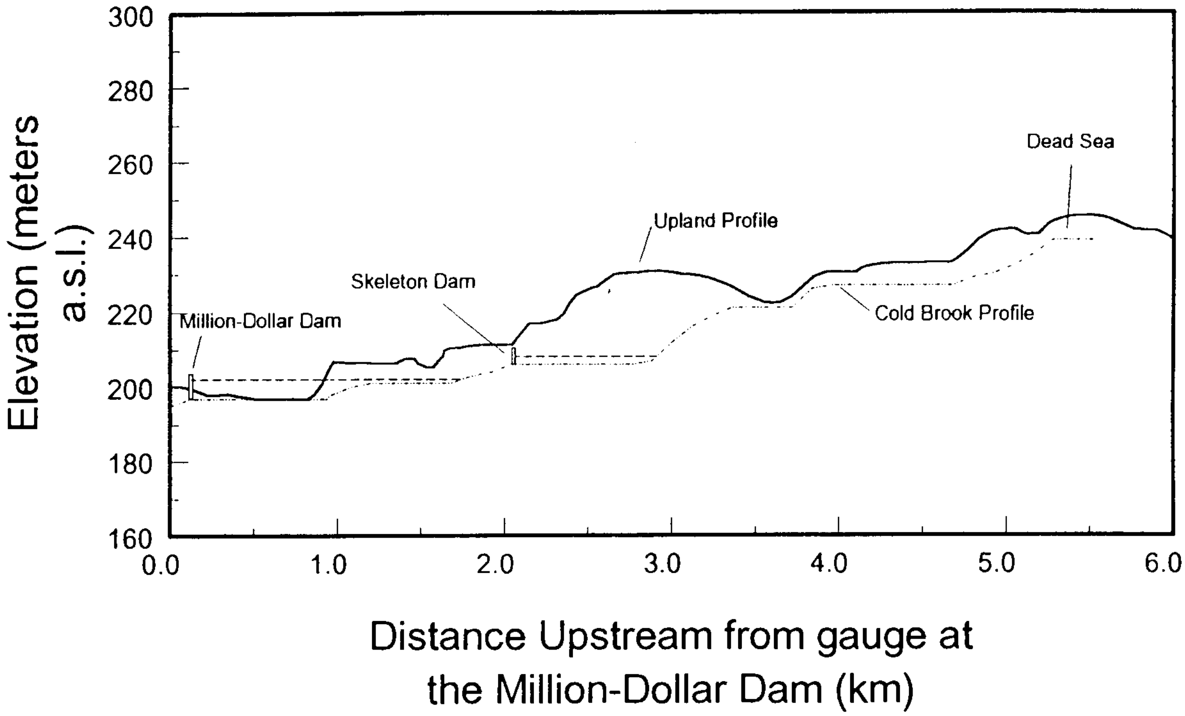

Figure 2. Topographic profile of Cold Brook and adjacent uplands on Altona Flat Rock showing the location of the Million-Dollar and Skeleton Dams and the approximate design pool elevations of their respective reservoirs. The upland profile represents the maximum land surface elevation within 0.5 kilometers of a line oriented N40°W through the Cold Brook Valley.

The central portion of Altona Flat Rock is drained by Cold Brook, a principal headwater tributary of the Little Chazy River that originates at the Dead Sea (Fig. 1). Cold Brook is an underfit stream that occupies a bedrock channel that may locally be more than 200 meters wide and 25 meters deep. The greatest channel incision generally occurs where the stream cuts across prominent southeast-facing bedrock risers. The generally southeastward drainage of Cold Brook is characterized by a subtle rectangular channel pattern that is probably related to bedrock fracture patterns.

Cobblestone Hill is a conspicuous, elongate ridge on the northern flank of Cold Brook at the southeastern margin of Altona Flat Rock. The ridge is more than 15 meters high, 500 m wide, and 2.5 kilometers long and is composed of angular boulders, almost exclusively Potsdam Sandstone, that range from 0.5 to 3 meters in diameter. The average size of surface boulders decreases to the southeast. Boulder and gravel terraces on the northeast flank of Cobblestone Hill represent beach ridges formed in Lake Vermont (Woodworth, 1905a; Chapman, 1937; Denny, 1974).

Geological History

The exposure of large areas of sandstone in the northwestern Champlain Lowland occurred more than 12,000 years before present by the erosional effects of ice-marginal streams during the catastrophic drainage of glacial Lake Iroquois and younger post-Iroquois lakes (Woodworth, 1905a, 1905b; Coleman, 1937; Denny, 1974; Clark and Karrow, 1984; Pair et al., 1988). Lake Iroquois occupied the Ontario Lowland and drained eastward across an outlet threshold near Rome in the western Mohawk Lowland (Coleman, 1937). The lake expanded northeastward into the St. Lawrence Lowland during deglaciation between the Adirondack Uplands to the south and the waning Laurentide Ice Sheet margin to the north. The former water level probably stood at a present elevation between 329 and 332 meters a.s.l. near Covey Hill, Québec (Fig. 1) (Denny, 1974; Clark and Karrow, 1984; Pair et al., 1988).

Eastward drainage of Lake Iroquois began as lower outlets were exhumed along the drainage divide between the Champlain and St. Lawrence drainage systems southwest of Covey Hill during ice recession. The initial drainage may have occurred through a channel approximately 1 km north of Clinton Mills (Fig. 1) that was controlled by a threshold between 329 and 332 meters a.s.l. (Clark and Karrow, 1984). The falling levels of proglacial lakes in the St. Lawrence and Ontario lowlands temporarily stabilized at the glacial Lake Frontenac level (308-311meters a.s.l.) as the ice margin receded northward and the col at The Gulf was uncovered (Clark and Karrow, 1984; Pair et al., 1988). Outflow from these lakes was directed southeastward along the ice margin where it crossed the English, North Branch and Great Chazy watersheds before eventually emptying into Lake Fort Ann which occupied the Champlain Lowland at an elevation between 225 and 228 meters a.s.l. (Denny, 1974). The outflow streams stripped large areas of their surficial cover and cut deep bedrock channels and plunge pools into the Potsdam Sandstone (Fig. 1), *e.g.* The Gulf (MacClintock and Terasme, 1960) and the Dead Sea (Woodworth, 1905a; Denny, 1974). The most intense scour (e.g. Stafford Rock, Blackman Rock, and Altona Flat Rock) generally occurred on major watershed divides. Cobblestone Hill (Fig. 1) is an accumulation of bouldery debris washed from the exposed rock areas by glacial lake outflow floods (Woodworth, 1905a; Denny, 1974).

The scour of the areas southeast of the St. Lawrence-Champlain divide continued as ice recession caused the drainage of Lake Frontenac around the northern flank of Covey Hill. Denny (1974) suggested that the ice margin may have oscillated in the area around Covey Hill causing the lakes in the eastern St. Lawrence Lowland to refill and empty several times. The lake-drainage episodes ended when the ice front receded from the northern flank of Covey Hill for the last time and the proglacial lake in the St. Lawrence Lowland, Lake Belleville, was lowered to the level of Lake Fort Ann in the Champlain Lowland (Pair et al., 1988).

The Jack Pine Barrens

The large areas of exposed bedrock on Altona Flat Rock (Fig. 3) provide habitat for one of the largest jack pine *(Pinus banksiana)* barrens in the eastern United States (Woehr, 1980). The Natural Heritage Program (NHP) classifies this barrens as a sandstone-pavement barrens. With fewer than 20 known sites in the world the NHP ranks the jack pine sandstone pavement barrens as a globally rare ecosystem (Reschke, 1990). The William H. Miner Agricultural Research Institute owns almost 1000 ha (hectares) of barrens on Altona Flat Rock and an additional 600 ha is owned by New York State.

Figure 3. Photograph of "typical" jack pine barrens on Altona Flat Rock.

Jack pine is a relatively short-lived (<150 years), shade-intolerant, boreal species that has maintained a relic community at Altona Flat Rock because of its adaptations to fire and ability to survive in an area with thin (or absent), nutrient-poor soils (Fig. 3). The Altona Flat Rock pine barrens is near the southern limit of the present natural range of jack pine (Burns and Honkala, 1990; Harlow, et al., 1991).

The relatively low species diversity in the barrens reflects low seasonal water availability and the thin, nutrient-poor soils on Altona Flat Rock. The barrens consists essentially of a single tree species, jack pine, with virtually no subcanopy or understory trees. The understory shrubs are predominantly lowbush blueberry *(Vaccinium angustifolium)*, black huckleberry *(Gaylussacia baccata)*, black chokeberry *(Pyrus melanocarpa)*, sweetfern *(Comptonia peregrina)*, and sheep laurel *(Kalmia angustifolia)*. Ground cover is primarily reindeer lichen *(Cladonia rangiferina)*, haircap moss *(Polytrichum commune)*, bracken fern *(Pteridium aquilinum)*, and *Sphagnum spp.* (Stergas and Adams, 1989).

Jack pine requires periodic crown fires for successful regeneration to occur (Ahlgren and Ahlgren, 1960; Cayford, 1971; Rowe and Scotter, 1973; Cayford and McRae, 1983; Rouse, 1986). Fire releases seeds from serotinous cones stored in the jack pine canopy, prepares a nutrient-rich ash seedbed, and reduces competition for the young seedlings. Since this barrens is a fire-dependent ecosystem, fire exclusion will ultimately cause the local extinction of jack pine and the deterioration of the major heath plants, blueberry and huckleberry.

