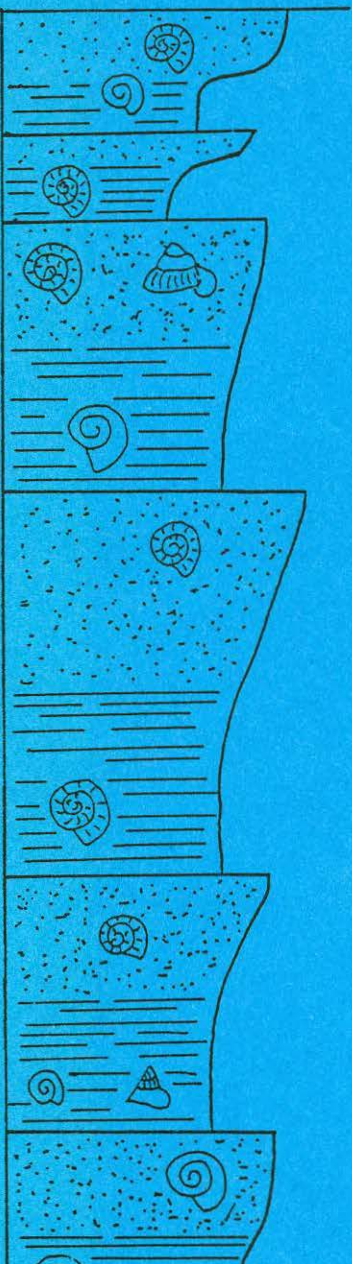

Department of Geology
Colgate University

New York State Geological Association Field Trip Guidebook

64th Annual Meeting
September 18-20, 1992

New York State Geological Association

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Richard H. April, Editor

Field Trip Guidebook

Department of Geology
Colgate University
Hamilton, NY 13346

Published by the New York State Geological Association.
Additional copies may be obtained from the permanent Executive Secretary, James F. Olmstead, Center for Earth and Environmental Science, State University of New York, College at Plattsburgh, Plattsburgh, New York 12901.

Step out onto the Planet.
Draw a circle a hundred feet round.

Inside the circle are
300 things nobody understands, and, maybe
nobody's ever really seen.

How many can you find?

Lew Welch

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The Trenton - Utica Problem Revisited: New Observations and Ideas Regarding Middle - Late Ordovician Stratigraphy and Depositional Environments in Central New York

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INTRODUCTION

The Middle to Upper Ordovician Trenton Limestone and Middle to Upper Ordovician black (Utica-type) shale which overlies it in New York State and Southern Ontario have been of considerable interest to geologists since the time of James Hall. This attention is due to the paleontological richness of the Trenton carbonates and early recognition of the facies equivalency of part of the Trenton with black shale deposits in the Mohawk Valley. More recently, heightened interest in Trenton-Utica stratigraphy, sedimentology, paleoecology, and tectonic processes reflects recognition that these sediments are synorogenic. The closely associated Taconic Orogeny represents an arc collision event involving subduction-related emplacement (overthrusting) of an accretionary prism (Taconic Allochthon) over the margin of the Laurentia craton. This emplacement was followed by orogenesis associated with collision, suturing, and uplift of the Vermontia Terrane (Bird and Dewey, 1970; Rowley and Kidd, 1981; Lash, 1988). This paleosetting is believed to be analogous to the present day subduction of the Australian Sahul Shelf under the Banda Arc with the Taconic basin correlating to the foredeep south of Timor. In this context, the Trenton to Utica transition records the collapse of a cratonic platform owing to subduction and thrust load-related down-bending of the craton as the orogeny proceeded. Evidence for this downbuckling is manifest in the Middle-Upper Ordovician succession in central and eastern New York as an upward- and eastward facies change from carbonate platform deposits to basinal facies.. Furthermore, numerous normal faults cross-cut the Trenton, which are believed to have been active mainly as synthetic (down to the east) and antithetic (down to the west) displacements reflecting lithospheric responses to subduction (Bradley and Kidd, 1991). Many of these faults are believed to have been active during the tectonic foundering of the Trenton shelf, resulting in a complex "patchwork" stratigraphy for upper Trenton and lower "Utica" units in the Mohawk Valley (see Cisne and Rabe, 1978; Cisne, *et al.*, 1982; Bradley and Kidd, 1991).

Ruedemann (1925), Ruedemann and Chadwick (1935), and Kay (1937, 1953) mapped Trenton and Utica deposits across this region and recognized that uppermost Trenton units (Rust, Steuben, and Hillier limestones) progressively disappeared to the east and southeast of Lowville, NY. Medial Trenton strata (Denmark Limestone) distinctly grade from richly fossiliferous carbonates at Trenton Falls (Poland and Russia members) through a belt of fossil-poor thin-bedded calcilutite facies (Dolgeville Member) to predominantly dark gray or black fissile Utica-type shale (Canajoharie Shale) near Canajoharie, New York. Lower Trenton units (Napanee, King's Falls, Sugar River Formations) were observed to remain fossil-rich across central New York. However, there is an eastward thinning of these lower strata east of Dolgeville, NY, possibly due to submarine erosion. West of Canajoharie, NY the Trenton Limestone is locally missing (see Kay, 1953; Fisher, 1977; Cisne, and Rabe, 1978).

In two important publications, Cisne and Rabe (1978); and Cisne *et al.* (1982), presented results of highly refined paleoecologic work on Trenton and Canajoharie deposits using volcanic ashes (metabentonites) as correlation isochrons (Figure 1). In so doing, they were able to subdivide the Trenton into thin, time-constrained divisions which could be followed eastward into black shale. Quantitative gradient analysis of fossil associations indicated that this was no steady eastward-sloping ramp; normal faults were believed to have been active in latest Trenton time such that a horst and graben topography was developed and submarine fault scarps existed locally (Cisne and Rabe, 1982).

Controversies concerning Trenton, Canajoharie, and Utica stratigraphy in New York arose early in the study of these units. Although some divisions can be correlated over long distances, abrupt facies changes, losses of units, and appearances of units over small distances have long been noted (see discussions in Ruedemann, 1925; Ruedemann and Chadwick, 1935; Kay, 1937, 1953; Cisne, *et al.*, 1982; Titus, 1986, 1988). The most prominent problem involves Trenton and lower Utica Shale stratigraphy in the interval between Trenton Falls and Middleville, a 16 km-wide region northeast of Utica characterized by thick surficial cover and incomplete sections. Another problem is the meaning of sharp lithologic contacts in sections where bioclastic Trenton carbonates are abruptly overlain by dark argillaceous strata (Canajoharie Shale, Utica Shale) or by sparsely fossiliferous ribbon limestones (Dolgeville facies). The presence of such contacts has long been noted, but there is disagreement concerning their extent, age, and significance (see discussions in Kay, 1953; Riva, 1969; Fisher, 1977, 1979; Cisne *et al.*, 1982; Titus, 1990). Very recently, parts of the metabentonite correlation scheme of Cisne and Rabe (1978) have been called into question owing to new "fingerprinting" correlation techniques for ash identification. This problem, coupled with the unstable existing stratigraphies of Kay (1953), Fisher (1977), Cisne *et al.*(1982), and Titus (1990), poses the need for continued field study of these rocks but with the application of new approaches and perspectives.

The interest of the present authors of the Trenton-Utica problem stems both from our interest in the relationships of black shales in foreland basins to surrounding facies and from our use of stratigraphic event-horizons in establishing refined correlations in the Silurian and Devonian as well as for Upper Ordovician strata of Ontario and northcentral New York. Important to this approach is the recognition of sedimentary cycles of varying magnitude in these younger deposits which owe their origin to eustatic sea level changes (Brett and Baird, 1985; Brett *et al.*, 1990b). Also relevant to Trenton mapping is the discovery of widespread submarine discontinuities of varying magnitude which define mappable stratigraphic packages of great lateral extent (Brett and Baird, 1986a; Brett *et al.*, 1990a, b). It is our goal to define, firstly, the widespread major discontinuities both within- and bounding the Trenton, Canajoharie, and Utica, and secondly, the sedimentary cycles and distinctive cycle-capping condensed units within these large discontinuity-bounded packages. Preliminary results of this work are presented below.

We are particularly attracted to the sharp erosional bases of the Ordovician black shales. Our work on Late Ordovician, Silurian, and Devonian black shale-roofed erosions contact shows that they reflect erosional processes occurring predominantly under basinal conditions of prevailing near-anoxia (Baird and Brett, 1986a, b; Baird *et al.*, 1988; Baird and Brett, 1990; Lehmann and Brett, 1991a, b). Preliminary work on Dolgeville, Canajoharie, and Utica discontinuities is presented herein; it shows that these contacts are mappable and significant. A very large and widespread discontinuity herein informally designated the "Sub-Utica Disconformity" may be an Ordovician analog of an erosional overlap surface which floors Middle and Late Devonian black shale deposits across numerous eastern states.

STRATIGRAPHIC NOMENCLATURE UTILIZED ON THIS TRIP

On this trip, we will examine primarily "middle" Trenton strata. Unfortunately, a plethora of stratigraphic nomenclature has been applied to these rocks by various workers (see Fisher, 1977; Kay, 1968, for overview). Stratigraphic schemes of different workers--and in some cases, in different publications by the same worker--are typically contradictory. Furthermore, some proposed lithostratigraphic units were based solely on biostratigraphic marker horizons, while

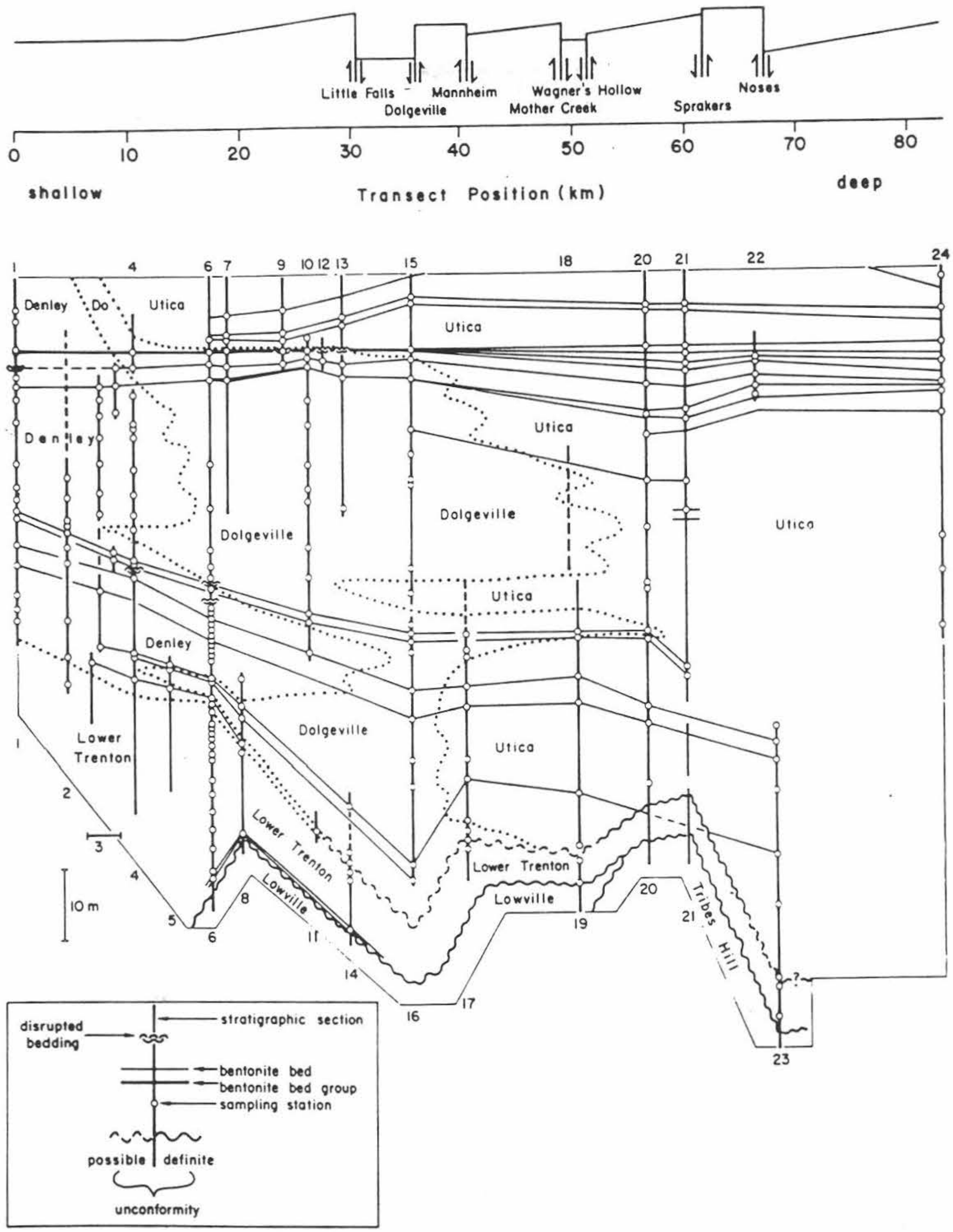


Figure 1. Stratigraphy of the Trenton Group and equivalent basinal facies based on work of Cisne and Rabe, 1978. Trenton sections as well as those for overlying and underlying units are represented by: 1 - Trenton Falls; 2 - Gravesville-Mill Creek; 3 - Poland; 4 - Rathbun Brook; 5 - Shedd Brook; 6 - Buttermilk Creek, Wolf Hollow Creek ("City Brook"), and creeks near the County Home; 7 - Farber Lane; 8 - Norway; 9 - Miller Rd.; 10 - North Creek; 11 - Gun Club Rd.; 12 & 13 - New York State Thruway; 14 - Burrell & Bronner Rds.; 15 - W. Crum Creek; 16 - Dolgeville Dam; 17 - E. Canada Creek; 18 - Mother Creek; 19 - Caroga Creek; 20 - Canajoharie Creek; 21 - Flat Creek; 22 - Currytown Quarry; 23 - Van Wie Creek; 24 - Chuctanunda Creek. From Cisne and Rabe, 1978. Note absence of discontinuities in upper Denley and Dolgeville as well as abrupt Denley - Utica facies transition west of Rathbun Brook (see upper left).

some others have not been recognized outside of very localized regions. The net result of this medusoid intertwining and confusion of nomenclature is that many workers no longer attempt to recognize or utilize highly resolved lithostratigraphic units: Titus and Cameron (1976), Titus (1982, 1986, 1988), and Mehrtens (1988) do not recognize members of Trenton formations in their contributions to Trenton stratigraphy; Bradley and Kidd (1991) generally treat the Trenton Group as an undifferentiated stratigraphic unit.

We feel that a refined, unified Trenton stratigraphy, in which stratigraphic units are bounded by key event beds and lithologic deviations is essential for two reasons. First, correlation of refined stratigraphic units, along with recognition of facies changes within those units, is vital for precisely timing and understanding major tectonic and eustatic events. Second, a unified Trenton stratigraphy will enable our colleagues and us to more easily communicate ideas concerning the dynamics of the Middle/Late Ordovician foreland basin of New York and Ontario. On this trip, we apply a modified version of Kay's (1968) stratigraphic nomenclature. Definitions and descriptions of formal and informal "middle" Trenton stratigraphic units (see Table 1) follows.

<u>Rathbun Member of the Sugar River Limestone:</u> The Rathbun Member includes massive, ledge-forming, packstones and grainstones with few, thin shale interbeds. The lowest packstone beds of this interval contain numerous large (>10 cm) specimens of the domal bryozoan, *Prasopora*. The Rathbun is bounded at its base by more shaley Sugar River wacke- to packstones and at its top by the stratigraphically condensed City Brook bed (discussed below and under Stop 4). The Rathbun Member is 2 m thick at its type locality (Kay, 1953) and can be traced southeastward to the Little Falls area and northwestward to Gravesville. Although Chenowith (1952) did not recognize the Rathbun Member in the Black River Valley, massive upper Sugar River strata, probably correlative with the Rathbun Member, crop out in numerous stream sections in the Lowville area (66 km northwest of Gravesville). In the Lowville area, the Rathbun grainstones are approximately 2 m above a distinctive horizon of large slump folds, suggesting that the abrupt change from typical Sugar River lithology to calcilutites of the Poland may reflect tectonic deepening (see below). Southeast of Little Falls, the Rathbun is absent; these strata have probably been removed by submarine erosion.

<u>Poland Member of the Denley Limestone:</u> The Poland Member includes calcilutites and interbedded dark shales. These calcilutites contain fossil faunas dominated by the trilobites *Isotelus* and *Flexicalymene*, as well as orthocerid nautiloids. Tops of beds show variable degrees of burrowing. An intensely bioturbated pack- to wackestone, the City Brook bed (*Trocholites* Bed of Kay, 1953), defines the base of the Poland and is exposed from the Trenton Falls area to the vicinity of Countryman, New York. Observations by the present authors show that the supposedly diagnostic coiled nautiloid *Trocholites* is rare. As a result, we herein informally designate this layer as the City Brook bed for the excellent exposure on Wolf Hollow Creek (City Brook) just above City Falls (see Stop 4). Southeast of Little Falls, the City Brook bed suffers the same fate as the Rathbun Member: it is cut out by erosion. The City Brook bed is apparently correlative with the Camp Member (basal Denley Limestone) of Chenowith (1952). If this is true, the Camp Member (which would have precedence over the name "City Brook bed") retains a uniform thickness (0.7 m) from City Brook to Lowville but then thickens to 3.5 m at Sackets Harbor (Chenowith, 1952), 55 km northwest of Lowville. The City Brook bed is abruptly overlain by variably fossiliferous micrites and interbedded dark shales, the more typical Poland lithology. This lithology is laterally persistent; an abrupt "kick" into Poland calcilutites is easily recognized in outcrops from Lowville to Little Falls, a distance of 97 km. Northwest of the Lowville area however, the calcilutites grade into bioclastic limestones (Chenowith, 1952). This abrupt vertical transition from storm-dominated shelf deposits (Sugar River Limestone) to deeper shelf deposits (Poland Member) may record an episode of tectonically-induced subsidence which was accommodated by movement along faults in the Lowville - Copenhagen region (Lehmann and Brett, 1991b). From

Kay, 1953		Kay, 1968		This paper			Key markers
Cobourg ↑	Steuben Mbr.	Steuben Ls.		Steuben Ls			Abrupt facies change
Cobourg	Rust Mbr.	Denley	Rust Mbr.	Denley	Rust Mbr.		Erosional contact
Denmark	Russia Mbr.	Denley	Russia Mbr.	Denley	Russia Mbr.	High Falls Tongue	Abrupt facies change
Denmark	Russia Mbr.	Denley	Russia Mbr.	Denley	Russia Mbr.	Brayton Corners Div.	Discontinuity/ erosional contact
Denmark	Russia Mbr.	Denley	Russia Mbr.	Denley	Russia Mbr.	Wolf Hollow Div.	
Denmark	Russia Mbr.	Denley	Russia Mbr.	Denley	Russia Mbr.	North Gage Rd. Bed	Discontinuity
Denmark	Poland Mbr.	Denley	Poland Mbr.	Denley	Poland Mbr.		Lag debris
Denmark	Trochelites Bed	Denley	Poland Mbr.	Denley	City Brook Bed		Abrupt facies change/ discontinuity
Shoreham	Rathbun Mbr.	Sugar River	Rathbun Mbr.	Sugar River	Rathbun Mbr.		Abrupt facies change
Shoreham ↓		Sugar River ↓		Sugar River ↓			

Table 1: Stratigraphic nomenclature utilized on this field trip and correlative nomenclature of Marshall Kay.

Gravesville to Kast Bridge, the calcilutitic Poland is capped by the North Gage Road bed, which marks the base of the overlying Russia Member.

Wolf Hollow Division, Russia Member of the Denley Limestone: The Wolf Hollow division includes nodular calcilutites to wackestones. It is marked by intervals of poorly bedded, conspicuously nodular, muddy carbonate which contain abundant three-dimensional nautiloid and endocerid steinkerns, as well as the trilobites *Isotelus* and *Flexicalymene*. The basal North Gage Road bed forms a 0.2 m ledge in its type locality (Rathbun Brook). This bioturbated packstone is recognizable from Gravesville to Middleville. However, correlation of this bed to Trenton Falls is only tentative. At Trenton Falls, a 0.7 m thick, shale-poor interval of wacke- to packstone crops out just above Sherman Falls. This resistant interval occupies approximately the same stratigraphic position as the North Gage Road bed, although, admittedly, the facies relationship between the resistant interval in the lower high falls of Trenton Falls and at Rathbun Brook (see Stop 3) is not entirely convincing. From Trenton Falls to Rathbun Brook, the Wolf Hollow division retains a uniform thickness (approximately 16 m) of cyclical progressions of tabularly to nodularly-bedded limestones. At Wolf Hollow Creek (City Brook); however, these strata are only 5.5 m thick. We attribute this thinning to stratigraphic condensation and submarine erosion. Two particularly nodular horizons thin dramatically between Rathbun Brook and Wolf Hollow Creek; the lower of these nodular horizons is capped by a carbonate conglomerate at the latter locality, supporting a submarine erosion hypothesis.

Brayton Corners Division, Russia Member of the Denley Limestone: In the Trenton Falls-to-Newport area, Wolf Hollow strata are disconformably overlain by an interval of well sorted calcisiltites to calcarenites and interbedded shales. We refer to this interval as the Brayton Corners division. It is 6 m thick at Trenton Falls and thins to 3.0 meters by Rathbun Brook. At Trenton Falls the the upper 3 m of this interval differs somewhat from the typical Brayton Corners lithology containing thin calcilutites, a few thin grainstones, and a 1 m thick amalgamated bed of nodular packstone. At Wolf Hollow Creek, the Brayton Corners division is represented by a 1.5 m interval of graded calcisiltites to calcilutites which contain phosphatic sand at their bases.

High Falls Tongue of the Dolgeville, Russia Member of the Denley Limestone: At Trenton Falls, the Brayton Corner lithology grades into a 4 meter interval of predominantly calcilutites and interbedded shales (Dolgeville lithology). The uppermost 0.5 m of the High Falls Tongue is calcarenitic and shows soft sediment deformation. This High Falls Tongue of the Dolgeville is disconformably overlain by basal Rust strata.

Rust Member, Denley Limestone: In the Trenton Falls-to-Poland area, a distinctive soft sediment deformed horizon of the uppermost Russia is erosionally truncated and is overlain by a 15 cm thick stratigraphically-condensed, amalgamated bed. The lower 12 cm of this bed is graded but fairly well sorted calcarenite. The top of this calcarenite is irregular and pitted, suggestive of submarine erosion. Overlying this calcarenite, and also filling erosional pits and crevices, is rusty-weathering calcareous silty mudstone. Abundant large molds of the brachiopod *Onniella* mark the top of this bed, which is, in turn, overlain by a 3 cm thick metabentonite. This volcanic ash is overlain by a 1 m thick interval of Dolgeville facies. Unlike typical Dolgeville, however, these calcilutites contain large vertical burrows. The interbedded calcilutites and shales are overlain by 22 m of more typical Rust lithology: nodular, bioturbated pack- to wackestones, intercalated with shales and rich in the brachiopods *Onniella* and *Rafinesquina? deltoidea*. Although Rust strata are typically poorly bedded, wave-rippled tabular beds are present and indicate sediment deposition within storm wave base. Three meters above the base of the Rust, strata are brecciated and slump folded; this is the "lower disturbed zone" of Prosser and Cummings (1897). Above the 22 m interval of typical Rust lithology is a 2.5 m interval of tabular, fossiliferous calcilutites and calcisiltites with interbedded shales. These somewhat Dolgeville-like strata contain exceptionally well preserved crinoids and trilobites and are probably correlative with the famed Rust Farm quarry beds of Walcott (1875 a, b, c, 1876, 1881). These finer-grained carbonates are overlain by 11.5 m of typical Rust lithology; the middle third of these upper Rust strata show severe slump folding ("upper disturbed zone" of Prosser and Cummings, 1897). Slump folding in the lower and upper disturbed zones are other possible correlatives of slump-folded upper Dolgeville seen at Stops 5, 6, and 7. At Stop 1, we will examine the fossiliferous calcilutites and the upper disturbed zone. The Rust strata are overlain by Steuben calcarenites, which lack interbedded shales. To the northwest of Trenton Falls, Rust shales die out, and differentiating Rust and Steuben strata is more difficult. However, the disturbed zones, as well as some distinctive fossil beds, carry through to at least Lowville, allowing correlation of these strata.

Steuben Limestone: The Steuben Limestone comprises thick beds of crinoidal grainstones. Nine meters of these strata are exposed at Trenton Falls; 18 m are exposed northwest of Booneville, where the strata have not been erosionally truncated. Some bed tops show two-dimensional wave ripples and interference ripples indicating sediment deposition near or above fairweather wave base. In the Trenton Falls area (and at South Trenton, Stop 1), Steuben grainstones are disconformably overlain by upper Utica black shales.

THE RATHBUN BROOK PROBLEM: THE RUST MEMBER VANISHING ACT

In the vicinity of Trenton Falls near Prospect, New York, the Middle Ordovician Trenton Limestone Group is approximately 130 meters thick, of which the upper 60% of the total succession is actually exposed at the falls (see Stop 1). In this region the lower 45 meter interval (Napanee, King's Falls, Sugar River Limestones) is dominantly calcilutites grading upward to tempestitic grainstones and packstones. The middle 70 meter interval is represented by the Denley which is subdivided into three members: Poland, Russia, and Rust (see Table 1). At Trenton Falls, 40 m of Poland and Russia strata are exposed (Kay, 1953). These shaley, nodular strata are dominantly a lower energy distal tempestitic wackestone to lime mudstone facies. The top five meters of the Russia Member contains a thin-bedded calcilutite interval recording low energy deeper subtidal conditions. The Russia, in turn, is overlain by the Rust Member which comprises 30 m of rubbly fossiliferous wackestones, packstones, and grainstones.

The upper regressive part of the Trenton Group (Steuben Limestone) is represented by up to 18 m of massive, high energy packstones and grainstones. Still higher Trenton deposits (Hillier Limestone) are present above the Steuben in the Watertown-Boonville area, but these beds are overstepped to the southeast by a discontinuity flooring the black, laminated and organic-rich Utica Shale and stratigraphically higher flysch. At South Trenton (Stop 1) the black Utica Shale yielding the Upper Ordovician graptolite *Climacograptus pygmaeus* directly overlies a largely-beveled Steuben section (Figures 7, 9).

Further southeast, near Poland, New York, the Steuben is apparently absent, but at least 25-30 m of Rust Member is recognized in sections north of the village. The lower units, although poorly exposed, appear little changed. Even to the south of Poland (1.5-2.5 km northwest of Rathbun Brook), some Rust is seen in sections. However, at Rathbun Brook (Stop 2) and at all sections southeast of there, none of the Rust Member is observed (Figure 2). On Rathbun Brook 33 m of Denley Limestone (17 m of Poland Member, 16 m of Russia Member) are followed by a 10 meter concealed interval. The concealed interval, in turn, is followed by 21 m of black Utica-type shale which contains numerous calcilutitic limestone bands in the lowest 5 to 7 m of the section (Figure 2). This black shale unit,which yields graptolites of the upper *C. spiniferus* zone (see zonation of Riva, 1969), is distinctly older than basal Utica to the west,.

Southeast of Rathbun Brook is a 9 km wide region of concealed Trenton deposits which ends at the classic "City Brook" section at the southwest end of Wolf Hollow Creek (Stop 4) north of Middleville, New York. At this locality, a 12 m Poland section is abruptly overlain by 8 m of sparsely fossiliferous "ribbon limestone" deposits of the Dolgeville Member with a distinctive corrosional discontinuity at the contact (Kay, 1953: Figure 2). At this section, upper Dolgeville and higher units are concealed. However, at the "County Home-South" section of Kay (1953) south of Middleville, approximately 18 meters of Dolgeville are observed above 17 meters of "Poland Member" and below black shale of the lower/middle Utica ("Upper Canajoharie Shale" of Kay, 1953 (see Figure 3) which yields the graptolite *C. spiniferus* (Kay, 1953; Riva, 1969). At this section, a conspicuous debris layer containing Dolgeville-type micritic intraclasts is observed at the base of metabentonite-rich and phosphatic hash-rich lower Utica beds. Riva (1969) believes a discontinuity is present here (Figure 4), although other authors (Kay, 1953; Cisne *et al.*, 1982; Titus, 1988) do not show any at this level (Figures 1, 3). This contact will be seen on this trip at Stops 5, 6, and 7.

The central problems posed by stratigraphic changes occurring near Rathbun Brook are: 1) the southeastward disappearance of 30 m of Rust strata across a distance of a few kilometers, 2) the southeastward appearance of the lower/middle Utica (*C. spiniferus*-bearing) black shale, and 3) the seemingly sudden appearance of significantly thicker "Russia-equivalent" Dolgeville facies at localities near Norway, City Brook, and Middleville (see Figures 2, 3, 7).

EXISTING INTERPRETATION OF DISJUNCT FACIES CHANGES

Kay's (1953) interpretation for this change is shown in Figure 3; he shows the Steuben disappearing to the southeast due to beveling below Utica. Furthermore, his correlations suggest an abrupt southeastward facies change from Rust bioclastic limestone at Poland to a "Dolgeville-Utica" facies at Rathbun Brook (the micritic "ribbon" limestone-bearing, Utica-type shale above the 10-meter covered interval). This Dolgeville-Utica facies, in turn, grades to black shale (his "upper Canajoharie Shale") further southeast. Although the "Poland" interval at Rathbun closely resembles its "correlative interval" at City Brook, Kay (1953) shows the Russia part of the Denley as having dramatically transformed to Dolgeville facies at this latter section. Subsequent workers (Fisher, 1977, 1979; Cisne *et al.*, 1982; Titus, 1988) also retain some or all of these lateral facies

Figure 2. Generalized Trenton and lower Utica sections, Trenton Falls - Little Falls region based on preliminary correlations. Note the prominent eastward thinning (condensation) of the interval between Units b and d (Wolf Hollow division of present authors) between Rathbun Brook and North Creek as well as the problematic abrupt eastward disappearance of the Rust Limestone Member near Poland. Facies symbols correspond to those explained in Figure 6. Letter units include: a, City Brook bed ("*Trocholites* Bed"); b, North Gage Road bed marking top of Poland Member and base of Wolf Hollow division (see text); c, base of prominent Wolf Hollow condensed cephalopodenkalk facies; d, submarine discontinuity (corrosion surface) marking top of Wolf Hollow condensed facies and flooring Dolgeville deposits (or Brayton Corners' division at loc. 5); e, nodular phosphate-rich concentration marking top of Brayton Corners division at localities 5 and 6; f, base of ribbon limestone facies of upper Dolgeville Member; g, regionally widespread levels of contorted Dolgeville Strata below top - Dolgeville disconformity (Thruway Discontinuity); h, folded limestone bed which marks the base of the Rust Member (*sensu* Kay, 1953); this unit is partially beveled and is overlapped by grainstone beds; i, conspicuous metabentonite and 0.2 meter of overlying dark, graptolite-and-ash-rich shale; j, base of typical rubbly, fossiliferous Rust Limestone deposits; k, corrasional discontinuity (Thruway Discontinuity) flooring lower Utica Shale. Localities include: 1, Trenton Falls Gorge; 2, Mill Creek at Gravesville; 3, creek section south of Brayton Road; 4, creek section paralleling Strumlock Road north of Brayton Corners; 5, Rathbun Brook; 6, Wolf Hollow Creek ("City Brook") east of Old City; 7-9, composite section for east-flowing tributaries of West Canada Creek near the old County Home (loc. 7,8) and northwest of Countryman (loc. 9); 10, creek northeast of Norway; 11a, North Creek east of Kast Bridge, 11b, North Creek south of Dillenbeck Corners.

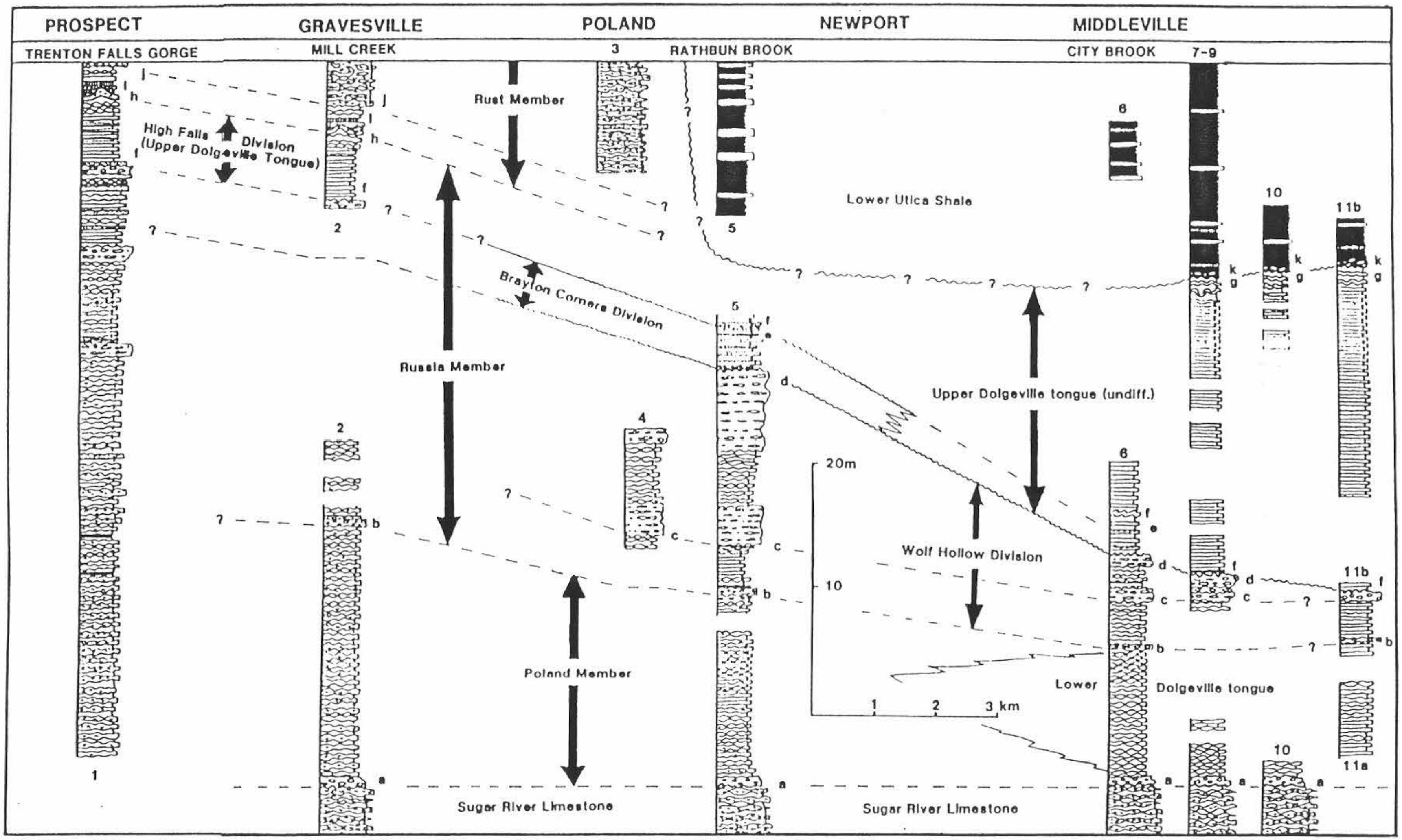

PROSPECT
GRAVESVILLE
POLAND
NEWPORT
MIDDLEVILLE
TRENTON FALLS GORGE
MILL CREEK
RATHBUN BROOK
CITY BROOK
Rust Member
High Falls Division
(Upper Dolgeville Tongue)
Lower Utica Shale
Braylon Corners Division
Russia Member
Upper Dolgeville tongue (undiff.)
Wolf Hollow Division
Poland Member
Lower Dolgeville tongue
Sugar River Limestone
20m
10
1 2 3 km

Figure 2

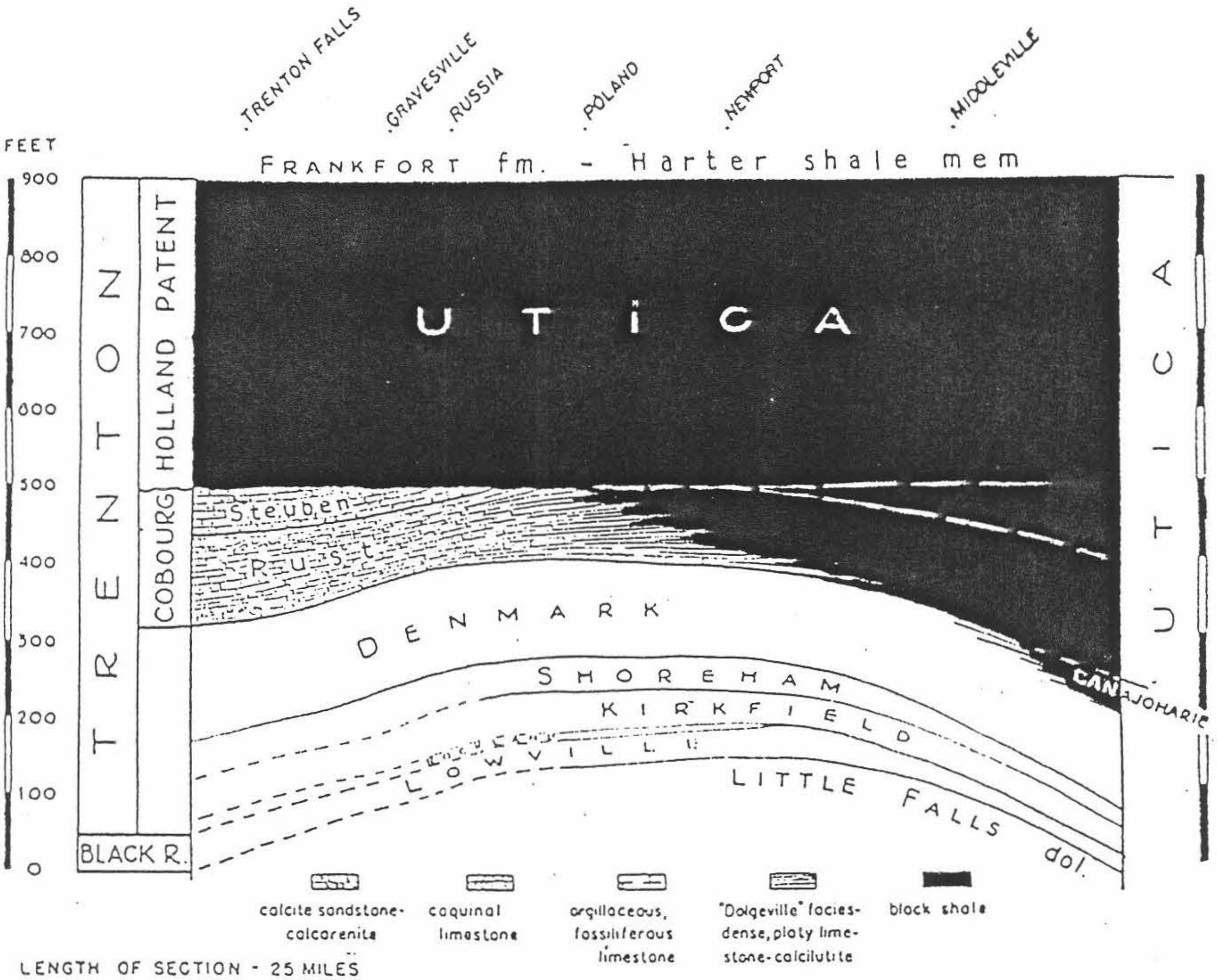


Figure 3. Inferred stratigraphic relationship of Trenton and Utica deposits based on work of Marshall Kay (from Kay, 1953). Note that uppermost Denmark Limestone (Russia Mbr.) and "Rust Limestone" grade eastward into black Utica facies across a very narrow Waltherian facies transition. Note also that the top-Trenton (Steuben-Utica) disconformity is shown to project cryptically into the Utica.

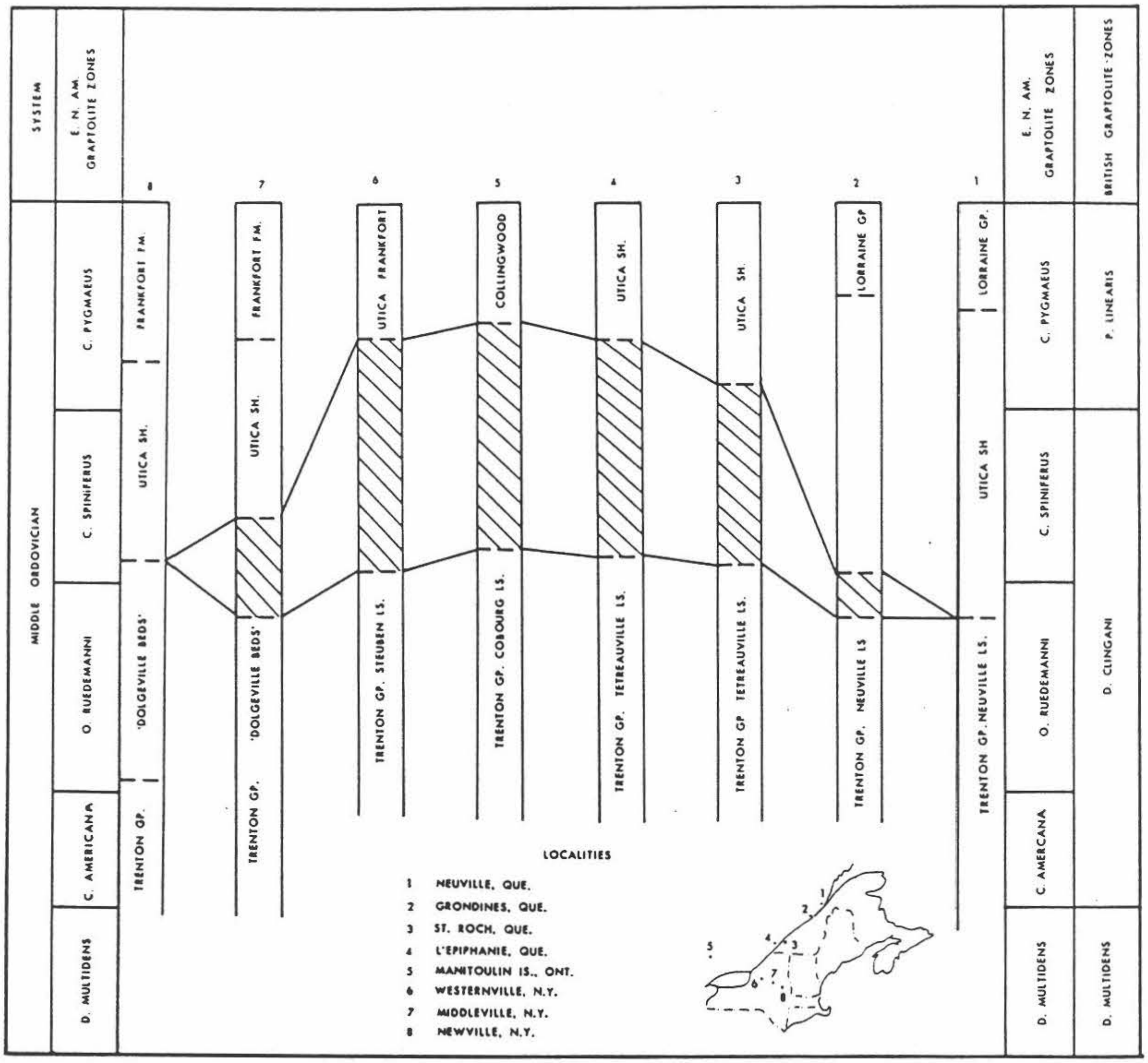


Figure 4. Chronostratigraphic diagram from Riva (1972a) illustrating a major hiatus separating Trenton carbonate deposits from basinal siliciclastic facies which overlap it. Note that this disconformity closes to continuity both northeastward across Quebec and eastward down the Mohawk Valley with the greatest gap separating the top Cobourg Limestone (upper Hillier carbonates) and the Collingwood Shale in Ontario. Yet, interestingly core and outcrop sections from the Toronto, Ontario region display an apparent gradation of facies between the Cobourg and overlying Collingwood strata; we have not recognized a Cobourg-Collingwood unconformity. This disconformity corresponds to our "Thruway Disconformity" and "Regional Sub-Utica Disconformity" discussed herein.