The pine barrens community is well adapted for the Altona Flat Rock environment. Mean annual precipitation from meteorological records for a 27-year period between July, 1963 to August, 1992 at the Miner Institute in Chazy, New York is approximately 80 cm. Mean monthly air temperature ranges from -11^{o}C in January to 20^{o} C in July (Stergas and Adams, 1989). Summer air temperature in bare rock areas, however, may be as much as 16^{o}C higher than in the surrounding areas, and midday temperatures commonly exceed 38^{o}C (Woehr, 1980). Preliminary data from observation wells on Altona Flat Rock indicate that, in many places, the water table lies well below the depth of root penetration.

The combined effects of anomalously high summer air temperature, low seasonal water availability, and flammable foliage produce a fire-prone environment at Altona Flat Rock. There have been four stand-replacing wildfires at Altona Flat Rock during this century (1919, 1940, 1957 and 1965). The oldest jack pine stand at Altona Flat Rock (ca. 73 years) is beginning to show signs of decline. Nearly 40 percent of the trees in this stand (1919 burn area) are dead (Hawver, 1992). The accumulation of dead tree biomass increases the probability of another fire in this stand. A fire management plan, that includes both planned-ignition and natural-ignition fires, is needed for the entire barrens.

The exclusion of fire from a fire-dependent ecosystem such as the jack pine barrens can result in a loss of biodiversity. At the landscape level, the complex mosaic of habitat types created and maintained by fire can benefit many species of plants and animals. The use of prescribed fire to preserve biodiversity in the barrens, however, will require several more years of ecological research and careful planning and coordination with local fire-control agencies and the New York Department of Environmental Conservation.

The Altona Flat Rock Hydroelectric Project

In the summer of 1910, William Miner, ignoring the advice of his engineers, began construction of a hydroelectric dam and generating station on the southeastern margin of Altona Flat Rock (Gooley, 1980). By the time of its completion in March, 1913, the concrete dam, known locally as the "Million-Dollar Dam," had a maximum height of over 10 meters and stretched more than 700 meters across the Cold Brook valley (Figs. 4 and 5). The design capacity of the reservoir was more than 3.5 million cubic meters. A second dam, the Skeleton Dam (Gooley, 1980), was constructed upstream to provide supplemental flow to the main impoundment (Fig. 2).

The dam and generating station were completed in 1913 but it took almost two years to fill the reservoir to near capacity. The inadequate flow of Cold Brook and ground water seepage through Cobblestone Hill, which formed the eastern flank of the reservoir, proved to be major design flaws. At one point, seepage beneath the dam was so great that it caused severe damage to the Stephen LaPierre residence, approximately 600 meters east of the dam (Gooley, 1980). A 10 to 15 cm thick layer of concrete grout was spread over more than 100,000 m^2 along the southwestern flank of Cobblestone Hill to mitigate the seepage loss (Figs. 4 and 6). A deep trench was excavated in the sandstone at the base of Cobblestone Hill behind the dam. A grout curtain was poured into this trench, presumably sealing the northeastern flank of the reservoir. The grouting effort was partially successful and the power generating plant began operation on January 21, 1915, more than four years from the beginning of the project (Gooley, 1980). The power plant produced electricity intermittently for seven years before mechanical problems forced the abandonment of the project.

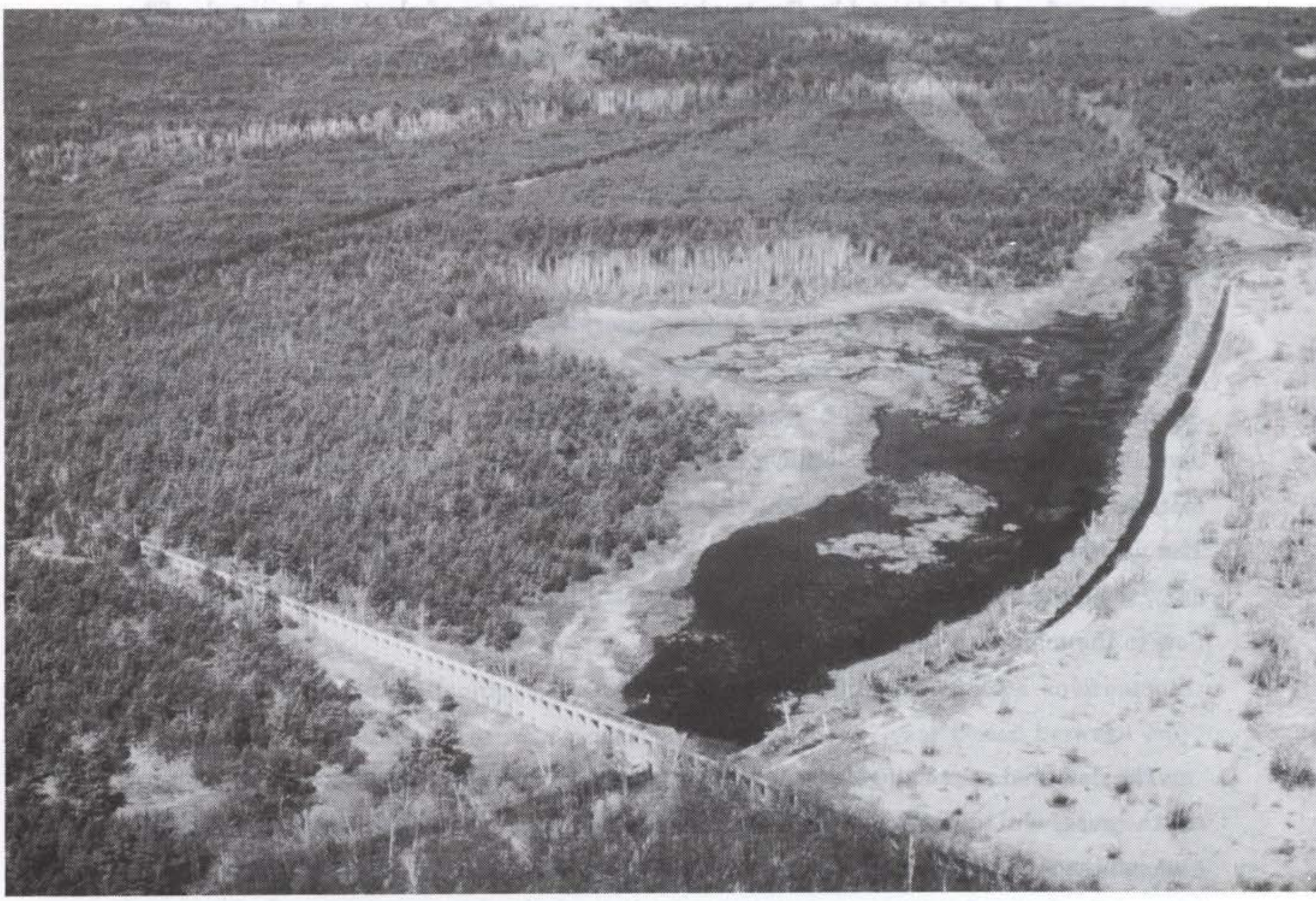

Figure 4. Oblique aerial photograph showing the Million-Dollar Dam and the northwestern flank of Cobblestone Hill (lower right) where it was covered with a concrete veneer to reduce seepage from the former reservoir. The grout-curtain trench can be seen on the right side of the photo.

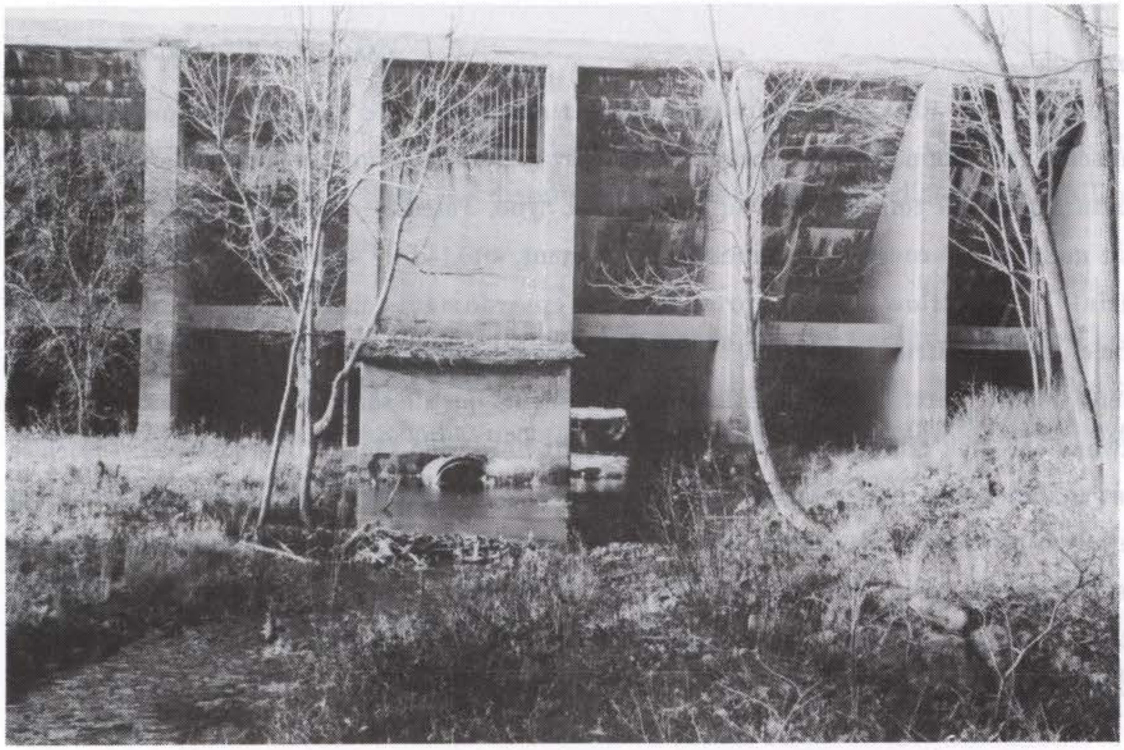

Figure 5. The Million-Dollar Dam looking northwest from the reservoir outlet.

Figure 6. The Scarpit.

Summary

Altona Flat Rock illustrates the impact of glacial and post-glacial processes on landscape development and contemporary ecosystem-level processes. The geology at Altona Flat Rock consists of a single bedrock type, Potsdam Sandstone, with relatively little mineral-soil cover. The sandstone pavement, created by erosion associated with late glacial lake-outflow floods, provides an environment characterized by extreme deficiencies in nutrients and soil moisture. Jack pine and its associated heath plants are among the few native species that can survive in this hostile setting. The abrupt transition between the jack pine barrens and the adjacent hardwood forests demonstrates the close relationship between vegetation distribution and hydrogeology. The combined effects of the harsh physical environment and its associated vegetation create an ecosystem that is adapted to and maintained by periodic fire. A fire-management program, based upon a detailed study of ecosystem dynamics and function, is needed if the uniqueness of the Altona Flat Rock jack pine barrens is to be preserved.

Acknowledgments

The authors would like to acknowledge the Center for Earth and Environmental Science and the Applied Environmental Science Program at SUNY Plattsburgh and the W.H. Miner Agricultural Research Institute for their support of our research and instructional efforts at Altona Flat Rock. Jamie Shanley, Jon Denner (U.S. Geological Survey, Montpelier, Vermont) and Marc Hult (U.S. Geological Survey, Bloomington, Minnesota) provided valuable technical assistance for the installation of the monitoring well network and the stream gaging station. Special thanks are also extended to Michael Parsons of Michael Parsons Well Drilling Company, who generously donated drilling services for well installation. Finally, we would like to thank our students Neil Gifford, Chris Lassell, and Kortney Brewster who have enthusiastically helped with equipment installation and monitoring during the early stages of our project.

FIELD TRIP DESCRIPTION

The road log begins at the Lake Champlain Ferry dock on Cumberland Head, Plattsburgh, New York. Road log distances are presented in English units. All other measurement are in SI units.

Persons using this log in the future should be aware that the field trip stops are located on private property that is owned and patrolled by the William H. Miner Agricultural Research Institute. A permit must be obtained from the Miner Institute (518-846-8020) to access this property.

START

Miles between points: 0.0 Cumulative Mileage: 0.0

Assemble in the Lake Champlain Ferry parking lot at Cumberland Head, Plattsburgh, New York.

Proceed west on N.Y. Route 314 toward Plattsburgh.

Miles between points: 0.8 Cumulative Mileage: 0.8

Route 314 bears sharply right (to northwest).

Miles between points: 3.2 Cumulative Mileage: 4.0

Intersection of N.Y. Route 314 and U.S. Route 9. Proceed straight ahead through the intersection to the northbound entrance ramp of Interstate 87 (Adirondack Northway).

Miles between points: 0.1 Cumulative Mileage: 4.1

Northbound entrance ramp. Turn right (north) and proceed to Interchange 40 (Spellman Road) in Beekmantown.

Miles between points: 3.8 Cumulative Mileage: 7.9

Exit ramp at Interchange 40. Exit right and proceed to Spellman Road.

Miles between points: 0.1 Cumulative Mileage: 8.0

Intersection of Northway exit ramp and Spellman Road. Turn left and proceed west to Beekmantown Corners.

Miles between points: 2.7 Cumulative Mileage: 10.7

Intersection of Spellman Road and U.S. Route 22. Turn right and proceed north on U.S. Route 22.

Miles between points: 3.4 Cumulative Mileage: 14.1

Intersection of U.S. Route 22, N.Y. Route 348, and West Church Street in West Chazy. Turn left and proceed west on West Church Street.

Miles between points: 0.7 Cumulative Mileage: 14.8

Intersection of West Church Street, Parker Road, and O'Neil Road. Bear left then right to remain on West Church Street.

Miles between points: 0.8 Cumulative Mileage: 15.6

Intersection of West Church Street and Barnaby Road. Turn right and proceed north on Barnaby Road.

Miles between points: 1.0 Cumulative Mileage: 16.6

Barnaby Road changes to a gravel surface at the farm just north of Slosson Road intersection.

STOP 1: LAKE FORT ANN BEACH RIDGES

Miles between points: 1.0 Cumulative Mileage: 17.6

Park at the gate at the entrance of the Miner Institute property and continue northward on foot along Barnaby Road approximately 100 meters (320 ft). Turn left into woods and proceed west for 150 to 200 meters (500–750 ft) up the eastern flank of Cobblestone Hill (Fig. 7). The beach ridges occur at elevations between 175 and 205 meters (580 and 670 ft) above sea level (Denny, 1974).

The beach ridges on Cobblestone Hill were first described by Woodworth (1905a) and later by Denny (1974). The beaches consist predominantly of moderately rounded to well rounded, pebble to cobble gravel that is deposited in multiple, elongate, low-relief ridges that extend along the northern and eastern flanks of Cobblestone Hill between 175 and 205 meters a.s.l. (Fig. 7). Individual deposits are typically as much as 1 meter high and 30 meters wide, and often extend laterally for more than 400 meters (Denny, 1974). The

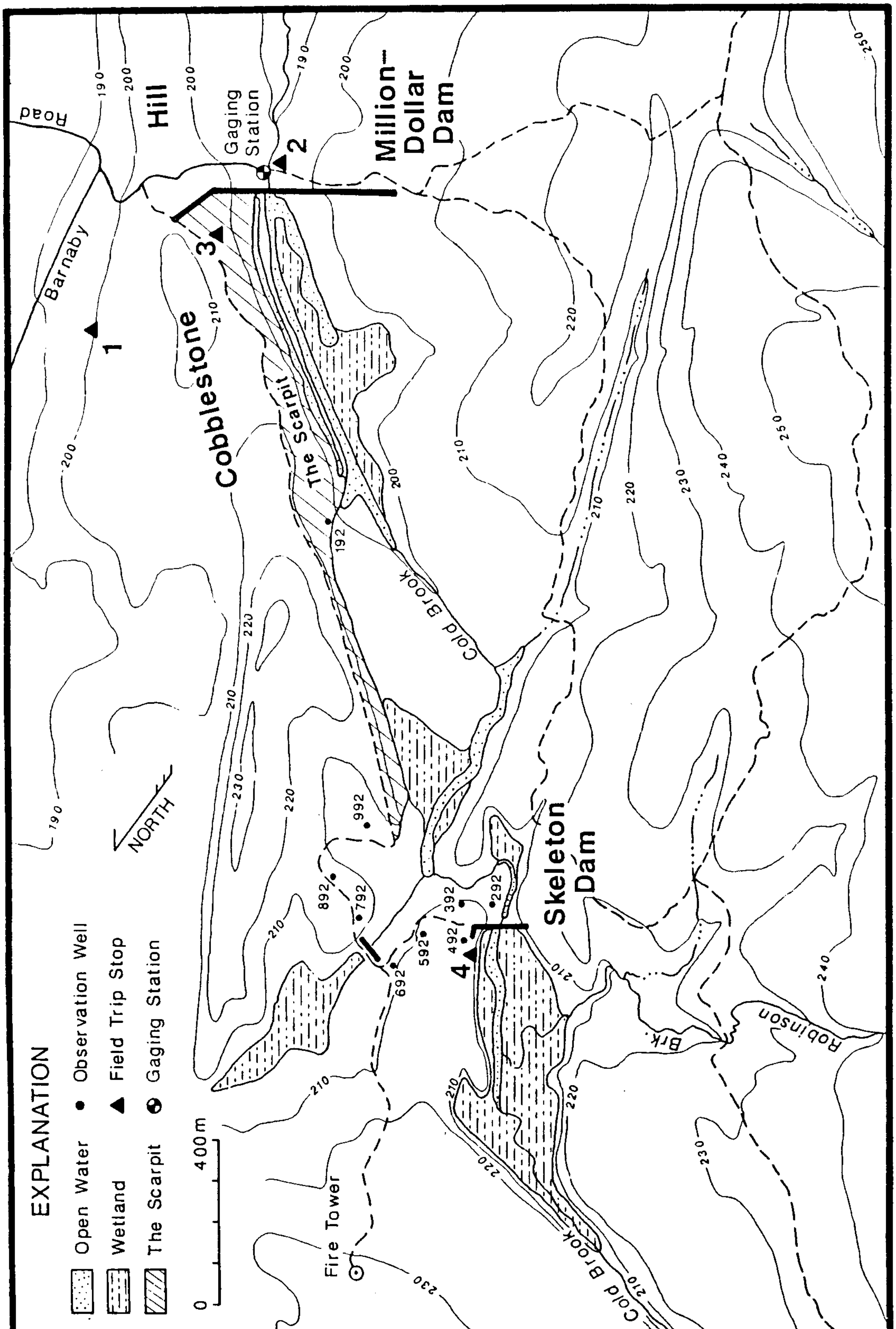

Figure 7. Topographic map of the southeastern portion of Altona Flat Rock showing locations referred to in text. (Topographic base from West Chazy Quadrangle, U.S. Geological Survey 7.5-Minute Series)

gravel is almost exclusively composed of Potsdam Sandstone that was presumably derived from the alluvial cobble to boulder gravel that composes Cobblestone Hill.