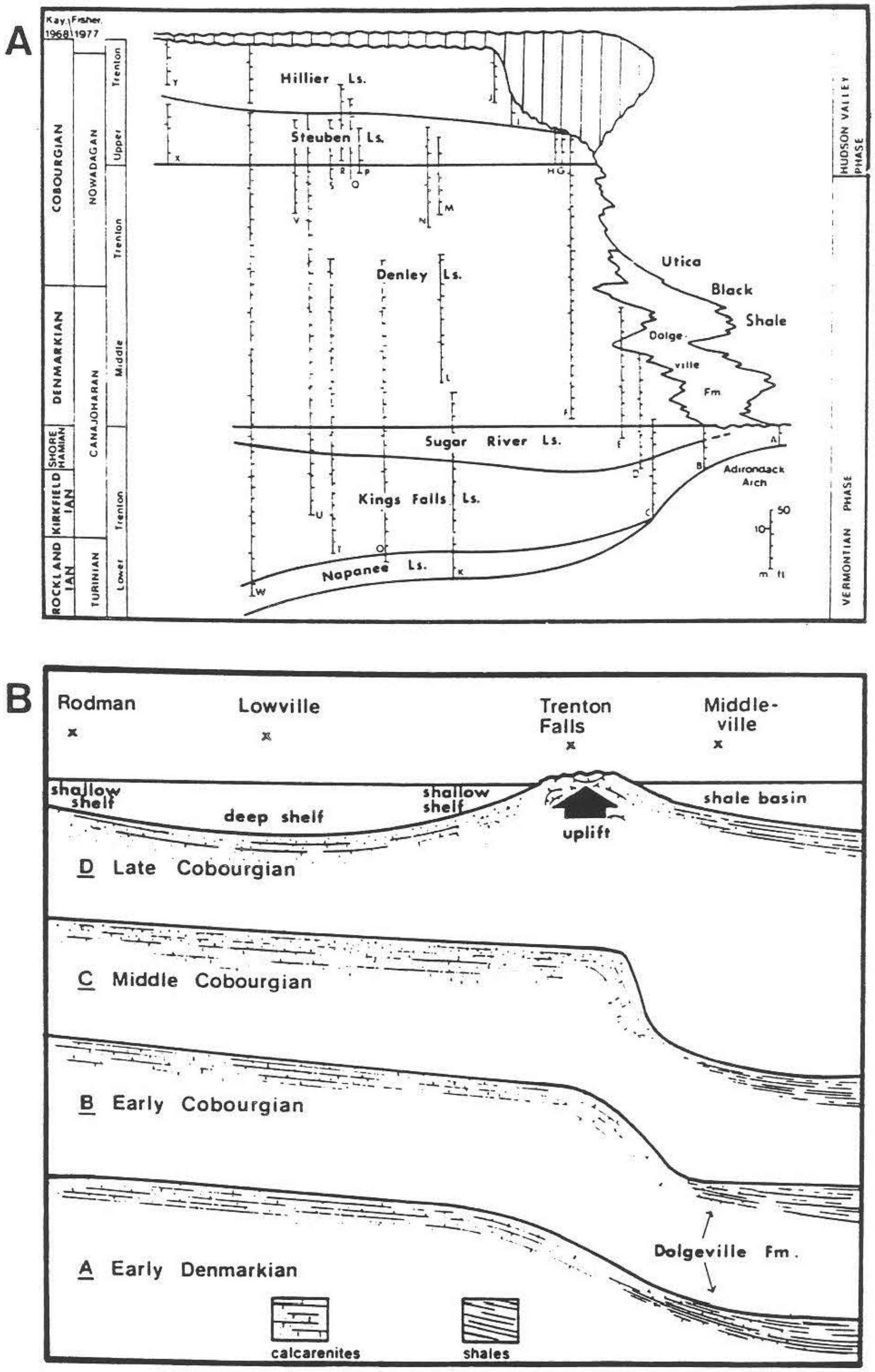


Figure 5. Late Trenton tectonism based on model of Titus (1986). a) Stratigraphy of Trenton Group; outcrops are as follows: A) Canajoharie Creek; B) Ingham Mills; C) Buttermilk Creek; D) Rathbun Creek; E) Mill Creek, Gravesville; F) Trenton Falls; G)Prospect quarry; H) Barnevald quarry; I) Big Brook, Frenchville; J) Pixley Falls; K) Sugar River, Booneville; L and M) Moose Creek; N) Roadcut, Talcotville; O) Mill Creek, Turin; P) Douglass Creek; Q) Whetstone Gulf; R) Atwater Creek; S and T) Roaring Brook; U) Mill Creek, Lowville; V) Black Creek; W) Deer River; X) Gulf Stream, Rodman. From Titus (1988). Note abrupt Denley-Utica facies boundary east of Trenton Falls (loc. F) and maximal development of top Trenton hiatus in same area. b) Evolution of the Trenton carbonate bank margin during Denmarkian (Poland-Russia) and Cobourgian (Rust-Steuben) time. From Titus (1988).

"jumps" in their stratigraphic reconstructions (see Figures 1, 5). If this correlation were correct, the width of the transition Russia into Dolgeville facies would be constrained to a few kilometers at most, and the Rust lower/middle Utica transition would be even more abrupt (Figures 2, 5).

It is possible that these abrupt "facies changes" may reflect synthetic "down-to-the-east" faulting of the type discussed by Bradley and Kidd (1991); such faults would have oversteepened the eastward-sloping submarine ramp through listric displacement, generating submarine escarpments over which storm-turbidite debris would have poured. Cisne and Rabe (1978), Cisne *et al.* (1982) as well as Mehrtens (1988) also evoke faulting as a mechanism for rapid and/or disjunctive facies changes.

Titus (1986) shows the upper Denley (Rust) to lower/middle Utica "transition" west of Rathbun Brook as a sharp facies jump with the implication that it represents a lateral shift from platform margin to basinal settings. (For an alternate platform model, see Surlyk and Ineson, 1992.) He argues that the Rust interval records regressive carbonate aggradation on this platform which culminated in the shallow Steuben platform. Titus (1986) envisions a subsequent episode of localized folding, uplift, and erosional beveling of the Steuben limestone prior to shelf drowning during Hillier time followed by deposition of Utica black muds across the entire region (see Titus, 1988: Figure 4).

PRELIMINARY OBSERVATIONS FROM CURRENT MAPPING PROJECT

Middle Trenton-Canajoharie Shale

The present authors have identified and traced the basal Poland marker bed (City Brook bed) from Lowville southeastward to a ravine near Countryman, north of Kast Bridge. This 0.7 to 0.2 m thick unit is a micritic nobbly limestone, rich in cephalopods and trilobites. The City Brook bed contrasts strongly with the tempestitic grainstones and packstones of the underlying Sugar River Limestone.

Beginning in the vicinity of Norway, Middleville, and Countryman (Figure 11), a thin, trilobite debris-and pelmatozoan-rich hash layer appears above the City Brook bed and below normal Poland Member micritic beds. At the creek northeast of Norway, this layer is sandy. At Countryman, it yields quartz granules and sand, as well as abundant phosphatic bioclasts. Southeast of Countryman, this hash layer yields abundant phosphatic debris and reworked limestone clasts; it marks a discontinuity which has beveled through the City Brook bed and into the Rathbun Member of the Sugar River Limestone. This discontinuity has now been traced from the vicinity of Little Falls (Rt. 5S roadcut east of Jacksonburg, NY) eastward to the vicinity of Cranesville, east of Amsterdam, New York. We will see the City Brook bed at Wolf Hollow Creek (Stop 4).

Herein, we present tentative revisions of Russia and Rust strata (Table 1; Figures 2, 6, 7). The basal Dolgeville corrosional discontinuity recognized at City Brook by Kay (1953) has now been traced southeastward to Gun Club Road, west of Little Falls, through three intervening outcrops. More significantly, this boundary is now traceable westward to Rathbun Brook (Figures 2, 6). At Rathbun Brook, this horizon is represented by a corrosion surface which is littered with phosphatized nautiloid steinkerns; this horizon is 3 m below the top of the visible Trenton section. At Trenton Falls, however, this contact is difficult to place.

These observations indicate that the Dolgeville Member of the Norway-Countryman area is equivalent to deposits no older than the Brayton Corners division, although the upper part of the Dolgeville may be Rust-equivalent (see discussion below). It also means that the Poland through Brayton Corners interval thins from approximately 37 m at Trenton Falls to 16.5 m at Wolf Hollow Creek. This occurs through southeastward stratigraphic condensation with most condensation occurring in the post-Poland part of the section (Wolf Hollow and Brayton Corners divisions; see

Figures 2, 6). Careful mapping of marker beds and sedimentary cycles in the Russia Member shows that stratigraphic units converge southeastward. In particular, the Wolf Hollow division grades southeastward from predominantly bedded bioclastic packstones and wackestones at Trenton Falls to predominantly nodular, cephalopod-rich, shaley micrites at Wolf Hollow Creek. These observations, thus, obviate the need for an abrupt Russia to Dolgeville facies change between Rathbun Brook and Wolf Hollow Creek (Stop 4) as inferred by Kay (1953); (compare Figure 3 and Figures 2, 6, 7).

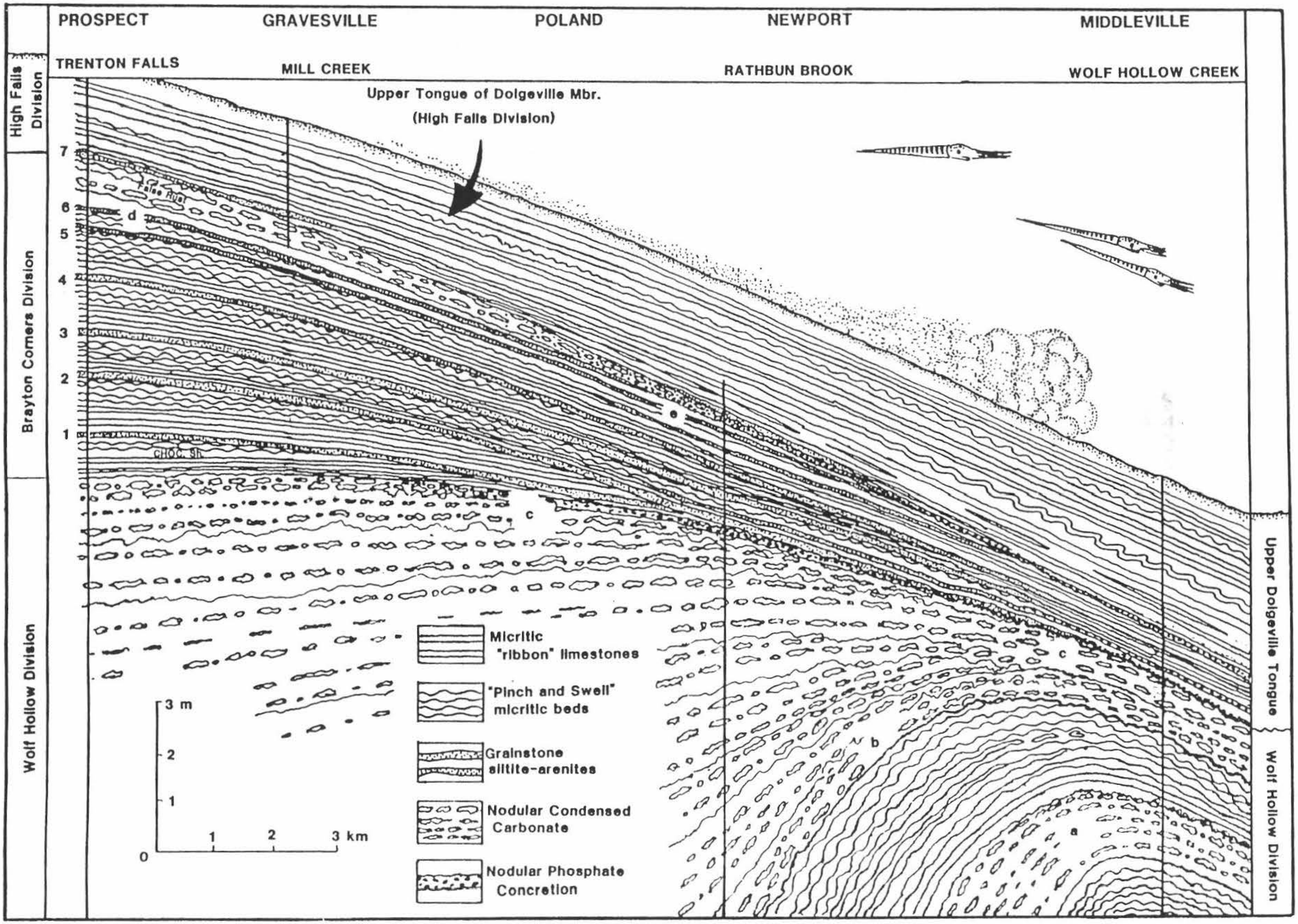
Conspicuous in the Poland - Russia interval are cyclic alternations of bedded biomicrites and nodular cephalopod-rich deposits. These constitute 4th or 5th order-scale sedimentary cycles which might record relative changes in water depth (Figure 8). Condensed nodular limestones record conditions of sediment-starvation and aggradation, possibly related to initial deepening episodes followed by early highstands, respectively. Overlying tabular calcilutites and metabentonite-bearing dark shales pass upward into burrowed and sediment-winnowed lowstand deposits. Lowstand deposits, in turn, pass upwards back into nodular, cephalopod-rich limestones. The North Gage Road bed, two higher condensed levels within the Wolf Hollow division, a condensed bed at the top of the Brayton Corners division, and a similar bed marking the base of the Rust Member constitute the caps of five such sedimentary cycles. These cycles, critical to correlation and facies interpretation, are the subject of ongoing study.

In addition to the pattern of southeastward thinning of the Poland through Brayton Corners deposits, we observe that the upper Dolgeville interval thickens from 5 meters at Trenton Falls to approximately 22-24 meters in the Middleville-Kast Bridge area with no overlying Rust Member visible. We do not know if this thickening is entirely due to stratigraphic expansion within Dolgeville or to eastward appearance of progressively higher Dolgeville beds from beneath a sub-Rust or sub-Utica unconformity. This question cannot be answered at present because no exposures of the Dolgeville-Rust contact exist southeast of Gravesville. Close study of the Dolgeville-Rust contact is needed at Trenton Falls and to the north of there (see "REMAINING PROBLEMS").

Upper Trenton Unconformity

The Steuben is definitely overstepped by Utica towards the southeast, and the underlying Rust probably suffers the same fate (Figures 2, 7). Although the Trenton-Utica contact is nowhere exposed between South Trenton (Stop 1) and sections near Norway and Middleville, 17 km east of there, visible Rust sections appear to thin dramatically southeast of Poland, and no Rust lithology is observed at Rathbun Brook. We believe that this loss of Rust is a continuation of southeastward overstep of Trenton beds, beginning with the Hillier "chop out" near Boonville and the Steuben overstep southeast of South Trenton. It is significant that a nearly complete 30-meter Rust section,

Figure 6. Inferred downslope facies change within upper Russia Member (Wolf Hollow division into upper Dolgeville stratigraphic interval) across West Canada Creek region (see Stops 3,4). Note prominent eastward-southeastward thinning (condensation) of nodular, cephalopod-rich Wolf Hollow division with appearance of top-Wolf Hollow discontinuity (corrosion surface) at Rathbun Brook (Stop 3) and Wolf Hollow Creek (Stop 4). The overlying Brayton Corners division, displaying a repetitive alternation of micritic limestone beds and tabular grainstone layers at Trenton Falls, thins southeastward through a stacked succession of ribbon grainstones at Rathbun Brook to a 1.3 meter interval of Dolgeville-type ribbon micrites with associated grainstone to calcisiltite layers and nodular phosphate concentrations. Lettered units include: a, lower condensed, cephalopod-rich interval in Wolf Hollow division; b, base of upper condensed, cephalopod-rich interval in Wolf Hollow division; c, phosphate-rich zone at top of Wolf Hollow which changes southeastward to a conspicuous corrosion surface near Middleville; d, recurrent condensed nodular limestone bed which marks top of Brayton Corners division; e, nodular, shaley limestone bed rich in nodular phosphate and phosphatized nautiloid steinkerns. This correlates upslope to unit d and downslope of phosphorite-rich micritic ribbon limestone bed in Dolgeville Member at Wolf Hollow Creek. Numbers correspond to possible 6th-order minor cycles in Brayton Corners division.



PROSPECT
GRAVESVILLE
POLAND
NEWPORT
MIDDLEVILLE
TRENTON FALLS
MILL CREEK
RATHBUN BROOK
WOLF HOLLOW CREEK
Upper Tongue of Dolgeville Mbr.
(High Falls Division)
High Falls Division
Brayton Corners Division
Wolf Hollow Division
Upper Dolgeville Tongue
Wolf Hollow Division
CHOC. Sh
Micritic "ribbon" limestones
"Pinch and Swell" micritic beds
Grainstone siltite-arenites
Nodular Condensed Carbonate
Nodular Phosphate Concretion
3 m
3 km

Figure 6

including Walcott's "zone of trilobites" and two characteristic zones of deformation (see Stop 1), is still present northwest of Poland. South of Poland one can observe small sections of characteristic Rust lithology which, presumably, represent lower Rust strata, but these outcrops are too poor to confirm this. In short, lines of evidence for southeastward overstep of Rust are circumstantial but strongly suggestive; these include: aforementioned overstep of Hillier and Steuben, lack of significant southeastward facies change within Rust to different facies, and rapid disappearance of Rust sections between Poland and Rathbun Brook (Figures 2, 7).

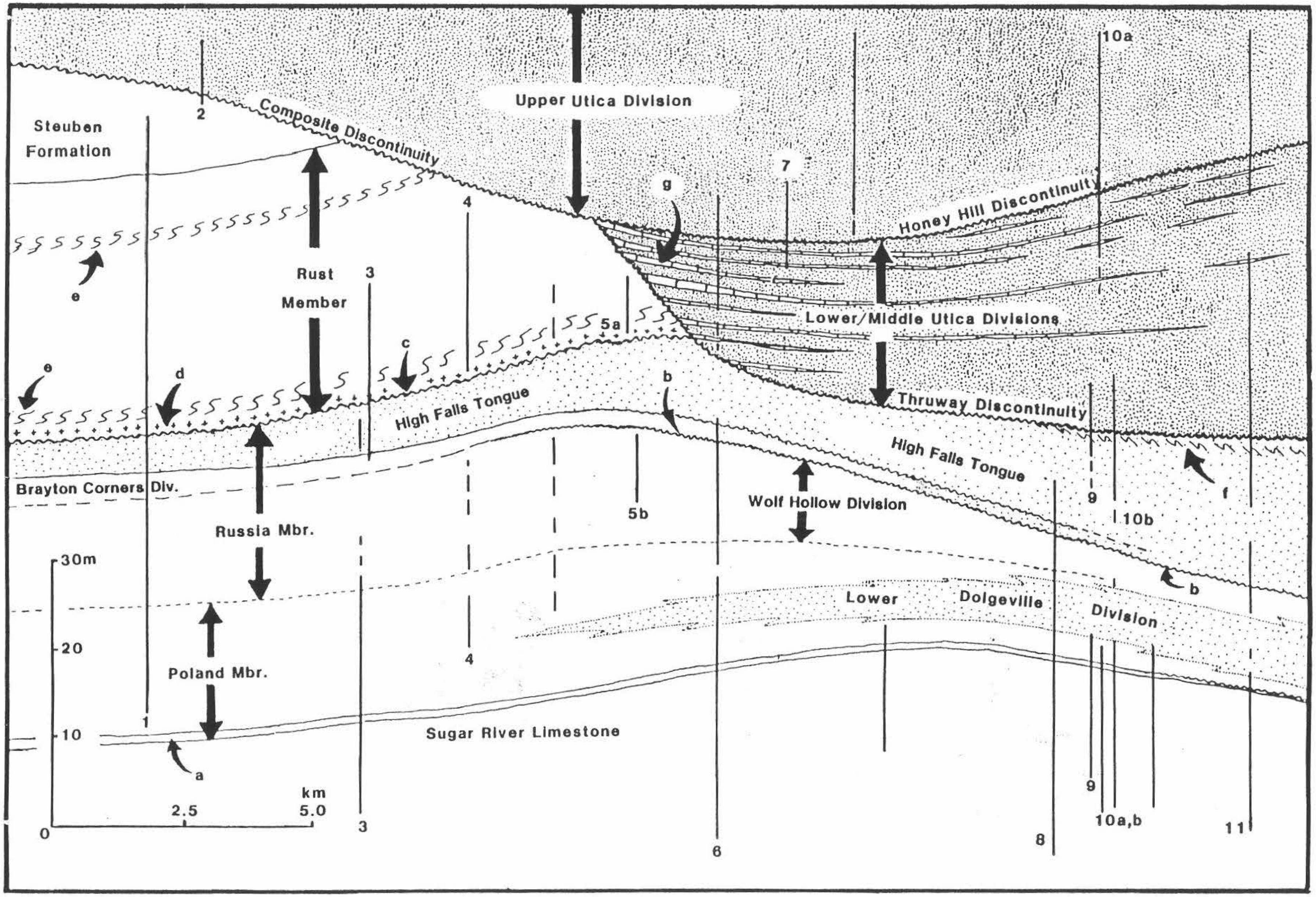
Utica Shale Divisions

It has long been known that an erosional contact floors Upper Ordovician black shale and siltstone deposits between Watertown and the vicinity of South Trenton (Stop 1). Similarly an aforementioned sharp contact above folded Dolgeville beds and below Utica-type facies has been noted for sections between Middleville and the New York State Thruway south of Little Falls (see Stops 5 and 6), but it has been relegated to greater or lesser significance by different authors (see discussions in Kay, 1953; Riva, 1969; Fisher, 1979; Cisne *et al.*, 1982). This latter contact which floors Utica facies yielding the zonal graptolites *Dicranograptus nicholsoni* and *Clmiacograptus typicalis* is now traceable from the small creek northeast of Norway to East Canada Creek and Nowadaga Creek 16-20 km to the southeast (Figures 2, 7). In most localities it overlies a conspicuous zone of penecontemporaneously deformed Dolgeville beds (see discussion: Stops 5, 6, and 7) and it is characterized by a phosphatic lag deposit with conspicuously shingled, corroded, and reworked Dolgeville micrite fragments.

This contact, herein informally designated the "Thruway discontinuity" for the long New York State Thruway section south of Little Falls (see Stop 6) is a major disconformity which marks a hiatus encompassing one or more graptolite zones (see Riva, 1969). Its eastern and western limits are unknown; we presently recognize it as extending from Newville and Dolgeville in the Mohawk Valley northwestward to the Watertown area (Figures 4, 7, 9).

At Rathbun Brook and in gullies to the southeast of it, the basal exposed Utica is a somewhat calcareous, laminated black shale characterized by buff-weathering micritic "ribbon" limestones which vary greatly in thickness and spacing within the black shale. The associated shale is rich in graptolites of the upper *Climacograptus spiniferus* zone, and bedding planes

Figure 7. Inferred relationship of Trenton and Utica stratigraphic divisions in region between Prospect and Kast Bridge. Eastward loss of Upper Trenton deposits (Rust and Steuben strata) is attributed to erosional overstep prior to deposition of Utica black muds. Anomalously abrupt overstep of Rust Member west of Rathbun Brook (Loc. 6) is problematic; thick ribbon limestones in lower/middle Utica deposits at Rathbun Brook may indicate the presence of a steepened east-facing, erosional ramp west of that locality which fed thick carbonate turbidites into the Utica basin as the ramp was being buried. A higher erosion surface within the Utica (Honey Hill Discontinuity) is shown to truncate both lower/ middle Utica beds as well as part of the upper Trenton (see text). This submarine erosion scenario for explaining the abrupt eastward loss of the upper Trenton is favored by the present authors (see text) but an alternative, fault-based explanation is presented in Figure 10. Lettered units include: a, City Brook bed (*"Trocholites "* Bed); b, corrosional discontinuity marking top of Wolf Hollow division; c, folded and partly truncated limestone bed marking base of Rust Member; d, prominent metabentonite in basal Rust interval; e, zones of structurally deformed strata in Rust Member; f, zone of disturbed beds in uppermost Dolgeville Member - this disturbance may correlated with units c, d, or e; g, unusually thick ribbon limestones observed at Rathbun Brook. Localities include: 1, Trenton Falls Gorge; 2, Ninemile Creek at South Trenton (Stop 2); 3, Mill Creek at Gravesville; 4, east fork of creek east of Beecher Road, 2 km southeast of Russia; 5a, creek section south Brayton Road; 5b, creek section paralleling Strumlock Road north of Brayton Corners; 6, Rathbun Brook; 7, second gully north of Honey Hill Road; 8, Wolf Hollow Creek ("City Falls"); 9, creek northeast of Norway; 10a, creek north of the County Home; 10b, creek south of the County Home; 11, east-flowing tributary of West Canada Creek at Countryman. Unpatterned= Trenton bioclastic limestones. Loose stippling= turbiditic ribbon limestone facies. Dense stippling= black shale deposits.



Upper Utica Division
Steuben Formation
Composite Discontinuity
Rust Member
Honey Hill Discontinuity
Lower/Middle Utica Divisions
Thruway Discontinuity
High Falls Tongue
High Falls Tongue
Brayton Corners Div.
Russia Mbr.
Wolf Hollow Division
Lower Dolgeville Division
Poland Mbr.
Sugar River Limestone
30m
20
10
0
2.5
km
5.0
1
2
3
4
5a
5b
6
7
8
9
10a
10b
10a,b
11
a
b
c
d
e
f
g

Figure 7

are often coated with phosphatic fossil hash. At Rathbun Brook the micritic limestones are anomalously thick and numerous, but thinner, equivalent beds can be seen in the Utica southeastward to the Little Falls area. At South Trenton, the Utica Shale yields the younger graptolite *Climacograptus pygmaeus* and it distinctly lacks ribbon limestones (see Stop 1).

The top of the ribbon limestone-bearing lower Utica division is now found to be marked by an erosion surface and an associated lag deposit which we informally designate the "Honey Hill Road bed" for exposures on an unnamed creek north of Honey Hill Road and northeast of the Newport Golf Course (Figures 7, 9). The Honey Hill Road bed is notable for locally thick lag concentrations of problematic phosphatic rod-like structures which are similar to those observed in the top-Trenton lag deposit at South Trenton (see Stop 1). The associated discontinuity is of unknown temporal magnitude but it may mark the *C. spiniferus - C. pygmaeus* zonal change. At Rathbun Brook this contact is probably 15-18 meters above the Thruway discontinuity which we believe is concealed under the 10-meter covered interval on that creek (Figures 2, 7, 9).

Above the Honey Hill Road bed at Rathbun Brook and along creeks near Middleville and at Countryman, the upper Utica is typically lacking in ribbon micrites; hence it is lithologically distinct from the lower/middle Utica succession below this contact. At the northern "County Home" section near Middleville and at the creek at Countryman, the lower/middle Utica interval is at least 30 m thick (Figure 7). Still further east it progressively thickens above the Thruway discontinuity. We believe that the micrite ribbon-bearing lower Utica division is overstepped by the Honey Hill Road discontinuity west of Rathbun Brook such that the regional sub-Utica disconformity at South Trenton reflects the additive merging of both hiatuses (Figures 7, 9).

INFERRED MIDDLE TRENTON AND UTICA EVENTS: UTICA-CANAJOHARIE REGION

Medial Trenton stratigraphy is suggestive of an eastward-sloping carbonate ramp which, at times, was oversteepened. Within the field trip area, conditions recorded by the Poland Member and the Wolf Hollow division of the Russia Member indicate a relative deepening of Trenton seas and increased steepening of the carbonate ramp from the earlier higher energy Sugar River shelf conditions. This sea level deepening brought on sediment-starvation, recorded by the condensed City Brook bed and the discontinuity which removed this bed in downslope areas. Many higher Russia strata resemble the City Brook bed and record similar processes. These nodular limestones are analogous to cephalopodenkalk facies present in younger systems (Tucker, 1973; Jenkyns, 1971; Baird and Brett, 1986a) Cephalopodenkalk facies typically record slow sediment accumulation, intense bioturbation, and cephalopod accumulation under predominantly quiet, minimally oxic to dysoxic conditions below storm wave base. Poland and Russia strata grade downslope to tabular micrites and interbedded dark shales (Dolgeville Member). For example, lower (Poland-equivalent) Dolgeville deposits east of Little Falls are turbidites of lime and siliciclastic mud which accumulated in a more basinal dysoxic environment. Further east, equivalent black shales (lower Canajoharie Shale) record dysoxic to anoxic basinal conditions presumably still further downslope. A similar facies gradation is present in lower Wolf Hollow-equivalent Dolgeville deposits. The presence of mildly bioturbated mudstone and muddy limestone beds yielding a sparse fossil benthos above Canajoharie Falls suggests that the lower strata of the Wolf Hollow division may have a lithologic signature in the Canajoharie Shale.

The vertical change from Wolf Hollow deposits through Dolgeville facies records an episode of relative sea level rise with consequent northwestward (up-ramp) advance of dysoxic bottom conditions. It is significant that a condensed interval of nodular phosphate, ribbon calcarenites and calcisiltites (Brayton Corners division) marks the boundary between Dolgeville facies and the Wolf Hollow interval at Rathbun Brook (Stop 2). In the Little Falls area, a corrosional-abrasional, phosphate-coated discontinuity underlies true Dolgeville facies at this horizon, and the Brayton Corners division is absent (Figures 2, 6, 7).

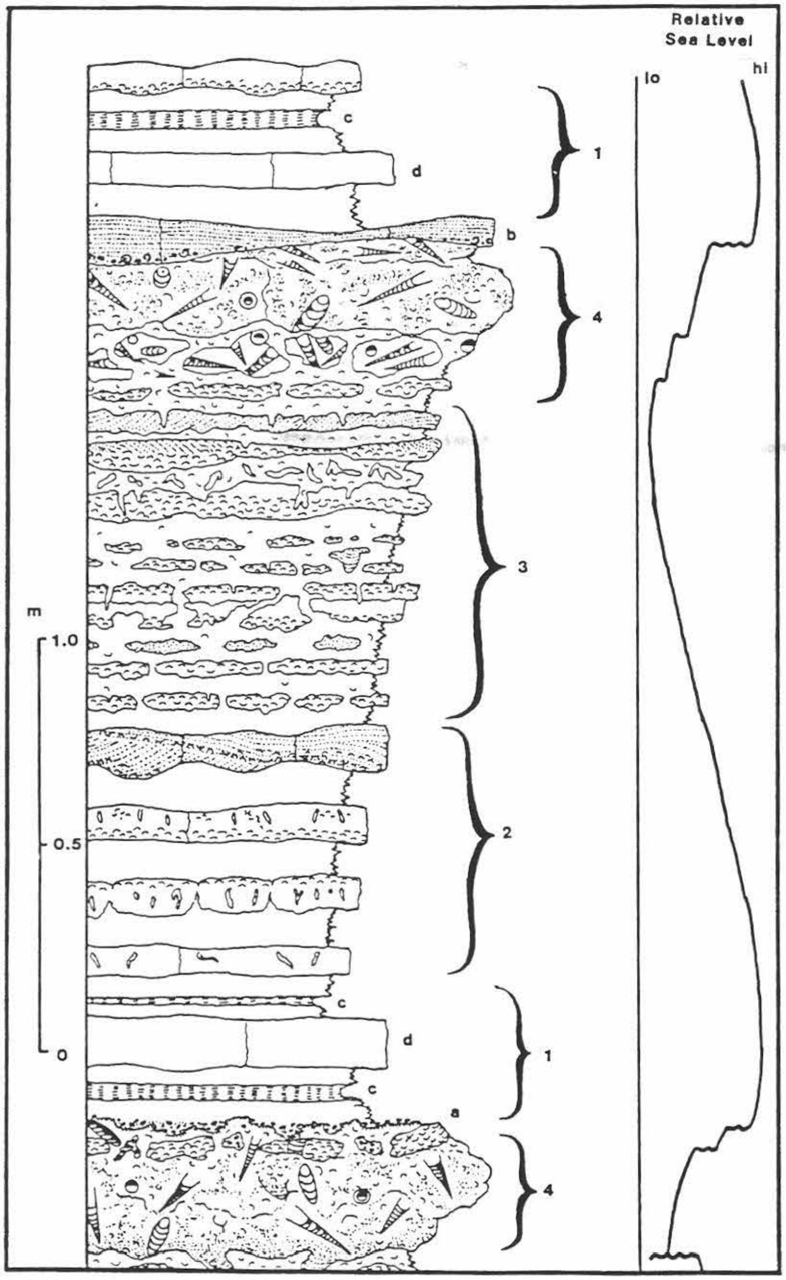


Figure 8. Generalized sedimentary cycle typical of the Wolf Hollow division and overlying Brayton Corners division (see Stops 3,4). Cycle magnitude shown is 5th order but complex larger-scale cycles show a similar facies spectrum. Sea level is relative; both eustatic and tectonic factors probably controlled cycle motif during Trenton deposition. Depth-related facies include the following: 1, early highstand, micritic limestones and dark shales yielding volcanic ashes and low diversity benthos (dysoxic conditions below storm wavebase); 2, late highstand - early sea level fall deposits showing greater evidence of benthonic activity; 3, late sea level fall - to - lowstand interval recording influence of storms, extensive bioturbation, richer benthos, and increased condensation due to sediment bypass; 4, condensed limestone facies recording relative sea level rise and associated conditions of sediment starvation. Erosion (and corrosion) surfaces capping condensed limestones correspond to surfaces of maximum flooding in cycle. Note that condensed limestones record a complex internal history of winnowing, erosional scour episodes, and multiple diagenetic events. Lettered units include: a) Phosphate - cemented corrosion surface littered with phosphatic nodules and phosphatized fossil steinkerns; b, grainstone - siltite - arenite lag bed on discontinuity which contains phosphatic gastropod steinkerns; c, metabentonites (volcanic ash units); d, prominent tabular calcilutite bed which is often observed above the maximum flooding surface.

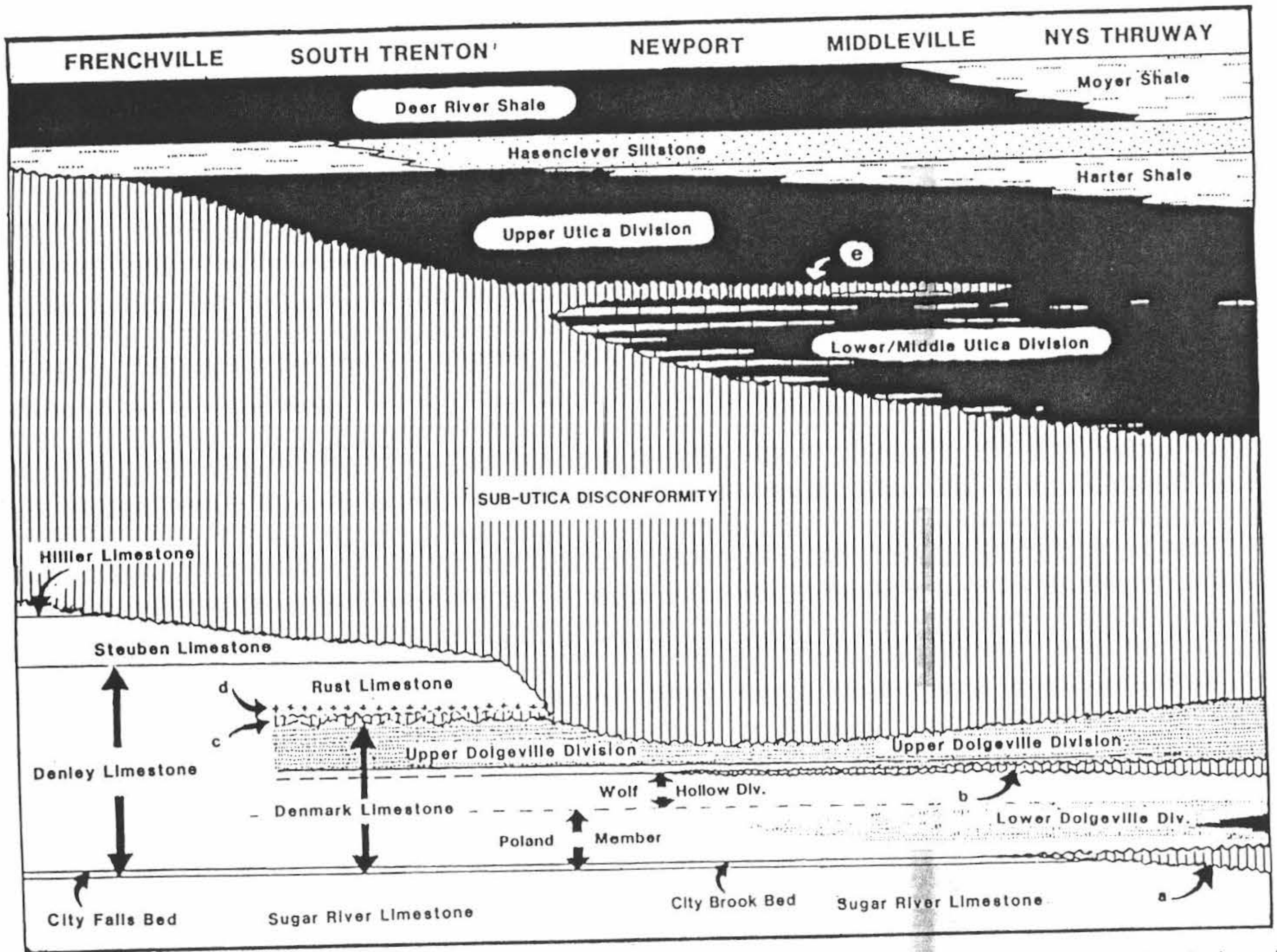


Figure 9. Middle Trenton to "Utica" chronostratigraphy, central New York region. Inferred time-magnitude of sub-Utica disconformity is conjectural but is believed to be large in this region. Note abrupt eastward erosional loss of upper Trenton deposits near Poland and progressive westward overlap of Utica deposits over progressively younger Trenton units across the region (see text). Hiatus associated with Honey Hill Discontinuity probably thins eastward to continuity into the foreland basin. Horizontal stipple pattern denotes turbiditic ribbon limestone facies of Dolgeville Member. Non-patterned area corresponds to bioclastic limestone deposits of Trenton Group. Black denotes laminated, organic-rich, black facies of Utica Shale. Discontinuities labeled as follows: a, sub-Denley discontinuity; b, Brayton Corners discontinuity; c, sub-Rust discontinuity; d, previously undescribed metabentonite in lower Rust; e, Honey Hill Discontinuity.

The downslope thinning of the Brayton Corners beds, followed by their truncation beneath basinal facies further downslope, is similar to phenomena we have observed in the Devonian Genesee Formation (Baird *et al.*, 1988). Our observations suggest that submarine erosion on Devonian sediment-starved slopes is most pronounced in the lower slope regime where dysoxic and anoxic basinal waters impinge upon the bottom. Downslope truncation of Devonian Wolf Hollow carbonate analogs--the Fir Tree and Lodi limestones--apparently occurred under conditions of transgressive sediment-starvation, carbonate undersaturation and consequent carbonate dissolution, as well as episodic current (storm- or internal wave-generated) impingement on substrate (Baird *et al.*, 1988; Baird and Brett, 1990). Contacts such as the Wolf Hollow - Dolgeville contact, the Trenton - Canajoharie discontinuity, the Thruway discontinuity, and the Honey Hill Road discontinuity probably all owe their origin to these processes.

The change from Dolgeville through Rust and into Steuben lithology marks a significant interval of marine shallowing from dysoxic conditions below storm wave base (Dolgeville) to oxic conditions within the reach of storm waves but below fair weather wave base (Rust) into high energy marine shoal facies (Steuben). This was not a uniform regression; in late Rust time, deposits resembling the proximal Dolgeville and some parts of the Wolf Hollow interval reappear within the normally rubbly, diverse-fossil facies of normal Rust. These beds, which were "mined" for trilobites by Walcott (1875a, b, c) and later workers, mark a minor transgression within the overall shallowing succession.

The eastward erosional loss of upper Trenton units across the study area leaves to speculation the character of the upper Trenton strata in eastern New York. We suspect that the Rust would have graded into downslope Dolgeville- and Canajoharie-type black shale facies reflecting persistence of the Albany Basin to the east. However, recent mapping by one of us (Lehmann) indicates that the Steuben becomes more grainstone-rich toward that unit's southeastern erosional limit. This opens up an interesting speculation that an epeirogenic upwarp (peripheral bulge?) may have developed in the western Mohawk Valley region at this time prior to its later collapse or corrasional beveling during late Hillier and early/middle Utica time. A variation of this theme is presented by Titus (1986, 1988) who argues that a discrete Trenton platform margin formed during Steuben time and that this margin was mildly folded during a disturbance timed with the Hudson Valley phase of the Taconic Orogeny. He argues that folded Steuben beds underlie flat-lying Utica shale at the Holland Patent locality on Nine Mile Creek west of Stop 2.

The problem of assessing the fate of upper Trenton strata east of the Poland - Newport area may be partly solved through future identification and location of Rust-, Steuben-, and lower Hillier-equivalent beds where they emerge beneath the Thruway discontinuity. Riva (1969) shows Trenton through Utica strata as nearing continuity at Nowadaga Creek, southeast of Little Falls (Figure 4). Yet, the discontinuity is still physically manifested on that stream, suggesting that discontinuity closes still further east. In any case, where continuity is approached, regressive shoal facies of the Steuben should perhaps be expressed as a tongue of Dolgeville facies which extends eastward into otherwise undifferentiated Canajoharie Shale.

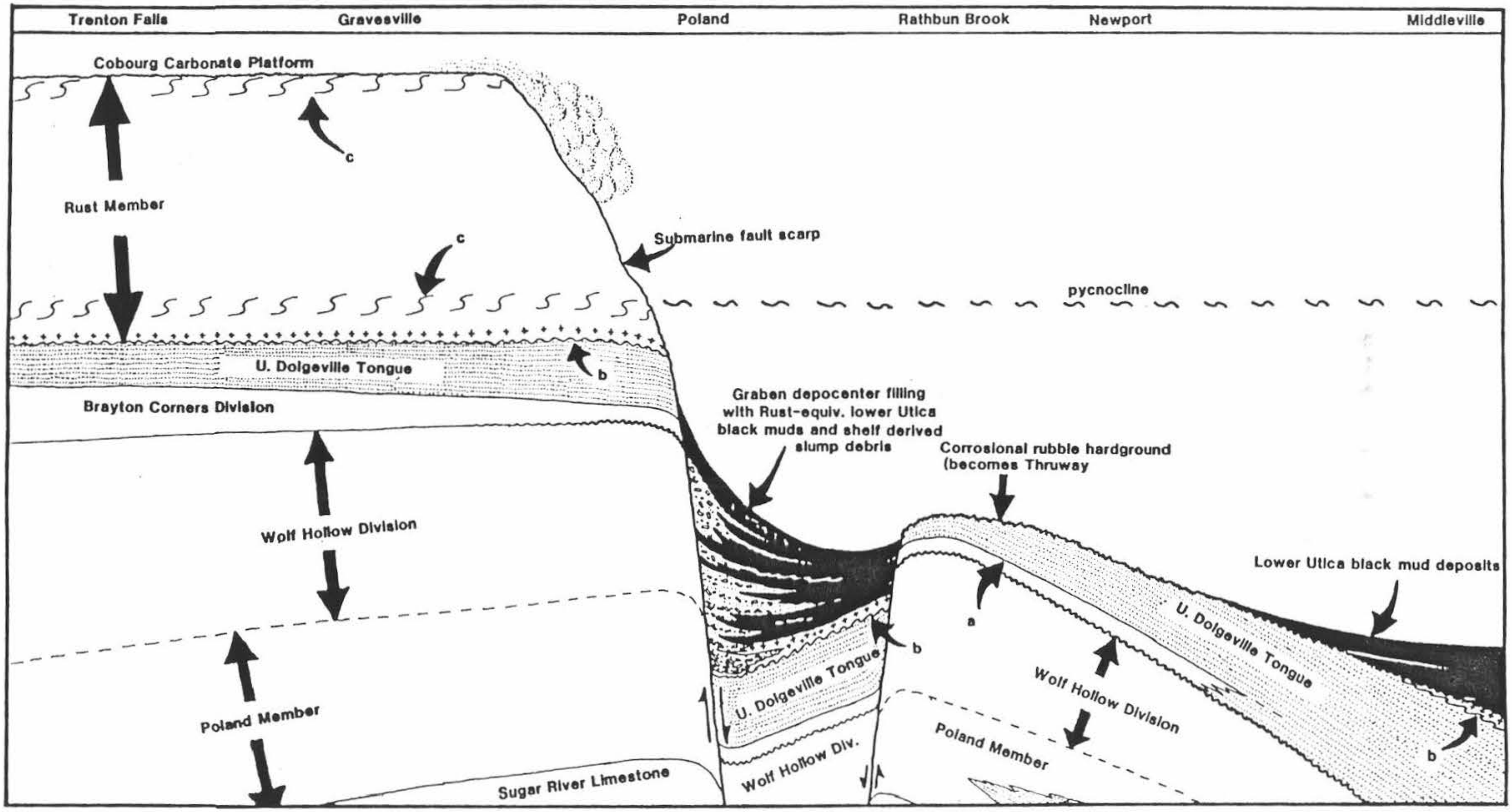
The Thruway discontinuity is characterized by westward overlap of lower/middle Utica beds suggesting that the erosion surface was a gently sloped east-facing ramp, which was progressively buried by black muds as the basin evolved. Sediment-starvation, episodic current activity, and pervasive anoxia and/or carbonate undersaturation sustained bottom erosion and corrosion to produce the distinctive lag of phosphatic debris and micrite intraclasts common this surface. Presumably this corrasion would have removed variable portions of the Upper Trenton over millions of years of time. Unusual steepness of this ramp near Rathbun Brook could explain both the rapid disappearance of lower/middle Utica beds west of that creek and the unusual development of thick ribbon micrites in the lower/middle Utica at that locality (Figures 2, 7); such thick beds would record turbiditic carbonate "wash off" into the Utica basin from the steeper slope. However, for an alternative tectonic explanation for the abrupt stratigraphic changes near Rathbun

Brook, see "REMAINING PROBLEMS" and Figure 10.