Return to the vehicles at the gate after the discussion at this stop.

Proceed through the entrance gate. Low roadside excavations approximately 75 meters (250 ft) west of the gate expose the cobble gravel that comprises the Lake Fort Ann beach ridges. Near the crest of the ridge the angular, 0.3 to 1.2 meter (1 to 4 ft) diameter boulders that comprise the core of Cobblestone Hill can be observed at the surface. The remains of the "Million-Dollar Dam" can be seen on the right just southwest of the hill crest.

STOP 2: THE "MILLION-DOLLAR DAM"

Miles between points: 0.3 Cumulative Mileage: 17.9

The "Million-Dollar Dam" and hydroelectric generation plant was completed on 11 March, 1913 and operated intermittently from 21 January, 1915 until its closure in 1922. A large hole was blasted in the dam shortly after William Miner's death in 1930 to permit Cold Brook to drain freely through the former reservoir. The Altona Flat Rock sandstone pavement is exposed southwest of Cold Brook. The change from mixed deciduous, primarily oak, forest on Cobblestone Hill to jack pine barrens on Altona Flat Rock is characteristically sharp at this location.

A gaging station was constructed at the reservoir outlet in the fall of 1992 to continuously monitor surface-water discharge (Fig. 7). The gaging station is part of a joint initiative by the Center for Earth and Environmental Science and the Applied Environmental Science Program at the SUNY Plattsburgh and the W.H. Miner Agricultural Research Institute to establish an instrumented field station for undergraduate research and instruction in geology and environmental science at the Miner Dam site. A monitoring-well network, consisting of nine wells ranging in depth from 10 to 25 meters, was completed in May, 1992 between the northeastern portion of the former Million-Dollar dam reservoir and the Skeleton Dam (Fig. 7). Water-level measurements were begun in late July, 1992 (Fig. 8). Future plans include the installation of a weather station, an inflow stream gaging station, and expansion of the monitoring-well network. The field station will provide an important linkage between traditional and applied educational opportunities that addresses some of the unique geological and ecological aspects of the Altona Flat Rock region.

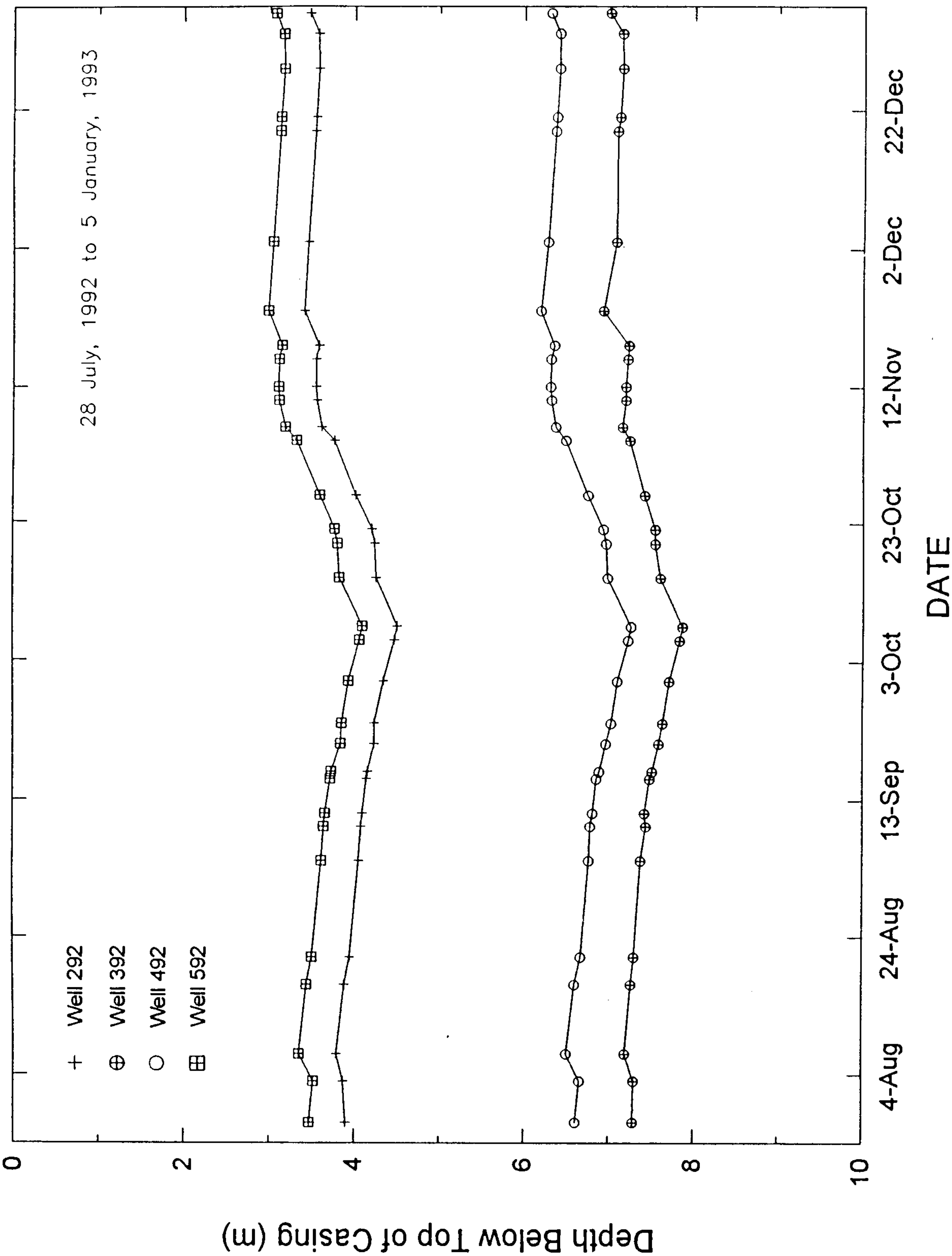

Figure 8. Hydrographs for monitoring wells near the Skeleton Dam.

Return to the vechicles following the discussions at this stop and proceed eastward toward Cobblestone Hill. Turn left onto a small road near the crest of the hill that leads northwestward along the flank of the former reservoir (Fig. 7).

STOP 3: THE "THE SCARPIT"

Miles between points: 0.2 Cumulative Mileage: 18.1

The "scarpit" is the local name given to the desolate landscape created by efforts to grout the porous boulder gravel slope of Cobblestone Hill (Figs. 6 and 7). The surface consists of a thin (1.2 to 2.5 cm) layer of cement that was poured and raked between large boulders composed predominantly of Potsdam Sandstone. The trench that was dug for the grout curtain (Fig. 4) can be observed approximately 100 meters west of the concrete road that parallels the former shoreline of the reservoir.

Return to the vehicles following the discussions at this stop and proceed northwestward on the concrete road.

Miles between points: 0.5 Cumulative Mileage: 18.6

The first of nine observation wells drilled in May 1992 can be observed to the left near the treeline at the edge of the grout surface. The wooded area beyond the well is part of minor southeast-facing bedrock riser. The slope of Cobblestone Hill steepens and the boulder size increases to the northwest.

Miles between points: 0.5 Cumulative Mileage: 19.1

The concrete road ends and the access road bears sharply northeast and continues on the bedrock surface through the jack pine barrens.

Miles between points: 0.1 Cumulative Mileage: 19.2

The road crosses a surface-water supported wetland. The road bed is deeply rutted where it crosses a wetland that contains 0.2 to 1.0 meters of organic soil. Observation wells located approximately 50 meters northeast and 70 meters southwest of the wetland indicate that the water table is more than 7.5 meters below the surface.

Miles between points: 0.2 Cumulative Mileage: 19.4

The road crosses a small channel that contains a large wetland. A concrete wall on the left (south) side of the road was constructed to prevent water impounded behind the "Million-Dollar Dam" to escape northward through this channel (Fig. 7).

The access road forks immediately west of the channel. The right fork leads to an abandoned fire tower on the top of Pine Ridge. Bear left and proceed southward.