Overlap of lower/middle Utica muds advanced for an unknown distance to the northwest of the Utica area before a later episode of erosion (associated with the Honey Hill discontinuity) removed them as well as additional Steuben beds (Figures 7, 9). This second discontinuity surface was then overlapped to the northwest by younger black muds of the lower part of the *Climacograptus pygmaeus* zone (upper Utica). This overlap is dramatic, continuing northwestward past the Watertown area such that siltstones equivalent to the Frankfort Formation lap onto the top-Trenton contact at Frenchville, NY, and even younger Utica-type facies of the Deer River Shale (upper *Climacograptus pygmaeus* zone) and the Blue Mountain Shale (*Climacograptus manitulinensis* zone) lap onto this surface still further northwest (Lehmann and Brett, 1991a). It is, as yet, unknown as to where the regional sub-Utica disconformity closes to continuity, but core and outcrop data suggest continuity in the Toronto, Ontario region, with closure occurring within the upper Hillier Limestone (upper Lindsay of Ontario) or above it.

In summation, the overall cratonward drift and facies succession from middle Trenton to upper Utica reflects the behavior of a peripheral foreland basin during orogeny; successive emplacement of allochthon slices and later thrust masses in the Hudson Valley region as well as synchronous filling of the basin with flysch would have led initially to sinking of the Trenton shelf from east to west. This was followed by progressive overlap of basinal muds (Canajoharie-Utica Shale facies) in the same direction. The development of a sediment-starved east-sloping ramp would have led to dissolution and abrasion of exposed Trenton carbonate as water depth, carbonate undersaturation, and depth-related anoxia increased; this would have produced the discontinuities discussed herein and it may have led to the removal of much of the upper Trenton strata. This mirrors the Taghanic onlap event of Middle to Late Devonian age where foreland basin migration and deltaic progradation were key responses to the Acadian Orogeny (Ettensohn, 1987). This makes the "Utica Shale" with its subjacent unconformities a close analog to the Devonian Geneseo, Ohio, Antrim, and Chattanooga black shale complex with its subjacent overlap surfaces (Baird and Brett, 1986a, 1986b, 1990).

Figure 10. Alternative model for explaining abrupt eastward disappearance of upper Trenton deposits (Rust Mbr.) by evoking movement of linked synthetic-antithetic faults (*sensu* Bradley and Kidd, 1991). Fault movement is purely conjectural as no fault system has yet been found between Poland and Rathbun Brook, but stratigraphic changes within the upper Trenton (see text) could be explained by the sense of fault movement depicted. Fault motion is shown to have commenced during deposition of the basal Rust disturbed layer and overlying thick metabentonite with ensuing shelf collapse and development of a southeast-facing submarine escarpment. Laminated black lower Utica deposits are shown as time-equivalent with Rust platform deposits; they accumulated under anoxic conditions along with turbiditic ribbon limestones and inferred olistostromes in the graben and with only minor allodapic carbonate further east in the foreland basin. Corrosion and abrasion of Dolgeville deposits on the southeastern upthrown block would have occurred under sediment-starved, oxygen-deficient conditions to produce an intraclast-littered rubble hardground; diachronous overlap of this surface would have produced the Thruway Discontinuity. This model explains both the abrupt eastward "replacement" of Rust by lower Utica and the seemingly thin 10-meter covered interval between the Brayton Corners division and lower Utica deposits at Rathbun Brook. Lettered units include: a, transitional Denmark - Dolgeville deposits (Brayton Corners div.) east of graben; b, basal Rust folded and truncated bed, plus overlying metabentonite which is marked by +++ symbol - note that these beds intertongue with lowest olistostrome and that the ash hypothetically reappears under the Thruway Discontinuity at the lower right; c, two key levels of disturbed strata (deformed zones) in the Rust Member - these would be linked to seismic activity on the adjacent active fault both during or immediately following Rust deposition.



Trenton Falls
Gravesville
Poland
Rathbun Brook
Newport
Middleville
Cobourg Carbonate Platform
c
Rust Member
c
Submarine fault scarp
pycnocline
U. Dolgeville Tongue
b
Brayton Corners Division
Graben depocenter filling with Rust-equiv. lower Utica black muds and shelf derived slump debris
Corrosional rubble hardground (becomes Thruway
Wolf Hollow Division
Lower Utica black mud deposits
a
U. Dolgeville Tongue
b
U. Dolgeville Tongue
Wolf Hollow Division
Poland Member
Poland Member
Wolf Hollow Div.
Sugar River Limestone
b

Figure 10

REMAINING PROBLEMS

Several outstanding stratigraphic problems exist which constrain our knowledge of upper Trenton depositional events. These questions include the following:

What is the fate of the Rust Member southeast of Poland? Apparently this unit is overstepped to the southeast by the lower/middle Utica Shale (Figures 2, 7) but others have hypothesized a rapid lateral facies change from Rust through Dolgeville to Utica near Rathbun Brook. Moreover, this change may also have a structural origin (see Figure 10). Rust correlative strata may: 1) abruptly grade into Dolgeville facies near Rathbun Brook, 2) not be present between Rathbun Brook and Canajoharie, due to sediment bypass or erosion, or 3) correlate with some of the calcareous lower/middle Utica strata. The first alternative is appealing in that no major erosional cut-out is required. Furthermore, the Rust bioclastic sediments would have provided a source for the thick Dolgeville ribbon-limestones, and some Dolgeville-like calcilutites are present in the Rust. However, such a rapid facies jump does seem unlikely. If Rust-correlative strata are absent between Rathbun Brook and Canajoharie, the absence is most likely the result of sediment bypass following ramp over-steenening associated with the slump folding of uppermost Dolgeville beds; it seems highly unlikely that erosional removal of carbonate strata would consistently cease at the same stratigraphic level. The third hypothetical explanation of the "Rust problem" is shown and described in Figure 10.

To date, correlating some of the key event horizons in, and associated with, the Rust has not been entirely successful. Folding at the top of the Dolgeville (seen at Stops 5, 6, and 7) *might* represent the same slump event which resulted in either the upper or lower disturbed zone; however, even if the Dolgeville folds could be correlated with one of the disturbed zones in the Rust, we could not say with certainty whether or not Rust-correlative strata were present in the Dolgeville. The discontinuity at the top of the Russia in the Trenton Falls - Poland area *might* correlate with the Thruway discontinuity, but this in itself would not explain the disappearance of the Rust.

An appealing explanation of the "Rust problem" involves erosional bevelling of Rust and Steuben strata, as well as some of their downslope facies equivalents due to formation of a peripheral bulge. Such a mechanism is invoked by others (Jacobi, 1981; Lash 1988) to explain major localized truncation of sub-Trenton units in the Appalachian Basin. Such locallized upwarp could have removed a sufficient thickness of upper Trenton strata to leave platform facies on its west flank (west of the erosional cut-out) and Dolgeville-type ribbons or Canajoharie Shale on the east flank (east of the discontinuity closure) of the bulge.

Improved graptolite control both for Rust and the lower Utica could aid in understanding the "Rust problem." Furthermore, core drilling through the covered interval on Rathbun Brook should show if the Thruway discontinuity is present and, if so, whether or not any Rust is still present beneath it (see Figures 7, 10). Core drilling through the small Rust section immediately to the north of Rathbun Brook would give a thickness for the Dolgeville in this area, and it may show whether or not a discontinuity surface is present at the Dolgeville-Rust contact. Drill cores and geophysical logs could offer vital information for eliminating some of the possible fates of the Rust.

The Rust problem is but one area of Trenton stratigraphy that we are currently examining. Other key questions which we hope to help answer in the near future include:

> How far east does the Thruway discontinuity extend? Do Rust-, Steuben-, and Hillier-equivalent beds reemerge under it, and, if so, where and in what facies form? We are currently attempting these correlations and, as of this writing, has traced this contact eastward to Nowadaga and East Canada Creeks.

Can the Wolf Hollow Limestone- Dolgeville (mid-Denmark) erosional contact be traced eastward into the Canajoharie Shale of the St. Johnsville-Amsterdam region? Recent mapping suggests that the Wolf Hollow interval is expressed as a sequence of dilute calcisiltite bands above Canajoharie Falls on Canajoharie Creek, but the top erosional contact has not been observed east of the Little Falls - North Creek area.

Can the Dolgeville interval at Trenton Falls (High Falls tongue) be traced northwestward from that area towards Watertown? Again, this work is ongoing.

Can zonal equivalents of the carbonate succession of Rust, Steuben, Hillier, and Lindsay be confidently matched with appropriate Dolgeville, Canajoharie, and Utica beds to constrain the timing and placement of carbonate accumulation, submarine erosion, and black mud onlap in the Lake Ontario-Mohawk Valley region during the Taconic disturbance? Currently, graptolite control is better for the basinal shales than it is for the limestones. However, current work on graptolite zones (Mitchell *et al.* in prep) as well as refined fingerprinting of ash beds holds great promise of answers.

ACKNOWLEDGEMENTS

This research has been supported by grants from the Petroleum Research Fund, A.C.S. (Carl Brett), Northeastern Science Foundation (David Lehmann), and New York State Museum Honorarium (David Lehmann). Sharon Ingrams and Talia Sher (University of Rochester) aided in manuscript preparation. R. Titus, C. Mitchell, and C. Mehrtens provided encouragement. Thanks.

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ROAD LOG

(Also applies to field trip led by C. Mehrtens)

0	0	Thruway - Rt. 12 junction, Utica. Head north on Rt. 8/12.
5.8	5.8	Routes 8 and 12 split. Continue north on Rt. 12.
8.5	2.7	Cross Ninemile Creek. Continue north on Rt. 12.
13.5	5.0	Exit onto 365 North.
14.5	1.0	Turn right onto Church Street, heading into "downtown" Prospect.
15.4	0.9	Turn right onto State Street (which becomes Military Road).
15.7	0.3	Cross West Canada Creek.
15.8	0.1	Pull into parking on left. Walk down into gorge of West Canada Creek via path on the northeast side of the bridge.

Stop 1: West Canada Creek, Prospect: Slump folds and obrution deposits in the Rust Member, Denley Limestone.

A nearly continuous, although faulted exposure, of middle and upper Trenton strata crops out in the West Canada Creek gorge from the village of Trenton Falls to the Prospect Dam (approximately 3.5 km of exposure). At this stop, we will examine the upper 15 m of Rust bioclastic limestones and the lower 5 m of Steuben calcarenites. A number of interesting and enigmatic sedimentary and taphonomic features are present in this interval.

Note the prominent slump fold horizon (the "upper disturbed zone" of Prosser and Cummings, 1897) in the upper Rust, approximately 5 m above stream level along the east wall of the gorge. Disturbed strata occur as intermittent pods or lenses. Within lenses, strata are folded, brecciated, and--in some cases--over-thrusted. Lenses of disturbed strata are typically bounded by, and merge laterally into, shaley horizons, some of which contain slickensides. Some of these shales form rusty, micaceous recessions in the outcrop; they may be metabentonites. This interplay of ductile and brittle deformation suggests that strata were fairly well lithified when they were deformed (Mehrtens, 1984).

This disturbed zone is clearly continuous from Poland to Remsen, a distance of 12 km. Furthermore, "disturbed" strata occur at approximately the same stratigraphic position from Remsen to Lowville, 45 km to the northwest. This widespread occurrence of upper Rust disturbed strata, suggests that deformation reflects some sort of regional event--probably oversteepening of the Trenton ramp/platform.

The timing and nature of this oversteepening event is problematic. Unlike the slump folds in the uppermost Dolgeville (see Stop 6), these slump folds seem to indicate a down-to-the-east paleoslope (Cisne and Rabe, 1978). Although these slump features are clearly post-depositional, sediment was not completely lithified during deformation. Potentially, water released during compaction of clays accommodated the ductile behavior of the limestone beds; the disturbed zone incorporates a particularly shaley horizon of the Rust. Clearly, water content of the muds, amount of overburden, and angle of slope all played important roles in this slump deformation.

How might this slumping have been influenced by tectonic activity? Cisne *et al*. (1982) and Mehrtens (1988) have suggested that middle Trenton sediments were deposited in a developing foreland basin in which subsidence was accommodated by movement along the numerous normal faults which are present in the Mohawk Valley region. One such fault, which displaces Trenton strata 35 m, is located 150 m upstream of the Prospect Bridge. A narrow gully occupies the

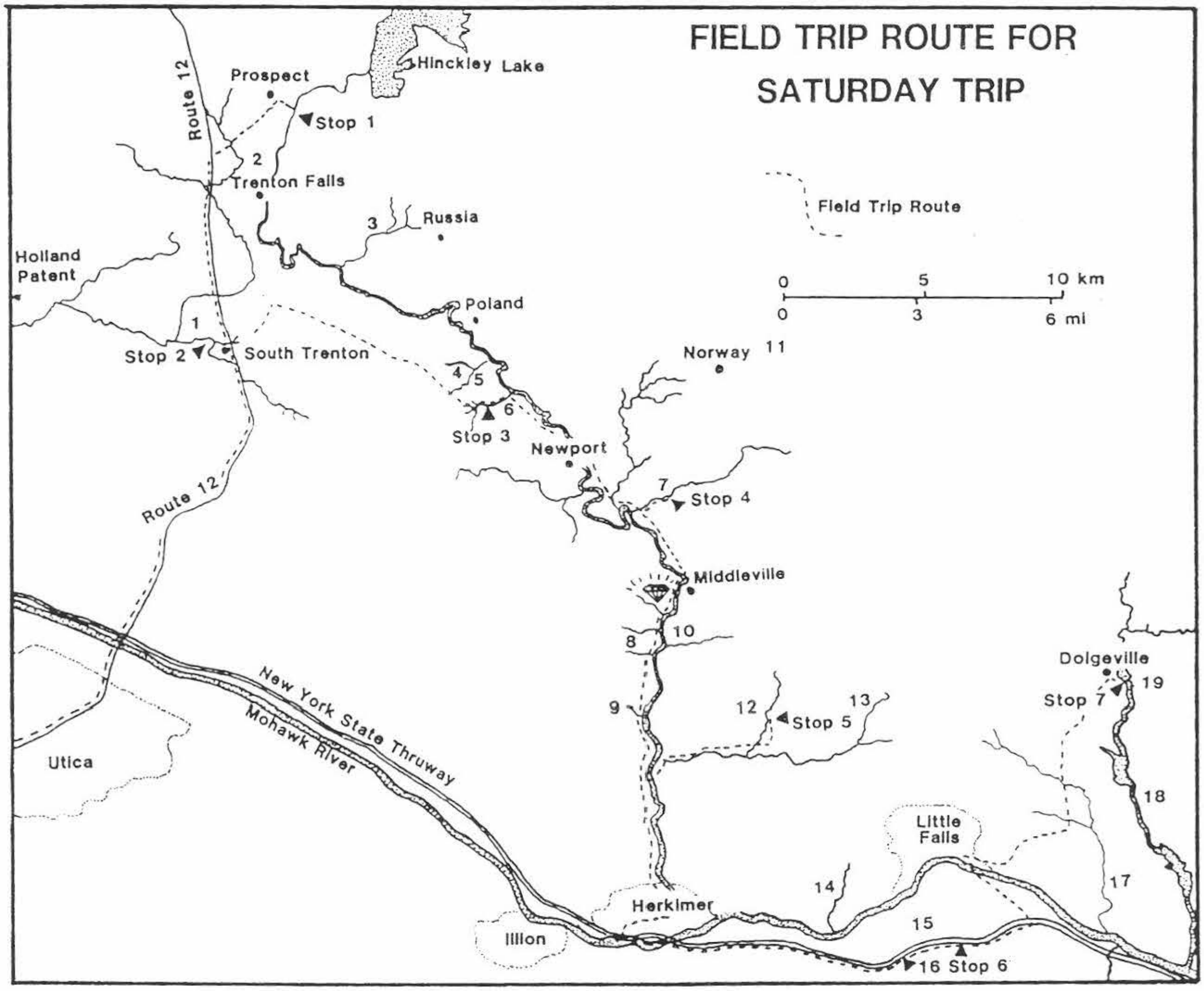

Figure 11. Study area showing field trip route. See road log for road and stop descriptions. Numbered localities include: 1, Ninemile Creek at South Trenton (Stop 2); 2, Trenton Falls Gorge (Stop 1); 3, Mill Creek at Gravesville; 4, creek south of Brayton Road; 5, creek parallel to Strumlock Road; 6, Rathbun Brook (Stop 3); 7, Wolf Hollow Creek (Stop 4); 8, creeks near the County Home; 9, creek northwest of Countryman; 10, Stony Creek; 11, creek northeast of Norway; 12, tributary of North Creek at Meyers Road (Stop 5); 13, tributary of North Creek south of Dillenbeck Corners; 14, creek at Gun Club Road; 15, Roadcut on Route 5s; 16, Roadcuts on New York State Thruway (Stop 6); 17, West Crum Creek; 18, East Canada Creek at Ingham Mills; 19, East Canada Creek below Dolgeville Dam (Stop 7).

surficial position of the fault. In the south wall of the gully, Steuben grainstones crop out. In the north wall of the gully, nodular limestones of the lower Rust are present. The northern exposure also contains abundant slickensides. The fault shows reverse polarity. However, drag folds associated with this fault indicate original normal, down to the east displacement; a net reversal of displacement apparently occurred at a later time (Bradley and Kidd, 1991). Tectonically-induced subsidence could have resulted in an oversteepened platform. Interestingly, the disturbed zone is cross-cut by a number of faults suggesting that either: 1) most motion along these faults occurred after slump folding, or 2) synchronous displacement along these faults--possibly related to thrust loading--resulted in widespread regional slumping.

Alternatively, Titus (1986) suggests that upper Denley (Rust) to Steuben deposition reflects upbuckling of the western Adirondack Arch region. In this scenario, oversteepening could accompany the basin modification. However, down-to-the-east slump features would suggest that the axis of uplift was to the west of Prospect.

Waltherian relationships suggest that the Rust strata at this outcrop represent deposition below normal wave base but possibly not below storm wave base. Lower Rust strata above High Falls at Trenton Falls contain wave ripple marks, while interference ripple marks and hummocky cross stratification occur in the Steuben grainstones at Cincinnati Creek (4 km to the northwest of here). Thus upper Rust strata are conformably bracketed by deposits representing storm-dominated shelf and wave base settings. Airy wave theory analysis of ripple marks of this stratigraphic interval at Booneville (25 km to the northwest) suggests sediment was deposited in water depths of between 1 and 27 m at that locality.

The occurrence of a diverse epibenthos and ichnofauna suggests well oxygenated seafloor and substrate conditions. The nodular bioclastic upper Rust Member contains abundant specimens of the brachiopods *Rafinesquina deltoidea*, and *Platystrophia sp*., the bryozoan *Prasopora orientalis,* and echinoderm plates. Nodular brachiopod- and trepostome-rich packestones and grainstones are typical Rust lithology. However, a 2 m interval of tabular bedded wackestones to fossiliferous lime mudstones are present below the disturbed zone. These limestones and interbedded shales--a tongue of deeper water carbonates, approaching Dolgeville facies--contain a distinctly different fossil fauna dominated by fenestrate and small ramose trepostome bryozoans, crinoids and the trilobites *Isotelus* and *Ceraurus*. These beds, probably correlative with the famed Rust Farm quarry beds of Walcott (1875a, b, c; 1876; 1881), contain abundant ceraurid trilobites at some localities. Brachiopods are much less common than in typical Rust strata. These tabular Rust strata include obrution deposits containing beautifully preserved specimens of *Glyptocrinus* and *Ectenocrinus*. These crinoids occur in both carbonate and siliciclastic mud tempestites/turbidites. Upright crinoid stems, up to 8 cm in length, attest to the severity of these sedimentation events.

15.8	0.0	Retrace route back to Rt. 12.
18.1	2.3	Head south on Rt. 12.
22.4	4.3	Take the Putnam Road/South Trenton exit.
22.6	0.2	Turn left at Putnam Road.
22.8	0.2	Turn left onto Old Rt. 12.
23.1	0.3	Turn right onto Church Street (at old white church).
23.4	0.3	Road forks. Stay to the left
23.5	0.1	Make a sharp turn to the right onto a private road (with multiple "No Trespassing" signs). Park. Follow path to below bridge. Walk upstream.

Stop 2: Ninemile Creek, South Trenton: Disconformity (regional composite unconformity) capping Steuben Limestone and flooring Utica Shale.

At this stop, we will examine a composite unconformity--the Thruway and Honey Hill discontinuities are superposed here--which separates Trenton carbonates from overlying Utica strata. This is the only outcrop between Frenchville (15 km to the northwest) and the Norway-Middleville area (15 km to the southeast) where the contact between Trenton carbonates and overlying siliciclastics (in this case, the upper portion of the Utica Shale) can actually be seen; near Holland Patent (2 km to the northwest), a mineralized Steuben top surface can be seen on this same creek, but the basal 0.5 m of Utica Shale is poorly exposed. Suprisinigly, although the Holland Patent stream section is commonly cited, the South Trenton section is not mentioned in most earlier reports--most notably Kay, 1953.

The South Trenton section results from a broad anticlinal fold which brings the Steuben Limestone up into view for a distance of 100 m along this stream; this fold is exposed 500 m downstream (west) of the private road bridge over the creek. Three beds (totaling 0.75 m) of Steuben are exposed at the breached fold axis, and over 10 m of Utica Shale are exposed in the stop vicinity.

Dark gray packstone facies of the Steuben is overlain by a 35 cm-thick interval of very dark gray bioturbated wackestone, rich in phosphatic nodules and small rod-like phosphatic bioclasts which forms the limestone bench above water on the creek bank. This is, in turn, overlain by a phosphate-cemented corrosion surface which is distinctly pitted and which contains infillings of Utica Shale and phosphatic debris in depressions between knobs and raised bosses. Locally, a 2 cm-thick micritic ledge is developed along the corrosion surface; it is distinctly perforated, roofing a horizontal crevice space which is filled with phosphatic lag debris. Note the complex pattern of burrow-infilling "let down" of lag debris into and under this uppermost Steuben carbonate bed. Resting on the mineralized surface is black laminated Utica Shale which yields the lower Upper Ordovician graptolite *Climacograptus pygmaeus*. At Stops 4, 5, and 6 (to the southeast), we will also view Utica Shale over the Thruway discontinuity, but in these latter sections, both the underlying carbonates and the overlying shales are distinctly older. At Frenchville (to the northwest), the Steuben carbonates are overlain by turbiditic siltstones and shales of the Frankfort Formation. In this respect, the South Trenton exposure offers a unique opportunity to see Utica Shale correlative with strata of the type section (Utica) in contact with underlying carbonates. Furthermore, these contact sections illustrate a pattern of northwestward overlap of black muds which continues from Canajoharie to near Toronto, Ontario: progressively younger siliciclastics disconformably overlie progressively younger carbonates (Lehmann and Brett, 1991a).

We interpret the contact displayed at this stop as a corrasional discontinuity produced by combined processes of corrosion and abrasion under conditions of sediment-starvation and dysoxic to anoxic bottom waters. Unconformities of this type typically underlie black shales in the Appalachian foreland basin; sediment-starvation, episodic impingement of deeper water current processes, carbonate undersaturation of bottom waters, and low pH conditions are all believed to contribute to the corrasion process (see Baird and Brett, 1990 for overview). Lag debris produced from the corrasion process consists of insoluble grains of pyrite, phosphate, and quartz. Middle and Upper Ordovician lags of this type are mainly composed of phosphatic nodules and bioclasts.

23.5	0	Retrace route back to Putnam Road.
24.4	0.9	Turn right onto Putnam Road. Continue on Putnam Road under Rt. 12 overpass.
25.9	1.5	Turn right onto road to North Gage.
27.5	1.6	Pass through intersection in North Gage (head straight).
28.7	1.2	Cross Rt. 8 (head straight).
29.9	1.2	Cross Herkimer County line.

30.6 0.7 Pass Strumlock Road. The furthest southeastern exposure of limestones of the Rust Member (Denley Limestone) occur along a small stream near this road.

31.5 0.9 Park at the old meat packing plant on the right.

Stop 3: Rathbun Brook: Key beds of the Medial Trenton succession.

We will examine bed succession from the top of the Poland Member, through a complete section of our informal Wolf Hollow division, to the lowest 3.5 meters of our Brayton Corners division. Above these strata is a 10 m-thick interval of concealed strata, which begins near the closed meat packing plant. This is one of the most critical and enigmatic sections with regard to the problems of southeastward disappearance of the upper Denley (Rust) strata, thickening of Dolgeville facies, and appearance of lower/middle Utica beds

Kay's (1953) Poland- Russia boundary on this creek is difficult to locate, as it is on several other creeks; he describes 33 m of "Denmark" at this section of which approximately half is Poland Member (which we will not examine) and half is Russia. We designate the base of Kay's (1953) Russia Member (and our Wolf Hollow division) as the base of a prominent 0.2 m-thick bioturbated packstone layer, herein designated the North Gage Road bed, which holds up a small waterfall on this creek and which will also be seen at Stop 4. This bed is approximately 12.5 m above the top of the Sugar River Limestone.

Above the North Gage Road bed is a 3.5 m interval of thin bedded, tabular to slightly nodular calcilutite beds which is, in turn, succeeded by a thick interval of nodular to massive shaly wackestone and lime mudstone which displays only weak bedding at many levels. This 15 m succession is rich in steinkerns of *Geisenoceras,* which are conspicuous on numerous bedding surfaces. Other common taxa include the trilobites *Flexicalymene* and *Isotelus*, the bryozoan *Prasopora*, and pelmatozoan debris. Two main zones of strongly nodular limestones in this interval are separated by more tabular-bedded calcilutites (Figures 2, 6); These intervals of nearly non-bedded Russia seem to correlate to two thinner massive intervals at City Brook (Stop 4) and correspond to times of sediment-starvation which produced a distinctive nodular, cephalopod-rich facies (cephalopodenkalk) that is thin (stratigraphically condensed) relative to correlative deposits.

Upstream from North Gage Road, nodular deposits give way to tabular bedded calcisiltites and calcarenites which display sharp bases, some with sole marks. This 3.5 m succession, best developed near the fork in the creek below the abandoned meat packing plant, is also characterized by nodular to sandy phosphorite concentrations both at its base and at higher levels. At its base is an irregular knobby contact overlain by a bioclastic debris layer yielding a profusion of nautiloid steinkerns as well as abundant phosphatic grains; this bed, in turn, is succeeded by a grainstone layer. This contact marks a submarine erosion surface that correlates to a corrasion surface at City Brook that Kay (1953) designated as the Poland-Russia boundary on that creek (Figure 3).

Herein, the strata between the top of the North Gage Road bed and the base of the 3.5 m-thick calcislitite/calcarenite interval are informally refered to as the Wolf Hollow division of the Russia Member (for the excellent exposure of strata which we will see at that creek), and the overlying 3.5 m interval is refered to as the Brayton Corners division. The Wolf Hollow beds are lithologically distinct and are roughly correlative with the lower 3/4 of Kay's (1953) Russia Member (Figures 2, 6, 7).

The phosphate-bearing, grainstone-rich interval (Brayton Corners division) capping the Wolf Hollow seen here is apparently the upslope facies equivalent of ribbon limestone deposits (upper Dolgeville tongue) which we will see at City Brook (Stop 4). The Brayton Corners division expands upslope to 6 or 7 m of tabular-bedded mixed grainstone-calcilutite facies at Trenton Falls and near Gravesville. In that region, the grainstones are overlain by a 3-6 m interval of thin calcilutites which we refer to as the High Falls Tongue of the Dolgeville Member and which consitute the uppermost Russia succession of Kay (1953). Thus, the Brayton Corners beds appear to be relatively condensed and grade to turbiditic ribbon carbonates (upper Dolgeville tongue) downslope and stratigraphically upwards (High Falls Tongue of the Dolgeville Member) (Figures 2, 6).

Above the critical 10 m covered interval is a 20 m-thick succession of lower/middle Utica black shale with widely spaced micritic beds yielding the graptolites *Dicranograptus nicholsonii* and *C. typicallis*. No Rust or Steuben lithology is observed. For a full discussion of the problematic southeastward disappearance of upper Trenton deposits as well as the southeastward appearance of thick Dolgeville and lower Utica divisions, the reader is referred to the text.

31.5	0	Make a right turn out of the meat packing plant onto North Gage Rd.
32.0	0.5	The City Brook bed caps the falls on the north side of the road. The type section for the Rathbun Member of the Sugar River Limestone is exposed in the face of the falls.
32.1	0.1	At the stop sign, turn right (south) onto Old State Road.
33.8	1.7	Turn left (east) at Bridge Road and cross West Canada Creek.
34.0	0.2	Turn right (south) on Rt. 28 in the town of Newport.
36.6	2.6	Turn left (east) on Castle Road.
36.8	0.3	Road forks; stay to the right. (The left fork leads to lower portion of the section--Little Falls Dolostone, Black River Group, and the lower Trenton Group--on Wolf Hollow Creek. However, the landowner for this stream section does not allow people on her property.)
37.0	0.2	Pull off to the right at the parking lot by the utility building, and walk up the left fork to the bridge over the creek.

Stop 4: Wolf Hollow Creek (City Brook), Old City: Condensation of Medial Trenton strata.

We will walk up the creek from the bridge below the Amish farm to examine, in ascending order: the uppermost Sugar River Limestone (Rathbun Member) and the Poland Member, the Wolf Hollow division, and the upper tongue of the Dolgeville Member of the Denley Limestone.

The lowest strata which we will examine are the massive calcarenites of the Rathbun Member of the Sugar River Limestone. Note the large domal bryozoans, *Prasopora*, in the lowest strata of the Rathbun Member. The Rathbun calcarenites are dramatically overlain by the "*Trocholites* Bed" of Kay (1937), which is the base of the Poland Member. The *Trocholites* Bed is herein designated the City Brook bed for the excellent bedding exposures in this interval above the road crossing on this creek. The nautiloid *Trocholites* is not sufficiently common for that name to have much meaning, and an alternative bed name is advanced here. This unit is widespread, marking the base of the Poland Member from the Lowville area (Chenowith, 1952) to the vicinity of Kast Bridge south of Middleville. At Wolf Hollow Creek this bed is 0.7 m-thick, and it is expressed as nodular, intensely bioturbated shaly wackestone-packstone facies which is extremely rich in orthoconic nautiloid and endocerid steinkerns as well as the trilobites *Flexicalymene* and *Isotelus* which are sometimes complete. A thin pelmatozoan packstone layer capping the condensed City Brook bed at this creek appears to be coextensive with a submarine discontinuity which oversteps the City Brook bed between the Kast Bridge area and Little Falls. This discontinuity may further correlate with a corrosion surface flooring the Canajoharie Shale at Canajoharie Creek and at a small stream southeast of Amsterdam.

Poland and Wolf Hollow deposits on this creek are represented by a stratigraphically condensed section which is much thinner than that developed at Rathbun Brook (Figures 2, 6). Again, at this section, the North Gage Road bed is identifiable as a 0.2 m compact, nodular limestone. This condensed interval is disconformably overlain by the sparsely fossiliferous ribbon carbonate facies of the Dolgeville Member of which only the lower 8 m are exposed at this section;

we will see the top of the Dolgeville at Stops 5 and 6. Careful mapping of middle Trenton beds shows that the Poland-through-Wolf Hollow interval of the Trenton Falls section thins negligibly from approximately 32 m in that area to 31 m at Rathbun Brook; however this interval thins dramatically from Rathbun Brook to City Brook, where only 15 m of section is present. Most of this southeastward thinning occurs in the upper (Wolf Hollow division) part of the section (Figures 2, 6); 15 m of nodular, cephalopod-rich wackestone at Rathbun Brook thin to 5.5 m at the present locality. Within the Wolf Hollow division, the two zones of nodular facies at Rathbun Creek thin respectively to form two massive nodular ledges which cap waterfalls on Wolf Hollow Creek and are separated by only 0.8 m of tabular calcilutites (Dolgeville facies). These meter thick benches, rich in nautiloid steinkerns and trilobite fragments, are an Ordovician expression of condensed cephalopodenkalk facies described from Devonian and younger systems (see Jenkyns, 1971; Tucker, 1973; Wendt and Aigner, 1985; Baird and Brett, 1986a, b).

Kay (1953) first reported the mineralized corrosion (discontinuity) surface which separates the upper nodular carbonate (Wolf Hollow division) from the overlying upper Dolgeville tongue. This conspicuous break, seen to best advantage upstream from our turnaround point, records a period of submarine erosion (and probably also carbonate dissolution) on the seafloor prior to deposition of Dolgeville sediment. The association of this break with condensed underlying facies is no surprise; extremely slow rates of sediment accumulation are often exceeded by rates of sediment removal by winnowing and dissolution. We saw this discontinuity in a more subtle development above the nodular interval at Rathbun Brook (Stop 3) where sediment-removal was apparently less severe. Still further west, at Trenton Falls, the top of the nodular interval is conformable with overlying beds. Preliminary study of sections between City Brook and Little Falls suggests that the condensed nodular beds are overstepped by this erosion surface towards the southeast such that some of the Wolf Hollow interval is missing in the latter area (Figures 2, 6, 7).

Kay (1953) correlated his Poland-Russia boundary at Rathbun Brook (near our North Gage Road bed position on that creek) with the corrosion surface level on City Brook. This resulted in the "abrupt facies change" of nodular Russia to Dolgeville between the two sections. Our recognition of the North Gage Road bed on Wolf Hollow Creek as well as successful matching of the nodular condensed limestone intervals (Figures 2, 6) provides a more palatable stratigraphic match between these sections, and it eliminates the disjunctive facies change. On Wolf Hollow Creek, the Poland interval, herein redefined, encompasses a 9.5 m-thick interval, dominated by tabular-to-wavy bedded micrite beds between the base of the City Brook bed and the base of the North Gage Road bed which caps a low waterfall on this creek. The Wolf Hollow division includes the remaining 5.5 m of Trenton section up to the discontinuity surface.

The overlying Dolgeville Member is composed of tabular bedded ribbon carbonate which alternates with very dark gray to essentially black shales. The Dolgeville strata yield a sparse fauna including graptolites and the small trilobites *Flexicalymene* and *Triarthrus*. We agree with other workers that this is a basin slope facies characterized by carbonate turbidites and an impoverished dysoxic fauna. Some beds show distinctive current markings (sole marks, ripple marks, and aligned fossils). Soft-sediment failure is represented by ball-and-pillow deformation at several levels.

The age of this Dolgeville sequence and its relationship to the Trenton Limestone is, in part, problematic (see text). The basal 1.5 m contains beds rich in nodular, sand-sized, phosphatic debris. This interval appears to be the downslope expression of the Brayton Corners division which we saw at Rathbun Brook and which also crops out at Trenton Falls (Figures 2, 6). However, higher, non-phosphatic Dolgeville deposits at Wolf Hollow Creek and near Middleville may correlate to the High Falls tongue of the Dolgeville at Trenton Falls and/or possibly to the Rust Member (see discussion in text).

37.0 0.0 Retrace route back to Route 28.

37.5 0.5 Turn left on Route 28.

38.0 0.5 Vuggy carbonates of the Little Falls Dolostone crop out on the left side of the road.

Many of these vugs contain doubly-terminated quartz crystal--the famed Herkimer Diamonds.

39.1 1.1 In the town of Middleville, turn right to continue following Rt. 28 south. Cross West Canada Creek.

39.3 0.2 Turn left to continue following Rt. 28 south.

39.6 0.3 Note the "Ace of Diamonds" Herkimer Diamond quarry and campground.We will pass another Herkimer Diamond quarry (Herkimer Development Corp.) over the next mile.

44.1 4.5 Turn left on West End Road.

44.3 0.2 Cross West Canada Creek on Kast Bridge.

44.8 0.5 Cross North Creek.

45.0 0.2 Turn right on North Creek Road.

45.3 0.3 Kay's Poland Mbr. of the Denley Limestone crops out along North Creek to the right.

47.5 2.2 Turn right (south) on Rt. 169.

47.7 0.2 Turn left onto Meyers Road.

48.2 0.5 Park by the farm on the left and walk down the hill to the stream.

Stop 5: Unnamed branch of North Creek, Eatonville: Upper Dolgeville - lower / middle Utica Shale contact

We will proceed to the falls above the lower end of the creek section and examine the top 4 to 5 m of the Dolgeville Member, the Dolgeville-lower/middle Utica contact (Thruway discontinuity) and metabentonite (volcanic ash) layers in the lower/middle Utica.

Turbiditic, ribbon limestones of the Dolgeville are visible at the base of the falls and in the downstream bank. In the downstream bank, these beds are of a nearly uniform thickness and spacing, which is typical for large portions of the upper Dolgeville. A low diversity, dysaerobic fauna of graptolites, orthoconic nautiloids, and the trilobite *Triarthrus* is characteristic of these strata, though diminutive *Onniella*, *Rafinesquina*, and *Flexicalymene* can be found in winnowed hash layers in some parts of the Dolgeville. Just below the discontinuity, and extending up to it, is a zone of folded and contorted Dolgeville beds. (Use caution as you ascend the falls; these beds can be slick!) This disturbed zone is seen to better advantage on the New York State Thruway and it is discussed under Stop 6.

The Dolgeville-Utica discontinuity is slightly undulatory here due to effects of differential compaction over beveled limestone beds, and/or, secondary diagenesis along the contact. A lag zone of phosphatic hash overlies this break with subsidiary phosphatic debris layers recurring through the next 10 m of section. Phosphatic debris includes orbiculoid fragments , phosphatic rod-like structures possibly belonging to *Sphenothallus*, diminutive *Onniella*, corroded pelmatozoan grains, and comminuted graptolites. On the Thruway and in sections near Middleville and Norway, a spectacular lag of imbricated,corroded ribbon limestone fragments is present on this contact. These clasts display pitted exteriors characteristic of dissolution; partly exhumed *in situ* limestone beds at the section northeast of Norway display interior crack systems which have been extended and widened by this same process, allowing lag debris to work its way into the cracks.

The lower/middle Utica Shale consists of very hard, laminated shale with thin ribbon limestones and metabentonite layers. Graptolites are common and are spectacularly current-aligned on several bedding planes. Two excellent metabentonites can be examined in the falls; these beds appear as pasty, pale gray layers. New "fingerprinting" techniques, which allow identification of specific ash beds (specific eruption events) as geochemically unique, are presently being applied to these layers in still another approach to working out Trenton-Utica stratigraphy.

The Utica Shale at this section is distinctly older than that observed at Stop 2; the upper beds of this lower/middle Utica strata yield the graptolites *Climacograptus typicalis* and *Dicranograptus nicholsonii* whereas the upper Utica Shale at South Trenton contains *C. pygmaeus*, a distinctly younger graptolite (Kay, 1953; Riva, 1969). This change reflects, in part, the northwestward overlap (termination) of progressively younger Utica beds against the discontinuity surface in that direction. It also reflects the additive merging of the Thruway discontinuity with the younger Honey Hill discontinuity due to erosional overstep (see text: Figures 2, 6, 7).

As indicated earlier, the age of the upper Dolgeville, seen at this section, and at Stops 4, 6, and 7, is problematical with respect to the Trenton Limestone. The lower part of Dolgeville (perhaps 10 to 13 m) and lower Canajoharie strata can be matched with the Poland Member using graptolites (Kay, 1953). However, the >22 m-thick upper Dolgeville tongue of the Norway - Kast Bridge area clearly overlies the top- Wolf Hollow corrosion surface, hence it cannot be linked with anything older than the Brayton Corners division at Trenton Falls. The great thickness of upper Dolgeville (and upper Canajoharie) strata relative to the upper Russia interval at Trenton Falls and at Gravesville opens the possibility that the upper Dolgeville may be time-equivalent with the Rust Member. This hypothesis remains to be tested.

48.2	0.0	Retrace route to Rt. 28.
52.3	4.1	Turn left (south) onto Rt. 28.
54.0	1.7	Enter the village of Herkimer.
55.2	1.2	Turn right to continue following Rt. 28 south.
55.7	0.5	McDonald's (a possible rest stop) is on the right.
55.9	0.2	Turn left to continue following Rt. 28 south.
56.1	0.2	Bear right on Rt. 28 towards the thruway entrance.
56.4	0.3	Turn left to the thruway entrance ramp. Bear right after going through the toll gates.
56.8	0.4	Merge with US Interstate 90 (east)
57.0	0.2	Cross the mighty Mohawk River.
60.0	3.0	Thick bedded Dolgeville carbonates crop out on the right (south) side of the highway.
61.4	1.4	Low outcrops showing Dolgeville and lower/middle Utica strata are on the right (south) side of the highway.
63.3	1.9	Pull over to the shoulder of the thruway.

Optional Stop 6A: New York Thruway: The Thruway discontinuity

This (along with optional Stops 6B and 6C) will be a relatively brief stop owing to hazardous traffic. We will stay in vehicles and observe outcrop features from van windows.

This outcrop exposes essentially the same section as at Stop 5. However, this roadcut is cleaner, more extensive, and much more dramatic. Only the upper succession of Dolgeville strata is visible at this locality. Note that the upper Dolgeville contains densely packed beds of calcilutite and calcisiltite. In this region, top Dolgeville strata are underlain by a middle, shale-rich Dolgeville interval which contains relatively few ribbon carbonate beds. This middle Dolgeville interval, in turn is underlain by a calcilutite-rich lower Dolgeville succession which correlates with the Poland Member and the lower part of the Wolf Hollow division; the base of the Dolgeville however, is quite shaley. This "double paired" pattern of shaley strata giving way to more calcareous strata continues east to the Canajoharie region. On Flat Creek for example, the uppermost Canajoharie Shale is actually a Dolgeville lithology and contains closely spaced ribbon carbonate beds.

Most notable in this roadcut are the numerous folds which are best developed in the top 1.5 m of the Dolgeville. This zone of deformed strata can now be traced for at least 20 km from a small creek northeast of Norway to Nowadaga Creek (north of Newville). Lower Dolgeville beds are also locally deformed, as can be seen on the Thruway approximately 2 km west of Stop 6, but the highest deformed zone is the most conspicuous and persistent.

The cause and timing of this folding is controversial, and it is central to hypotheses concerning the inferred sequence of events associated with the Taconic Orogeny. Fisher (1979) noted that axial planes of these slump folds typically dipped to the east, indicating a down-to-the-west paleoslope, and postulated that the paleoslope reversal was the result of compressional upbuckling of the Adirondack Arch. Work by Bradley and Kidd (1991) outlining the dynamics of this sort of "flexural extension" largely supports this interpretation. Clearly, a number of features indicate that this deformation is the result of post-depositional slumping. Firstly, although most deformation is ductile, fracturing and brecciation does occur in the deformed zone. Furthermore, even ductiley deformed strata contain wedge-shaped fractures suggesting that this sediment was partly lithified. Secondly, most folds incorporate numerous beds.

At West Crum Creek, 5 km to the east of Little Falls, slump features indicate a more complex scenario. At that locality, most slump features indicate a northeastern dipping paleoslope; however, some folds indicate slumping towards the southwest. Perhaps these slump features represent a number of episodes of fault block adjustment. Similarly, directions of fold axes tend to be polydirectional at Nowadaga Creek, suggesting the possibility of muliple deformation events.

Above the folded beds is a sharp horizontal contact marking the change to lower Utica deposits. As we saw at Stop 5 and will see at Stop 7, this is an erosional discontinuity marked by reworked phosphatic lag debris, disseminated pyroclastic material, and localized concentrations of variably corroded, reworked Dolgeville limestone intraclasts. Furthermore, the uppermost micritic ribbons of the Dolgeville are coated by a pyritic crust. Because the contact between the Dolgeville and the lower/middle Utica is so well exposed at this locality, we refer to this contact as the "Thruway discontinuity."

The Thruway discontinuity appears to breach many of the underlying folds such that many of the anticlines are "bald headed" structures. This indicates that post-Dolgeville, pre-Utica submarine erosion would have occurred after the folding event. Since we envision the episode of erosion as having taken place during a period of relative sealevel rise and consequent sediment-starvation to the region, it is interesting to speculate that the folding event may have been a seismic harbinger of tectonic deepening of the basin. Again, we must note that the relationship of the fold horizon to the upper Trenton Limestone of the Poland-Prospect region is problematic; these folds may link to one the deformed zones in the Rust Member or to Brayton Corners strata (see text, Stop 1).

64.1 0.8 Pull over to the shoulder of the thruway.

Optional Stop 6B: New York Thruway: Lower/middle Utica strata.

Along this portion of the roadcut, the lower/middle Utica Shale, which overlies the Dolgeville, is well exposed. As noted at Stop 6A, the contact between the Utica Shale and underlying Dolgeville strata is unconformable. Note, the metabentonite which forms a prominent rusty recession in the outcrop approximately 1.75 m above the base of the Utica. This particular metabentonite was seen at Stop 5 and will be seen at Stop 7. Other thin metabentonites can be seen in this roadcut and also in the "Black Canyon" (Stop 6C). Qualitatively, we have noticed that metabentonites typically are present in condensed intervals.

Also note that the lower/middle Utica strata in this outcrop contains thin, rusty (pyritic?), calcilutites. Locally, the lower/middle Utica succession between the Thruway discontinuity and the Honey Hill discontinuity is crudely divisible into three lithostratigraphic successions: 1) a lower calcilutite-rich interval, 2) a middle calcilutite-poor, black shale interval, and 3) an upper calcilutite-rich interval.

Finally, note that this outcrop contains numerous small faults. These faults which displace both Dolgeville and overlying Utica strata belong to the Little Falls fault zone. The major fault of this fault zone, the Little Falls fault (> 1 km to the southeast), has a vertical displacement of 155 m (Bradley and Kidd, 1991).

65.2 1.1 Take the Little Falls exit to leave the thruway.

65.4 0.2 Pull over to the shoulder of the exit lane.

Optional Stop 6C: Little Falls exit, New York Thruway: Utica strata of the "Black Canyon"

As we drive through the "Black Canyon," note the apparently monotonous nature of the Utica black shales. A number of thin metabentonites (which form rusty recessions in the outcrop) are present at this locality. The metabentonites along with graptolite zonation offer key--yet sometimes controversial--stratigraphic information for regional correlation of Utica strata (see Cisne and Rabe, 1978; Mitchell and Goldman, 1991).

65.9 0.5 Toll booth. Go through toll gate (after paying toll) and follow Rt. 169 north.

66.4 0.5 Junction of Rts. 5S and 169. Continue north on Rt. 169.

67.9 1.5 Take bridge across the Mohawk River.

68.3 0.4 Turn right on Rt. 5W. Drive through the graffiti-enhanced canyon of Precambrain Grenville basement.

68.6 0.3 Turn left (north) on Rt. 167.

71.0 2.4 Turn left to continue following Rt. 167.

74.3 3.3 Enter the village of Dolgeville.

74.9 0.6 Turn right on Faville Avenue. Follow Faville Avenue to Niagara-Mohawk power station.

75.6 0.6 Park outside of the power station gates.

Stop 7: East Canada Creek, Dolgeville: Drag-folded Dolgeville and Utica strata.

Along East Canada Creek, below the Dolgeville Dam, the upper 15 m of Dolgeville strata and a large thickness of the overlying shales of the lower/middle Utica are spectacularly exposed in a drag fold along the east side of the creek. This drag fold is associated with the Dolgeville Fault, another of the syntectonic normal faults of the Mohawk Valley region.

Many of the features discussed and seen at Stops 5 and 6 are also readily examined at this outcrop:

1) The upper Dolgeville strata are slump-folded, but, due to the drag folding process, slump folding is not as evident as at Stop 6.

2) The Dolgeville strata are overlain by a 5 cm-thick phosphatic lag which we will sample if the stream level is low.

3) A metabentonite forms a prominent recession in the outcrop approximately 1.5 m above the Dolgeville-Utica contact.

4) The lower/middle Utica contains rusty-weathering calcilutites. However, calcilutite beds are less numerous than in lowermost Utica strata at Stop 5. This may be due to distance from a carbonate source, proximity to a siliciclastic source, and accommodation space (subsidence). Alternatively, we may be seeing a slightly lower stratigraphic interval at Dolgeville which is not represented at Stop 5 or is represented by condensed facies in that area.

Well developed fine-grained turbidites (T_{bcd}) are present in the Utica. Winnowed shell lags containing trilobite hash, small brachiopods, and current-aligned graptolites within black shales, however, indicate that not all downslope-directed currents resulted in deposition.

75.6 0.0 Retrace route to the thruway (US Interstate 90).

85.3 9.7 Take the thruway west.

Non-Fossiliferous and Fossil Rich Beds in the Hamilton Group; Indicators of Sedimentation Rates

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The Hamilton Group siliciclastic rocks are part of the progradational sequence formed at the onset of the Acadian Orogeny. Mountain building occurred to the east of New York state on a NNE/SSW trend. Paralleling the mountains to the west was the axis of the depositional basin which ran approximately through the Finger Lakes region. This is reflected by the thickening trend of Hamilton Group rocks from west (100 m) to east (1000 m). Throughout most of the Givetian Stage, eastern New York was close to the depositional source. This can be seen in the Albany area and immediately to the south where Hamilton Group equivalent rocks such as the Plattekill and Monorkill Formations preserve evidence of fluvial depositional environments (Willis 1990). To the west of the fluvial facies is the Panther Mountain Formation which contains nearshore marine sandstones which extend to the Unadilla valley area. The Panther Mountain Formation is the eastern equivalent of the Skaneateles and Ludlowville Formations of Central New York. In western New York the Hamilton group is predominantly dark shales and siltstones with rare carbonate units. The dark shales represent environments distant from the depositional source and in deeper water conditions than the sandstones of the Panther Mountain Formation. The rocks of central New York are a mixing and interfingering of the basinal and nearshore marine environments (Fig 1).

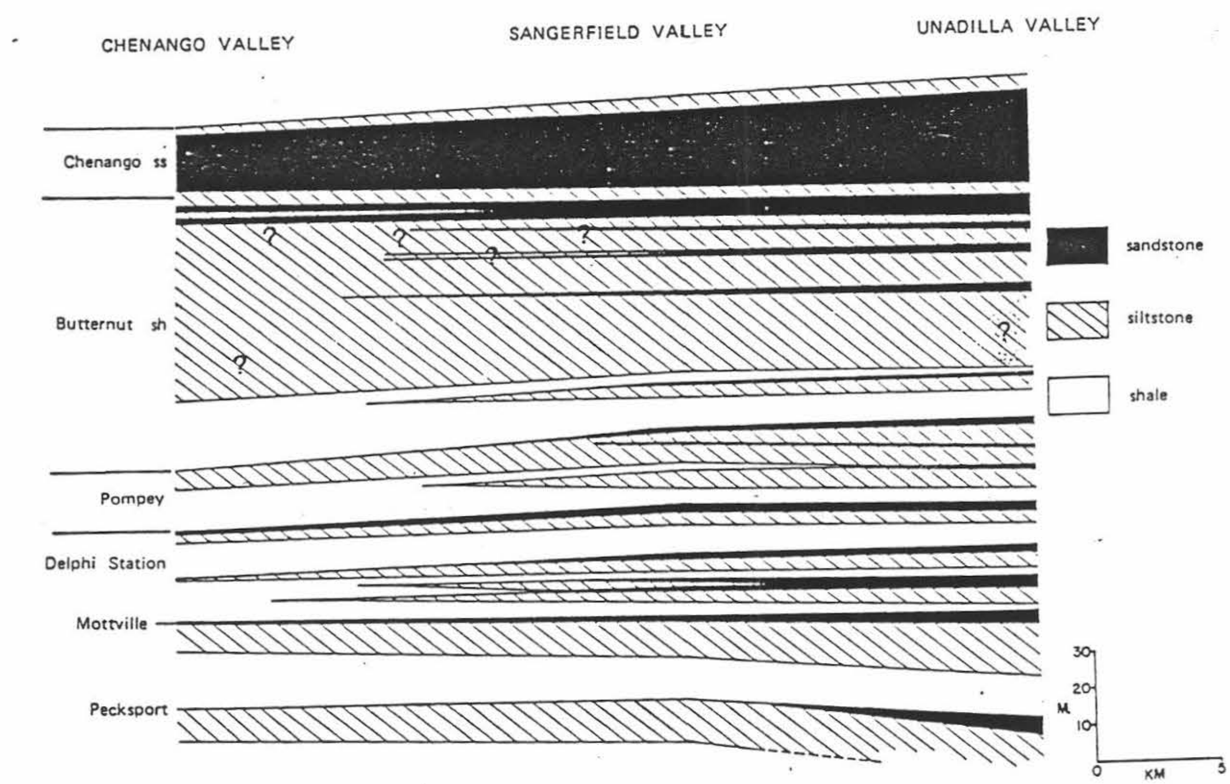


Figure 1. Interfingering of Panther Mountain Sandstone of Eastern New York with basinal shales of Western New York from the Chenango Valley to the Unadilla Valley.

The formations of the Hamilton Group were defined in part by Lardner Vanuxem and James Hall during the geologic survey of the state begun in 1836. The members the Hamilton Group were defined by Cooper (1929, 1930, 1934). In central New York Cooper defined members on the basis of rock type (Chenango sandstone, Butternut shale, Sollville sandstone) and on a recognizable coarsening interval such as the Delphi Station and Pompey members (Fig 2).

The stratigraphy of the Hamilton Group is comprised of a series of coarsening upward cycles. In central New York the base of the Marcellus through the top of the Skaneatelles Formations could be considered a single large-scale coarsening cycle. Within this large coarsening sequence smaller coarsening sequences can be observed; examples of these are the Chittenango Shale through Solsville Sandstone, or the Butternut Shale through Chenango Sandstone. Within each of these intermediate coarsening sequences are the coarsening sequences visible at the outcrop scale such as the Pecksport, Delphi Station and Pompey Members. Finally, meter- scale coarsening sequences on a scale equivalent to the punctuated aggradational cycles of Goodwin and Anderson (1986) can be seen within a single outcrop. These depositional cycles are heirarchical (Busch and Rollins 1984, Busch and West 1987). Johnson and others (1985) consider the Skaneateles to the top of the Moscow to be a third order transgressive-regressive sequence within the Vail (1977) system. The Marcellus can be divided into two third order cycles, one from the base of the Unions Springs to the top of the Cherry Valley and the second one from the Chittenango to the top of the Pecksport, cycles Id and Ie of Johnson and others (1985). Within the third order cycles are the fourth order cycles of Bush and Rollins (1984). These would correspond to a coarsening sequence from the Butternut through the Chenango Sandstone or the entire Ludlowville Formation. Fifth order cycles are what can be seen as coarsening cycles on the outcrop scale and in central New York have thicknesses of 10-20 m. Sixth order cycles are meter scale coarsening sequences which can be seen within one fifth order or outcrop scale cycle. Each scale of cyclicity has an approximate duration. Sixth order cycles are thought to represent tens of thousands of years, fifth order cycles a few hundred thousand years and so on. The coarsening sequences observed on this trip will be fifth and sixth order cycles.

The coarsening upward cycles are interpreted to represent shallowing upward depositional environments. The top of each coarsening sequence is followed by an abrupt return to deeper water conditions or by an effect equivalent to a rapid distancing from the sedimentary source. Many of the larger scale cycles are traceable over hundreds of kilometers indicating that the cause of cyclicity is operating over a very large area of the basin. Some workers have attempted to tie the larger-scale cycles to global sea level curves (House 1983, Johnson and others 1985). If the cycles are allocyclic and are driven by climate change or tectonism, they could be used to correlate with a much finer resolution of time than is currently available with biostratigraphy.

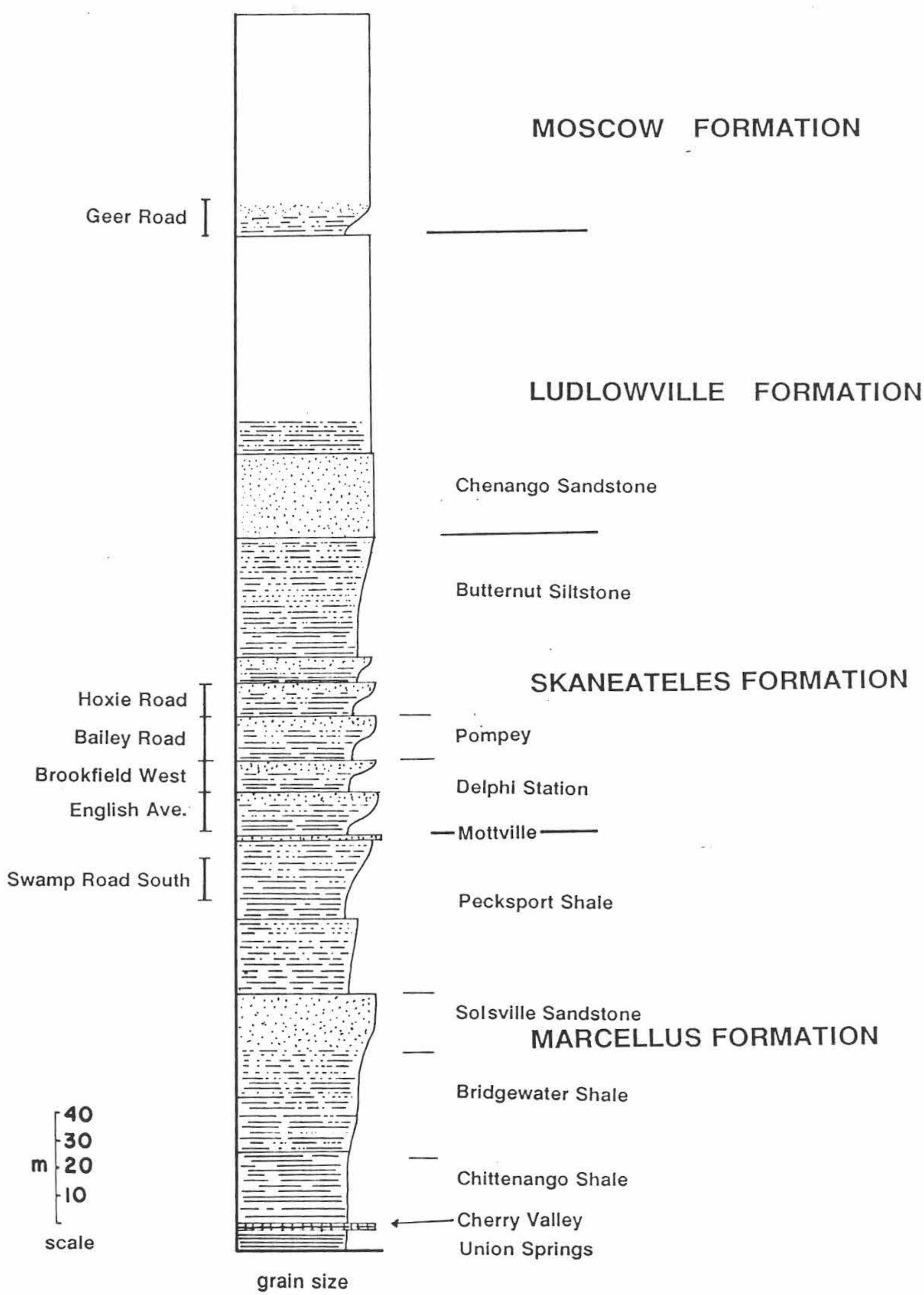
The Hamilton Group is famous for its well preserved and abundant fossils. Fossils occur in rock types from shale through sandstone and in rare beds of bioclastic packstone. In central New York it is unusual to find stratigraphic intervals where fossils are absent. Since non-fossiliferous units are the exception rather than the rule, by examining vertical and lateral changes from fossiliferous to non-fossiliferous zones it may be possible to determine what factors control the diversity and abundance of benthic faunal assemblages in the Middle Devonian of central New York.

The coarsening sequences are interpreted to be shallowing sequences. The basal shales contain a faunal assemblage different from the sandstones at the top of the cycles, and the mudstones in between the shales and sandstones have their own characteristic faunal assemblages. The faunal composition of these assemblages is presumably controlled by different water depths, turbidity conditions, and substrates.

Non-fossiliferous zones can be observed in both the basal shales and the mid-cycle mudstones. There are several possible explanations for the presence of non-fossiliferous zones within otherwise fossiliferous rocks. Exclusion of benthic organisms may result from anoxic or dysoxic conditions, from rapid sedimentation rates which produce water saturated and highly unstable substrate, or from hypo- or hypersaline conditions.

Figure 2

HAMILTON GROUP STRATIGRAPHY IN THE SANGERFIELD VALLEY AREA



The non-fossiliferous, organic-rich, black shale of the Oatka Creek and Levanna Members of western New York are stratigraphically equivalent to the highly fossiliferous mudstone and sandstone members of the Marcellus and Skaneateles formations of central New York. The model of a density stratified basin (Woodrow and Isley 1983, Ettonsohn and Elam 1985) has been used to explain the absence of benthic fauna and presence of abundant carbon in the Marcellus and Skaneateles of western New York. This model postulates that the Appalachian Basin may have been density stratified due to differences in temperature and salinity. The lower colder bottom waters were anoxic and higher in the water column were followed by dysaerobic zone called the pycnocline. The upper part of the water column maintained fully oxygenated normal marine conditions. In the Upper Devonian of Kentucky in the central Appalachin Basin the following sequence is interpreted for different water depths (Ettonsohn and Elam 1985): black non-fossiliferous shales were deposited below the pycnocline in the deepest part of the basin, gray bioturbated shales with no hard shelled fossils in the dysaerobic zone of the pycnocline and abundant benthic fossils above the pycnocline. In western New York the black non-fossiliferous shales of the Oatka Creek and Levanna members grade eastward into the gray mudstones and fossil rich siltstones of the Marcellus and Skaneateles Formations. Western New York can be interpreted as the deeper water dysaerobic zone while central New York could be the transition from the dysaerobic zone into the oxygenated zone.

If this model is applied to the shallowing up cycles of central New York then the deposition of black shales must have occurred below the pycnocline where anoxic conditions made habitation by benthic organisms impossible. Progradation of the shoreline and an infilling of the basin to a depth above the pycnocline then allowed benthic organisms to inhabit the substrate. From here on up in the shallowing cycle benthic fauna are present unless some other variable prevented them from colonizing the substrate. The lower Marcellus of central New York is probably an excellent example of basinal infilling producing a shallowing upward sequence. The anaerobic, dysaerobic and aerobic zones are clearly observed in the shallowing trend formed by the Chittenango, Bridgewater and Sollsville members and again in the Pecksport through Mottville sequence. It is not so clear whether this model is applicable to the Skaneateles cycles because water depths as interpreted by the thickness of the sedimentary cycles are too shallow. In the coarsening cycles of the Skaneateles Formation the basal shales may be highly fossiliferous and yet fossils may be absent from the mudstones in the middle of the cycle. Fossils are found both above and below the non-fossiliferous zones and often in rare beds within them. Deposition rates which create unstable substrates may be a better explanation for the absence of benthic organisms but do not provide a complete solution.

Hoxie Road Quarry

This coarsening cycle is interpreted to be at the base of the Butternut Member in the Skaneateles Formation. The quarry exhibits the repeated coarsening-upward trend observed at most localities in the Hamilton Group. The rock type at the base of the quarry is shale interlaminated or thinly interbedded with fine siltstone. Close inspection reveals repeated 1-2 cm thick couplets of fine siltstone and shale. The siltstone contains low angle cross laminae and planar laminae. The base of the siltstone units is abrupt and the contact with the shale above is gradational. In the middle part of the exposure the silt beds become thicker (5 cm) and preserve rare articulated crinoid columnals typically on the upper surface of the siltstone beds. No fossil lag deposits are observed at the base of the siltstone beds and fossils are otherwise rare in this part of the outcrop. The upward increase in thickness of the siltstone beds indicates a closer proximity to the source of sediment and an increase in energy conditions on the substrate, both of which reflect the shallowing upward nature of the depositional cycles. The upper part of the outcrop is bioturbated very fine sandstone and contains a normal Hamilton assemblage of bivalves and brachiopods.

Fossils in the lower part of the exposure are rare and when observed are concentrated on single bedding planes. At vertical intervals of 1-2 m discontinuous lenses, meters in width and 1-2 cm in

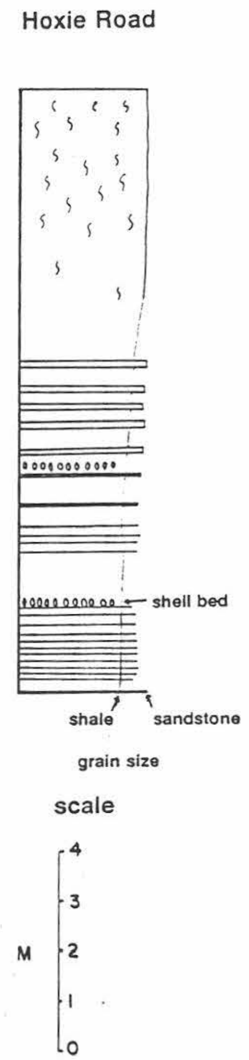


Figure 3

height, contain articulated bivalves, brachiopods, and the crinoids Gilbertsocrinus and Acanthocrinus. The bivalves Leiopteria conradi, Actinodesma erectum and Actinopteria boydi are numerically the most abundant taxa and are all epifaunal, possibly byssally attached forms. The brachiopods Tropidoleptus carinatus and Camarotoechia congregata are associated with the bivalves but are less abundant. The crinoids are articulated and appear to be using the bivalve shells as a substrate. The unusual aspect of these shell beds is that they are not storm accumulations and the fossils are articulated and in life position. Rocks above and below the fossil beds are nearly barren and preserve only rare brachiopods and bivalves.

The presence of small-scale sedimentary structures in fine grained sediments is rare in Hamilton Group rocks in central New York. Infaunal and semi-infaunal organisms disturb the substrate and destroy sedimentary structures. If sedimentation rates are low enough for the organisms to continually rework the sediment, no sedimentary structures will be preserved and the mud and silt layers are homogenized into the bioturbated mudstone commonly observed in Hamilton Group rocks. At this and other localities there is a conspicuous absence of fossils associated with the stratigraphic intervals where sedimentary structures are preserved. An absence of fossils may be explained by rapid deposition rates which create an unstable substrate, by a lack of oxygen, or by non-normal marine salinities. The pycnocline model is applicable in relatively deep water settings; however central New York is closer to the depositional source and in presumably shallow water depths.

An alternate explanation for non-fossiliferous units could be viewed as follows: assuming that the depositional cycles are shallowing up sequences one method of deposition may have been basinal infilling. If sea level is stable and sediment is transported to the basin, progradation of the shoreline will occur. The part of the basin proximal to the shoreline will become infilled with sediment and a shallowing upward depositional sequence will be produced. By this model the muddy siltstones would be less than ten meters below the sandstones. The water depth of the sandstones is equivocal but they are definitely within storm wave base, and given the proximity to the shoreline, probably in less than ten meters water depth.

Brett (1985) has postulated that periods of maximum deposition occur in the middle of the depositional cycles in the siltstone/mudstones and that while the sandstones are deposited in the shallowest water, they are also periods of low deposition rates. The sandstones are possibly reworked or winnowed over long periods of time. If the siltstones and shales were deposited rapidly this could have created unstable water saturated sediment which would have kept epifaunal or semi-infaunal filter feeders from colonizing the substrate, thus creating non-fossiliferous intervals. However evidence from this locality suggests otherwise due to the presence of rare shell beds with epifaunal bivalves. Furthermore, the centimeter-scale siltstone beds, which should have been a relatively stable substrate, have no fossils associated with the upper surface. This suggests that a factor other than anoxia or substrate instability was preventing the colonization of the substrate by epifaunal benthic organisms.

The proximity of central New York to the fluvial environments in the east suggests that hyposaline conditions could be a factor in controlling faunal assemblages on the nearshore shelf. The northern end of the Appalachian Basin during the Devonian was a relatively enclosed basin. The uplands created to the east by the Acadian Orogeny were the major source of siliciclastic sediment.

Brookfield West

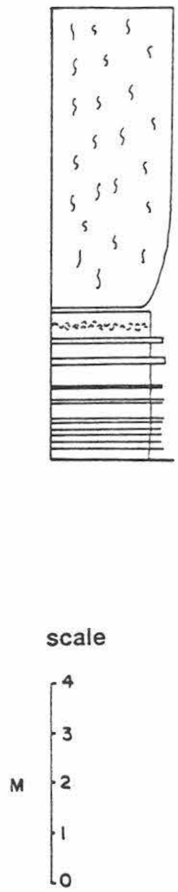


Figure 4

Fluvial depositional systems have been identified in Hamilton Group equivalent rocks in eastern New York. Rivers feeding the sediment to the basin may have had the effect of reducing salinity in the shallow, nearshore regions of the basin. If the siltstone/mudstone intervals were deposited during periods of maximum sedimentary influx, and the position of the ancient shoreline was less than 100 km to the east, it is possible that the nearshore or shallow part of the basin experienced periods of hyposalinity. During periods of low rainfall normal salinities would return to the nearshore part of the shelf and a normal marine fauna would colonize the substrate. If the sandstones were deposited or reworked during periods of low sedimentary influx this may also have been a time of relatively low rainfall and consequently normal marine salinities.

Fossil evidence in the form of thin lenticular shell beds containing in place articulated bivalves and crinoids suggests episodic colonization of the substrate followed by an abrupt event which preserved the fossils without disruption. The rare fossils, typically Camarotoechia congregata, Tropidoleptus carinatus, Grammysia bisulcata and Pterochaenia fragilis in the intervening layers between shell beds are found on single bedding planes with only Grammysia appearing to be in life position. In the middle part of the outcrop siltstone beds preserve articulated crinoids on the upper surface of the beds. The fossils between the shell beds in the interbedded siltstones and shales are the best indicators of the typical paleoenvironment. Fossils such as Camarotoechia, Tropidoleptus and Grammysia could have been salinity tolerant. Only Grammysia and Pterochaenia are found articulated and in place. The brachiopods could have been transported although they are not found associated with sedimentary lag deposits. The crinoids associated with siltstone layers could have been transported, although not over long distances since they are still articulated. The shell beds contain dominantly bivalves which may have been salinity tolerant (Brower and Osborne 1991) but also contain articulated crinoids which are much less likely to have been salinity tolerant.

The rarity of both fossils and bioturbation in the thinly interbedded siltstones and shales suggests that some factor was preventing colonization of the substrate. Dysoxia has been suggested as a possible factor but the fossils that are present tend to be salinity tolerant species rather than species typical of the deeper water dark shales associated with dysaerobic conditions. When periods of normal marine conditions occur the substrate is apparently rapidly colonized by bivalves, brachiopods and crinoids. These brief periods of colonization are equally rapidly ended by an event which preserves the fauna intact and articulated. The proximity of central New York to the ancient shoreline and accompanying river systems suggests that fresh water flooding of the nearshore shelf may account for some non-fossiliferous intervals in the interbedded silts and shales.

Brookfield West Quarry

This section is tentatively placed as the fourth or uppermost coarsening cycle of the Delphi Station member of the Skaneateles Formation. The lower two meters of section contain 5-20 cm thick interbeds of shale and siltstone. Siltstone beds have an abrupt base and grade upward from planar laminae into small scale, high angle cross laminae. The upper surfaces of some siltstone beds preserve symmetrical ripples. The siltstone units are interpreted to be event beds, probably storm deposits. The absence of infaunal and epifaunal organisms with the related bioturbation is discussed in the section on the Hoxie Road quarry. The shale separating the siltstones is fissile, lacks lamination and contains no fossils with the exception of rare cephalopods.

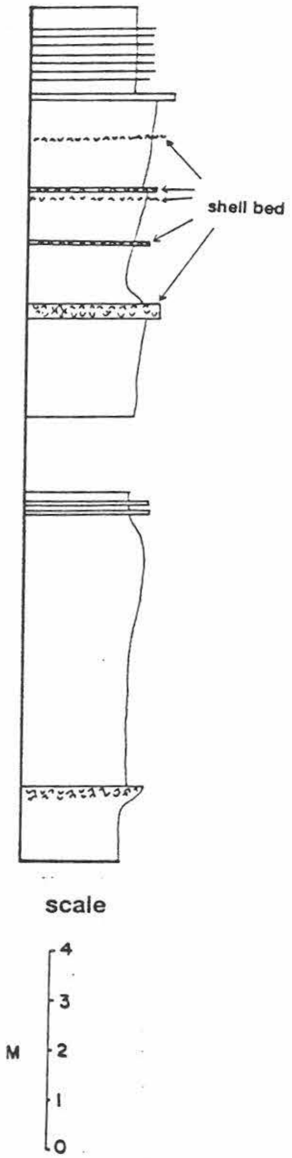


Figure 5

In the middle of the exposure is a noticable notch in the quarry wall. Above this the siltstone/shale interbeds disappear and bioturbated sandstone, with a normal Hamilton faunal assemblage of epifaunal bivalves and brachiopods, is observed. The notch in the quarry wall is a light gray clay, varies in thickness from 0.5-3.0 cm, and is tracable the length of the outcrop. X-ray analysis reveals the presence of illite and chlorite but no expandable clay minerals. The coarse fraction of the clay does contain some euhedral crystals. This bed is tentatively identified as a meta-bentonite.

Ten centimeters below the clay bed is a densly packed fossil-rich zone 1-4 cm thick. This is the first appearence of fossils in the lower part of the quarry. The assemblage is unusual in that it contains the monoplacophoran Cyrtonella and abundant infaunal bivalves, Nuculoidea, Nuculites and Modiella. Infaunal bivalves would not normally be found in a packed shell bed like this since when they die they are already buried. The presence of these shells together indicates that the sediment has been reworked and that this bed may be a sediment-starved omission surface. Phosphate pebbles are found with the shells in this bed which also suggests that the accumulation of shells took place during a depositional hiatus.

This section is somewhat unusual in that it does not follow the typical pattern of a gradual coarsening upward trend. The fossil rich sandstones at the top of the locality appear relatively abruptly following the appearance of the clay bed. Whether there is a tectonic connection between the two events is highly ambiguous.

Bailey Road Quarry

The seventeen meters of section at this locality represent the Pompey Member of the Skaneateles Formation. From the base of the section to the top there is an overall coarsening trend from shale through fine sandstone. Within this coarsening trend there are four, meter-scale coarsening upward units. In the terminology of Bush and Rollins (1984) the entire outcrop would be a fifth order cycle and the smaller coarsening sequences would be sixth order cycles. The base of the smaller cycles is indicated by topographically level areas within the quarry.

The lowest small-scale coarsening cycle is bioturbated shale with abundant, well preserved, articulated bivalves and brachiopods. The bivalve assemblage is dominated by Nuculites oblongata, N. triquiter, Modiella pygmaeaand Paracyclas lirata. The top of the first cycle grades upward into a packed silty shell bed containing Longispina, Mucrospirifer, and rare septate rugose corals. The second depositional cycle grades upward from a siltstone through a coarse siltstone. The finer grained interval is bioturbated but contains no fossils with the exception of rare cephalopods. The coarser part of the cycle contains a chonetid/Mucrospirifer assemblage. The third cycle is progressively coarser than the first two, grading from mudstone through very fine sandstone. The base is again bioturbated but relatively non-fossiliferous. The top of the cycle is capped by a packed shell bed containing the brachiopod genera Rhipidomella, Pseudoatrypa and Athyris. These brachiopods, particularly Pseodoatrypa, are relatively rare in the rocks of central New York and have been interpreted to represent clear non-turbid water conditions. The fourth cycle is again coarser than the previous three, grading from siltstone to fine sandstone. Brachiopods dominate the fauna though bivalves, particularly Cypricardinea indenta, Paracyclas lirata and Nyassa arguta, are quite abundant. Bryozoa and zoophycus are common in the upper part of the cycle but appear only occasionally below this.

Several packed shell beds are located within this cycle. Two are similar to the one capping the third cycle (i.e. dominated by Rhipidomella and Athyris) while several others are dominated by Spirifers and Ambocoelia, with lesser amounts of Chonetes and other organisms.

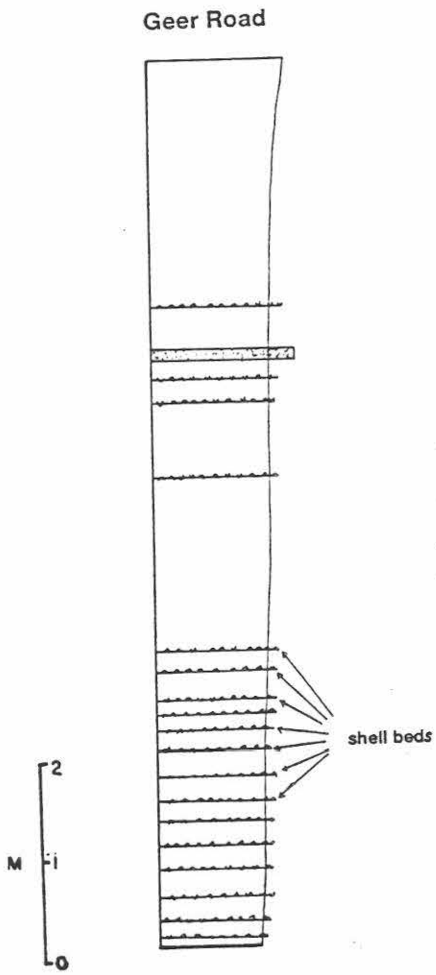


Figure 6

The depositional sequences at this locality provide some clue to sedimentation rates if the assumption is made that fossil preservation potential and biogenic productivity, in terms of shell production, is constant through time. The fossiliferous shaley interval of the lowest cycle can be interpreted to have relatively low sedimentation rates due to the presence of numerous infaunal and epifaunal organisms. The shell bed which caps this interval is a period of low siliciclastic deposition where fossils accumulated gradually through time. Evidence in the form of disarticulated shells suggests that the shell bed was affected by storm events; however, no sedimentary structures are visible to support wholesale transport or reworking of the shells as a lag deposit. The base of the shell bed is gradational with the underlying shale, further supporting an in place formation of the bed. The base of the second cycle is nonfossiliferous yet it is bioturbated. This interval probably represents a period of rapid deposition where epifaunal filter feeders would have had difficulty with an unstable and rapidly aggrading substrate. The absense of infaunal deposit feeding bivalves cannot be easily explained since the presence of bioturbation and rare cephalopods suggests that organisms were present and that fossilization was possible. The upper part of the cycle is coarser grained and preserves abundant brachiopods. This indicates a decrease in sedimentation rates. The third cycle is again non-fossiliferous at the base and is capped by a shell bed containing an unusual assemblage of brachiopods. The Athyris/Pseudoatrypa/Rhipidomella assemblage has been interpreted to represent a shallow, clear water depositional environment. This shell bed represents an absence of siliciclastic deposition. The abundant brachiopods of the fourth cycle again suggest fairly low sedimentation rates, while the shell beds, similar to those of the third cycle, most likely represent periods of non-deposition.

Geer Road Quarry

The nine meters of section at this locality represent the lowermost of the four coarsening upward cycles in the Moscow Formation (Selleck, personal comm.). Grain size ranges from very fine siltstone at the base of the quarry through very fine sand at the top, with most of the coarsening occurring in the upper three meters. The faunal assemblage is dominated by brachiopods (80%) and bivalves (10%) with gastropods, cephalopods, trilobites, bryozoans and crinoids also present.

Fossils are distributed abundantly and fairly uniformly throughout the section; however, flat bedding planes of densely packed shells, predominantly of the brachiopod Chonetes, also occur. In the lower three meters of the section these Chonetes shell beds occur at intervals of 20-25 cm. Above the first three meters Chonetes beds become rare. Orientation of the shells varies between the upper and lower beds. In the lower shell beds Chonetes are found 50% in life position and 50% overturned while in the upper beds 90% of the Chonetes are overturned. In the siltstones between the shell beds different taxa are present in approximately the same ratio as seen on the shell beds, but the density of fossils is much less. No sedimentary structures are visible.

The Chonetes beds may represent storm events during which transported siliciclastic sediment smothered the brachiopods. However, the sedimentary characteristics normally associated with tempestites, such as fining upward sequences or planar or cross laminae, are not visible. The presence of fossils in the sediment between the shell beds suggests that they are not winnowed lag deposits.

A more probable theory is that the shell beds represent periods of non-deposition allowing for the build-up of many shells over time. If the coarsening upward sequence reflects shallowing, the lower shell beds would have been in deeper water than the upper beds. The differing orientations of the Chonetes may then be explained; the lower, deep-water shell beds would have remained relatively unaffected by wave action, allowing half of the brachiopod shells to remain, even after death, in the less stable life position. The Chonetes of the upper shell beds, being in shallower water, would have been 90% overturned into a hydrodynamically stable position.

The most definitive evidence for storm deposited sediments is a 5-8 cm fine to medium grained sand bed located six meters above the base of the quarry. The sand bed has an abrupt base and contains small scale cross and planar laminations. Both shelled fossils and trace fossils are absent from this bed, appearing neither as storm lag deposits nor as colonizers of the post-storm surface. The position of this bed is near the top of the section, in the coarser part of the cycle. Assuming that the cycles represent shallowing upward, this part of the quarry may have been within the influence of storm wave base. The lower, finer grained part of the cycle was likely deposited in deeper water below storm wave base.

English Avenue Quarry

This quarry is situated in the Delphi Station member of the Skaneateles Formation. The gastropod species Bembexia sulcomarginata is abundantly represented at this locality and all specimens are either of the ornamented variety or the intermediate B variety. (See the discussion of Bembexia following the locality 6 description.) They are distributed from the finer grained siltstones at the bottom of the quarry through the coarser material at the top.

Peterboro South (Swamp Road South)

This locality is in the lower part of the Pecksport Member of the Marcellus Formation. The rock type is mudstone with very few sedimentary structures and abundant fossils. The preservation at this locality is quite exceptional for the Hamilton Group in that molluscs have their shells intact and are not preserved as composite molds as at Geer Road. Small amounts of original aragonite may still be found in some shells and the replacement of aragonite by calcite is so fine that the original shell microstructure is still preserved (Carter 1978).

All the specimens of Bembexia sulcomarginata that have been found in this quarry are of the unornamented type. They may be found throughout the entire quarry but are most abundant in the upper surface of the lower quarry face.

Ornamentation in Bembexia sulcomarginata

Bembexia sulcomarginata (Conrad 1871), one of the most common gastropods in the Hamilton Group, is characterized by great intraspecific variation in the amount of ornamentation on its shell. Workers have observed variation within the species but have failed to describe the range of variation or document the stratigraphic distribution of variants. An explanation for the observed variation is problematic due to an inability to determine variables such as temperature variations, food supply and predatory relationships. The only direct information available is from the sedimentary record and evidence of the depositional environment preserved in the coarsening up cycles.

Knight (1944) was the first to document variation in B. sulcomarginata. He described one variant with strong collabral and spiral ornamentation throughout ontogeny, and a selenizone always raised above the suture. Other specimens were unornamented, exhibiting faint growth lines across whorl faces parallel to the aperture. On these specimens, the selenizone was immediately subjacent to the suture and in some cases was completely or partially covered by it. Intermediate varieties exhibited collabral and spiral ornamentation early in ontogeny, loosing all ornament later in life. Rollins et. al. (1971) described variations concerned with overall shell form, position of the suture relative to the selenizone, strength of collabral and spiral ornamentation and duration of ornamentation throughout ontogeny. Rollins et. al. (1971) characterized B. sulcomerginata as ornamented, unornamented and intermediate.

Field work involved collecting specimens from localities distributed throughout the Marcellus and Skaneateles Formations of the Hamilton Group. Collections of approximately 600 specimens from these localities were analyzed and four morphologic variations of B. sulcomarginata were recognized:

1) Ornamented shells have collabral and spiral ornament throughout ontogeny, the selenizone is always raised above the suture and the shells tend to be more high spired than other forms.

2) Unornamented shells show faint growth lines and rarely subtle carinae growth. The selenizone is typically partly or completely covered by the suture.

3) Intermediate A has both collabral and spiral ornamentation on the early whorls with the collabral ornament being the first to dissappear. Spiral carinae typically extend past the early whorls.

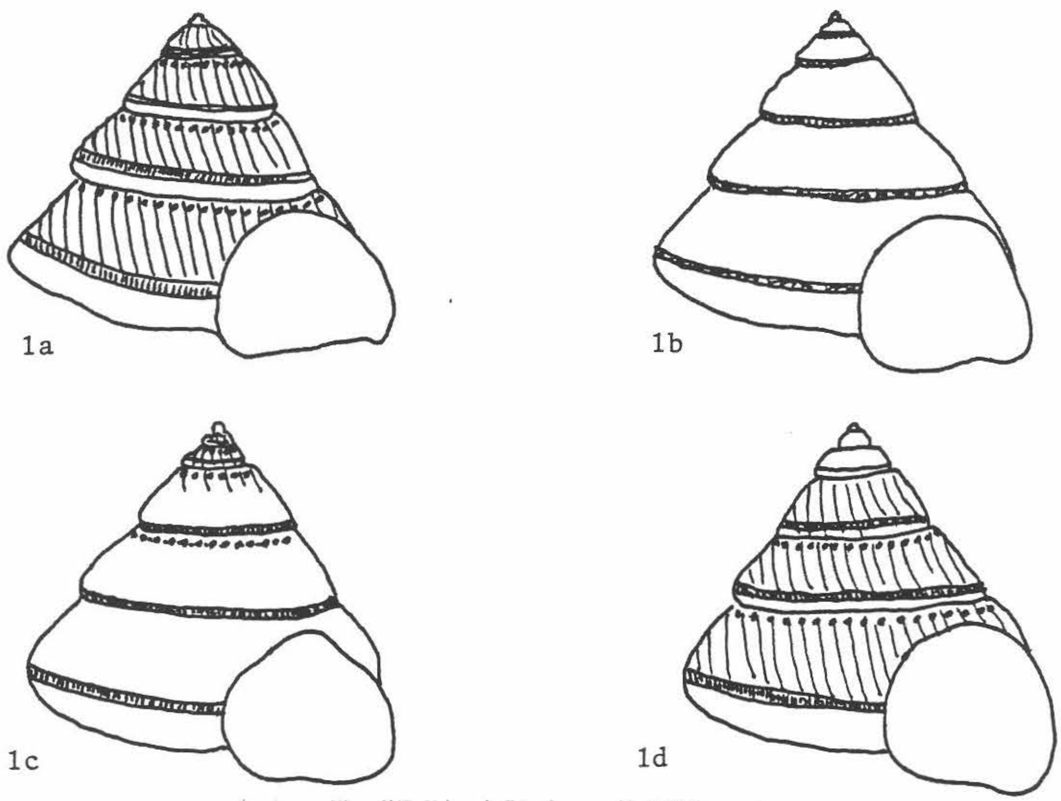
4) Intermediate B has the ornamentation reversed from that of intermediate A. The early whorls are smooth with collabral and spiral ornament developing on later whorls and continuing through ontogeny.