STOP 4: THE "SKELETON DAM" & CHASM LAKE

Miles between points: 0.2 Cumulative Mileage: 19.6

The partially completed "Skeleton Dam" was designed to augment flow to the reservoir impounded behind the "Million-Dollar Dam" (Figs. 7 and 8). The dam impounds "Chasm Lake" (Gooley, 1980), presumably named for the deep gorge cut into a prominent sandstone riser at its northwestern edge.

The water level of Chasm Lake dropped more than 2 meters below the spillway of the Skeleton Dam during the summers of 1991 and 1992. What little surface flow reached the basin during the summer months was lost by evaporation and ground-water seepage from the basin. Water level measurements from nearby observation wells since late July, 1992 indicate that steep, eastwardly directed hydraulic gradients exist at the southeastern flank of Chasm Lake, providing support for the hypothesis that some water is being lost from the reservoir by groundwater seepage.

Return to vehicles after the discussion at this stop and follow the rod log in reverse order to the Lake Champlain Ferry dock at Cumberland Head.

Miles between points: 19.6 Cumulative Mileage: 39.2

Lake Champlain Ferry dock.

End of road log.

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The Champlain thrust fault, Lone Rock Point, Burlington, Vermont

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LOCATION

The 0.6 mi (1 km) exposure of the Champlain thrust fault is located on the eastern shore of Lake Champlain at the north end of Burlington Harbor. The property is owned by the Episcopal Diocesan Center. Drive several miles (km) north along North Avenue (Vermont 127) from the center of Burlington until you reach the traffic light at Institute Road, which leads to Burlington High School, The Episcopal Diocesan Center, and North Beach. Turn west toward the lake and take the first right (north) beyond Burlington High School. The road is marked by a stone archway. Stop at the second building on the west side of the road, which is the Administration Building (low rectangular building), for written permission to visit the field site.

Continue north from the Administration Building, cross the bridge over the old railroad bed, and keep to the left as you drive over a small rise beyond the bridge. Go to the end of this lower road. Park your vehicle so that it does not interfere with the people living at the end of the road (Fig. 1). Walk west from the parking area to the iron fence at the edge of the cliff past the outdoor altar where you will see a fine view of Lake Champlain and the Adirondack Mountains. From here walk south along a footpath for about 600 ft (200 m) until you reach a depression in the cliff that leads to the shore (Fig. 1).

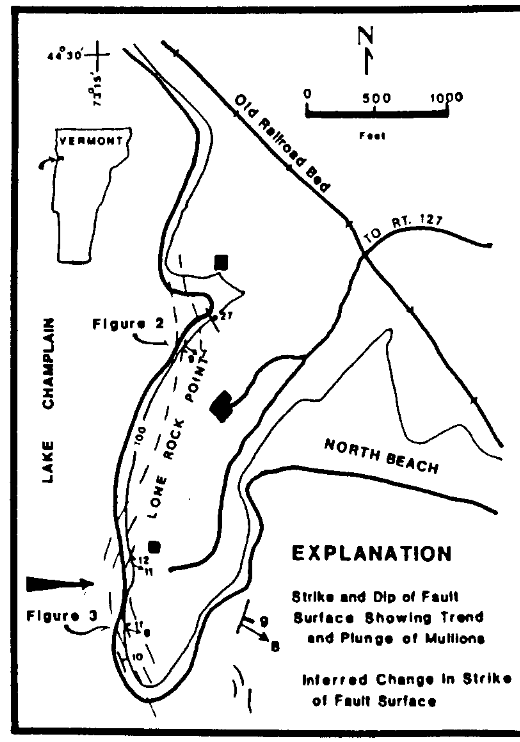

Figure 1. Location map of the Champlain thrust fault at Lone Rock Point, Burlington, Vermont. The buildings belong to the Episcopal Diocesan Center. The road leads to Institute Road and Vermont 127 (North Avenue). The inferred change in orientation of the fault surface is based on measured orientations shown by the dip and strike symbols. The large eastward-directed arrow marks the axis of a broad, late syncline in the fault zone. The location of Figures 2 and 3 are shown to the left of "Lone Rock Point." The large arrow points to the depression referred to in the text.

SIGNIFICANCE

This locality is one of the finest exposures of a thrust fault in the Appalachians because it shows many of the fault zone features characteristic of thrust faults throughout the world. Early studies considered the fault to be an unconformity between the strongly-tilted Ordovician shales of the "Hudson River Group" and the overlying, gently-inclined dolostones and sandstones of the "Red Sandrock Formation" (Dunham, Monkton, and Winooski formations of Cady, 1945), which was thought to be Silurian because it was lithically similar to the Medina Sandstone of New York. Between 1847 and 1861, fossils of pre-Medina age were found in the "Red Sandrock Formation" and its equivalent "Quebec Group" in Canada. Based on this information, Hitchcock and others (1861, p. 340) concluded that the contact was a major fault of regional extent. We now know that it is one of several very important faults that floor major slices of Middle Proterozoic continental crust exposed in western New England.

Our current understanding of the Champlain thrust fault and its associated faults (Champlain thrust zone) is primarily the result of field studies by Keith (1923, 1932), Clark (1934), Cady (1945), Welby (1961), Doll and others (1961), Coney and others (1972), Stanley and Sarkisian (1972), Dorsey and others (1983), and Leonard (1985). Recent seismic reflection studies by Ando and others (1983, 1984) and private industry have shown that the Champlain thrust fault dips eastward beneath the metamorphosed rocks of the Green Mountains. This geometry agrees with earlier interpretations shown in cross sections across central and northern Vermont (Doll and others, 1961; Coney and others, 1972). Leonard's work has shown that the earliest folds and faults in the Ordovician sequence to the west in the Champlain Islands are genetically related to the development of the Champlain thrust fault.

In southern Vermont and eastern New York, Rowley and others (1979), Bosworth (1980), Bosworth and Vollmer (1981), and Bosworth and Rowley (1984), have recognized a zone of late post-cleavage faults (Taconic Frontal Thrust of Bosworth and Rowley, 1984) along the western side of the Taconic Mountains. Rowley (1983), Stanley and Ratcliffe (1983, 1985), and Ratcliffe (in Zen and others, 1983) have correlated this zone with the Champlain thrust fault. If this correlation is correct then the Champlain thrust zone would extend from Rosenberg, Canada, to the Catskill Plateau in east-central New York, a distance of 199 mi (320 km), where it appears to be overlain by Silurian and Devonian rocks. The COCORP line through southern Vermont shows an east-dipping reflection that roots within Middle Proterozoic rocks of the Green Mountains and intersects the earth's surface along the western side of the Taconic Mountains (Ando and others, 1983, 1984).

The relations described in the foregoing paragraphs suggest that the Champlain thrust fault developed during the later part of the Taconian orogeny of Middle to Late Ordovician age. Subsequent movement, however, during the middle Paleozoic Acadian orogeny and the late Paleozoic Alleghenian orogeny can not be ruled out. The importance of the Champlain thrust in the plate tectonic evolution of western New England has been discussed by Stanley and Ratcliffe (1983, 1985). Earlier discussions can be found in Cady (1969), Rodgers (1970), and Zen (1972).

REGIONAL GEOLOGY

In Vermont the Champlain thrust fault places Lower Cambrian rocks on highly-deformed Middle Ordovician shale. North of Burlington the thrust surface is confined to the lower part of the Dunham Dolomite. At Burlington, the thrust surface cuts upward through 2,275 ft (700 m) of the Dunham into the thick-bedded quartzites and dolostones in the very lower part of the Monkton Quartzite. Throughout its extent, the thrust fault is located within the lowest, thick dolostone of the carbonate-siliciclastic platform sequence that was deposited upon Late Proterozoic rift-clastic rocks and Middle Proterozoic, continental crust of ancient North America.

At Lone Rock Point in Burlington the stratigraphic throw is about 8,850 ft (2,700 m), which represents the thickness of rock cut by the thrust surface. To the north the throw decreases as the thrust surface is lost in the shale terrain north of Rosenberg, Canada. Part, if not all, of this displacement is taken up by the Highgate Springs and Philipsburg thrust faults that continue northward and become the "Logan's Line" thrust of Cady (1969). South of Burlington the stratigraphic throw is in the order of 6,000 ft (1,800 m). As the throw decreases on the Champlain thrust fault in central Vermont the displacement is again taken up by movement on the Orwell, Shoreham, and Pinnacle thrust faults.

Younger open folds and arches that deform the Champlain slice may be due either duplexes or ramps along or beneath the Champlain thrust fault. To the west, numerous thrust faults are exposed in the Ordovician section along the shores of Lake Champlain (Hawley, 1957; Fisher, 1968; Leonard, 1985). One of these broad folds is exposed along the north part of Lone Rock Point (Fig. 2). Based on seismic reflection studies in Vermont, duplex formation as described by Suppe (1982) and Boyer and Elliot (1982) indeed appears to be the mechanism by which major folds have developed in the Champlain slice.

North of Burlington the trace of the Champlain thrust fault is relatively straight and the surface strikes north and dips at about 15° to the east. South of Burlington the trace is irregular because the thrust has been more deformed by high-angle faults and broad folds. Slivers of dolostone (Lower Cambrian Dunham Dolomite) and limestone (Lower Ordovician Beekmantown Group) can be found all along the trace of the thrust. The limestone represents fragments from the Highgate Springs slice exposed directly west and beneath the Champlain thrust fault north of Burlington (Doll and others, 1961). In a 3.3 to 10 ft (1 to 3 m) zone along the thrust surface, fractured clasts of these slivers are found in a matrix of ground and rewelded shale.