A series of ten measurements for each snail was developed based on different morphologies observed in hand specimen. Figures obtained from measurements were then entered into a spread sheet and graphed on a series of x-y plots and bar graphs. Significantly different trends resulted for unornamented and ornamented shells. These trends indicate the differing position of the selenizone relative to the suture between unornamented and ornamented varieties. As would be predicted from their morphology, both intermediate A and B samples showed characteristics of both unornamented and ornamented varieties.

Field observations demonstrate that the unornamented and ornamented varieties of B. sulcomarginata alternate in their occurence throughout the Hamilton Group (see Fig. 7). We may thus conclude that the differences are not the result of an evolutionary trend of unornamented to ornamented or vice versa. However, within any given horizon all of the individuals are either unornamented or ornamented morphotypes. Combinations of the two end member morphotypes were never observed on the same bedding plane. If intermediate forms are present, intermediate A is always associated with the unornamented variety. Thus whether the juvenile is ornamented or unornamented they all end up looking alike. (Likewise intermediate B is always associated with the ornamented variety). This suggests that the degree of ornamentation is an ecophenotypic adaptation to some environmental parameter. Perhaps the intermediates began life in one area and then migrated into an area which caused a change in their ornamentation. The shift in ornamentation type matched that of the individuals who had lived there all their lives.

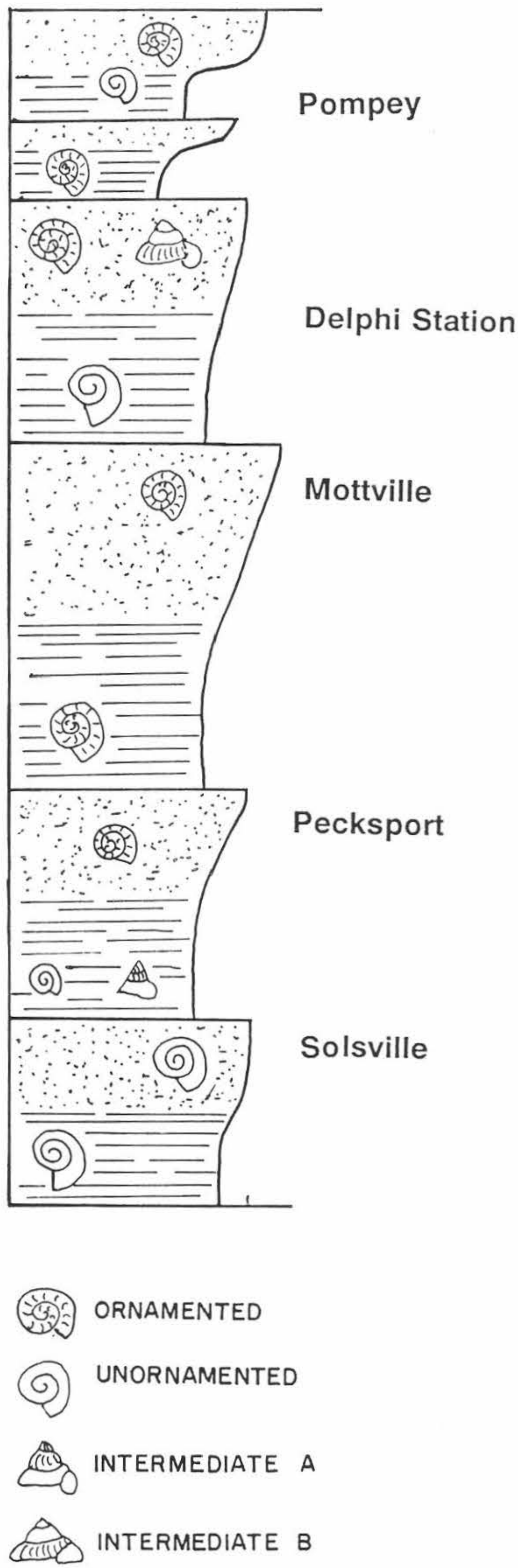
The obvious (and testable) environmental parameter which could account for the differences in ornamentation would be sediment grain size with its inferred parameter of turbulence. While there is a crude correlation between unornamented shells and fine grain size sediments (as will be seen at Peterboro south) it is not an invariant relationship (Fig. 7). At the English Avenue quarry all the individuals of B. sulcomarginata are either the ornamented variety or intermediate B, even though they may be found throughout the entire coarsening up cycle. Thus we feel compelled to reject the testable model of substrate grain size control on ornamentation and sugest that the causal factor is a variable such as salinity, food supply or a predator which we have not been able to document.

Figure 7



1a - ornamented B. sulcomarginata; 1b - smooth B. sulcomarginata; 1c - intermediate A; 1d - intermedaite B.

Figure 8



Pompey
Delphi Station
Mottville
Pecksport
Solsville
ORNAMENTED
UNORNAMENTED
INTERMEDIATE A
INTERMEDIATE B

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ROAD LOG

Mileage Miles	Cum	
		From Cutten parking lot turn right (north) onto Broad Street (Rt. 12B)
.1	.1	Turn right at stoplight onto the Colgate Campus (College Street)
.3	.4	Stops sign with Oak Drive, turn right.
.1	.5	Turn left past Merrill House to leave campus.
.1	.6	Stop sign, turn right onto Hamilton Street. Continue straight on Hamilton Street.
.8	1.4	Gorton Road is on the right. Continue straight on Hamilton.
1.5	2.9	Hamilton Street curves right, continue straight on Kiley.
.6	3.5	Intersection with Poolville Road. Stop sign, continue straight on Larkin Road.
1.2	4.7	Stop sign. Intersection with Route 12, continue straight on Larkin Road.
3.2	7.9	Stop sign. Turn left onto Moscow Road.
.5	8.4	Turn right onto Skaneateles Turnpike.
3.0	11.4	Junction with Ouleout. Keep going straight on Turnpike.
1.0	12.4	Stop sign. Junction with Ouleout Road. Go straight.
.6	13.0	Village of Brookfield.
.7	13.7	Turn right onto Dougway Road.
1.4	15.1	Stop sign. Intersection with Hoxie Road, turn right.
.3	15.4	Settlement of Five Corners.
1.3	16.7	**Locality 1 on the right. Hoxie Road Quarry.** Leaving quarry, turn left onto Hoxie Road. Retrace steps to Brookfield.
1.3	18.0	Back through Settlement of Five Corners.
.3	18.3	Junction with Dougway. Turn left.
1.4	19.7	Stop sign. Turn left onto Main Street (Skaneateles Turnpike).
.9	20.6	**Locality 2. Brookfield Quarry.** Leaving Locality 2, turn right onto Skaneateles Turnpike.
.3	20.9	Junction with Ouleout Road. Continue straight on Turnpike.
1.1	22.0	Junction with Ouleout Road. Turn right.
3.3	25.3	Turn right onto Barnes Road.
.5	25.8	Turn left onto Main Street.
.8	26.6	Stop sign. Junction with Route 12. Continue straight onto Swamp Road.
1.3	27.9	Stop sign. Turn left onto Cole Hill Road.
	27.95	Turn right onto Bailey Road.
1.0	28.9	**Locality 3. Bailey Road Quarry.** Leaving Bailey Road turn right (east). Back to Cole Road.
1.0	29.9	Stop sign. Turn right onto Cole Hill Road.
2.3	32.2	Turn right onto Rhoades Road.
.7	32.9	Turn left onto Quarterline Road.
.8	33.7	Village of Hubbardsville. Keep straight on Quarterline.
1.7	35.4	Turn right onto Kiley Road.
.7	36.1	Stop sign. Junction with Hamilton Street. Continue straight.
2.2	38.3	Turn left onto Colgate Campus.
.1	38.4	Turn right onto Oak Drive
.1	38.5	Turn left onto College Street.
.4	38.9	Stop light. Junction with 12B. Continue straight on College Street.
.5	39.4	Stop sign. Turn left onto Lebanon Street (Randallsvilel Road).
1.2	40.6	Turn right onto Armstrong Road.
1.0	41.6	Stop sign. Turn right on River Road.
1.2	42.8	Turn left onto Chamberlain Road.
.8	43.6	Stop sign. Turn left onto Lebanon Hill Road.
1.0	44.6	Turn right onto Geer Road.
1.0	45.6	**Locality 4. Geer Road Quarry.** Leaving quarry, turn left (east) onto Geer Road.
.9	46.5	Stop sign. Turn left onto Lebanon Hill Road.

1.0	47.5	Intersection with Chamberlain Road. Continue straight on Lebanon Hill Road.
2.3	49.8	Stop sign. Junction with Rt. 26 in Eaton. Turn right onto Rt. 26.
.5	50.3	Turn left onto English Avenue.
.7	51.0	Bear left on English Avenue.
.5	51.5	**Locality 5. English Avenue Quarry.** Leaving quarry, turn right onto English Avenue.
1.4	52.9	Stop sign. Continue straight.
1.2	54.1	Junction with Route 20. Turn left onto Route 20 (note that Route 20 is a divided highway).
.7	54.8	Entering Village of Morrisville.
.6	55.4	Stop light. Turn right onto Swamp Road.
2.3	57.7	**Locality 6. Peterboro South.** Leaving quarry, turn right. Continue along Swamp Road.
.5	58.2	**Locality 6a. Peterboro North.**

COMPARATIVE SEDIMENTOLOGY OF THE LOWER DEVONIAN MANLIUS FORMATION NEAR HAMILTON, NEW YORK

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INTRODUCTION

The paleo-environments of ancient sedimentary deposits are diagnosed from the vertical and lateral distribution of elemental rock units (**subfacies**), characterized principally by their assemblages of sedimentary structures, using analogs established from observations of processes and the sediments in modern depositional environments (the "*comparative sedimentology*" method of Ginsburg, 1974). Modern shallow marine carbonate environments carry a particularly rich inventory of primary and early diagenetic sedimentary structures, such as current and wave bedforms, trough and tabular cross-stratification, "herring-bone" cross-stratification, flat lamination, wavy and crinkled lamination, thin bedding, stromatolites, thrombolites, mudcracks, sheet cracks, prism cracks, flat pebble gravels, fenestrae, burrows, evaporite minerals, early cements, hardgrounds, caliche crusts, tepee structures, soils, and so on. And most significantly, the stratigraphic record back at least into the Proterozoic is replete with carbonate deposits that delicately preserve these sedimentary structures (see, for example, Ginsburg, 1975; Wilson, 1975; Hardie and Shinn, 1986; Grotzinger, 1989), demonstrating the existence through much of geologic time of environments and environmental processes analogous to those of modern shallow marine carbonate platforms and shelves.

There are other areas of study beyond reconstruction of paleo-environments where the environmental information preserved in the primary sedimentary structures and early diagenetic features of shallow water carbonates is of considerable value, for example, in the fields of paleo-oceanography, paleo-climatology and cyclostratigraphy. In particular, tidal flat facies are unsurpassed "sea level gauges," "tide gauges," and "climate recorders" (Hardie, 1977, p.188-189). An unambiguous record of the position of ancient mean sea level is engraved within the intertidal subfacies of all ancient shallow marine carbonate deposits. And the same intertidal subfacies carry a record, quantitatively determinable, of the tidal range in the depositional environment (cf. Klein, 1971). At the same time, the subtidal subfacies record the ambient and storm wave energy levels across the buildup, and this information in turn reflects, at least in a qualitative way, the prevailing weather patterns. The nature of the supratidal subfacies is a direct response to the prevailing climate in the region (Hardie and Shinn, 1986). If the climate is arid, like the modern Persian Gulf carbonate environments, then the supratidal subfacies will carry evaporite and aeolian features (Shinn, 1983). If the climate is rainy, like the modern Andros Island tidal flats, then the supratidal deposits will carry freshwater marsh and lake features (Hardie, 1977; Shinn, 1983).

The stratigraphic record abounds with shallow water carbonate deposits characterized by meter-scale vertical successions of subfacies that are organized into shallowing-upward "cycles", as revealed by analysis of primary sedimentary structures and early diagenetic features (Wilson, 1975; James, 1984; Hardie and Shinn, 1986). Vertical stacks of such shallowing-upward cycles record repeated fluctuations in relative sea level. In some cases these sea level fluctuations appear to be periodic oscillations, driven by Milankovitch astronomical rhythms (e.g. Goldhammer and others, 1987, 1990). Clearly such cyclic shallow water carbonates take on a special significance as storehouses of information about global climatic and eustatic variations in the past.

Overall, an understanding of the origin and significance of primary sedimentary structures and early diagenetic features is vital in our quest to unravel the origin and significance of carbonate deposits in the geologic record. Without such an understanding at the individual sedimentary structure scale we cannot hope to accurately reconstruct the large scale accumulation history of carbonate buildups or to decipher the roles of sea level changes, sedimentation rates, subsidence rates, and tectonics in determining the facies stratigraphy, cyclostratigraphy and sequence stratigraphy of these buildups. In summary, it could be said that *unless we get the little things right we may not be able get the big things right.*

With this in mind, the purpose of this trip is to examine two or three sections of the Manlius Formation (Lower Devonian) near Hamilton, New York. We are principally concerned on this trip with describing the **subfacies** that comprise the Manlius Formation in this area. We shall then use the sedimentary features preserved in these rocks to interpret their depositional significance. Defining cycles of subfacies that record relative sea level changes in the Manlius Formation is a controversial matter. One need only compare: (1) Laporte's (1975) original ideas about migrating tidal flat islands; (2) the shallowing-upwards cycle definitions of Kradyna (1992); and (3) the allocyclic stratigraphy of Anderson and Goodwin (1991). Literally every possible kind of cycle and interpretation of cycle significance have been suggested for this unit. It is not our intention to review this controversy. However, it is our contention that a thorough analysis of the primary and early diagenetic sedimentary structures that are so well preserved in this unit **must** be the **starting point** for any paleo-environmental analysis of these rocks including any analysis of their cyclostratigraphic significance.

GEOLOGIC SETTING

The Manlius Formation is the lowermost unit of the Helderberg Group, a succession of Lower Devonain carbonates exposed across central and eastern New York. Rickard (1962) interpreted the Manlius Formation as time transgressive from east to west. Figure 1 is a portion of Rickard's (1962) stratigraphic chart of the outcrop belt of the Helderberg Group across central

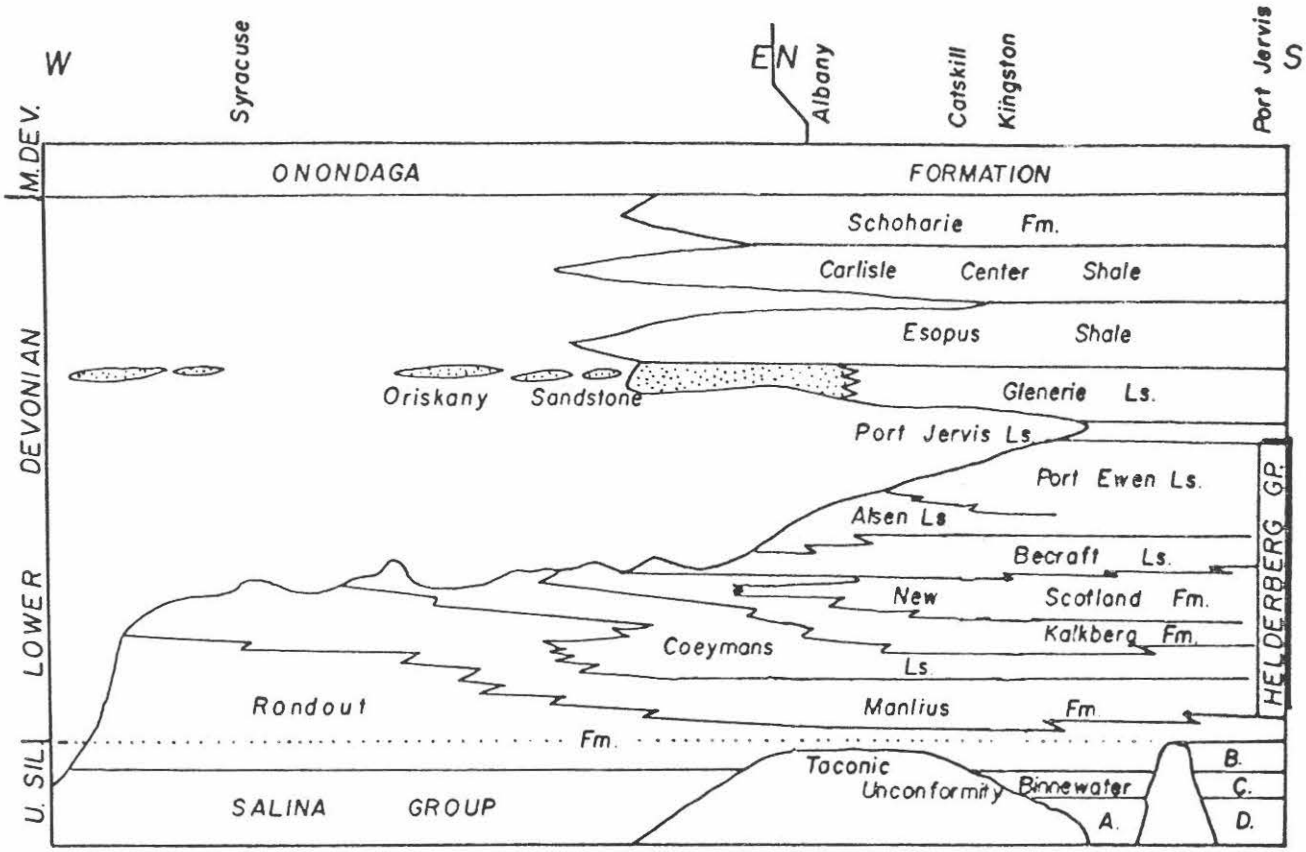

Figure 1. Stratigraphic section and lateral distribution of the Helderberg Group and associated rock units in New York State. From Rickard (1962).

New York from Cherry Valley on the east to Syracuse on the west. Here the Manlius Formation is very nearly flat-lying and is divided into a number of members (Fig. 1).

DESCRIPTION OF MANLIUS SUBFACIES

Figures 2 and 3 are measured stratigraphic logs of the Manlius Formation from two stops of this fieldtrip: (1) Clockville and (2) the Jamesville Quarry (locations given in Figure 4). A third optional stop is the quarry at Munnsville. At these three locations the Manlius Formation can be divided into 6 subfacies defined by their assemblages of primary and diagenetic sedimentary structures: (1) *grainstone subfacies*; (2) *microbial bioherm* subfacies; (3) *wavy to lenticular thin beddded* subfacies; (4) *laminite* subfacies; (5) *disrupted mudstone* subfacies; and (6) *thin beddite* subfacies. These are described below.

A note on the significance of the colors of the rocks is in order. In outcrops in this part of the Appalachians the color of a well weathered surface is a fairly reliable guide to the mineralogy of the rock. Limestone (composed of low magnesium calcite) generally weathers blue or grayish blue. On the other hand, dolomite ($CaCO_3 \cdot MgCO_3$) commonly weathers a tannish yellow. Dolomite takes on this color upon weathering as most natural dolomites have small amounts of iron substituting for Mg. Upon weathering the iron oxidizes and stains the rock with iron oxide.

Grainstone Subfacies

The grainstone subfacies is composed of intraclastic, bioclastic, and peloidal grainstones and conglomerates. The grainstones are generally well sorted and individual sets vary from coarse to fine sand-sized. Intraclastic conglomerates have rounded clasts up to 20 mm in diameter. In two outcrops of the Manlius Formation we will examine, the main primary sedimentary structure of the grainstones and conglomerates is planar stratification. The sets of planar strata are tens of millimeters thick and are separated by dolomite-rich seams a few millimeters thick. Cross-stratified grainstones are rare in these two outcrops; one probably example is 4.5 m above the base of the Jamesville Quarry section.

Microbial Bioherm Subfacies

The microbial bioherm subfacies includes thrombolites and stromatolites found at the Clockville and Munnsville Quarry sections. The thrombolites of the Manlius Formation are described in Browne and Demicco (1987). Aitken (1967, p. 1164) proposed the term thrombolite: "(from the Greek thrombos , bloodclot) ... for cryptalgal structures related to stromatolites, but lacking lamination and characterized by a *macroscopic clotted fabric* (italics ours)." Other terms that are commonly used to refer to these and similar structures are "bioherm" and "biostrome" (e.g. thrombolitic bioherm or biostrome). The former term refers to discrete mound or coalesced mounds (interpreted to be of organic origin) embedded in rocks of a different lithology whereas a biostrome is a layer consisting of and built mainly by organisms. Manlius Formation thrombolites occur as discrete mounds, as coalesced mounds and as continuous biostromes. Figure 5 is a bedding diagram of a portion of the Clockville section illustrating the three layers that contain thrombolites. The basal layer (0 to 1 m) contains discrete mounds up to 1 m thick and coalesced mounds where laterally adjacent thrombolites have welded upon upward growth. The thrombolites in the bottom layer are surrounded by planar-stratified grainstones. The upper most layer at Clockville (3.8 to 5 m above the base) is a

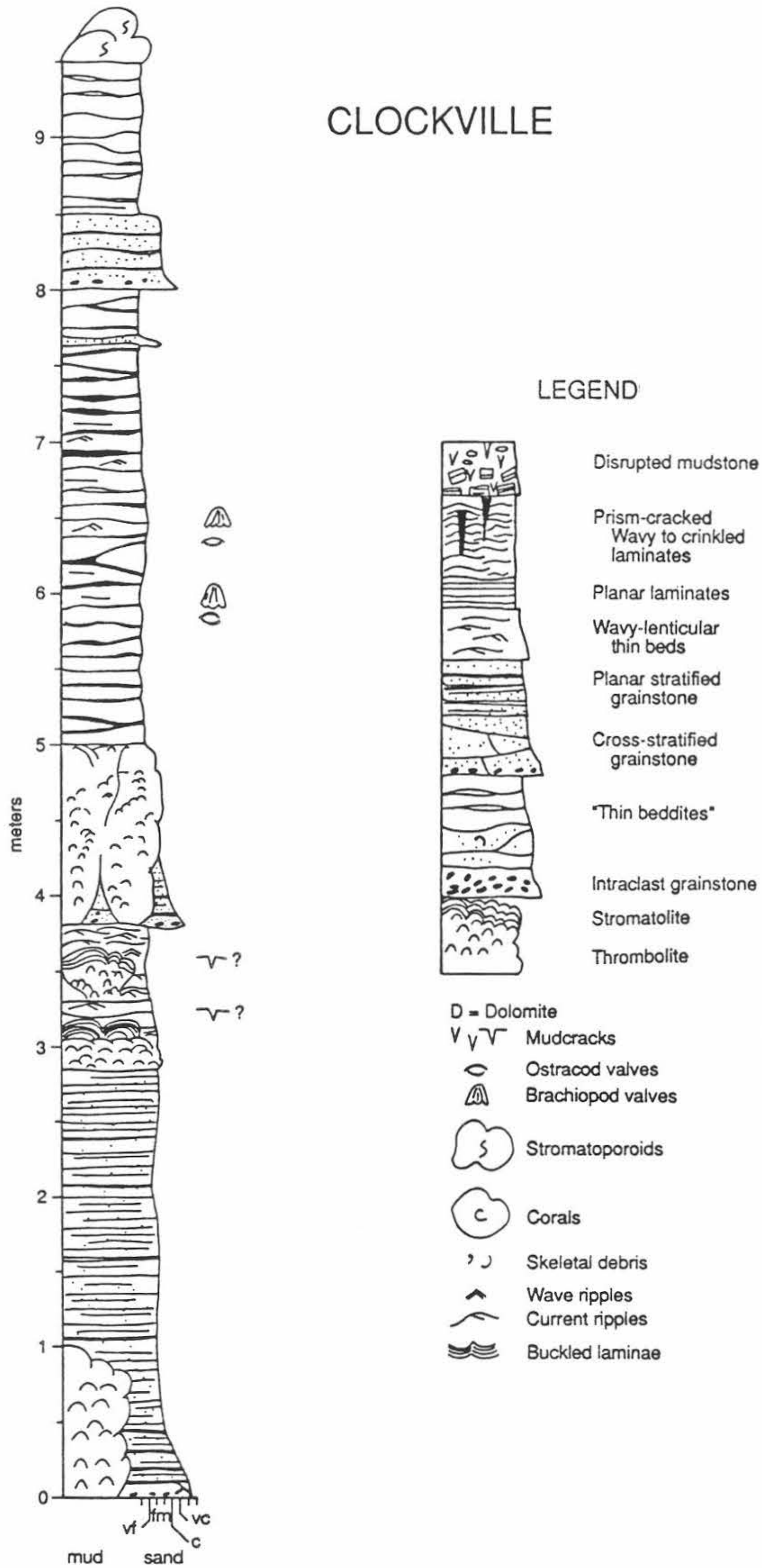

Figure 2. Measured stratigraphic section of the Lower Manlius Formation at Clockville.

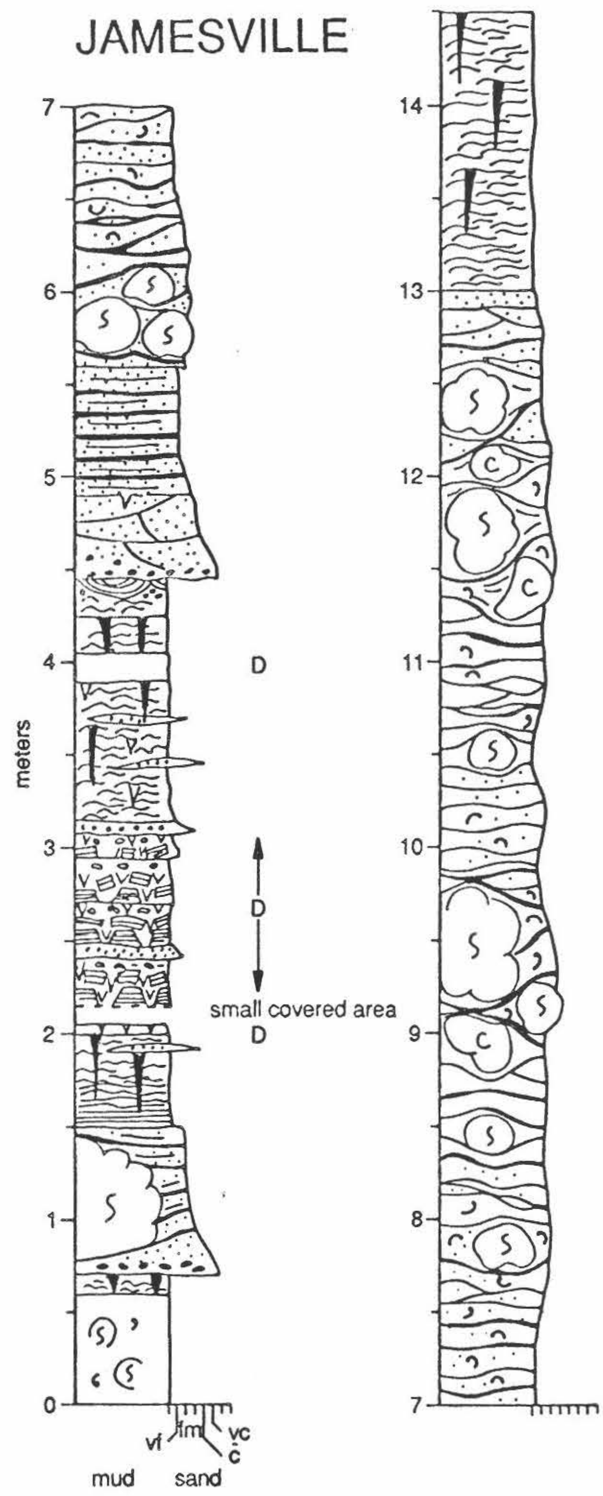

Figure 3. Measured stratigraphic section of the Manlius Formation exposed in the Jamesville Quarry.

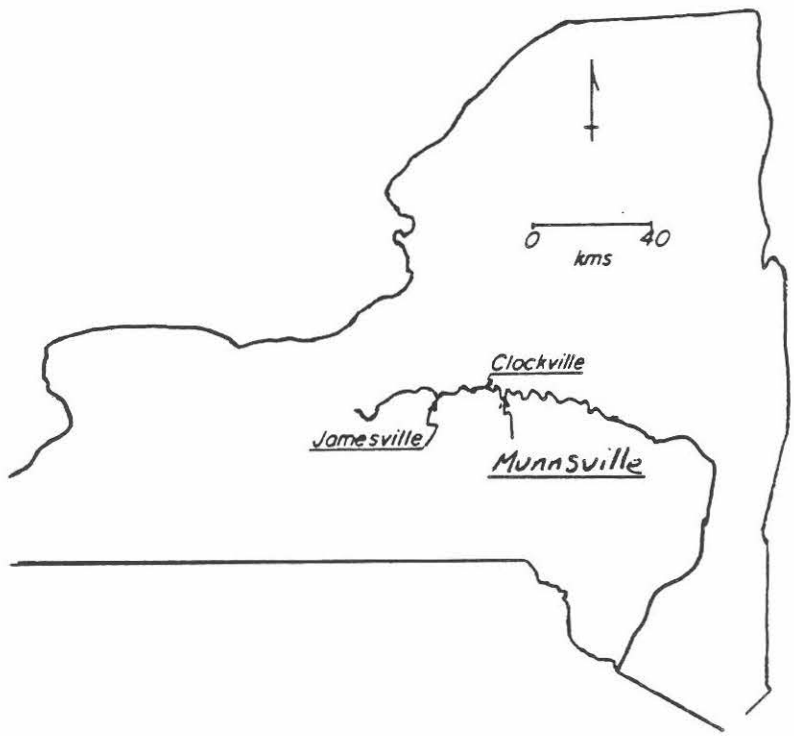

Figure 4. Outcrop belt of the Helderberg Group in New York State. Field trip stops indicated.

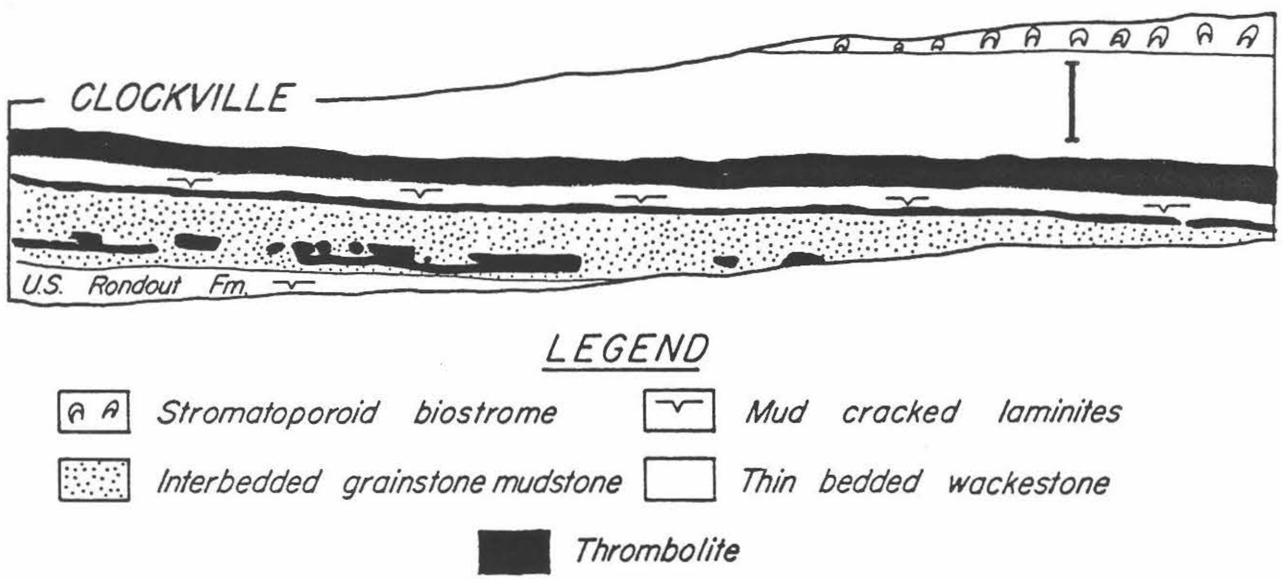

Figure 5. Scaled diagram of thrombolites and associated rock types at Clockville, New York. Vertical scale = horizontal scale = 2 m.

biostrome. Irregularly- shaped pockets of grainstone and intraclastic conglomerate are found within this layer. The middle layer of thrombolites between 2.9 and 3.8 m is an interesting and complicated one. The lowermost 200 mm of this layer is a continuous thrombolite biostrome with stromatolites nucleated on its upper surface. Higher up in the layer are scattered, discrete, delicate thrombolites composed of upward-directed fingers of mudstone. These also are capped with stromatolites. All of the stromatolites in this layer have "gravity defying" mudstone laminae. However, surrounding the gravity defying laminae are wavy laminae of fine peloidal sand that are quite different. These wavy laminae are part of the wavy and lenticular thin bedded subfacies and are more fully described below.

The internal structures of thrombolites of the Manlius Formation at the hand-sample scale are complicated. Structures are visible in sawn slabs and in those outcrops which have been delicately etched by selective weathering. Internally, Manlius Formation thombolites comprise millimeter- to centimeter-scale mudstone masses ("clots" or "mesoclots") surrounded by skeletal mudstones, skeletal wackestones, skeletal packstones, or rare sparry patches. The mudstone masses typically make up 30 to 50 % of the rock and vary from highly irregular shaped to vertically-oriented columns with circular cross-sections in bedding plane views. Columns can branch upward. There are two types of thrombolites at the Clockville section: (1) those found between 0-1 m, 2.8-3 m, and 3.75-5 m above the base; and those found at 3.5 m above the base.

Mudstone masses in the thrombolites found between 0-1 m, 2.8-3 m, and 3.75-5 m above the base at Clockville are made up of 5 components: (1) millimeter-scale hemispheroids with radial and concentric structure which Browne (1986) interpreted as the problematic fossil *Keega*; (2) variable amounts of the problematic fossil *Renalcis* and masses of micrite that superficially resemble this form; (3) stromatolitic mudstone laminae; (4) peloidal laminae with micron-diameter filament molds; (5) rare, submillimeter-scale brachiopod, ostracod and gastropod fragments. In some cases Browne (1986) reported possible framework-building skeletal red algae encrusting the sides of mudstone clots. Structureless skeletal mudstones, wackestones and packstones that are lighter in color fill the spaces between the mudstone masses. Millimeter-scale burrow-tubes filled with peloids disrupt the fills among thrombolitic mudstone clots and fingers but do not cross-cut them.

The thrombolites in the layer at approximately 3.5 m differ from other Manlius Formation thrombolites insofar as the digitate fingers that make up these thrombolites are dense mudstone and they *do not contain* microfossils. Also, the fingers of these thrombolites are surrounded by mudstone with no skeletal fragments. Finally, the thrombolite mounds themselves (and the ones immediately below) are capped by *stromatolites* and surrounded by a unique subfacies: the wavy to lenticular thin bedded subfacies described below.

Wavy to Lenticular Thin Bedded Subfacies

The wavy to lenticular thin bedded subfacies is only found in the outcrops of this field trip surrounding thrombolites and stromatolites 3.5 m above the base of the Clockville section. The rocks comprise dolomite fine-grainstones alternating with dolomite mudstones. Mudcracks are rare. The grainstones have flat bottoms and wavy tops. The wavy tops of the grainstone layers appear to be oblique cuts through wave ripples and, indeed, internally the grainstone layers are sets of wavy fine lamination typical of oblique cuts through wave ripples (see de Raaf and others, 1975). The mudstone layers also are sets of fine wavy laminae. These mudstone layers drape the ripple-marked surfaces of the grainstones, thinning over crests and thickening

The *geometry* of the limestone layers and sets varies from planar and continuous through discontinuous lenses to notably *wavy* and *nodular* forms similar to "nodular", "irregular", "flaser", and "lumpy bedding" described by Wilson (1975), Wilson and Jordan (1983), Matter (1967), and Schwarz (1975), among others. At the lateral boundaries of the nodules, internal graded layers within the limestone patches thin by compaction and pass laterally into dolomitic shaley partings. Other volume reduction features are very common in these rocks, especially in the mudstones and wackestones. Stromatoporoids and corals of various scales are common in the thin beddite subfacies of the Manlius Formation. In many cases, these rigid skeletons are encased in dolomitic shaley seams which bend around the skeleton from both above and below (drag or penetration effects of Pray, 1960). The affected seams thin laterally directly above and below the bioherm and show increasing dips and thicknesses down the sides of the skeleton. Significantly, seams are thickest along the flanks of the skeleton and thin not only over and under the bioherm but laterally away from the bioherm as well. It is important to note that nodules of mudstone show the same drag effects of shaley dolomitic seams around them just as the stromatoporoids and corals do. This, in turn, suggests that the limestone nodules were hard at the time of compaction as well.

COMPARATIVE SEDIMENTOLOGIC INTERPRETATIONS

By far the easiest rocks to interpret are the *laminites*. The crinkled geometry of many of the laminae, a number of which are clearly composed of detrital fine-sand and silt-sized peloids, suggests that sediment laminae were agglutinated by a sticky microbial mat. The desiccation cracks suggest periodic exposure. Moreover, the prism-cracks imply the periodic rise and fall of a ground-water table (Ginsburg, 1991, pers. comm.), and drying out in the vadose zone. Modern analogs of this subfacies are found beneath high intertidal to supratidal subenvironments of low-energy tidal flats where there is an organic mat dominantly composed of cyanobacteria ("blue-green algae"). Modern examples of laminated sedimentary deposits directly influenced by a surface mat of cyanobacteria have been described from : (1) the supratidal islands of Florida Bay (Ginsburg and others, 1954); (2) the intertidal and supratidal mud flats and coastal marshes of Andros Island by Black (1933), Monty (1967, 1972, 1976), Monty and Hardie (1976), and Hardie and Ginsburg (1977); (3) the intertidal and supratidal mud flats of the Trucial Coast of the Persian Gulf by Kendall and Skipwith (1968) and Kinsman and Park (1976); (4) the intertidal sand and mud flats of Shark Bay in Western Australia (Davies, 1970; Logan and others, 1974) and (5) the siliciclastic mud flats accumulating behind barrier islands of the Delmarva Peninsula, Virginia, by Harrison (1971). Subaerial desiccation of thick fleshy surface mats results in disruption into polygonal cracks (e.g. Kinsman and Park, 1976) the upturned edges of which become preferential sites for growth of the succeeding generation of cyanobacteria to produce oversteepened "stromatolitic" layering (e.g. the type C algal heads of Black, 1933). The anticlinal tepee-like buckles of the laminite 4.3 m above the base of the Jamesville Quarry section have two possible origins. Lateral expansive growths of cyanobacterial mats can produce buckles in soft sediment (see Figure 8A and 8B, p. 178, in Shinn, 1983). Alternatively, expansive growth of *cements* within crusts forming in the high intertidal to supratidal zone can produce anticlinal buckles where crusts overthrust and break.

The laminite subfacies of the Manlius Formation records deposition on a tidal flat. However, it should be noted that "tidal" is used in a rather loose way. The periodic introduction of sediment onto most of the modern examples cited above is by storms. Indeed, for many modern tidal flats, the normal tidal process is to inundate the cyanobacterial mats with clear water. Many modern examples are more properly called wind tidal flats insofar as the prevailing wind direction can have as much, if not more, to do with the flooding of the flats as the tides do.

The *disrupted mudstone subfacies* also has a quite elegant modern analog: the mudcracked soils developing on modern playas in closed basins and supratidal flats (Smoot, 1983; Smoot and Katz, 1982; and Smoot and Lowenstein, 1991). These modern playa muds are riddled with mudcracks of various sizes and spacings that are open at the surface but at depth are filled with mud, muddy sand, or sand. Mudcracks branch and connect via a complicated network of sheet cracks that surround irregularly-shaped patches of finer mud. Within these modern muds are irregularly shaped fenestrae that are the result of entrapped air bubbles and desiccation. Typically, laminated Pleistocene lake clays grade up through a brecciated zone into a chaotic mud disrupted by complex, superimposed mudcracks. This sequence strongly resembles the smaller cycles that comprise the disrupted mudstone subfacies. It is interesting to speculate as to the *time value* of each of the small cycles in this subfacies because Holocene/Pleistocene examples may record thousands of years of slow aggradation and intense disruption by desiccation.

The *thrombolites* of the Manlius Formation are similar to their more common Cambrian and Lower Ordovician counterparts and are similarly interpreted as small, in-place mounds that were hard, rigid bioherms with a biogenic framework when deposited; in other words, they were biological reefs. The clotted mudstone that makes up Manlius Formation thrombolitic fingers is interpreted as calcified cyanobacterial filaments similar to *Girvanella.* The problematic microfossil *Renalcis* is generally interpreted as calcified coccoid cyanobacteria or an encrusting foraminiferae. The enigmatic micrite clots that give thrombolites their name have also been interpreted to be the result of calcification of cyanobacterial mats and colonies.

Oddly enough, thrombolites of the Paleozoic resemble in some respects porous carbonate mounds that were deposited in Pleistocene and Holocene pluvial lakes collectively known as *tufa.* Pleistocene to Modern lacustrine tufas comprise mounds and coalesced mounds that make large encrustations up to 30 m high. Internally, lacustrine tufas may be composed of: (1) rigid centimeter-scale clots and fingers composed of micrite that strongly resemble *Renalcis*, *Girvanella*, etc.; (2) stromatolitic laminae and small stromatolites; and (3) dendritic masses and arborescent millimeter- to centimeter-scale "shrubs" interpreted by Chafetz and Folk (1984) to by calcified bacterial colonies. Surrounding these framework elements in lacustrine tufas are a variety of detrital sediments, many of which show evidence of penecontemporaneous cementation.

The *wavy to lenticular thin bedded subfacies* surrounds thrombolites and stromatolites 3.0-3.75 m above the base of the Clockville section. We are not sure of the origin of this subfacies and critical to any interpretation is whether these rocks are mudcracked. They probably represent alternating layers of wave-ripple cross-stratified fine sands and suspension settle-out of mud, implying an on-off wave regime. Furthermore, these rocks are not bioturbated, they contain stromatolites which are otherwise rare in the Manlius Formation, and associated thrombolites contain no microfossils. These observations suggest that this is the deposit of a subenvironment with waters inimical to organisms. Our best guess is that this subfacies represents some kind of restricted pond or lagoon with elevated salinities developed behind a lateral barrier.

The *grainstone subfacies* surrounds thrombolites at the base of the Clockville section and is likewise interpreted as a shallow subtidal shelf deposit. The common planar stratification implies that deposition commonly occurred beneath an upper-stage plane bed generated by either unidirectional or oscillatory (wave) currents. The common fining-upwards sets capped by finer-grained dolomites suggests these are storm deposits.

The *thin beddite subfacies* has long been recognized as subtidal shelf deposits (Laporte 1967; Walker and Laporte 1970). We concur. However, we feel that there are two features of this rock type worthy of note.

First, these rocks have been severely effected by diagenesis. There is ample evidence of differential cementation and differential compaction in the common drag effects of layering around both hard skeletal elements and nodules of limestone. Moreover, the obvious association of the shaley dolomitic partings with compacted sediment opens the door to consideration of pressure-solution dolomitization (Wanless, 1979; and Logan and Seminiuk, 1976). Indeed, most of the preserved layering here might be due to early, layercake diagenesis argued for by Bathurst (1987). The nature and significance of dolomite and shaley dolomitic partings (which, after all apparently define the main layering style) in these rocks must await formal study of the diagenesis of this formation.

The second point about the thin beddite subfacies of the Manlius Formation is that we see absolutely no compelling reasons to subdivide these rocks. The nature of the diagenetic overprints, the nature of the internal sedimentary structures, and the nature of their preserved fossils suggest that these rocks are rather insensitive indicators of water depth. Moreover, there is no one no parameter in these rocks that is sensitive to water depth. Wave ripple marks preserved on bedding surfaces would be an obvious place to start considering their potential for quantitatively giving water depth (Komar, 1974; Clifton, 1976). However, much of the waviness of bedding surfaces in this subfacies may be the result of diagenesis. As an example of the difficulty in interpreting water depth we would like to draw attention to the grainstone coset 8 m above the base of the Clockville section. At first consideration, it may seem that this would be a good place to "draw a line" separating deposits of different water depth. However, examination of the thin beds above and below this layer will reveal similar, thinner sets of coarse grainstones. In this respect, and in its overall geometry and crude internal organization, this grainstone most likely represents a storm deposit on the shelf similar to those from siliciclastic shelves. We do not think that such "event layers" are necessarily good candidates for major changes in depositional environments.