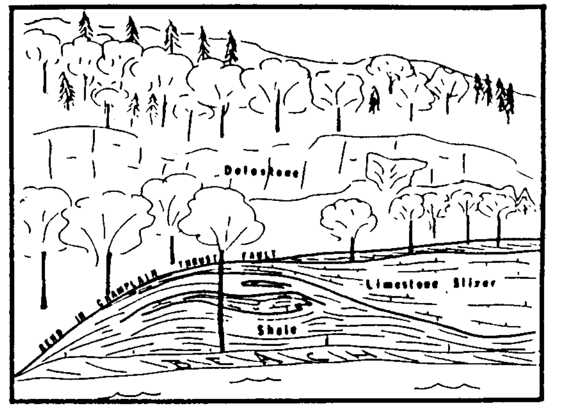

Figure 2. A sketch of the Champlain thrust fault at the north end of Lone Rock Point showing the large bend in the fault zone and the slivers of Lower Ordovician limestone. The layering in the shale is the S1 cleavage. It is folded by small folds and cut by many generations of calcite veins and faults. The sketch is located in Figure 1.

Estimates of displacement along the Champlain thrust fault have increased substantially as a result of regional considerations (Palmer, 1969; Zen and others, 1983; Stanley and Ratcliffe, 1983, 1985) and seismic reflection studies (Ando and others, 1983, 1984). The earlier estimates were less than 9 mi (15 km) and were either based on cross sections accompanying the Geologic Map of Vermont (Doll and others, 1961) or simply trigonometric calculations using the average dip of the fault and its stratigraphic throw. Current estimates are in the order of 35 to 50 mi (60 to 80 km). Using plate tectonic considerations, Rowley (1982) has suggested an even higher value of 62 mi (100 km). These larger estimates are more realistic than earlier ones considering the regional extent of the Champlain thrust fault.

Lone Rock Point

At Lone Rock Point the basal part of the Lower Cambrian Dunham Dolostone overlies the Middle Ordovician Iberville Formation. Because the upper plate dolostone is more resistent than the lower plate shale, the fault zone is well exposed from the northern part of Burlington Bay northward for approximately 0.9 mi (1.5 km; Fig. 1). The features are typical of the Champlain thrust fault where it has been observed elsewhere.

The Champlain fault zone can be divided into an inner and outer part. The inner zone is 1.6 to 20 ft (0.5 to 6 m) thick and consists of dolostone and limestone breccia encased in welded, but highly contorted shale (Fig. 3). Calcite veins are abundant. One of the most prominent and important features of the inner fault zone is the slip surface, which is very planar and continuous throughout the exposed fault zone (Fig. 3). This surface is marked by very fine-grained gouge and, in some places, calcite slickenlines. Where the inner fault zone is thin, the slip surface is located

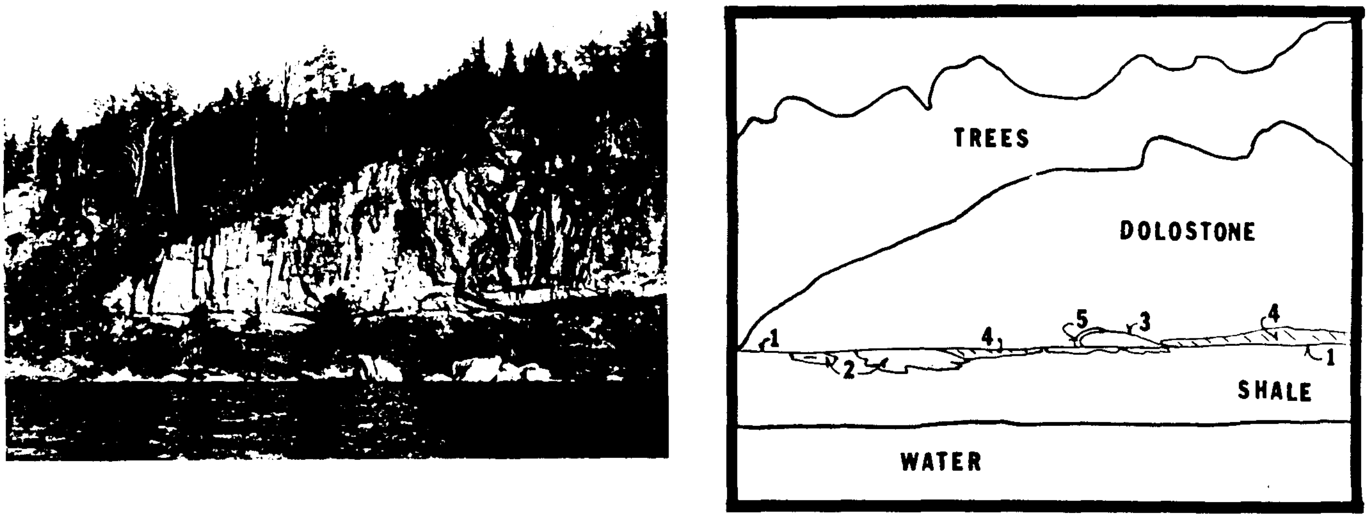

Figure 3. View of the Champlain thrust fault looking east at the southern end of Lone Rock Point (Fig. 1). The accompanying line drawing locates by number the important features discussed in the text: 1, the continuous planar slip surface; 2, limestone slivers; 3, A hollow in the base of the dolostone is filled in with limestone and dolostone breccia; 4, Fault mullions decorate the slip surface at the base of the dolostone; 5, a small dike of shale has been injected between the breccia and the dolostone.

along the interface between the Dunham Dolomite and the Iberville Shale. Where the inner fault is wider by virtue of slivers and irregularities along the basal surface of the Dunham Dolomite, the slip surface is located in the shale, where it forms the chord between these irregularities (Fig. 3). The slip surface represents the surface along which most of the recent motion in the fault zone has occurred. As a consequence, it cuts across all the irregularities in the harder dolostone of the upper plate with the exception of long wave-length corrugations (fault mullions) that parallel the transport direction. As a result, irregular hollows along the base of the Dunham Dolomite are filled in by highly contorted shales and welded breccia (Fig. 3).

The deformation in the shale beneath the fault provides a basis for interpreting the movement and evolution along the Champlain thrust fault. The compositional layering in the shale of the lower plate represents the well-developed S1 pressure-solution cleavage that is essentially parallel to the axial planes of the first-generation of folds in the Ordovician shale exposed below and to the west of the Champlain thrust fault (Fig. 4). As the trace of the thrust fault is approached from the west this cleavage is rotated eastward to shallow dips as a result of westward movement of the upper plate (Fig. 4). Slickenlines, grooves, and prominent fault mullions on the lower surface of the dolostone and in the adjacent shales, where they are not badly deformed by younger events, indicate displacement was along an azimuth of approximately N60°W (Fig. 4; Hawley, 1957; Stanley and Sarkesian, 1972; Leonard, 1985). The S1 cleavage at Lone Rock Point is so well developed in the fault zone that folds in the original bedding are largely destroyed. In a few places, however, isolated hinges are preserved and are seen to plunge eastward or southeastward at low angles (Fig. 4). As these F1 folds are traced westward from the fault zone, their hinges change orientation to the northeast. A similar geometric pattern is seen along smaller faults, which deform S1 cleavage in the Ordovician rocks west of the Champlain thrust fault. These relations suggest that F1 hinges are rotated towards the transport direction as the Champlain thrust fault is approached. The process involved fragmentation of

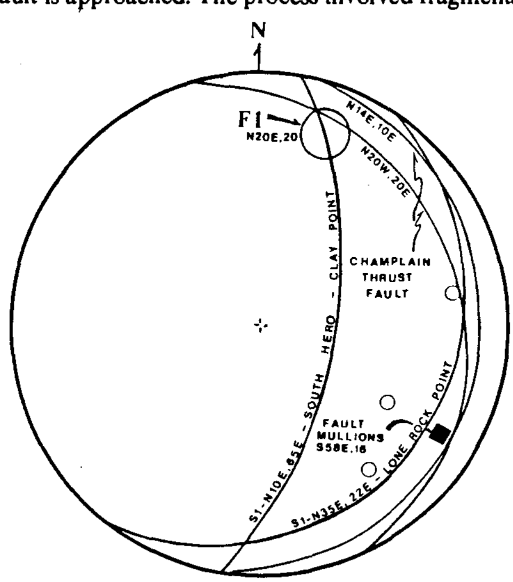

Figure 4. Lower hemisphere equal-area net showing structural elements associated with the Champlain thrust fault. The change in orientation of the thrust surface varies from approximately N20°W to N14°E at Lone Rock Point. The orientation of S1 cleavage directly below the thrust is the average of 40 measurements collected along the length of the exposure. S1, however, dips steeply eastward in the Ordovician rocks to the west of the Champlain thrust fault as seen at South Hero and Clay Point where F1 hinges plunge gently to the northeast. Near the Champlain thrust fault F1 hinges (small circles) plunge to the east. Most slickenlines in the adjacent shale are approximately parallel to the fault mullions shown in the figure.

the FI folds since continuous fold trains are absent near the thrust. Much of this deformation and rotation occurs, however, within 300 ft (100 m) of the thrust surface. Within this same zone the S1 cleavage is folded by a second generation of folds that rarely developed a new cleavage. These hinges also plunge to the east or southeast like the earlier F1 hinges. The direction of transport inferred from the analysis of F2 data is parallel or nearly parallel to the fault mullions along the Champlain thrust fault. Stanley and Sarkesian (1972) suggested that these folds developed during late translation on the thrust with major displacement during and after the development of generation 1 folds. New information, however, suggests that the F2 folds are simply the result of internal adjustment in the shale as the fault zone is deformed by lower duplexes and frontal or lateral ramps (Figs. 1, 2). The critical evidence for this new interpretation is the sense of shear inferred from F2 folds and their relation to the broad undulations mapped in the fault zone as it is traced northward along Lone Rock Point (Fig. 1). South of the position of the thick arrow in Figure 1, the inferred shear is west-over-east whereas north of the arrow it is east-over-west. The shear direction therefore changes across the axis of the undulation (marked by the arrow) as it should for a synclinal fold.

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Stratigraphy of the Cambrian platform in northwestern Vermont

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LOCATION

The Cambrian to Lower Ordovician stratigraphic sequence in northwestern Vermont outcrops in a north-south trending belt bordered on the east by the Green Mountain anticlinorium and on the west by deformed Middle Ordovician shales. Significant facies changes, both parallel and perpendicular to depositional strike, can be observed between Burlington and U.S. 2, 8 mi (13 km) to the north (Fig. 1). Stops 1 to 5 in the Lower Cambrian Dunham Dolomite, Middle Cambrian Monkton Quartzite, and Winooski Dolomite are located on a 3 mi (5 km) west-to-east series of outcrops along U.S. 2 near Milton, starting in an abandoned quarry on the north side of the road, 5 mi (8 km) east of the Sand Bar State Park. The section continues 0.9 mi (1.5 km) farther east along U.S. 2 at a large roadcut on the north side of U.S. 2 (Stop 2), and again another 0.8 mi (1.3 km) farther along Route 2 at a roadcut on the south side of the road (Stop 3). The final two stops along U.S. 2 are another 0.8 mi (1.3 km) (Stop 4) and 0.8 mi (1.3 km) (Stop 5, Chimney Corners). Total mileage along U.S. 2 is 3.3 mi (5.3 km). Stop 6 is located along the Winooski River in downtown Winooski, Vermont, where the supra-to-intertidal and shallow subtidal facies of the Monkton, Winooski, and Danby Formations are exposed. All outcrops are accessible by car; access does not require permission.

SIGNIFICANCE

Cambro-Ordovician siliciclastic and carbonate sediments in western Vermont were deposited on a tectonically stable shelf following late Precambrian rifting of the Iapetus Ocean (Rodgers, 1968). The alternating siliciclastic and carbonate units record sedimentation in supra-to shallow subtidal platform environments, which pass laterally into platform margin and basinal deposits. The basal Cheshire Quartzite, not included in this site, represents a shallow siliciclastic "blanket" over the Eocambrian rift topography (Myrow, 1983). Subsequent Cambrian units record the vertical upbuilding of the carbonate platform characteristic of the early Paleozoic continental margin in the Appalachians.

SITE INFORMATION

Background. The Lower Cambrian Dunham Dolomite lies in gradational contact with the Cheshire Quartzite (Fig. 2) and represents the initial carbonate deposition on the newly formed shelf (Gregory, 1982). The Dunham Dolomite is important because it records the initial carbonate facies development and the establishment of the platform geometry that are continued in subsequent units in northwestern Vermont. Upward building of the carbonate platform during Dunham time resulted

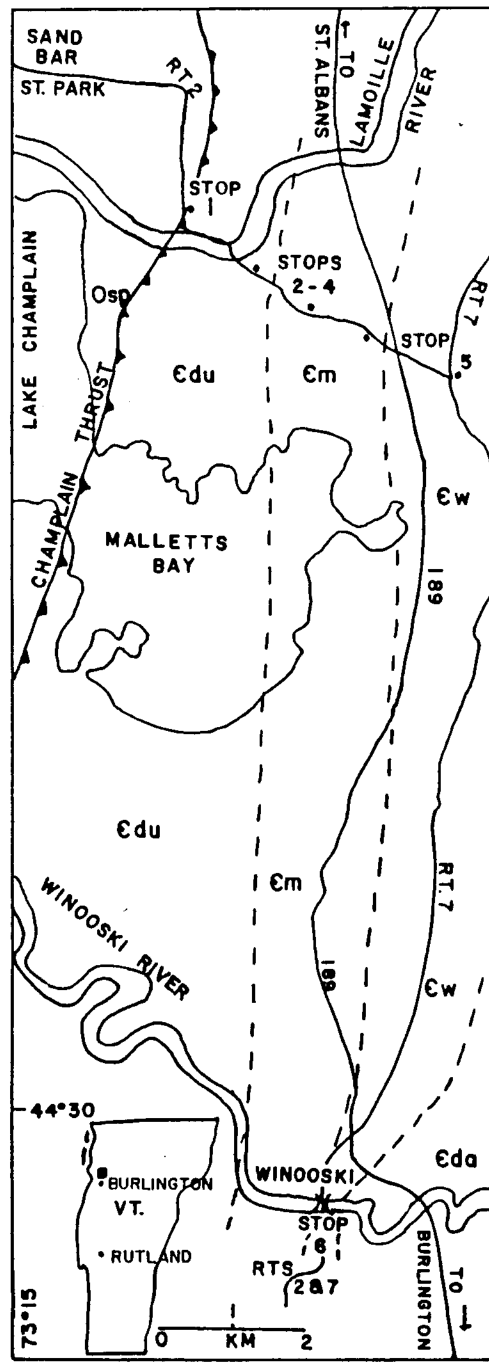

Figure 1. Locality map for stops in northwestern Vermont. Stops 1–5 occur along U.S. 2 between Sand Bar State Park and the intersection of U.S. 2 and 7 (Chimney Corners). Stop 6 occurs along the banks of the Winooski River in downtown Winooski, starting at the first ledges downstream of the bridge and extending upstream past the mill to the last ledges. Lithic designators: Cdu-Dunham Dolomite, Cm-Monkton Quartzite, Cw-Winooski Dolomite, Cda-Danby Quartzite, Osp-Stony Point Shale.

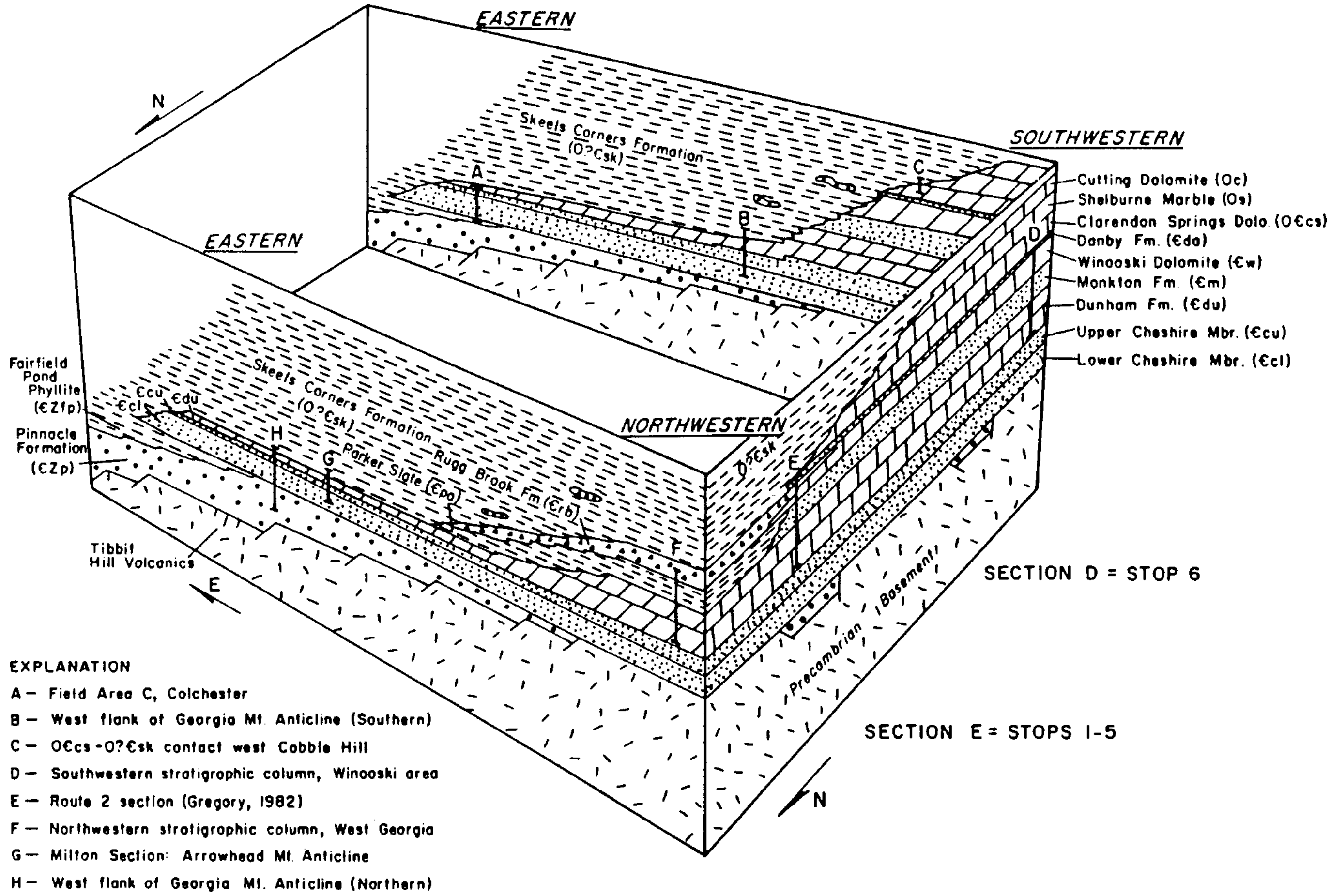

Figure 2. Block diagram from Dorsey, and others (1983) illustrating the platform stratigraphy and the eastward to northward facies changes associated with the platform margin transition. Sections represented by Stops 1-5 and 6 are sections E and D, respectively.