CONCLUDING REMARKS

We hope that we have shown that the sedimentary and diagenetic features of the Manlius Formation are a storehouse of information that is vital to any attempt to unravel the depositional significance of these rocks. Questions need to be answered before the larger-scale significance of these rocks can be addressed. What was the nature of the original deposits that were diagenetically altered into the thin bedded subfacies? What is the sequence of diagenetic events that effected these deposits? Are there storm deposits preserved in this subfacies or are there significant depositional breaks? What is the nature of the small-scale, desiccating upward cycles preserved in the disrupted mudstone? How much time do they represent? Are they caliche soils like those capping Pleistocene carbonates of the Florida - Bahama Banks Province? Are there tepee structures preserved that imply early cementation?

ACKNOWLEDGEMENTS

The discussion of Manlius Thrombolites is from an M.A. thesis by Kathy Browne, Geological Sciences Dept., State University of New York at Binghamton.

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SEDIMENTOLOGY OF THE DENLEY LIMESTONE (MIDDLE ORDOVICIAN, TRENTON GROUP) IN CENTRAL NEW YORK

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Introduction

The Denley Limestone of the Trenton Group (Middle Ordovician) in central New York is characterized by nearly 90 meters of alternating beds of limestone and shale. The origin of limestone and shale couplets in general is problematic, however, in the case of the Denley Limestone, the limestones can be shown to be bioclastic turbidites in origin. These bioclastic turbidites are recognized by Bouma divisions within the limestone beds. Interbedded shale horizons represent the *in situ* sediment. This sedimentologic interpretation has implications for the paleoenvironmental setting of the Denley Limestone.

This field trip will ask you to address the following questions: (1) How can resedimented carbonates be distinguished from *in situ* sediment? (2) How can tempestites be distinguished from turbidites? (3) Why should one bother with the tempestite/turbidite distinction?

Regional Geology

Outcrops of the Middle Ordovician Trenton Group extend through the St. Lawrence Valley of southern Quebec, the Camplain Valley of New York and Vermont, and eastward through the Mohawk and Black River Valleys of central New York before extending once again into Canada above Lake Ontario (Figure 1). In addition to being so widespread, these rocks have long attracted attention for their abundant and diverse marine faunas. Early models by Kay (1953) for the depositional setting of the Middle Ordovician Trenton Group described an eastward (present coordinate) dipping carbonate ramp which passed into an adjacent shale basin. More recent publications by Cisne and Rabe (1978) and Cisne, et al. (1982a) describe a revised interpretation for the depo-tectonic setting of the Middle Ordovician sequence. These authors suggested that these sediments accumulated in a outer shelf to trench slope

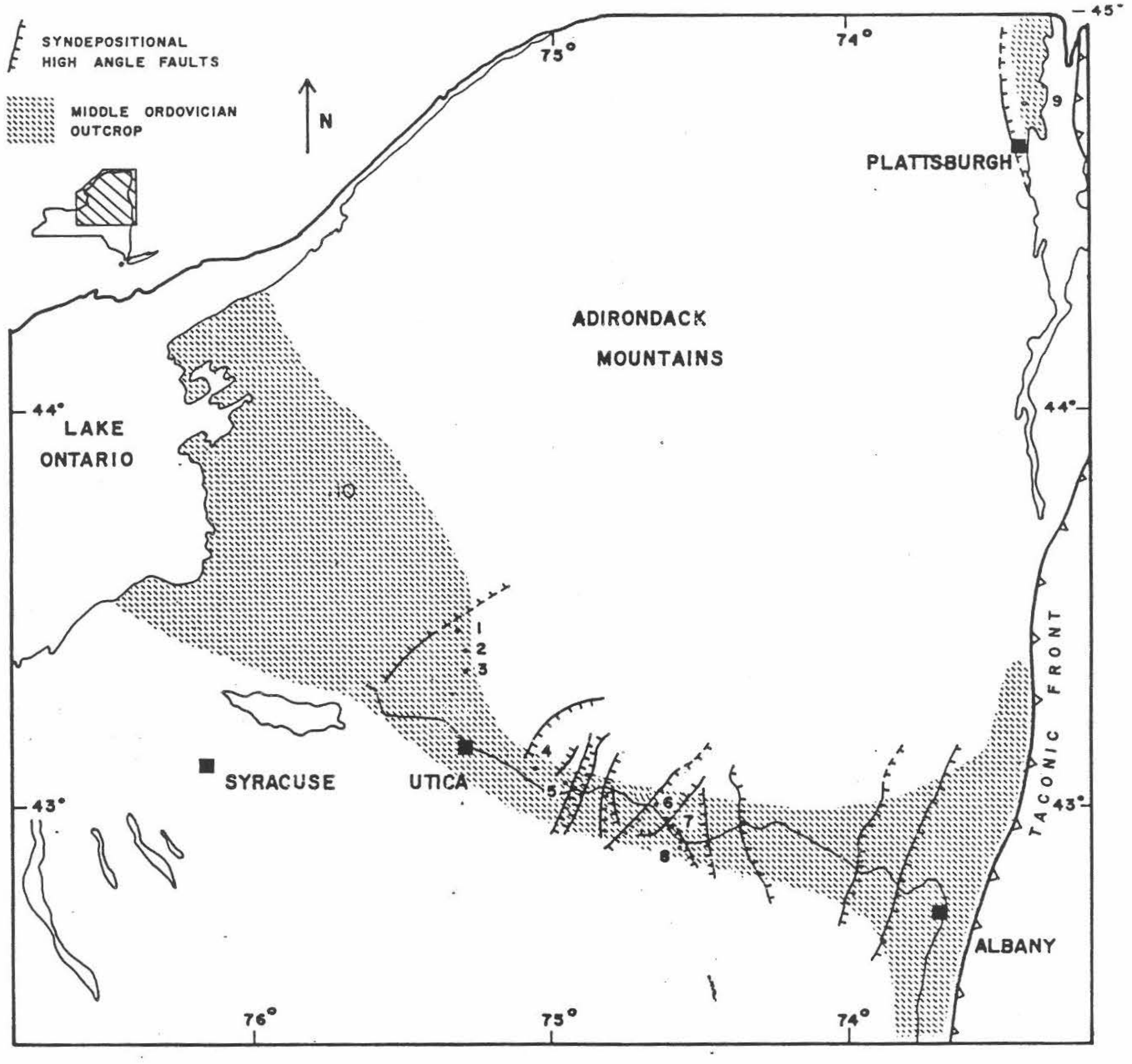

Figure 1. Outcrop map of the Trenton Group in New York State. Key locations include: 1-Trenton Falls Gorge; 2-Buttermilk Creek; 3-City Brook; 4-Herkimer; 5-Little Falls; 6-Caroga Creek; 7-Palatine Bridge; 8-Canajoharie Creek; 9-Beekmantown; 10-Roaring Brook.

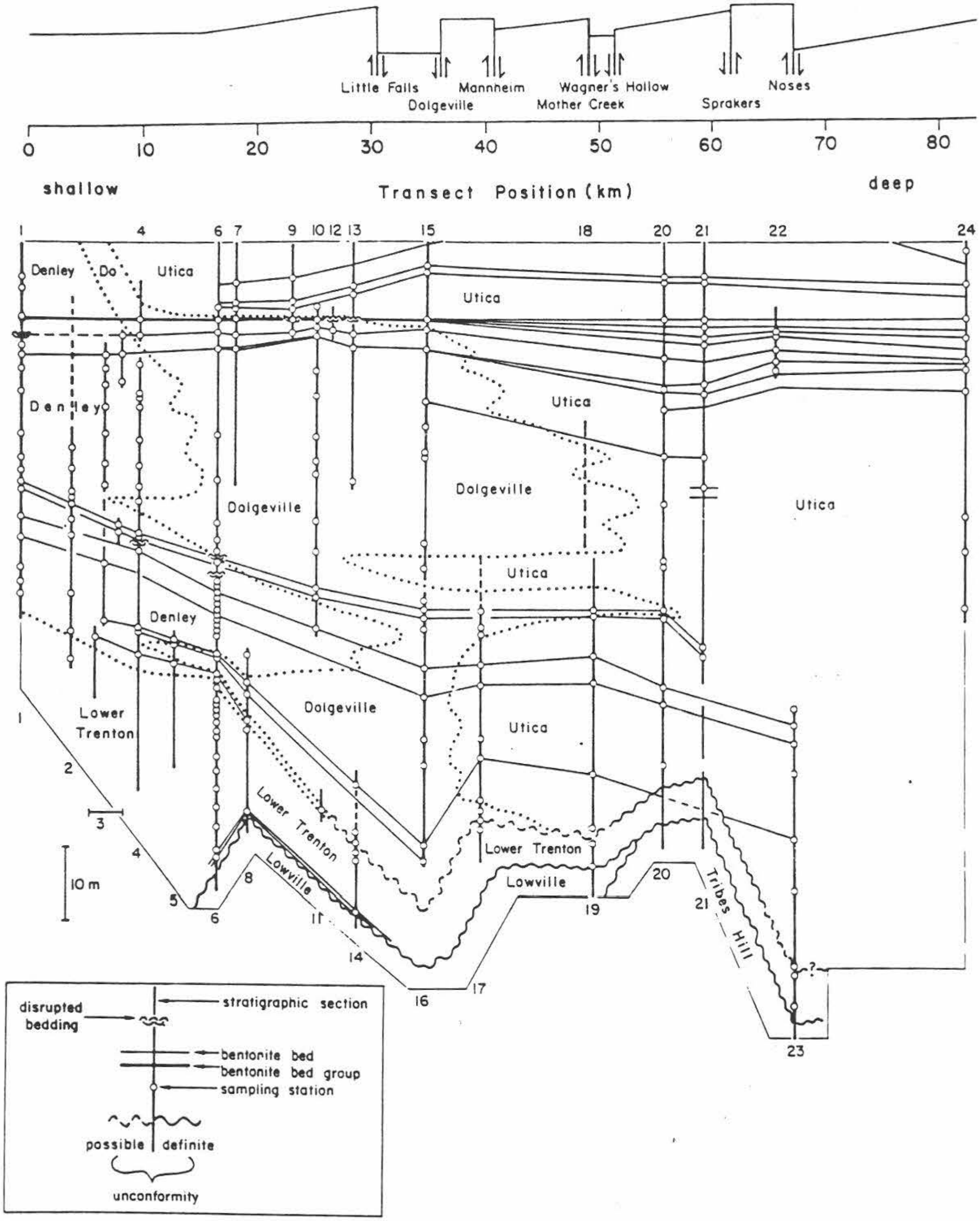

Figure 2. Stratigraphic sections of the Trenton Group and over-and underlying units are represented by vertical lines and have been correlated by bentonites (sub-horizontal lines). Sampling horizons for faunal studies are indicated by circles on vertical lines. The structural cross section at the top of the diagram indicates that the Trenton strata thicken across basins and thin across basement highs, suggesting that faulting was syn-depositional. Sections: 1-Trenton Falls Gorge; 2-Gravesville-Mill Creek; 3-Poland; 4-Rathbun Brook; 5-Shedd Brook; 6-Buttermilk Creek; 7-Farber Lane; 8-Norway; 9-Miller Rd.; 10-North Creek; 11-Gun Club Rd.; 12 & 13-New York State Thruway; 14-Burrell and Bronner Rds.; 15-W. Crum Creek; 16-Dolgeville Dam; 17-E. Canada Creek; 18-Mother Creek; 19-Caroga Creek; 20-Canada Creek; 21-Flat Creek; 22-Currytown Quarry; 23-Van Wie Creek; 24-Chuctanunda Creek. From Cisne and Rabe, 1978.

setting, adjacent to the Taconic Arc. An analogy would be the present day Australian flank of the Timor Trough, where the northwestern Australian carbonate shelf is dissected by numeorus high angle block faults which control the thickness and distribution of carbonate lithofacies. Sediments accumulating in these isolated basins are primarily bioclastic turbidites in origin (Veevers, et al., 1978), derived from the adjacent shallow water carbonate shelf. Cisne and Rabe (1978) and Cisne, et al. (1982a) suggested that the Middle and Upper Ordovician strata of central New York were deposited on a fragmented shelf in the foredeep where each fault block accumulated a unique stratigraphy. By using bentonite horizons to correlate fault blocks, Cisne and his coworkers constructed paleotopographic maps which illustrated the existence of local topographic highs and lows superimposed on an overall easterly dipping slope into the trench (Figure 2). A possible mechanism for producing extensional block faults within a compressional tectonic regime was presented by Chapple (1973), who suggested that the formation and passage of a peripheral bulge through the foreland basin would generate extensional faults. Mehrtens (1988) described the existence of syn-depositional block faults in the Taconic foreland basin of southern Quebec and also documented their effect in controlling the stratigraphy and sedimentology of these Middle and Upper Ordovician strata.

Stratigraphy

Kay (1937 and 1953) was the first to present a stratigraphic framework for these units which was later modified by Titus and Cameron (1976) and Titus (1988).

The sedimentology of the basal units of the lower Trenton Group were described in detail by Titus and Cameron (1976) and are interpreted to represent a continuum of progressively deepening environments from lagoonal (Napanee and Selby Formations) to shoal (lower Kings Falls Formation) to shelf facies (middle and upper Kings Falls Formation) and ultimately into deeper water shelf (Sugar River Formation).

The Denley Limestone, the focus of this study, overlies the Sugar River Limestone (Figure 3). At its type locality at Trenton Falls (Figure 1, locality 1) the Denley has been divided into three members, in ascending order the Poland, Russia and Rust Members.

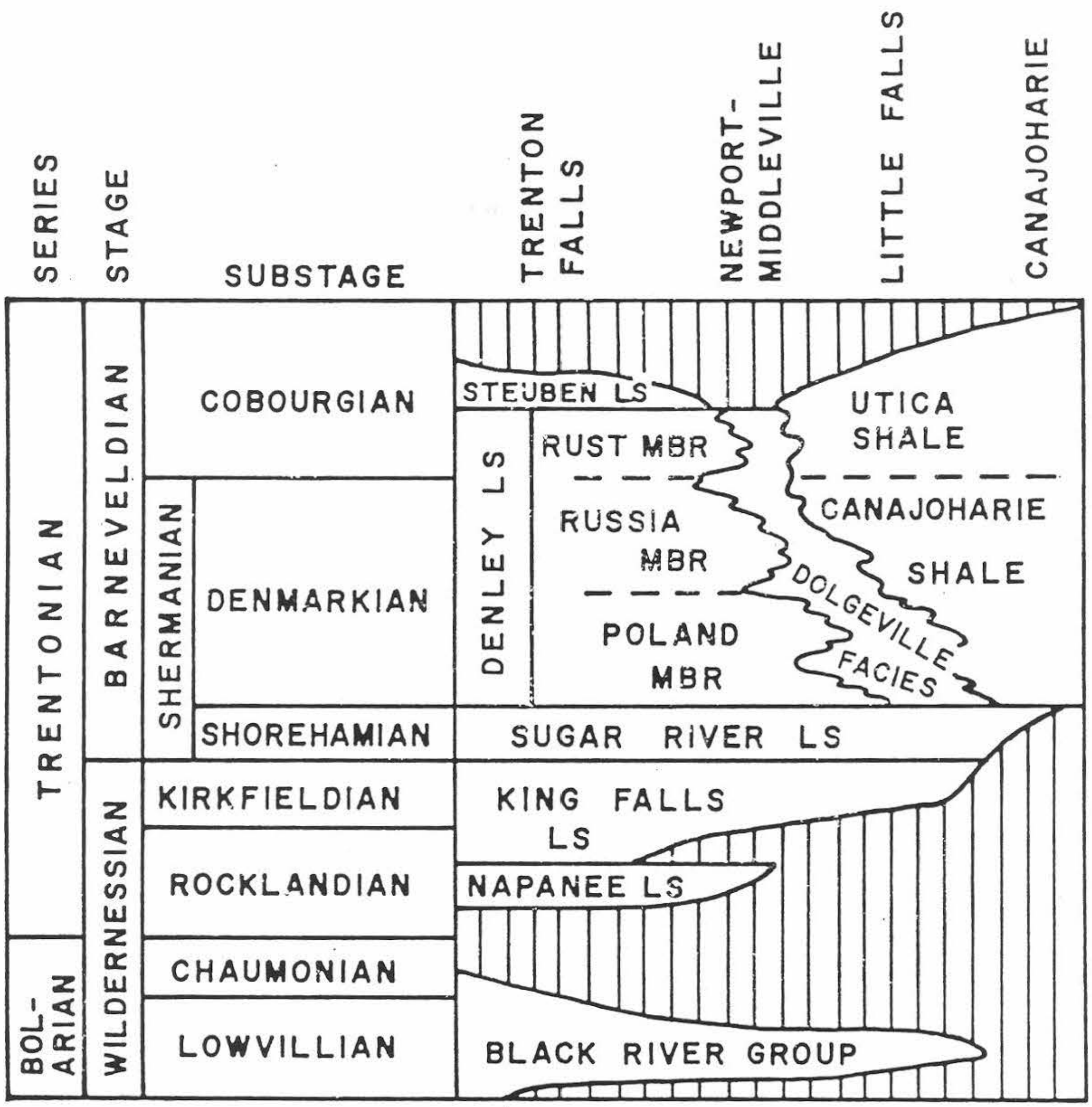

Figure 3. Correlation chart for the Black River and Trenton Groups in central New York, modified from Titus and Cameron (1976).

The contact of the Sugar River and Denley Formations is visible at several localities and appears gradational (Figure 4).

The Denley Limestone is overlain by a variety of units, depending upon position within the basin. At its type section at Trenton Falls the Denley is overlain by the Steuben and Hillier Limestones. Titus (1988) describes the Steuben as a coarse-grained calcarenite bearing wave-generated structures such as cross bedding which he interpretes as a shoal facies. The Hillier Limestone is a finer-grained calcsiltite lacking evidence of wave reworking and is interpreted by Titus as a shelf to outer shelf deposit. These limestones are in turn conformably overlain by the Utica and Canajoharie Shales (Riva, 1972, his figure 16). Eastward of the Trenton Falls type section the Denley is overlain by the Dolgeville Facies, a laminated mudstone with thick shale interbeds, which Kay (1953) described as the transition beds from the Denley to the Canajoharie Shale.

The deepening-shallowing-deepening sequence recorded by the Trenton Group in central New York has been interpreted in various ways. Titus (1988) attributes this oscillation to facies mosaicing accompanying base level changes generated by epeirogeny of the Adirondack Arch. An alternative model proposed by Cisne and Rabe (1978) suggests that bathymetric changes, irregular bottom topography, and condensed stratigraphic sequences result from the activity of syndepositional block faults within the Taconic foreland basin which subsequently exerted control of sedimentation patterns.

Sedimentology of the Denley Limestone

The Denley Limestone occupies a stratigraphic position in the Trenton Group at the culmination of the deepening trend which initiated with the Napaee and Selby Formations. The Denley contact with the underlying Sugar River Limestone is gradational and is defined as the first appearance of an unfossiliferous, "lumpy-bedded" mudstone lithology. This lithology is interbedded with fossiliferous wackestone and packstone limestone beds and shale interbeds bearing diverse and abundant assemblages of benthic marine organisms. Figure 4 illustrates the stratigraphy at two outcrops where the Sugar River/Denley contact is well exposed. Although the contact is gradational, the Denley and Sugar River can

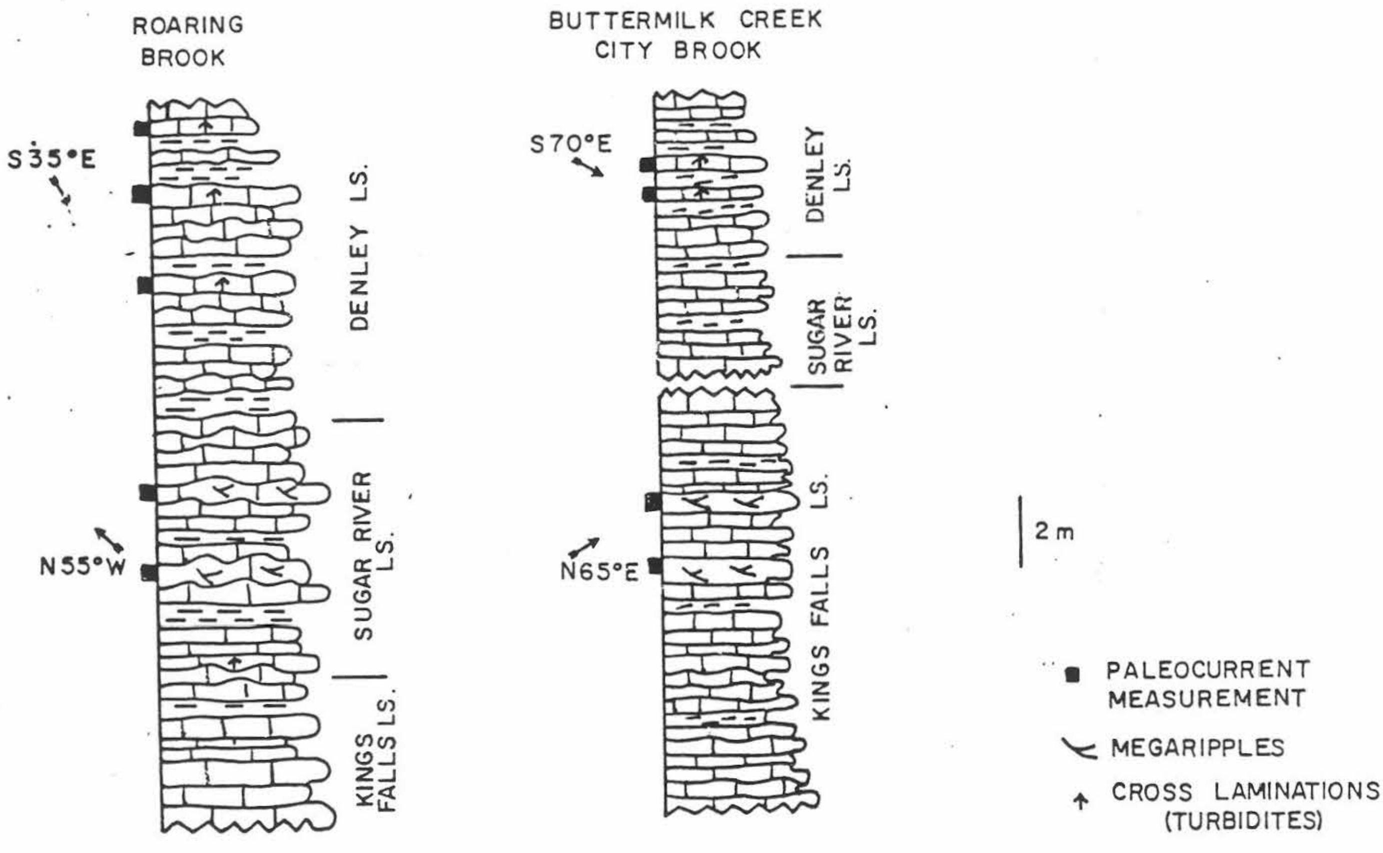

Figure 4. The contact between the Sugar River and Denley Limestones is gradational, for example, as seen at Roaring Brook (Figure 1, locality 10) and Buttermilk Creek/City Brook (Figure 1, locality 3).

be differentiated on the following criteria: (1) the Denley is a wackestone to packstone in composition while the Sugar River is a packstone to grainstone; (2) the Denley has thicker shale interbeds, locally exceeding 1cm in thickness; shale is often absent between limestone beds in the Sugar River; (3) the Denley contains limestone horizons which contain a predictable sequence of sedimentary structures which are interpreted as Bouma divisions; (4) the Denley contains horizons of slump folds and breccias (Figure 5): (5) the Sugar River and underlying Kings Falls Limestones display features which suggest deposition at or near wave base, including beds that pinch and swell, cross bedded grainstones, and tempestite horizons; (6) Paleocurrent measurements from ripple crestline trends are approximately 90 degrees different between the Kings Falls and Sugar River and Denley Limestones, reflecting currents oriented relative to a shoreline or shoal (Kings Falls) versus those oriented downslope (Denley).

The sedimentology of the Denley Limestone was initially undertaken as part of a study on the evolution of the trepestone bryozoan *Prasopora,* whose colonies cover bedding planes throughout the unit (Mehrtens and Barnett, 1979). Initially, the 90-m thick Trenton Group type section at Trenton Falls gorge (Figure 1, locality 1) was sampled and large thin sections were made of entire beds. Petrographic analysis revealed that limestone beds often contained sequences of sedimentary structures, in ascending order: layers of skeletal fragments, mm-scale thickness of planar laminae composed of finely crushed skeletal debris; mm-scale laminae of micrite; bioturbated micrite; and shale (Figure 6). These sequences were identified as Bouma divisions within bioclastic turbidites, suggesting that, at least in part, the Denley Limestone represented turbidite deposition. Following recognition of a bioclastic origin for many of the limestones, sampling to determine the stratigraphic and paleogeographic distribution of the turbidites was undertaken at other Trenton Group exposures. Outcrops sampled included Upper West Canada Creek, Rathbun Brook, Shed Brook, Roaring Brook, City Brook, Caroga Creek, and outcrops around the Champlain Valley of New York and Vermont. In the course of this study it became clear that a comprehensive study of bioclastic turbidites was lacking and that great potential exists for misidentification of bioclastic turbidites with tempestites (see Kreisa, 1981).

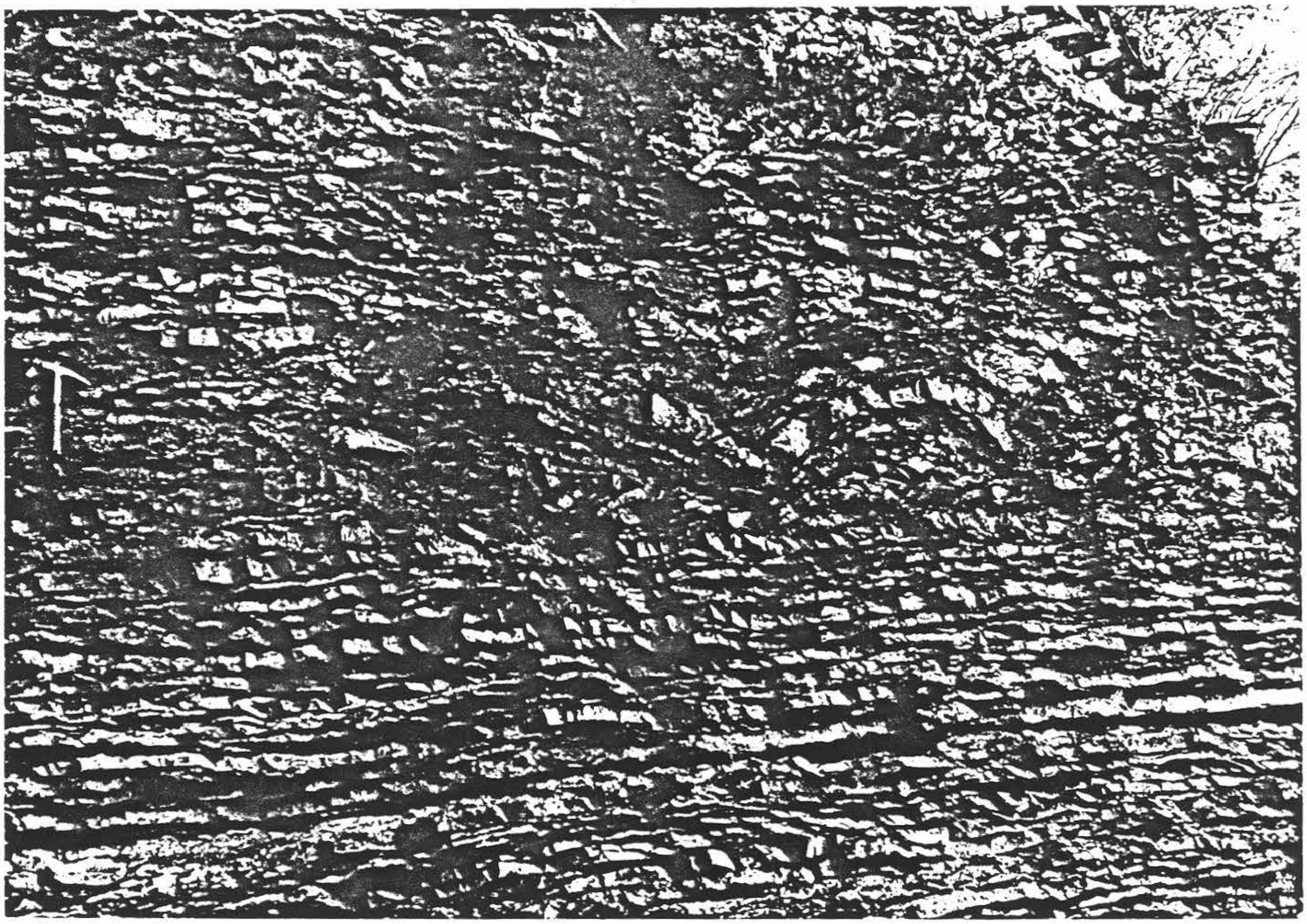

Figure 5. Slump fold and breccia horizon (at hammer level) in the Denley Limestone along West Canada Creek. The presence of brecciated beds indicates that the sediment was at least partially lithified prior to movement, which suggests early, possible submarine, lithification occured.

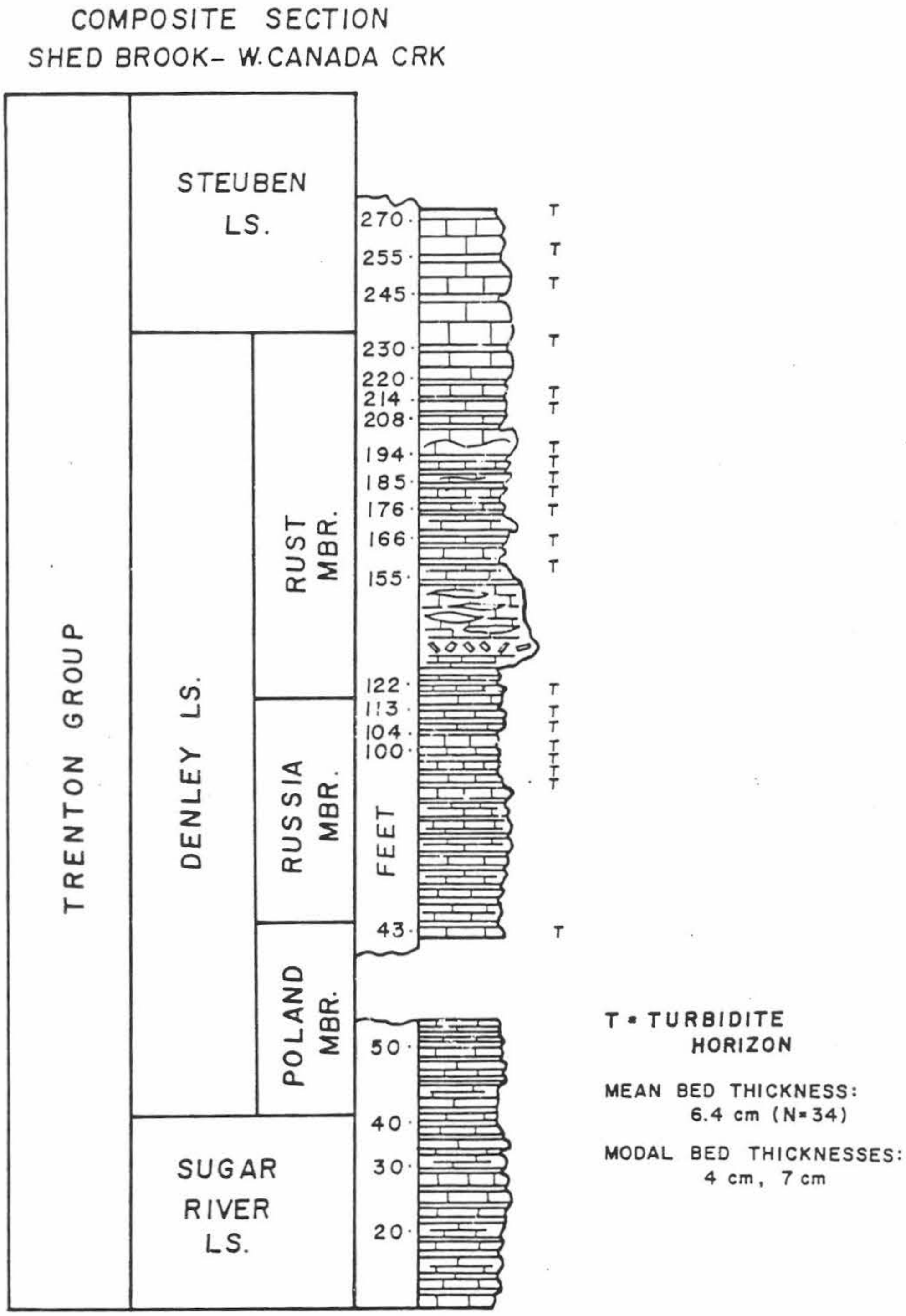

Figure 6. Measured stratigraphic section of the Sugar River-Denley and Steuben Limestones at Shed Brook and West Canada Creek (Trenton Falls Gorge). Horizons which upon thin sectioning were identified as turbidites are indicated by a "T".

Approximately 70% of the Denley Limestone horizons sampled at Trenton Falls Gorge (n=34) are interpreted as resedimented detritus, most likely turbidites, however there are several other lithologies present within this unit (Figure 7). This variety of lithologies are sufficiently different to suggest multiple origins. Three non-turbiditic lithologies present include: (1) wackestone-packstone beds lacking any sedimentary structures; (2) fossiliferous shale interbeds between the limestone horizons; (3) unfossiliferous mudstone beds. Given the abundance of bioclastic turbidite horizons, these three lithologies can be interpreted in light of the turbidite model.

The wackestone-packstone limestone beds contain disarticulated whole and broken skeletal fragments in a matrix containing variable amounts of micrite. No sedimentary structures, such as grading, laminations, or micrite infiltration are seen. Fossil fragments are unsorted in size. This lithology is interpreted to represent an *in situ* accumulation, and, lacking any evidence of resuspension and settling, is thought to have been deposited below wave base. Care must be taken to not confuse these *in situ* accumulations with top or bottom cut-out turbidites. These are turbidites which, due to an absence of either the necessary grain size or hydraulic conditions, do not exhibit complete Bouma divisions. Thin section examination should reveal any internal structures. Also, if there are long periods of time between bioclastic turbidite emplacement, bioturbation can obliterate Bouma structures. A second *in situ* sediment accumulating within the Denley Limestone is the shale present between limestone beds. These beds can be quite thin, forming a veneer on top of the limestones, however they often support diverse benthic communities (Cisne and Chandlee, 1982; Cisne, et al., 1982b; Hay and Cisne, 1988). The presence of articulated fossils, including trilobites and articulated crinoid stems, attests to the probable quietude of the waters during the deposition of this material.

The final limestone lithology is the unfossiliferous mudstone whose most diagnostic features are the extensive horizontal burrow traces which cover bedding planes and extend several centimeters into beds and the "lumpy bedding" which can produce up to 10cm of relief on a bedding plane surface. The origin of the "lumpy bedding" is attributed to the burrowing activity of the trilobite *Isotelus,* followed by submarine cementation which preserved the high angles of respose of the sediment. The carbonate mud

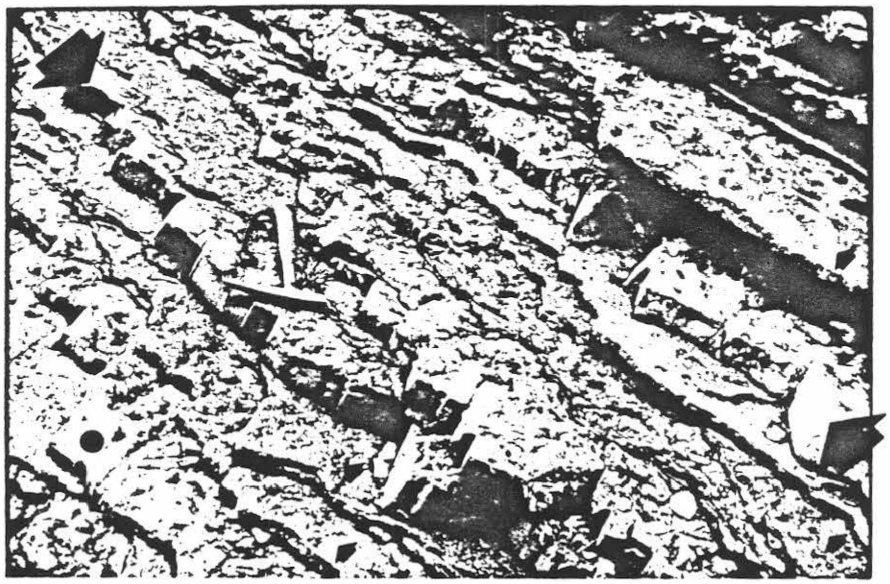

Figure 7. Photograph of the four important lithologies in the Denley Limestone along West Canada Creek. Tubidite beds (largest arrows) are clearly recognizable by their planar bases and bioturbated tops. A bed of the "lumpy mudstone" is labeled with a black dot. Thin beds of packstone-grainstone are indicated with small arrows. The fourth lithology, shale, is present between all limestone beds.

composing this lithology is thought to represent accumulations of peri-platform ooze derived from the adjacent shelf which was resuspended and transported into deeper water in a series of low density tubidity current flows.

The interbedding of these four lithologies suggests that the Denley Limestone was deposited in an outer ramp setting. The bioclastic turbidites and peri-platform ooze were resedimented downslope into an environment where the dominant *in situ* sediments were shale with lesser amounts of in situ fossiliferous limestones. Although turbidites can form in shallow water, an outer ramp, deep water environment for the Denley Limestone is supported by the presence of slump fold and breccia horizons, both of which require a slope on which to form. Orthocone cephalopods oriented parallel to slump fold axes indicate the presence of currents which flowed downslope.

Description of Bioclastic Turbidites

Although bioclastic turbidites have been described by many authors (see Mehrtens, 1988 for references). In all of these numerous examples there was ample supporting information for placing the limestones in a slope or basin environments. Until the recent work of Cisne and his coworkers there was no such model for the Denley Limestone. Furthermore, the Denley Limestone is in gradational contact with a unit (Sugar River Limestone) which has attributes of shallow water deposition. Finally, a great deal of attention has been given to the recognition of limestone tempestites which outwardly appear similar in structure to bioclastic turbidites.

The general characteristics of bioclastic turbidites include: (1) division of beds into a lower graded and upper mudstone portion (Figure 8); (2) mudstone which contains pelagic faunal elements versus shallow water faunas in the lower, graded horizons; (3) graded carbonate skeletal sand units capped by laminated or cross laminated units (Figure 9); (4) planar bases and bioturbated tops; (5) partial to complete Bouma divisions. Table 1 provides a more complete list of the characteristics of bioclastic turbidites compared to those of limestone tempestites. Bioclastic turbidites are structurally similar to siliciclastic turbidites described by Bouma (1962) and hydraulically interpreted by Walker (1965), however there are several noteworthy differences, described below. The coarsest-grained sediment deposited from the initial stages of

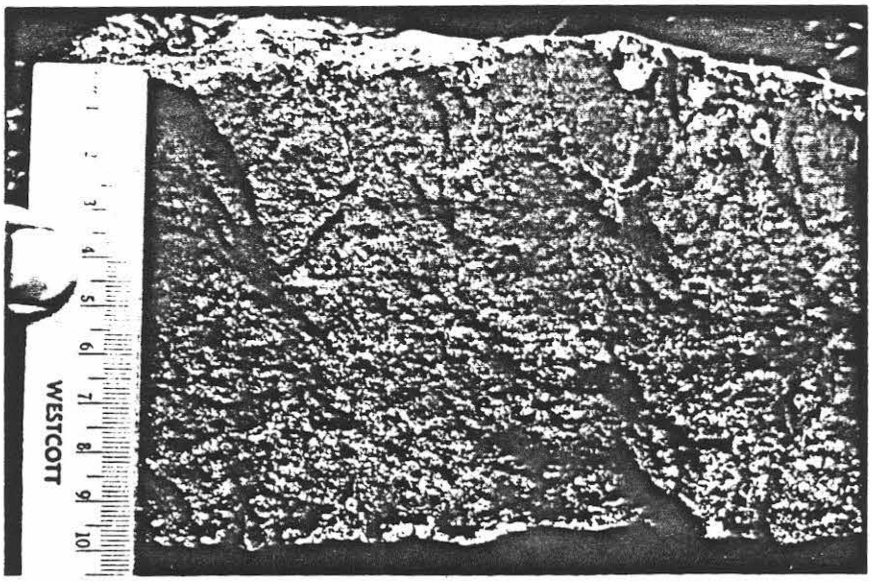

Figure 8. Outcrop photograph of a graded limestone bed. Note the sharp base to the limestone bed, with a concentration of the coarsest skeletal grains, and fining-upwards in skeletal size.

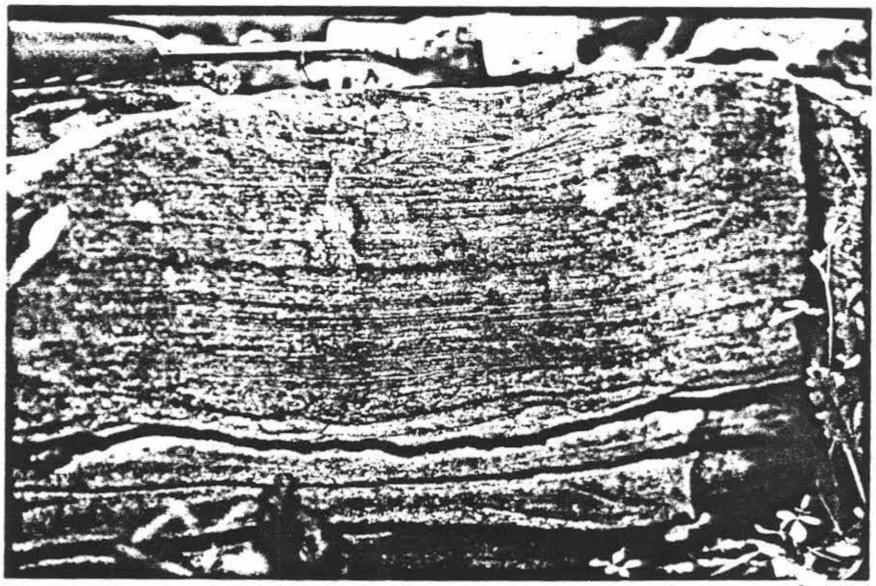

Figure 9. The hammer head rests on a limestone bed which exhibits well defined planar (large arrow) and cross (small arrow) laminations.

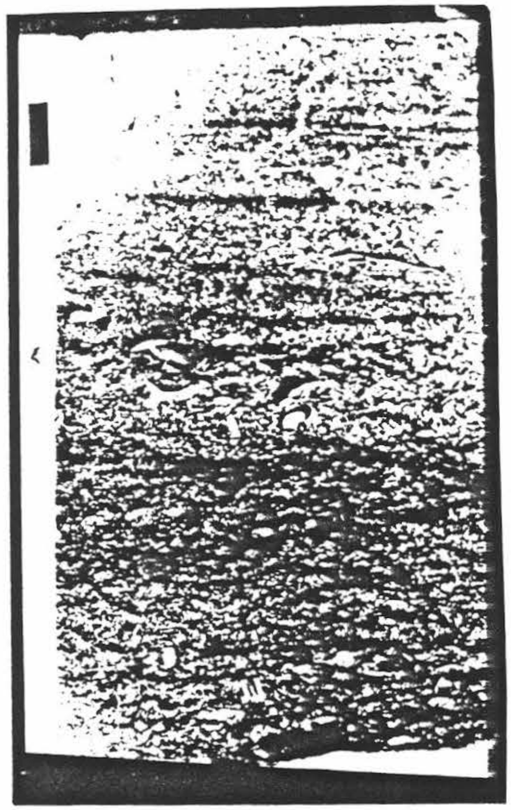

Figure 10. Thin section photomicrograph of a turbidite which exhibits grading in skeletal fragments (Ta). Sample is from the uppermost Rust Member of the Denley Limestone. Bar equals 1 cm.