in the development of a platform-to-platform margin transition which was characterized by the abrupt pinch-out of shallow water facies into basinal shales and proximal talus slope breccias. This platform-to-basin transition remained localized in roughly the same paleogeographic position throughout the remainder of the Cambrian and Early Ordovician, probably as a result of movement on underlying rift-related faults. The Cambro-Ordovician platform was characterized by vertical upbuilding and little-to-no progradation into the adjacent basin.

The nature and distribution of lithofacies in the Monkton Quartzite (lower Middle Cambrian) can be compared to the underlying Dunham Dolomite; these also serve as models for interpreting the youngest siliciclastic unit, the Danby Quartzite. As seen in the Dunham Dolomite, the Monkton also records deposition in tidal, shallow subtidal, and platform margin environments. The distribution of lithofacies on the platform is also similar, recording both east-to-west and north-to-south facies changes into the adjacent shale basin. The lithofacies and environments of deposition of the Monkton Quartzite were studied and summarized by Rahmanian (1981). Rahmanian recognized seven lithofacies: three comprise mixed siliciclastic and carbonate sediments associated with shallowing-up cycles, three are pure siliciclastic deposits, and one is a pure carbonate (oolitic dolomite) facies. The 1,000-ft (300-m)-thick Monkton is composed of cyclic shallowing-up cycles characterized by repetitive packages of: (1) basal subtidal siliciclastic sand shoals and channels (Stop 4) overlain by, (2) interbedded siliciclastic sand and silt and carbonate intertidal flat sediments (Stop 6), capped by (3) carbonate muds of the high intertidal and supratidal flat (Stop 6). These shallowing-up cycles are interpreted to represent prograding tidal flat deposits, structurally similar to those in the underlying Dunham Dolomite (Mehrtens, 1986).

The Winooski Dolomite (Middle Cambrian) is a structureless dolomite with disseminated quartz sand and cryptalgalaminites common throughout. Little detailed sedimentology has been done on the Winooski, but based on its stratigraphic relationship with over- and underlying units as well as lithofacies analysis, it is interpreted to represent shallow subtidal (Stop 6) to platform margin (Stop 5) environments.

The Upper Cambrian Danby Quartzite is a mixed siliciclastic and carbonate unit (Butler, 1986) that represents deposition in intertidal to shallow subtidal as well as platform margin settings. Butler (1986) has interpreted much of the Danby as recording shallow subtidal, storm-influenced sedimentation.

The Cambro-Ordovician sequence seen in Vermont records alternating carbonate and siliciclastic sedimentation not seen elsewhere in the Appalachian-Caledonide Orogen. The environments of deposition represented by each formation are similar, regardless of composition, and contacts between units are gradational. These three characteristics: regionally localized cyclic carbonate or siliciclastic sedimentation, similarity of platform paleogeography, and consistently gradational contacts, suggests that the cyclicity is not a result of large-scale regressive events (Rowley, 1979), but rather local variations in siliciclastic sand supply and distribution.

Stop 1. Stop 1, an abandoned quarry, lies immediately above the Champlain Thrust, as the floor of the quarry is Middle Ordovician Stony Point Shale overlain by Lower Cambrian Dunham Dolomite. The Dunham Dolomite is approximately 1,300 ft (400 m) thick and is exposed along U.S. 2, near Milton (Stops 1–3), where rocks characteristic of the peritidal, subtidal/open shelf, and platform margin can be seen. Stop 1 exhibits fresh exposures of the peritidal facies, characterized by the rhythmic interbedding of white dolomite and red dolomitic siltstone in a "sedimentary boudinage" bedding style. Intraformational conglomerate, imbricated clasts, and burrows are locally common. It is strongly recommended that visitors to this stop avoid climbing on the quarry walls but confine themselves to large blocks lying around the quarry floor.

Stop 2. The overlying subtidal/open shelf facies of the Dunham Dolomite is characterized by shallowing-up cycles up to 30 ft (10 m) thick that have at their base massive beds of bioturbated and mottled dolomite with disseminated quartz and feldspar sand throughout. This lithology is capped by packages of the rhythmically interbedded dolomite and dolomitic siltstone of the peritidal facies. These shallowing-up cycles are interpreted to represent tongues of tidal flat sediments that prograded into the adjacent subtidal shelf. Shallowing-up cycles make up the bulk of the 1,000-ft (300-m)-thick subtidal facies, and these pass upsection into structureless, subtidal, burrowed muds before passing into the platform margin facies.

Stop 3. Stop 3 exhibits massive beds of polymictic breccia interpreted as proximal debris flows interbedded with graded dolomitic sandstone beds interpreted as turbidites. Clasts within the breccia are poorly sorted and angular. Beds are poorly developed and structureless. At other localities the Dunham platform margin lithofacies can be seen to pass laterally into basinal shales of the Parker Slate.

Stop 4. Stop 4, along U.S. 2, exhibits the subtidal and platform margin facies of the Monkton Quartzite. At this stop, subtidal/tidal channel, crossbedded sands pass up section into thickly bedded, crossbedded platform margin sand bodies. A smaller outcrop immediately to the east of Stop 4 exposes horizons of the platform margin polymictic breccia in a matrix of coarse-grained quartz sand. Note the variable clast composition and the angularity and poor sorting of the clasts. Compare the sedimentary structures and bed thickness at this exposure to the inter- and supratidal Monkton seen at Stop 6.

Stop 5. The Winooski Dolomite exposed at Stop 5 (Chimney Corners) consists of structureless dolomite which is environmentally nondiagnostic, but it is capped by horizons of polymictic breccia in a sand-rich dolomite matrix which has been interpreted as a platform margin breccia. The breccia can be compared to those seen in the underlying Monkton and Dunham Formations.

Stop 6. Stop 6, downstream of the bridge (southeast bank) over the Winooski River in Winooski, exhibits shallow-water facies of the Monkton, Winooski, and Danby Formations. The base of the section is in the Monkton Quartzite, where inter- and supratidal facies are exposed in shallowing-up cycles. Bedforms diagnostic of supra-, inter- and shallow subtidal environments are exposed on broad bedding planes. Several different ripple morphologies can be identified, along with mudcracks and vertical and horizontal burrows. Structureless beds of buff dolomite are interpreted as supratidal deposits.

The Winooski Dolomite is also exposed on the north side of the Winooski River (Stop 6) from below the bridge to 300 ft (100 m) upstream. The basal contact with the Monkton is underwater here, but in a quarry a few miles away it can be seen to be gradational over a 30-ft (10-m) interval. The upper contact of the Winooski with the Danby Quartzite (Stop 6, upstream on ledges above the mill) is also gradational, characterized by increasing sand content until the first quartzite bed of the Danby is reached. The Winooski here consists of structureless dolomite, which, with the exception of horizons of cryptalgalaminites, is environmentally nondiagnostic. The Winooski is interpreted as shallow subtidal in origin, based on the abundance of stromatolites(?) and its stratigraphic position relative to the underlying inter- and supratidal Monkton facies.

The Danby Quartzite (Stop 6) is exposed in a series of ledges upstream of the bridge over the Winooski River in Winooski. The outcrop at Stop 6 exhibits beds with a diverse assemblage of ripple morphologies, laterally discontinuous bedding, hummocky cross-stratification, stromatolites, and oncolites. The platform margin facies of the Danby is not exposed along Route 2, but occurs in the woods nearby. It is also characterized by polymictic breccia clasts in a crossbedded sand matrix, similar to those described for the Monkton Quartzite.

SUMMARY

The distribution of facies on the Cambrian platform records deposition on a flat-topped, low-gradient platform bordered on the east and north by a deep basin in which shale was deposited (Fig. 2). Regardless of siliciclastic or carbonate composition, all platform deposits appear to record similar facies: supratidal to shallow subtidal in the platform interior and platform margins characterized by carbonate or siliciclastic shoal deposits and talus slope breccias. Significant lateral migration of facies between units is not seen, suggesting that sedimentation on the platform was continuous and able to keep pace with subsidence.

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