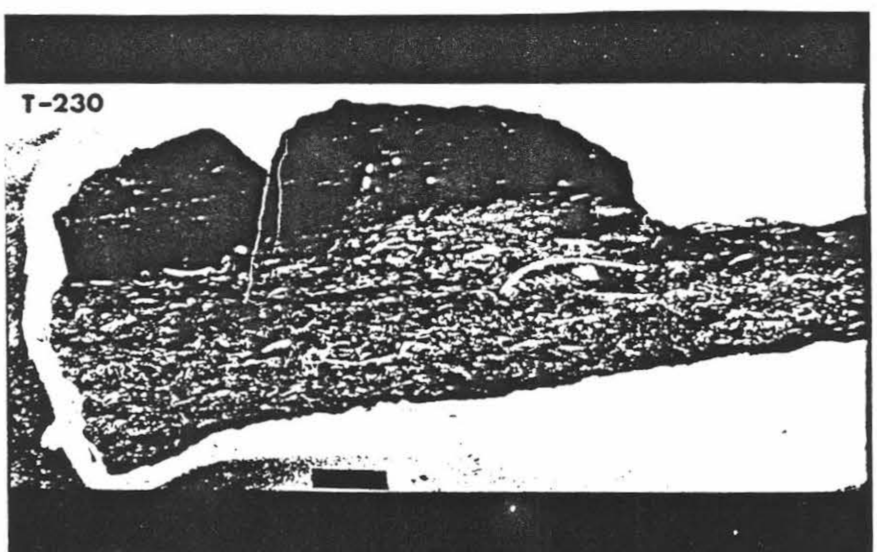

Figure 11. Thin section photomicrograph of a turbidite from the Rust Member of the Denley Limestone which exhibits a coarse-grained skeletal lag (Ta) overlain by structureless micrite (Td'). Bar equals 1 cm.

deceleration of the current form a graded unit (Ta) composed of skeletal material (Figure 10). The high velocity planar laminated unit (Tb) is composed of mm-thin laminae of diminuted skeletal material. The third Bouma division, ripple cross laminae (Tc) is present, but uncommon in bioclastic turbidites. These structures are thought to form from either unidirectional waves behind the nose of the turbidity current of by reworking of planar laminations (Tb) into ripples (Walker, 1965), however the formation of ripple laminations requries cohesionless sand and silt-sized material. It is possible that bioclastic material is too cohesive, and the absence of a silt-size equivalent in carbonate sediment is well known. The origin of the fine-grained laminae (Td) in terrigenous mud turbidites is still enigmatic. (Piper, 1972 a & b; Piper, 1978; Rupke, 1975) but in bioclastic turbidites the laminations can be seen to consist of alternations of very finely crushed skeletal grains and their interparticle cement alternating with micrite. Overlying the finely-laminated Td unit should be Te, a structureless pelagic mud. Piper (1978) has summarized the controversy surrounding the origin of this mud which led him to subdivide the mud fraction into E1 (laminated mud), E2 (graded mud), E3, (ungraded mud), and F (hemipleagic mud= shale cap). The barren and frequently bioturbated carbonate mud present at the top of bioclastic turbidites (Figure 11) is designated Td' (Td prime), a name chosen to most accurately reflect the genetic association with the finely-laminated turbidite material below than with the inter-turbidite material above. Overlying the micrite of Td' is the terrigenous shale (Te) which is often colonized by benthic organisms. Burrowing activity of these organisms often extends well into the turbidite beneath, destroying primary sedimentary structures. Several authors (Piper, 1978; Hesse, 1975; Scholle, 1971) have described this type of bioturbation as diagnostic of post-turbidite recolonization of substrates.

Terminology for describing the internal structures of bioclastic turbidites follows that proposed by Einsele (1982) where "T" designates a turbidite origin (as opposed to "S" for storm, or tempestites), and lower case letters a through e designate the specific bedform. Therefore, Ta = turbidite layer of coarse-grained graded material; Tc = cross laminated turbidite layer, and Tac = vertical sequence of graded to cross laminated layers (Figure 12).

Although the internal structures present within bioclastic turbidites are clearly identical to Bouma divisions there still exists

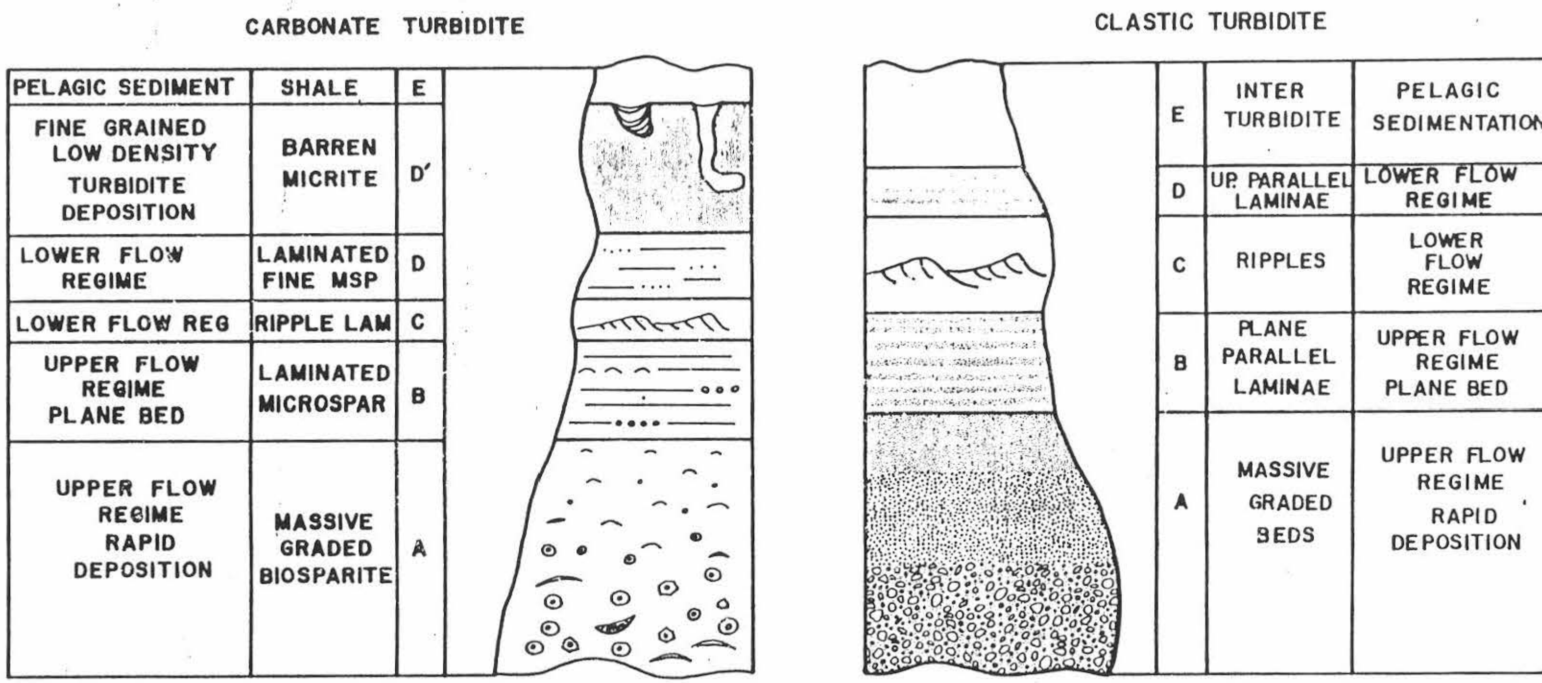

Figure 12. Comparison of the terminology and hydraulic interpretation of bioclastic (left) and clastic (right) turbidites. The primary difference lies in the addition of a structureless micrite (Td′) in the limestone.

the possibility of mis-identification of turbidites with tempestites. Table 1 shows how subtle some of these differences can be. One analytical technique which has proven to be extremely useful in identifying turbidites has been employment of a simple statistical analysis of the relationship between grain size and bed thickness. Crevello and Schlager (1980) and Kelling and Mullin (1975) employed similar strategies, which are based on the hypothesis that a relationship should exist between the grain size of the sand (in this case skeletal sand) material comprising the turbidite and the size of the turbidity current. Since the storm resuspension of sediment and subsequent settling is the process identified for producing graded tempestites, there is no reason why these types of deposits would exhibit this relationship. A linear regression analysis of skeletal grain size versus the thickness of the Ta layer shows a high degree of correlation (Figure 13).

A second valuable indicator of turbidite origin of a graded or cross laminated limestone bed is the absence of infiltration structures in the uppermost micrite layer (Td'). Kreisa (1981) describes the internal structures produced by storm resuspension and settling, including the late stages of fallout of the finest-grained carbonate mud fraction. The mud settles out and infiltrates around the underlying skeletal grains, which, in addition to producing many shelter fabrics, produces micrite drapes around grains. This type of fabric is not present within turbidites as the sedimentation of micrite is part of the continuum of deposition of coarser through finer grain sizes.

Finally, the presence of escape burrows are highly diagnostic of tempestites. Organisms picked up and deposited during resuspension and settling quickly burrow to the surface, producing traces which rise from the base to top of the bed. Turbidites, on the other hand, scour across the surface as they travel downslope, incorporating skeletal material into the traction carpet. Upon coming to rest the newly-formed bedding plane surfaces are recolonized and burrows descend from this upper surface into the layers below . No escape burrows have been described from modern bioclastic turbidites, so apparently organisms which become incorporated into turbidite flows do not survive transport.

Is the distinction between bioclastic turbidites and tempestites worth making? Certainly there are bathymetric associations for the two types of resedimented limestones. Although bioclastic

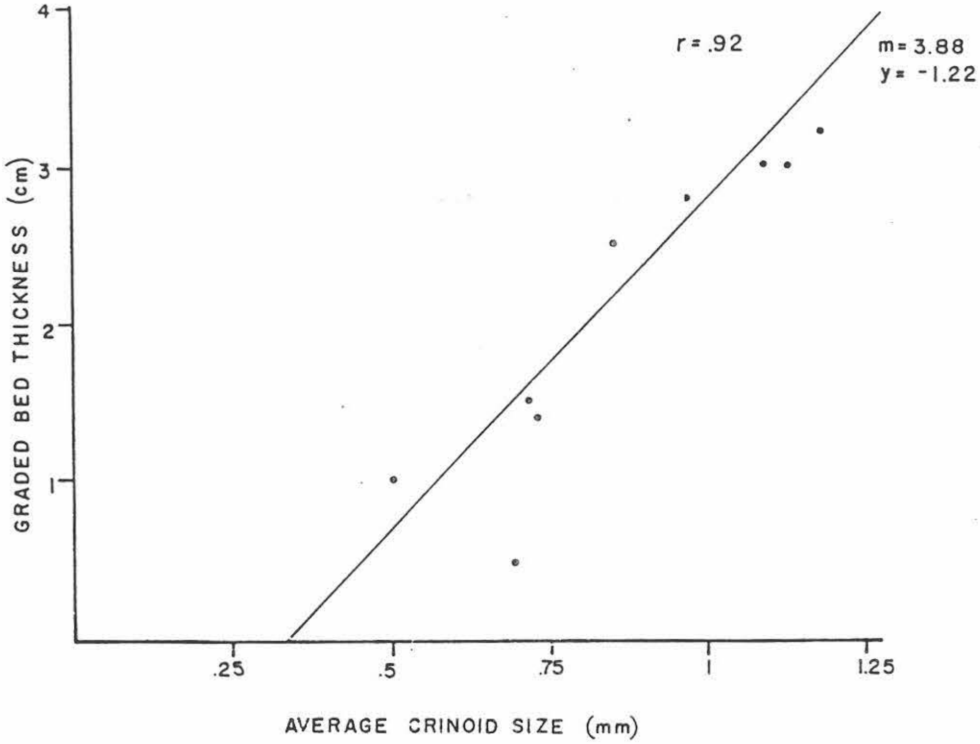

Figure 13. In order to demonstrate the turbidite origin of the graded limestones the diameter of crinoid ossicles were measured and plotted versus the thickness of the graded beds in which they are found. The best fit line and correlation coefficient (=0.92) produced by a linear regression analysis indicates that their is a relationship between these two variables, a documented property of turbidites.

turbidites need not form in deep water, their preservation is greatly enhanced at depths below wave base and in the absence of concentrated numbers of burrowing organisms. Tempestites, on the other hand, are often associated with depths near wave base (Kriesa, 1981) and therefore record a specific range of bathymetries between fairweather and storm conditions. Thus, paleoenvironmental reconstructions would depend on the accurate identification of turbidites versus tempestites. As Hamblin and Walker (1979) demonstrated so successfully in a siliciclastic sequence, it is possible to identify the transition from the storm-dominated to turbidite-dominated portions of a shelf. If the turbidites accumulated in a slope apron setting, the "traditional" facies associations of deep sea submarine fan turbidites would be lacking, thus making the distinction between turbidites and tempestites even more difficult.

Depositional Model for the Denley Limestone

Any model for the depositional setting of the Denley Limestone must take into account several different lines of evidence: (1) The stratigraphic position of the unit between the underlying shallow water Kings Falls and Sugar River Limestones and overlying black shales of the Utica and Canajoharie Shales. This evidence suggests that the unit is transitional in its bathymetry from shallow ramp to deep water basin. (2) The depo-tectonic setting of the Denley Limestone proposed by Cisne and his coworkers places the unit in a fragmenting foreland basin flanked to the east by a foredeep and to the west by a shallow water ramp. (3) The Denley Limestone consists of four lithologies of which bioclastic turbidites and interbedded shale are the most abundant.

Turbidites can accumulate in two very different settings, one associated with submarine fans and the other with slope aprons. Submarine fan deposits would contain evidence of channelized and interchannel flow, lobate distribution patterns from point sources, and coarsening-upward sequences, none of which are found in the Denley Limestone. Slope aprons, on the other hand, can form on slopes of low angle along a line source, exhibit great lateral continuity of beds, contain few breccia or slump horizons, and lack evidence of channelization. For these reasons, the sediment of the Denley Limestone is interpreted as having formed in a slope apron setting on the distal portion of a fault-dissected, eastward dipping ramp.

Characteristics of Tempestites and Turbidites

Tempestites

1. internal structures, although similar to Bouma divisions lack the repetitive consistency[1,2]
2. association with Skolithos burrows[2]
3. lateral continuity of beds on the order of 10's meters[2]
4. strongly lenticular bedding[1]
5. multiple paleocurrents: fairweather and storm[1,3,4]
6. association with wave-generated bedforms[5,4]
7. association with hummock cross stratification[6,3,4]
8. lateral facies grading into normal marine lithoa and biofacies[2]
9. escape burrows[7]
10. amalgamated beds[4]

Turbidites

1. planar bases, often with scours[8,9]
2. ubiquitous grading and horizontal laminations[8,10,11,12,13,14]
3. multiple internal Bouma divisions[8]
4. burrowing restricted to the upper few cm[11,12,13,14]
5. load casts and convolute laminations[11]
6. association with slump fold horizons and allochthonous blocks[9]
7. resedimented shallow water allochems[9]
8. oriented fossils[14]
9. lateral continuity of beds over 100's meters[11]
10. interbedding with pelagig sediments and faunas[12,13]

Distinguishing Features of Turbidites

1. absence of escape burrows
2. consistency of internal structures
3. association of couplets with deep water facies
4. extreme lateral continuity of beds
5. multiple internal Bouma divisions
6. unidirectional paleocurrents
7. association with slump folds

Sources:
1-Kelling and Mullin, 1975
2-Benchley, Newell and Stanistreet, 1979
3-Seilacher, 1982
4-Aigner, 1982
5-Nelson, 1982
6-Hamblin and Walker, 1979
7-Reineck and Singh, 1972
8-Tucker, 1969
9-Piper, 1972b
10-Piper, 1972a 1972
11-Scholle, 1971
12-Rupke, 1975
13-Hesse, 1975
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LATE WISCONSINAN GLACIAL LAKES OF THE WESTERN MOHAWK VALLEY REGION OF CENTRAL NEW YORK

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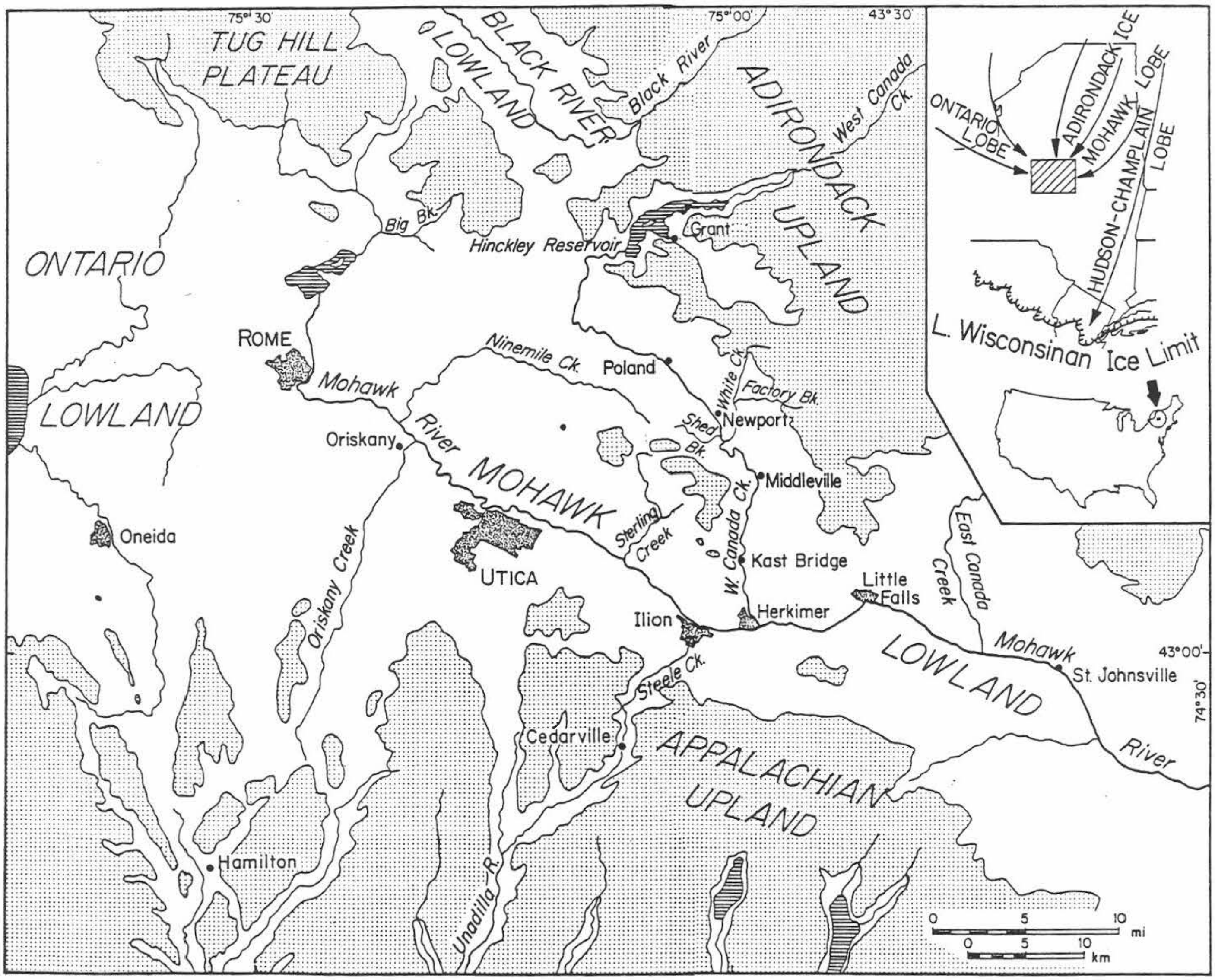


FIGURE 1. Location map of the western Mohawk Valley region of central New York. Patterned areas indicate uplands above an elevation of 400 m.

INTRODUCTION

The glacial history of the western Mohawk Valley region (Fig. 1) is critical to evaluating the contemporaneity of glacial events in the Great Lakes region, the Hudson-Champlain Valley, and New England. Understanding the composition and stratigraphy of glaciolacustrine deposits is particularly important to defining the positions and advances of impounding ice lobes, the pattern of ice recession in the Ontario Basin, Mohawk Valley, and Adirondacks, and the timing of the eastward release of water from the eastern Great Lakes region during Late Wisconsinan time.

Until recently, considerable controversy existed over the character of ice recession and the existence of regional glacial lakes and their levels in the Mohawk Valley. Several early workers recognized the widespread evidence for eastward (Ontario Lobe) and westward (Mohawk Lobe) ice flow in the Mohawk Valley (Dana, 1863; Chamberlin, 1883, 1888; Brigham, 1898, 1908; Cushing, 1905; Miller, 1909). A lack of deposits that clearly define ice-marginal positions in the Mohawk Valley led Cushing (1905), Brigham (1911, 1929), and Fairchild (1912) to conclude that ice in the lowlands receded by stagnation and downwasting, rather than by backwasting of an active ice margin. Fairchild's (1912) reconstructions depict lowland ice lobes that surrounded an ice-free Adirondacks during deglaciation. Fairchild inferred that deep glacial lakes, which drained southward across the Appalachian Upland, were impounded between the Mohawk and Ontario Lobes (Fig. 1) during deglaciation. Brigham (1911, 1929) opposed this idea based on a scarcity of lacustrine deposits above an elevation of 200 m. Recent stratigraphic studies (Fullerton, 1971; Krall 1977; Franzi, 1984; Ridge, 1985; Muller and others, 1986; Ridge and others, 1990; 1991) have provided evidence that deep, regional proglacial lakes formed at the margins of active lowland ice lobes at elevations up to 475 m. The lowland lobes formed prior to the deglaciation of most of the central Adirondack highlands. Regional deglaciation occurred primarily by calving at deep water, active ice margins, but was interrupted by several episodes of lowland ice readvance.

The purpose of this excursion will be to examine the evidence and history of glacial lakes in the western Mohawk Valley region. The lithostratigraphic model for the West Canada Creek valley, the application of paleomagnetism, and the genetic interpretation of sediments as related to glacial dynamics will be discussed.

METHODS AND ANALYSIS

A brief review of the fundamental concepts and nomenclature that provide the basis for interpretations of glaciolacustrine history is necessary before discussing the glaciation of the western Mohawk Valley. Essentially, most interpretations are based on detailed lithostratigraphic analysis of glacial sediments in the West Canada Valley and morphologic analysis of deltaic landforms and spillways. Perhaps the most important recent development is the formulation of regional time-stratigraphic correlations based upon radiocarbon-dated paleomagnetic stratigraphy.

Lithostratigraphy

Because of its completeness and lateral continuity the lithostratigraphic section of the West Canada Valley (Fig. 2) has been the foundation for the interpretation of the glacial history of the western Mohawk Valley (Franzi, 1984; Ridge, 1985; Muller and others, 1986; Ridge and others, 1990, 1991). The definitions and nomenclature of stratigraphic units in the West Canada Valley are non-genetic to avoid the confusion, misinterpretation, and ambiguity associated with genetic terms. Continuous packages of sediment bounded by discontinuities are defined as formations that can be subdivided into members. Members include stratified units, signified as "beds" that range from clay and silt to bouldery

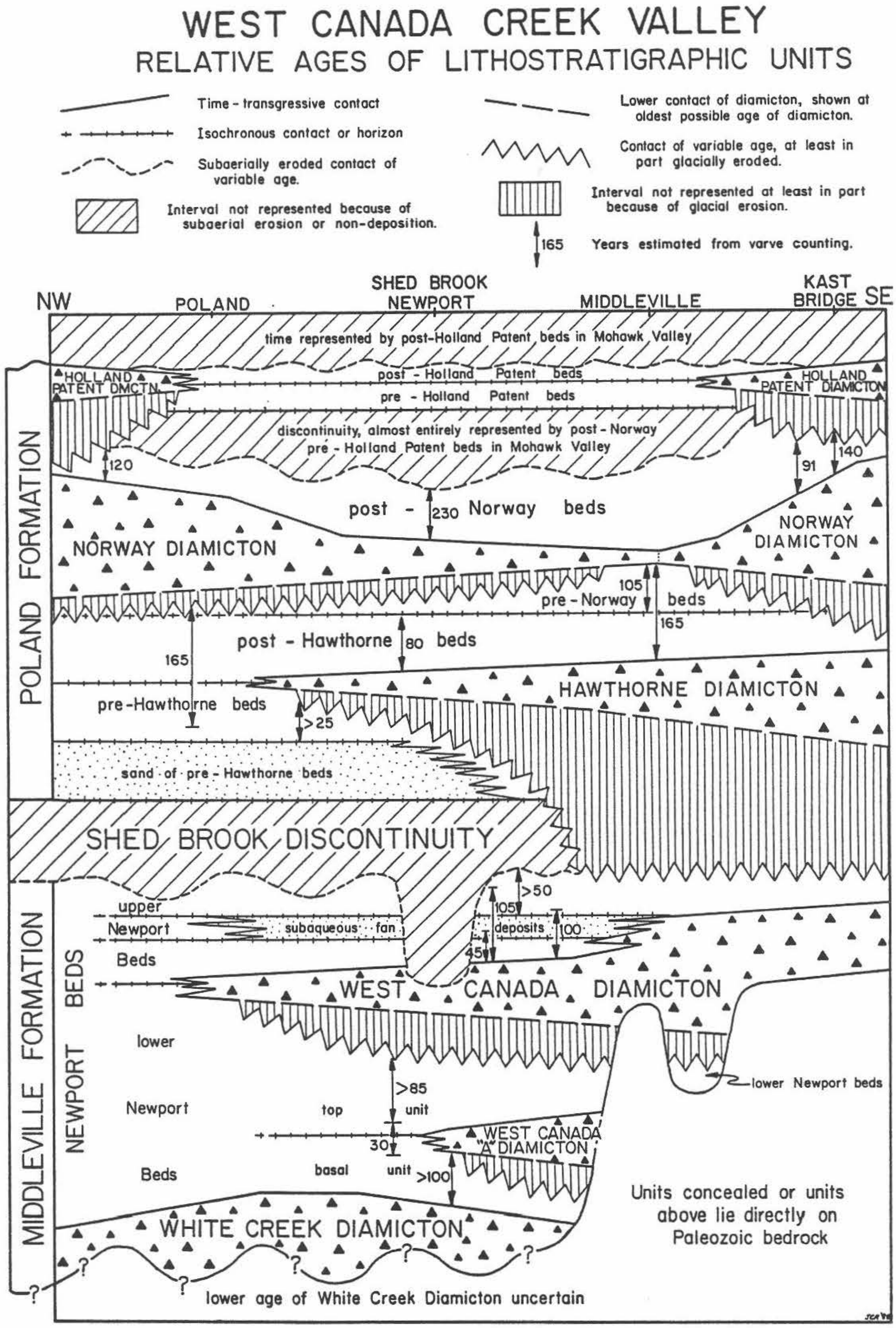


FIGURE 2. Summary of late Wisconsinan lithostratigraphy along the axis of the West Canada Valley. Lithostratigraphic units are plotted versus relative time on the vertical axis and position in the West Canada Valley on the horizontal axis to show the time-transgressive properties of some of the units. Place names are given on Figure 1.

gravel, and diamicton units. The non-genetic term "diamicton", following the usage of Frakes (1978) as a nonlithified, poorly sorted mixture of clay to boulders, has been chosen, instead of "till" because it more accurately describes units which are composed of till and poorly sorted sediment of other origins.

Provenance and Ice Flow Direction

The source and flow direction of ice lobes associated with individual diamicton units is recorded by striations on bedrock, subglacial grooves on the upper surfaces of till, elongate landforms, till fabrics, the sense of displacement on deformation structures in glacially overridden sediment, pebble provenance, and the color and composition of diamicton matrix material. Diamicton clast and matrix compositions are dominated by local rock types. In many places provenance differences are not distinct in the field because opposing ice lobes may have overridden the same rock types, or glacial readvances reworked older diamicton units. Detailed pebble counts and geochemical data, combined with ice flow indicators, provide the most accurate and consistent identification of ice lobe source for all the diamicton units (Fig. 2; Franzi, 1984; Ridge, 1985; Ridge and others, 1991). In general, Ontario Lobe diamictons contain the highest percentage of clastic sedimentary rocks and may have a matrix with a tannish gray to red color reflecting a western source. Mohawk Lobe diamictons are generally very dark gray to black and are dominated by black calcareous shale and dark gray limestone pebbles. Adirondack ice deposited sandy diamictons with high percentages of pebbles composed of metamorphic rock types that become more abundant as one approaches the Adirondacks. Locally derived materials may complicate the general compositional trends outlined above. For instance, in the West Canada Valley, Ontario Lobe ice transported non-calcareous shale pebbles from the summits of the Deerfield Hills to positions to the east while the Mohawk Lobe deposited diamictons with little non-calcareous shale.

It has been possible to distinguish the sources of fine-grained, varved lacustrine sediment based on field observations of color, grain-size, and the composition of ice-rafted debris. Ice-rafted debris includes pebbles and pellets (Ovenshine, 1970) that melted out of debris-laden icebergs. The iceberg debris probably originates as basal debris bands or subglacial sediment adhered to basal ice after calving. Lacustrine beds associated with the Ontario Lobe are light to medium gray in color with subtle pink to red hues and contain both gray and red ice-rafted pellets. Mohawk Lobe beds are generally dark to medium gray in color and contain only gray pellets. Varved sediment of an Adirondack provenance contain very thin bluish to greenish gray clay laminae with coarser beds that contain a high percentage of light gray fine sand.

Paleomagnetic Stratigraphy

A paleomagnetic record has been formulated for the Late Wisconsinan lithostratigraphy of the western Mohawk Valley from the detrital remanent magnetization of clayey and silty lacustrine beds that record the declination of the geomagnetic field at the time of deposition (Fig. 3; Ridge, 1985; Ridge and others, 1990). The secular variation record of remanent declination can be used to characterize intervals of time represented by different stratigraphic units. Paleomagnetism allows a time-dependent test of stratigraphic correlations of laminated lacustrine clay and silt at different outcrops based upon declination values and systematic stratigraphic changes in declination. Paleomagnetic declinations are unique in some stratigraphic units and provide strong supporting evidence for their identification. Examples of units with unique remanent declinations are the upper Newport Beds, which have a 20-30° East declination, and the upper part of the post-Holland Patent beds which have a 20-30° West declination. A gap in the declination record (Fig. 3) between the upper Newport Beds (20° East)

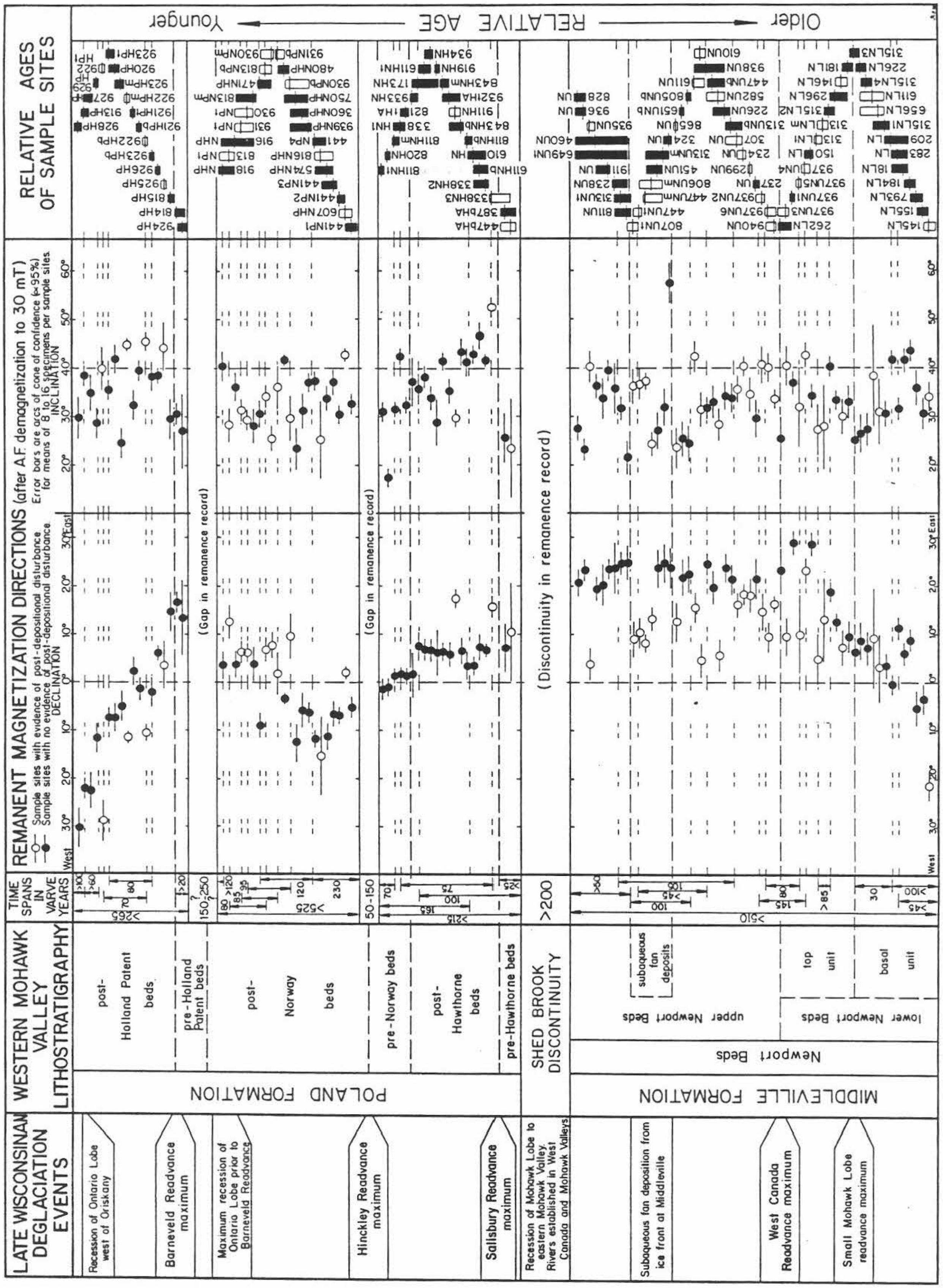


FIGURE 3. Plot of mean remanent declination and inclination for sampling sites in the western Mohawk Valley from Ridge and others (1990). Precise locations of sampling sites are given in Ridge (1985). The possible overlap of relative ages of sampling sites are shown with a bar graph on the right side of the diagram.

and the overlying pre-Hawthorne beds (8° East) provides additional evidence for an unconformity at the Shed Brook Discontinuity (Fig. 2). Remanent inclination is not a useful correlation tool because it tends to be inconsistent and underestimates the inclination of the geomagnetic field at the time of deposition (Fig. 2; Ridge and others, 1990). The apparent flattening of inclination is probably due to depositional or post-depositional processes in different sediment types.

Chronologic Inferences

Paleomagnetic records have been formulated from glaciolacustrine deposits from central New York (Brennan and others, 1984; Ridge, 1985; Ridge and others, 1990) and glacial Lake Hitchcock in the Connecticut Valley of western New England (Johnson and others, 1948; Verosub, 1979; Ridge, unpublished data; Fig. 4). The varve sediments of glacial lakes Hitchcock, Merrimack, and Ashuelot in New England, and Lake Albany in the Hudson Valley were studied by Antevs (1922) in his formulation of the New England varve chronology (Fig. 4). The paleomagnetic records permit a crude regional correlation of glacial events, but more importantly, radiocarbon calibration of the New England varve chronology (Ridge and Larsen, 1990) provides linkage to a numerical time scale. Radiocarbon-calibrated paleomagnetic records that are tied to the New England varve chronology may provide an important new tool that can be applied to the Late Wisconsinan lithostratigraphy and events of central New York. For example, the only existing radiocarbon dates from the New England varve chronology of about 12.4 ka (Varve yr 6150) are from sediments that have a unique greater than 30° West declination (Fig. 4). These sediments correlate paleomagnetically with pre-Lake Iroquois sediment in the eastern Ontario Basin (Brennan and others, 1984) that has a radiocarbon age of 12.8-12.3 ka (Fullerton, 1980; Muller and Prest, 1985). At present, this is the only independent test of the paleomagnetic correlation between central New York and New England and thus should be considered preliminary.

Glacial Lake Reconstruction and Isostatic Rebound

Two major obstacles have historically prevented accurate reconstructions of glacial lakes in the western Mohawk Valley region. First, glacial lakes associated with older events in the valley are not usually represented by landforms which can be used to delineate ancient strandlines. Strandline features may have been obliterated by later glacial readvances or the glacial lakes were simply too deep in the central part of the Mohawk Valley. The minimum levels for older lakes are inferred from the maximum elevations of buried lake floor deposits and subaqueous sand and gravel deposits.

A second problem has been the uncertainty in the positions of spillways for high level lakes because of uncertainties in the direction and magnitude of isostatic tilting in the region. The degree of tilting in the region is at least 0.8 m/km (4.3 ft/mile) based on the areal distribution and elevations of deltas associated with glacial Lake Miller in the upper West Canada Valley. Reconstructed glacial lake planes in the eastern Ontario Basin generally slope southwest with isobases (lines of equal uplift) oriented west-northwest to east-southeast (Fairchild, 1916, 1917; Pair, 1986; Pair and others, 1988). Isobases in New England trend east-northeast to west-southwest (Koteff and Larsen, 1989). By inference, isobases in the western Mohawk and Hudson valleys must be oriented between those in the Ontario Basin and New England. An inferred direction and magnitude for tilting of lake planes in the western Mohawk Valley region (south-southwest at 0.8-1.0 m/km; Fig. 5) is used to reconstruct the extent and elevation of former glacial lakes.

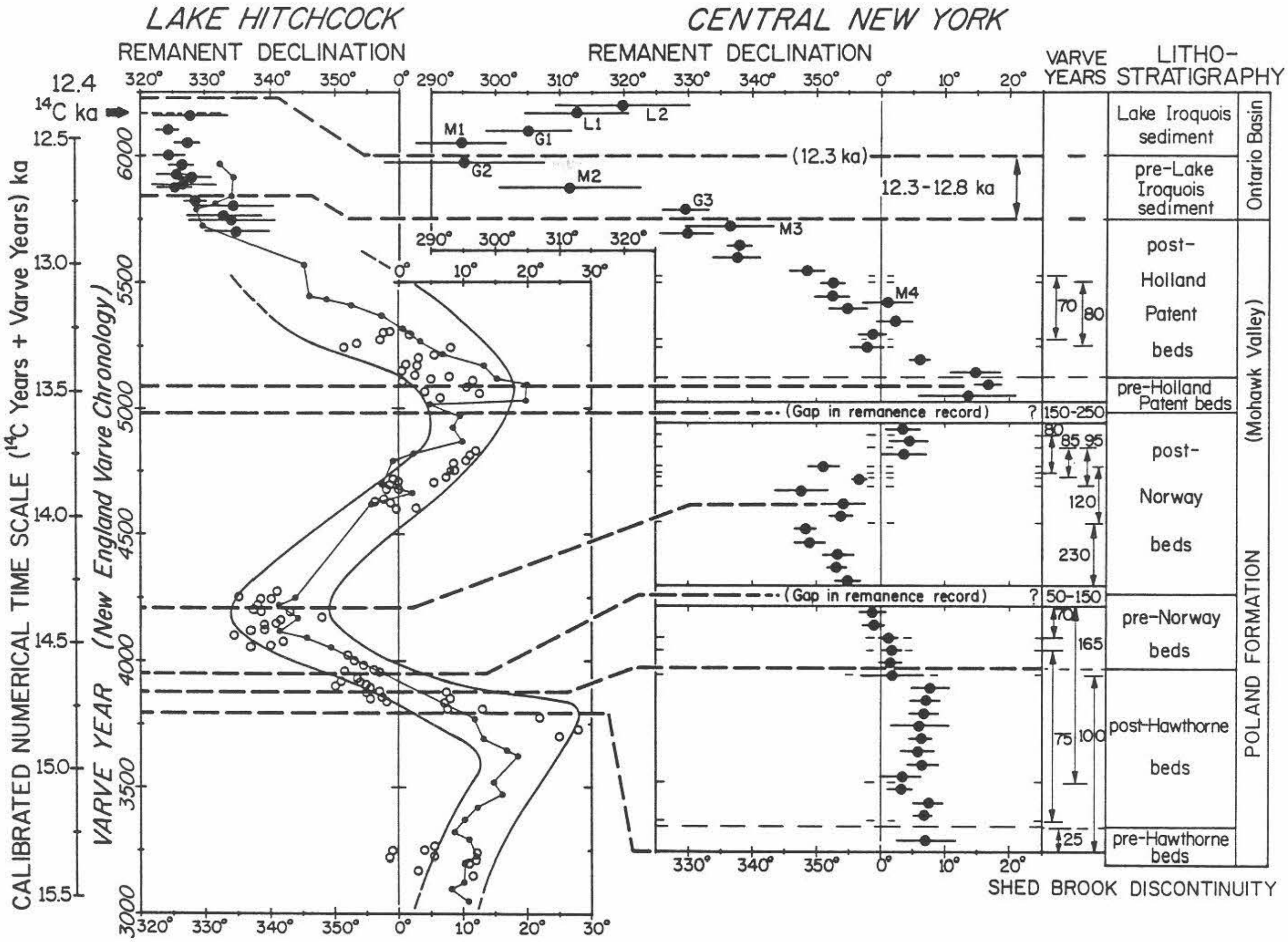


FIGURE 4. Correlation (heavy dashed tie lines) of remanent declination records from lacustrine deposits of Lake Hitchcock in western New England and central New York. Lake Hitchcock records are from sections matched with the New England varve chronology (varve yrs 3000-6250; Antevs, 1922). Radiocarbon calibration is based on radiocarbon ages of 12.4 ka from Canoe Brook, Vermont (Ridge and Larsen, 1990). The area between the two smoothed curves on the Lake Hitchcock record represents the region within which geomagnetic declination lies as interpreted by Verosub (1979) from natural remanences (Johnson and others, 1948; solid circles with tie line) and remanences (Verosub, 1979; open circles). Remanent declination means from sites in couplets 5700-6200 (new data from Canoe and Mill Brook sections in Vermont) are shown as large solid circles with alpha-95 confidence intervals. From central New York, unlabeled declination means and alpha-95 confidence intervals (Ridge and others, 1990) and labeled sites (Brennan and others, 1984) are plotted by relative age as indicated by position in superposed lithostratigraphic units or morphologic successions in the western Mohawk and West Canada Valleys (Figs. 2 and 3; Ridge and others, 1990) and the eastern Ontario Lowland (Fullerton, 1980). Numerical ages of pre-Iroquois and Lake Iroquois sediments are based on Fullerton (1980) and Muller and Prest(1985).

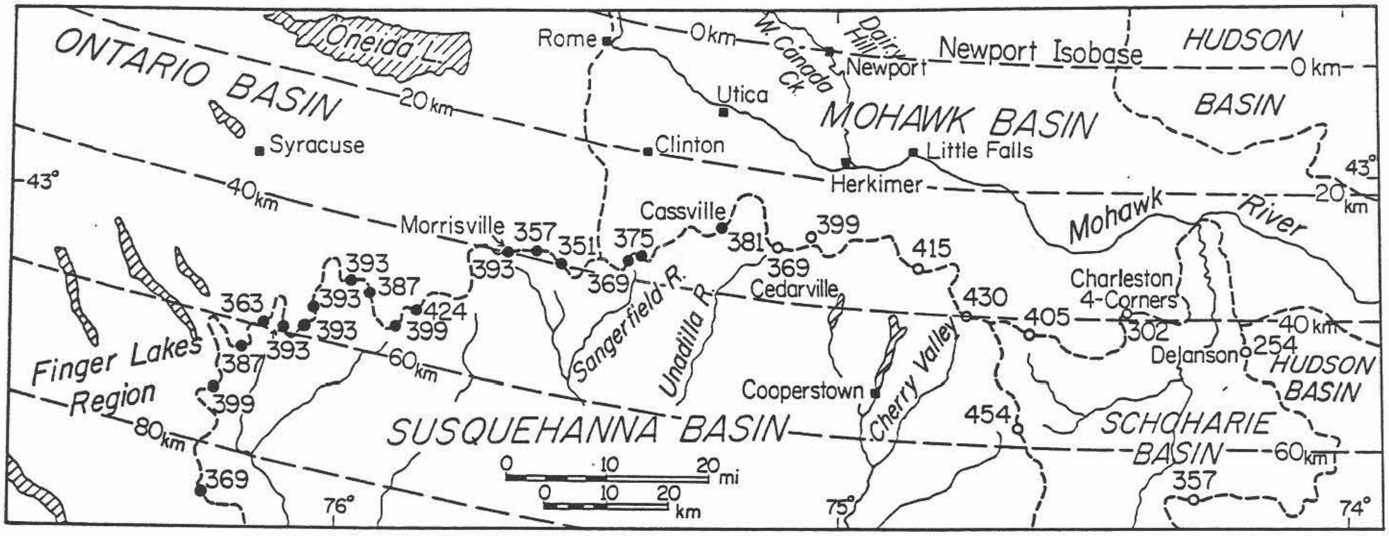


FIGURE 5. Potential lake-draining cols on the Ontario, Mohawk, and Schoharie basin boundaries. All elevations are in meters as converted from USGS 1:24,000-scale topo maps. Cols known to be filled with Valley Heads drift (Randall and others, 1988) are indicated with solid symbols. Inferred isobase trends are shown with isobases spaced 20 km apart, starting in the north with an isobase through Newport.

LATE WISCONSINAN GLACIATION AND GLACIAL LAKES

The Late Wisconsinan glaciation of the western Mohawk Valley region is mostly recorded by two sequences of glacial sediment, separated by an unconformity, that represent major periods of Mohawk and Ontario lobe oscillation. The older sequence is the Middleville Formation, dating from prior to 16 ka, which represents the pre-Valley Heads glaciation (Fig. 2; Ridge, 1985; Ridge and others, 1991). The younger sequence is the Poland Formation (Fig. 2) which represents Valley Heads glaciation (Mickelson and others, 1983) and dates from about 15-13 ka as inferred from paleomagnetic correlations with radiocarbon-dated varves in New England (see previous section: Chronologic Inferences).

Early Pre-Valley Heads Glaciation

The only known complete record of pre-Valley Heads glaciation in the western Mohawk Valley is the Middleville Formation in the West Canada Valley (Ridge, 1985; Ridge and others, 1991). The earliest records of pre-Valley Heads glaciation are southwest ice flow indicators and sandy till containing abundant metamorphic rock in the base of the White Creek Diamicton (Fig. 2). These features may represent the maximum Late Wisconsinan advance to or recession from the Terminal Moraine in Pennsylvania (Lewis, 1884; Crowl, 1980; Crowl and Sevon, 1980; Cotter and others, 1986). Later deposition of till in the White Creek Diamicton indicates a shift in ice flow direction to the south to south-southeast as local subglacial topography exerted a greater influence on ice flow (Fig. 6).

Development of Lake Newport and Mohawk Lobe Readvances

Initial ice recession in the West Canada Valley produced an opening in the retreating ice sheet which is recorded by lacustrine deposition of the lower Newport Beds in the beginning of Lake Newport (Figs. 2 and 6). Sedimentation in Lake Newport continued throughout pre-Valley Heads glaciation. For much of its existence the level of Lake Newport was too high, considering southward projection of isostatically tilted water planes, to have drained freely across channels on the

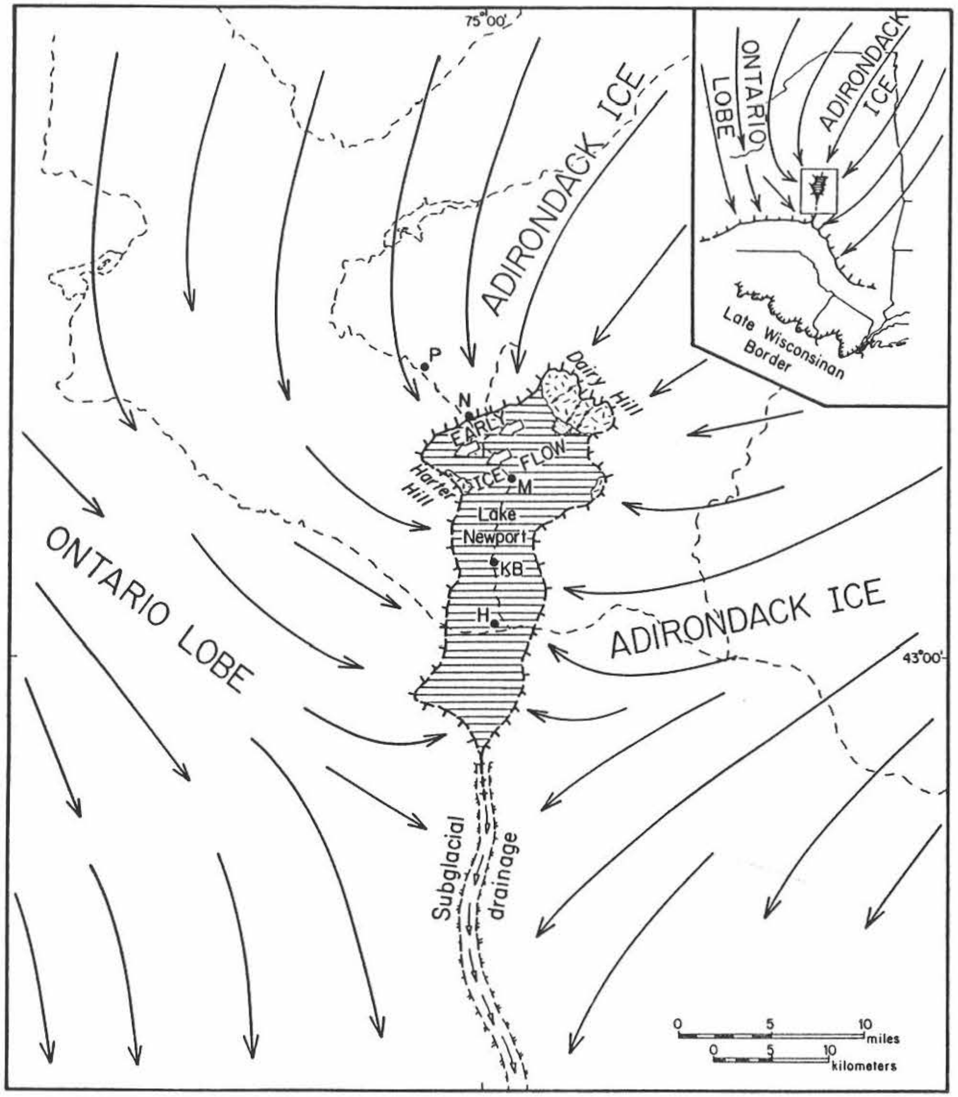


FIGURE 6. Early deglaciation during pre-Valley Heads time and the beginning of Lake Newport.

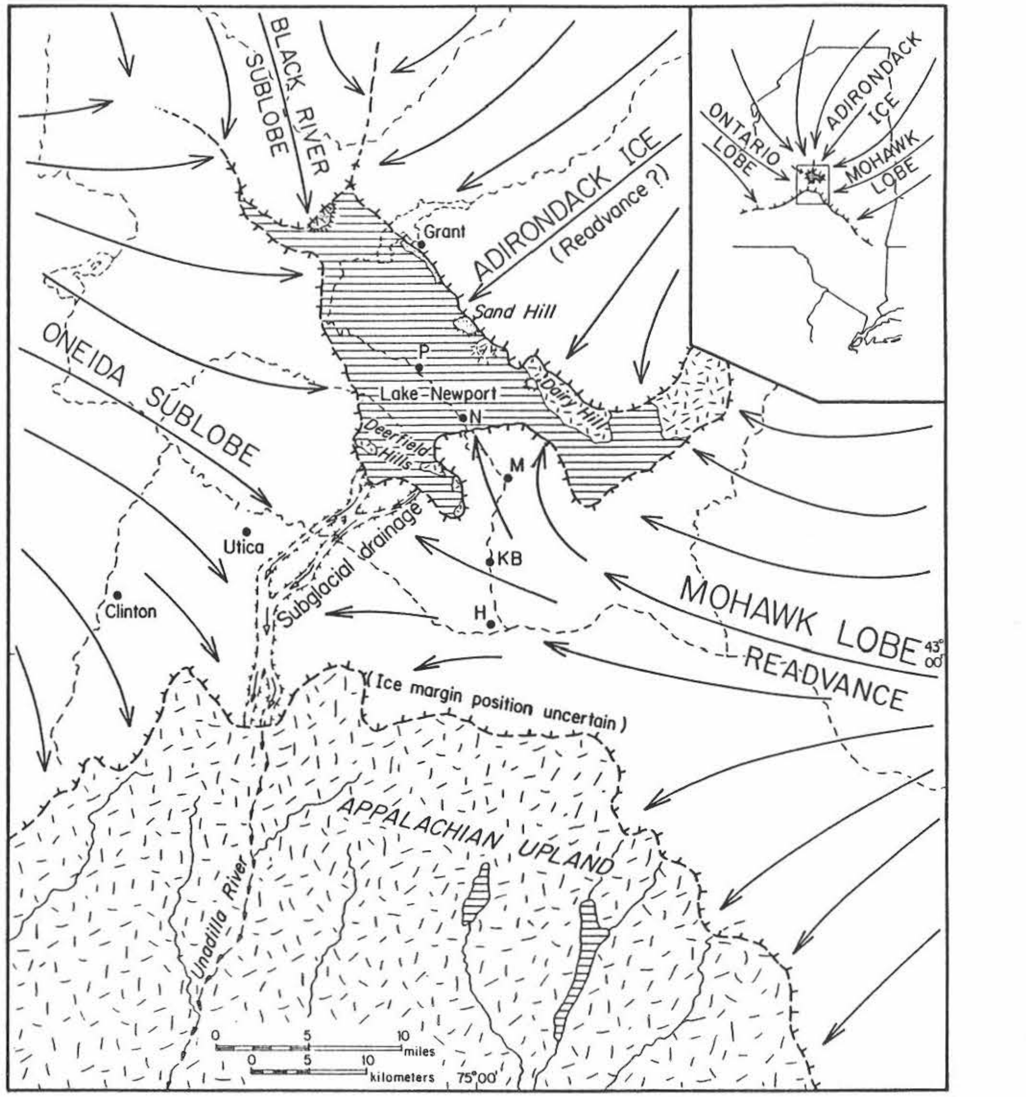


FIGURE 7. The first pre-Valley Heads readvance of the Mohawk Lobe.

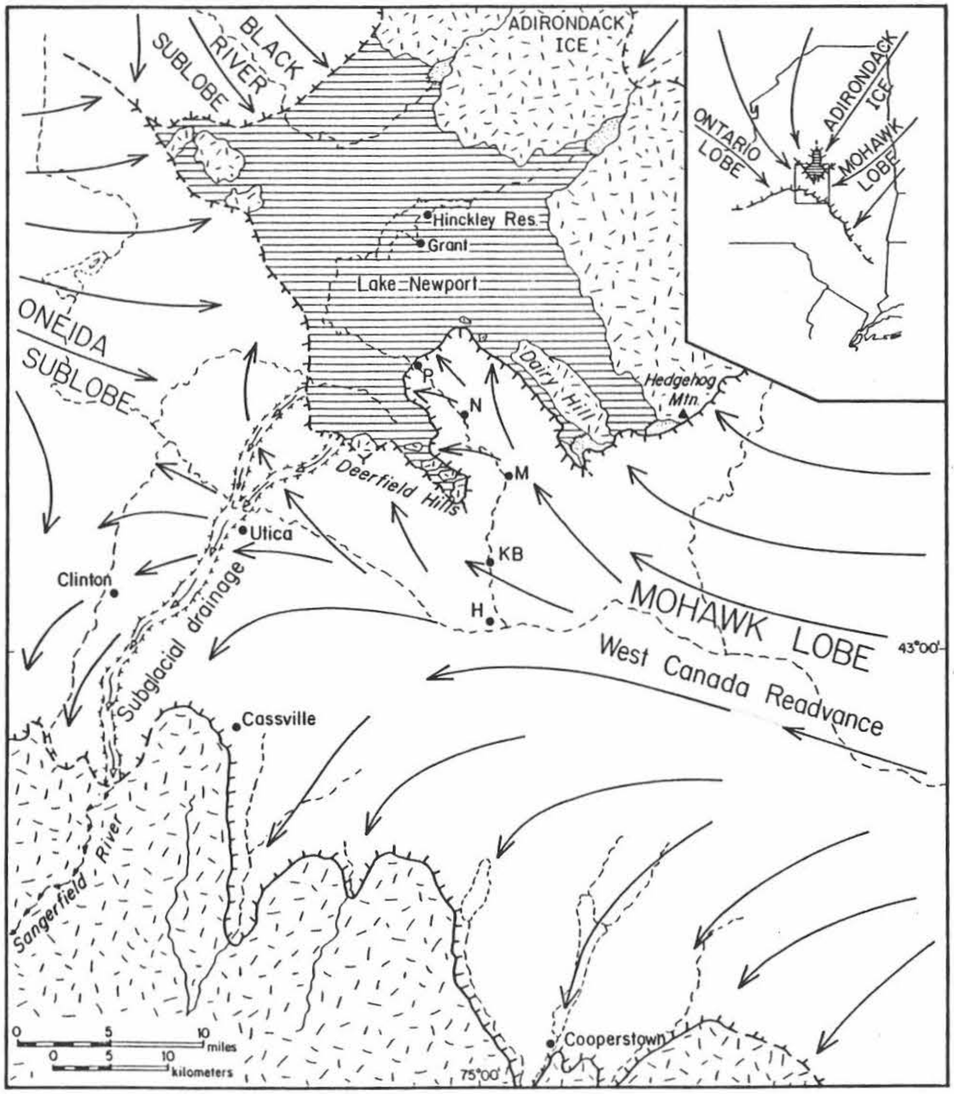


FIGURE 8. The West Canada Readvance.

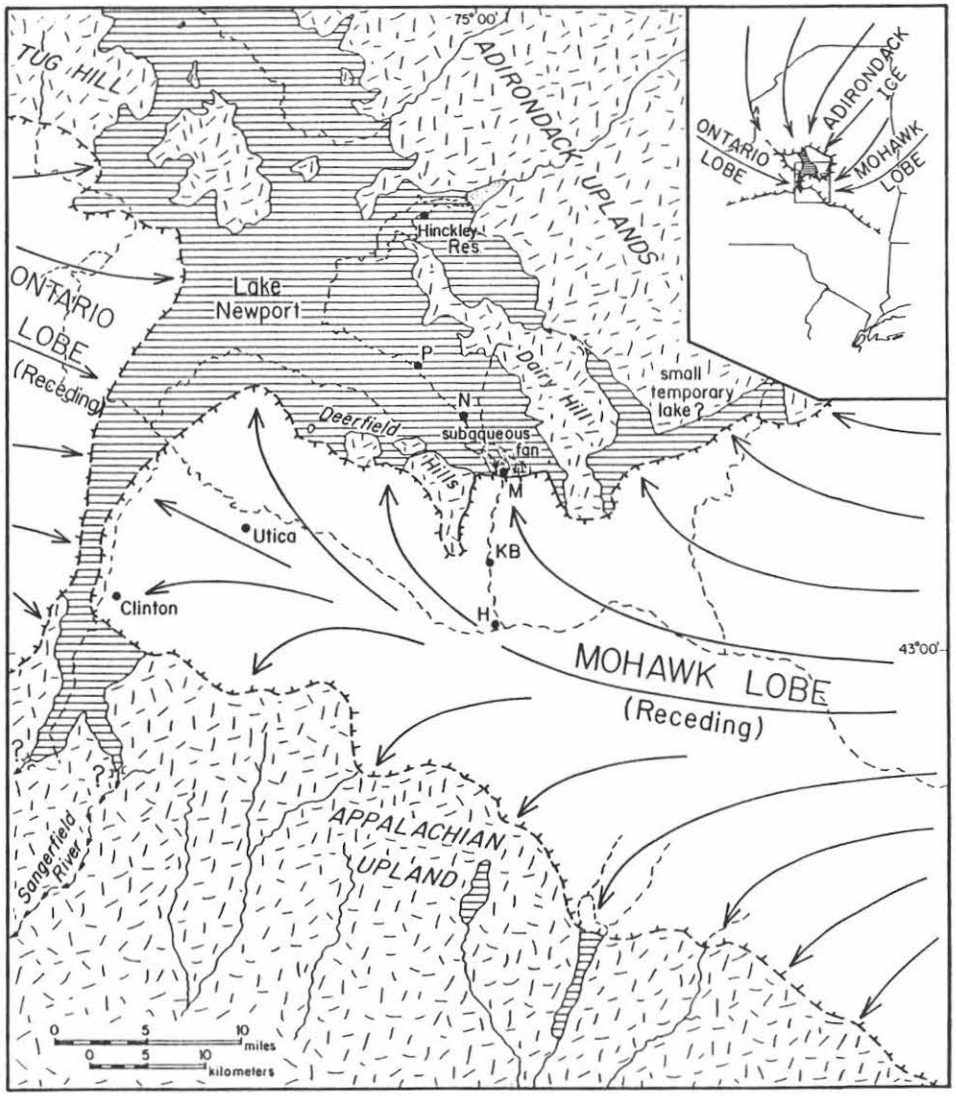


FIGURE 9. Final pre-Valley Heads recession of the Mohawk Lobe.

Appalachian Upland (Figs. 6-8; Ridge and others, 1991). A subglacial, and possibly seasonal or ephemeral, channel system may have drained Lake Newport throughout most of its history.

Two Mohawk Lobe readvances occurred in the West Canada Valley during the history of Lake Newport. The first readvance is thought to be a relatively minor oscillation of the Mohawk Lobe that deposited the West Canada 'A' Diamicton (Figs. 2 and 7; Ridge and others, 1991). This event may be synchronous with the development of a series of ice-marginal subaqueous gravel deposits and a sheet of sandy till that occur across the north flank of Dairy Hill and the upper West Canada Valley, and possibly represent a readvance of Adirondack ice. The later West Canada Readvance represents a major oscillation of the Mohawk Lobe which deposited the West Canada Diamicton. The readvance corresponds to a thick varve sequence which spans the lower and upper Newport Beds exposed in bluffs along the Hinckley Reservoir (Fig. 8; Fullerton, 1971; Franzi, 1984) and deposits that require water levels of at least 463 m (1520 ft) in the West Canada Valley (Fig. 8).

Mohawk Lobe Recession and the Demise of Lake Newport

Final pre-Valley Heads recession of the Mohawk Lobe in Lake Newport left the upper Newport Beds, a thick (up to 40 m) varve and turbidite sequence which is traceable throughout the West Canada Valley (Fig. 2; Ridge and others, 1991). Recession of the Mohawk Lobe triggered a sudden drop in the high levels of Lake Newport, to a lower level which drained freely across the Appalachian Upland (Fig. 9). An esker and subaqueous fan complex in the base of the upper Newport Beds at Middleville may represent this drainage event. Rapid drainage of meltwater from the base of the Mohawk Lobe, in response to dropping lake levels and steepening of subglacial hydraulic gradients, may account for the rapid deposition, great thickness, and lateral extent of the fan complex at Middleville.

Recession of the Mohawk Lobe to the eastern end of the Mohawk Valley eventually caused complete drainage of Lake Newport. This interval of ice recession, which may be an Erie Interstadial equivalent in central New York, is represented by a river gravel in the Mohawk Valley (Little Falls Gravel) and subaerial erosion of the upper Newport Beds in the West Canada Valley (Shed Brook Discontinuity; Figs. 2 and 10; Lykens, 1983; Ridge, 1991).

Early Valley Heads Glaciation and Mohawk Lobe Readvance

At the beginning of Valley Heads glaciation both the Mohawk and Ontario lobes began to advance, eastward river drainage in the Mohawk Valley became blocked by the Mohawk Lobe at the eastern end of the valley, and the Little Falls Gravel and Shed Brook Discontinuity were overlain by lacustrine sediment of the pre-Hawthorne beds (Figs. 2 and 10). Before the Salisbury Readvance of the Mohawk Lobe reached its maximum extent (Fig. 11), lake levels in the Mohawk Valley rose to an elevation of at least 215 m as marked by lacustrine sand in the base of the pre-Hawthorne beds near Poland in the West Canada Valley (Fig. 2).

Valley Heads Impoundment of Lake Cedarville

A sudden change from lacustrine sand to dark gray varves in the pre-Hawthorne beds at about the time the Salisbury Readvance reached its maximum position records a sudden increase in lake level (Figs. 2 and 10). This rise in lake level represents the formation of Lake Cedarville which drained across Cedarville col (369 m, 1210 ft) on the Appalachian Upland into the Unadilla River valley (Fig. 11). The Mohawk Lobe, which advanced into a deepening and widening water body in the Mohawk trough, may have experienced accelerated calving which retarded its advance. The level of Lake Cedarville is recorded at the Salisbury Readvance limit on the north side of the Mohawk Valley by an ice-contact delta (Fig. 11) with an elevation of

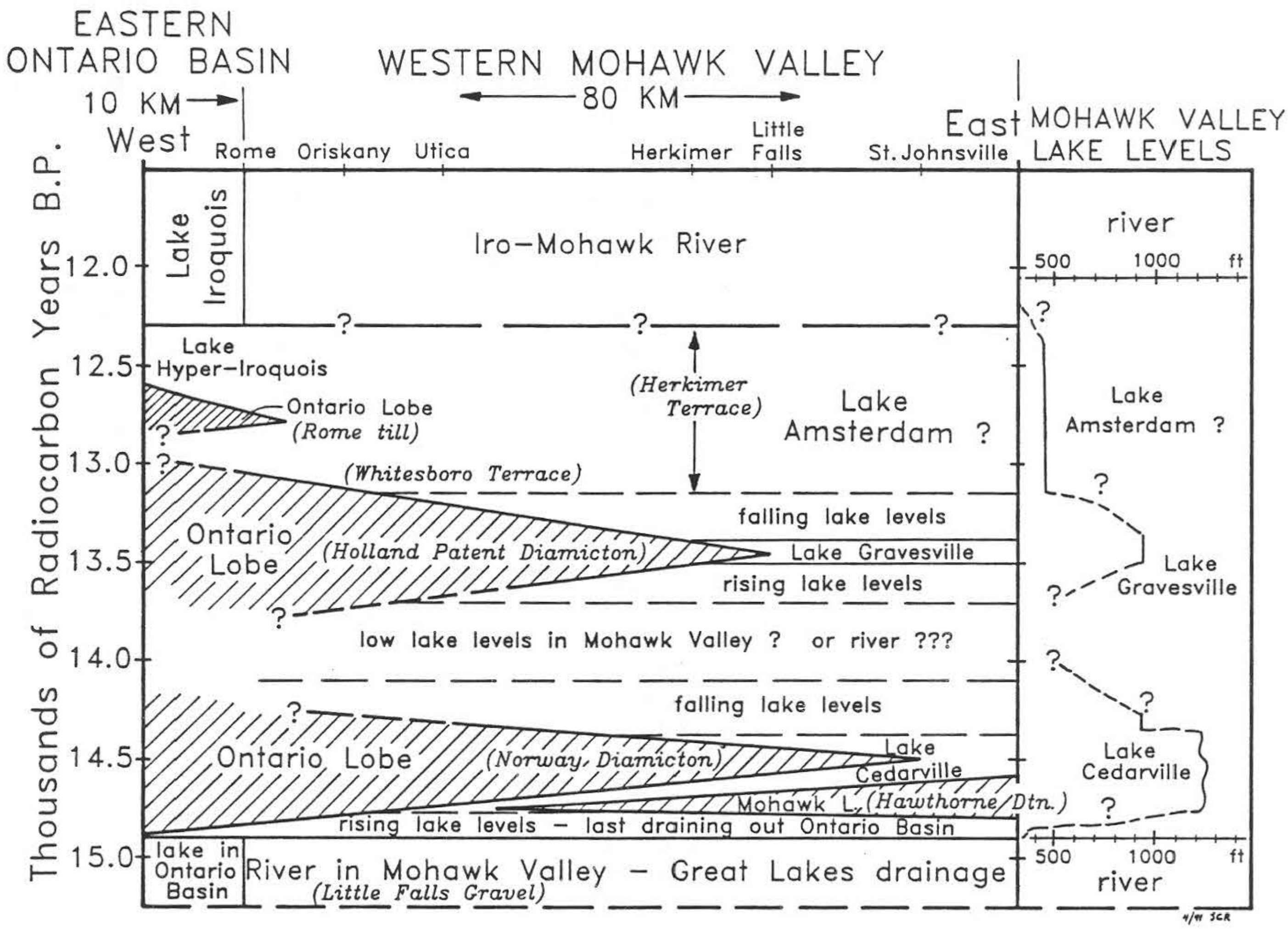


FIGURE 10. Time-distance plot of Valley Heads through post-Valley Heads glacial events in the western Mohawk Valley. Litho- and morphostratigraphic units are shown in italics.

393-396 m (1290-1300 ft). Isostatic tilting of about 0.9 m/km (4.8 ft/mile) accounts for delta elevations which are higher than the Cedarville spillway. Lake Cedarville was formed within 25 years of the Mohawk Lobe attaining its maximum position and during the last several tens of kilometers of advance the Mohawk Lobe did not block off any new outlets that could have caused the impoundment of Lake Cedarville. Therefore, the initial impoundment of Lake Cedarville appears to be the result of the closure of an outlet to the west by the advance of the Ontario Lobe. Outlets to the west (Fig. 5) are today filled with younger deposits of the Valley Heads moraines (Fairchild, 1932; Randall and others, 1988) and would have been lower at the time of the impoundment of Lake Cedarville.

The Maximum Ontario Lobe Readvance into Lake Cedarville

Within 200 years of the Salisbury Readvance the Mohawk Lobe was replaced by the Hinckley-St. Johnsville Readvance of the Ontario Lobe (Figs. 10 and 12; Ridge, 1985; Muller and others, 1986). Initial recession of the Mohawk Lobe allowed deposition of varves of Mohawk Lobe provenance in the post-Hawthorne beds which were overlain by varves of Ontario Lobe provenance in the pre-Norway beds (Fig. 2). The Norway Diamicton was deposited over the varve section as the Ontario Lobe advanced toward the maximum extent of the Hinckley-St. Johnsville Readvance (Fig. 10 and 12). The Ontario Lobe surrounded the Deerfield Hills, creating a nunatak, and Lake Miller

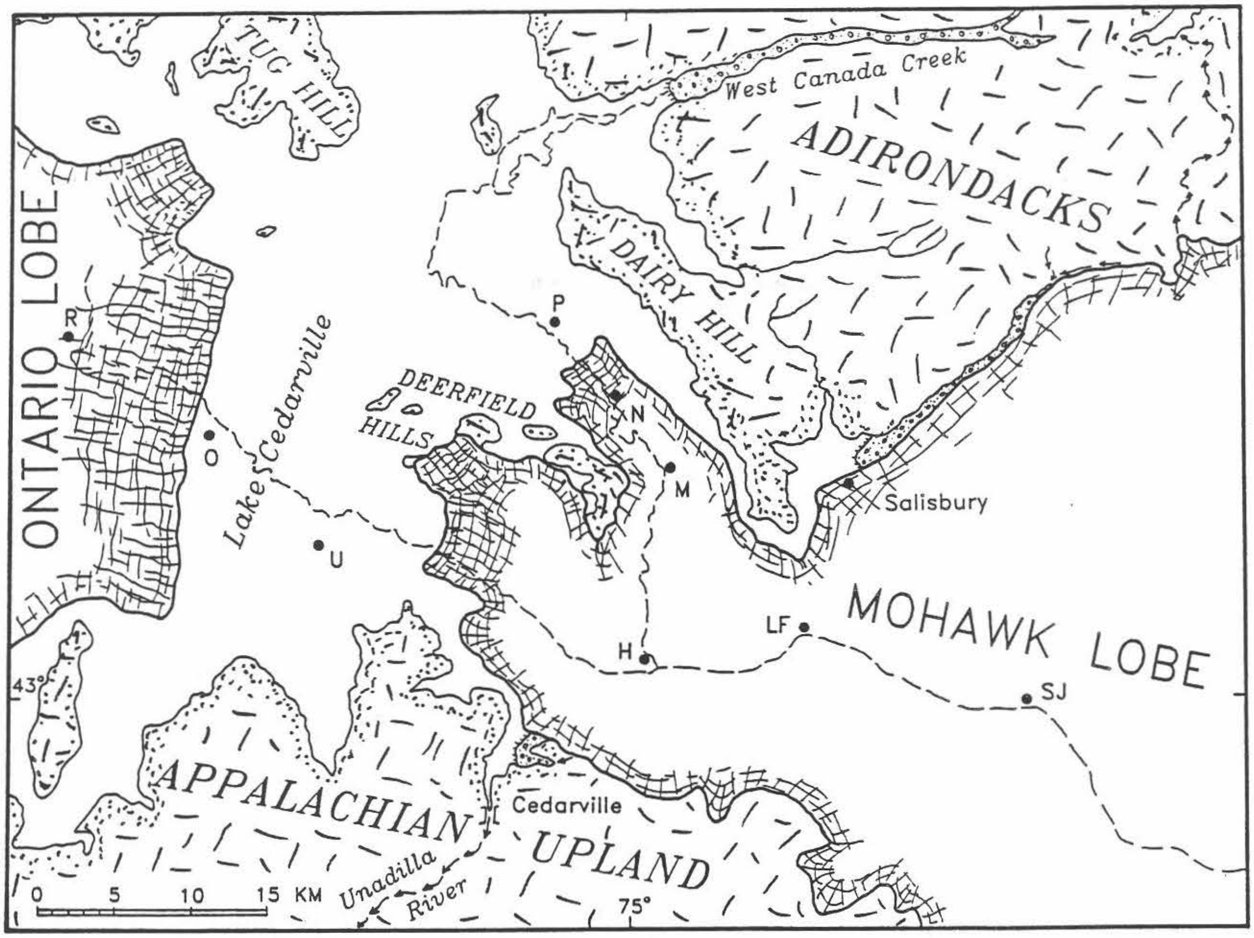
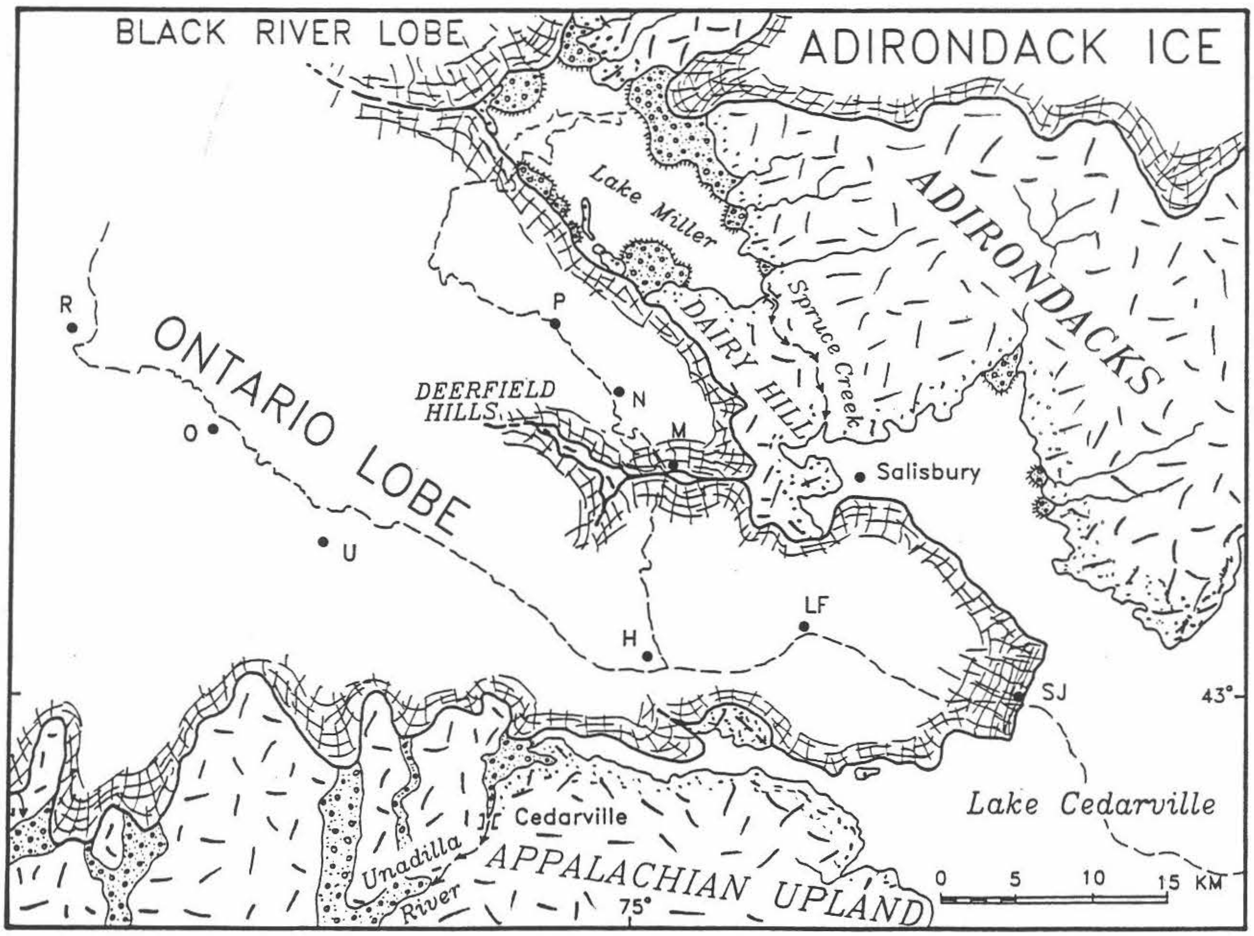


FIGURE 11. The Salisbury Readvance and early Lake Cedarville.

FIGURE 12. The Hinckley-St. Johnsville Readvance and Lake Cedarville.



(Fullerton, 1971; Franzi, 1984) was impounded in the upper West Canada Valley. The position of the Ontario Lobe along the Appalachian Upland is marked by an ice-contact delta near the entrance to the Cedarville channel. The advancing Ontario Lobe probably did not collide with the receding Mohawk Lobe because the Hawthorne and Norway diamictons are separated by lacustrine sediment of the post-Hawthorne and pre-Norway beds throughout the central Mohawk trough.

Recession of the Ontario Lobe is known to have occurred as far west as Oriskany where lacustrine deposits occur beneath the later Holland Patent Diamicton. Early during recession, Lake Miller drained into Lake Cedarville along the northern flank of the Ontario Lobe. Lake Cedarville eventually dropped to a succession of short-lived lake levels in the Mohawk Valley that probably drained across the divide between the Mohawk and Schoharie basins near Charleston Four Corners (302 m, 990 ft; Fig. 5). These events are recorded by large deltas in the upper West Canada Valley that represent Lake Prospect 347-365 m (1120-1200 ft; Franzi, 1984).

<u>The Barneveld-Little Falls Readvance and Glacial Lake Gravesville</u>

The second Valley Heads readvance of the Ontario Lobe, the Barneveld-Little Falls Readvance, deposited the Holland Patent Diamicton (Figs. 2, 10, and 13). This advance was matched by blockage of the eastern Mohawk Valley by the Mohawk Lobe and the creation of a valley-wide lake known as Lake Gravesville. The impoundment of Lake Gravesville appears to closely coincide with the halt of the Barneveld-Little Falls Readvance which may have temporarily stabilized on a bedrock high at Little Falls as water levels rose in front of the Ontario Lobe. Lake Gravesville is marked by the deposition of ice contact deltas along the northern flank of the Ontario Lobe and a delta at the mouth of East Canada Creek (292-310 m, 960-1020 ft; Fig. 13). After consideration of isostatic tilting in the region the most likely stable spillway for Lake Gravesville is Delanson channel at 254 m (835 ft; Fig. 5) which is located on the Schoharie-Hudson divide in eastern New York.

<u>Ontario Lobe Recession and Impoundment of Lake Amsterdam</u>

Lake Gravesville did not persist long after Ontario Lobe recession began because Gravesville deltas in the West Canada Valley are limited to ice-contact features and they do not occur along the southwest side of the valley from Barneveld to Poland where high discharge from the Adirondacks continued to deliver large volumes of sediment. Lake levels dropped as the Mohawk Lobe receded and lower outlets were opened in the southeastern Mohawk Valley, but the exact pattern of lake drainage is not known from evidence in the western Mohawk Valley. Water levels in the Mohawk Valley dropped to a minimum elevation of 171 m (560 ft) by the time the Ontario Lobe receded to just west of Herkimer (Figs. 10 and 14). High fluvial terraces in the West Canada Valley may have formed at this time.

Ontario Lobe recession continued to Oriskany (Figs. 10 and 15) where the Whitesboro Terrace represents the first deposition of ice-contact deposits graded to a water level lower than 152 m (500 ft) in the Mohawk Valley in what is probably the beginning of Lake Amsterdam. The Whitesboro Terrace at first glance appears to be an ice-contact delta, but several exposures in this feature show a cap of fluvial beds truncating lacustrine sand along a regionally tilted contact. The fluvial beds emanate from a kettle and esker complex in the Oriskany Valley and probably represent deposition on a surface trimmed by fluvial erosion in response to falling lake levels in the Mohawk Valley. The Whitesboro Terrace surface as well as inwash deltas further east in the Mohawk Valley may be graded to Lake Amsterdam which had an elevation of 139 m (457 ft) at Herkimer. Recession of the Ontario Lobe continued to Rome while Lake Amsterdam remained impounded in the Mohawk Valley.

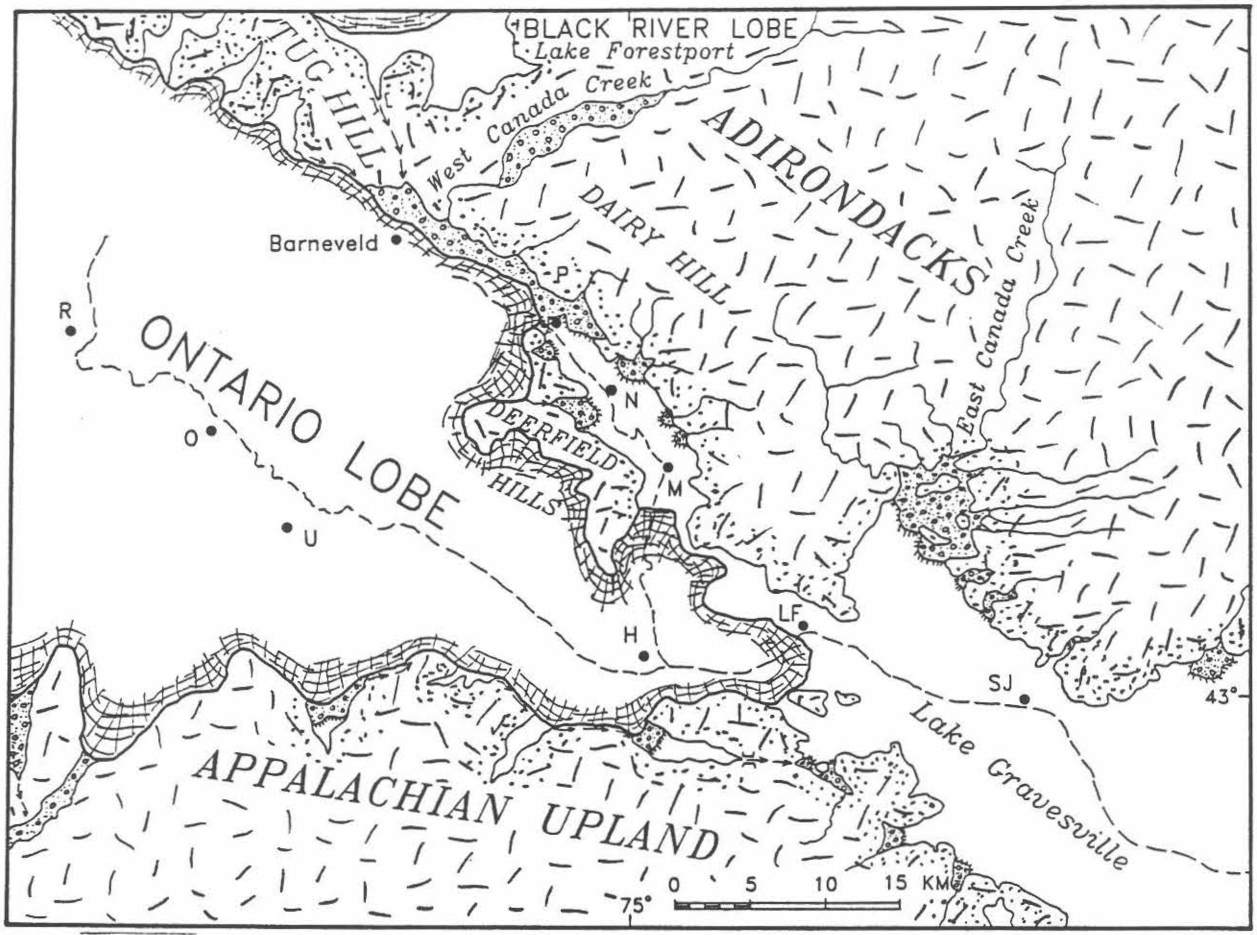
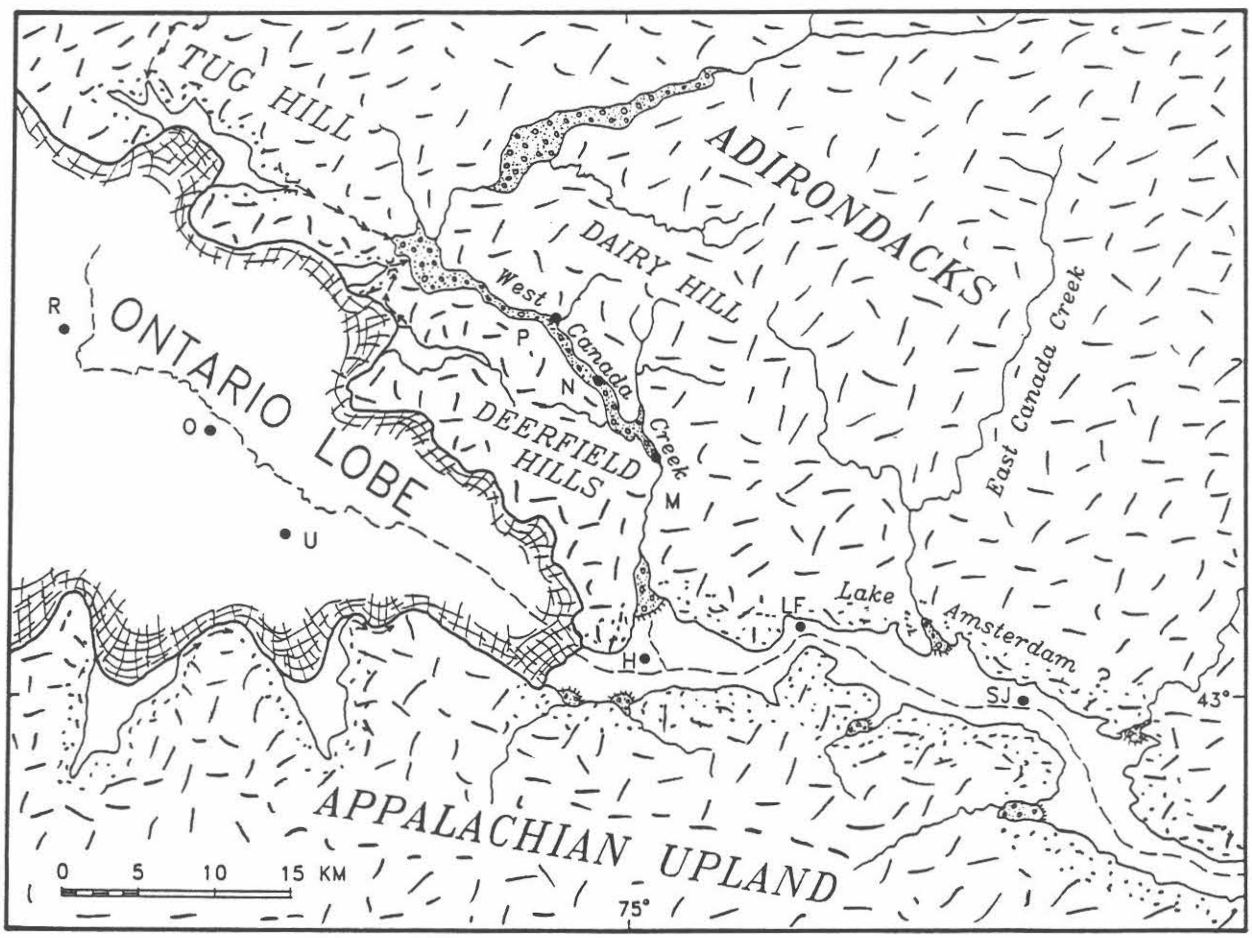


FIGURE 13. The Barneveld-Little Falls Readvance and Lake Gravesville.

FIGURE 14. The final recession of the Ontario Lobe west of Herkimer.



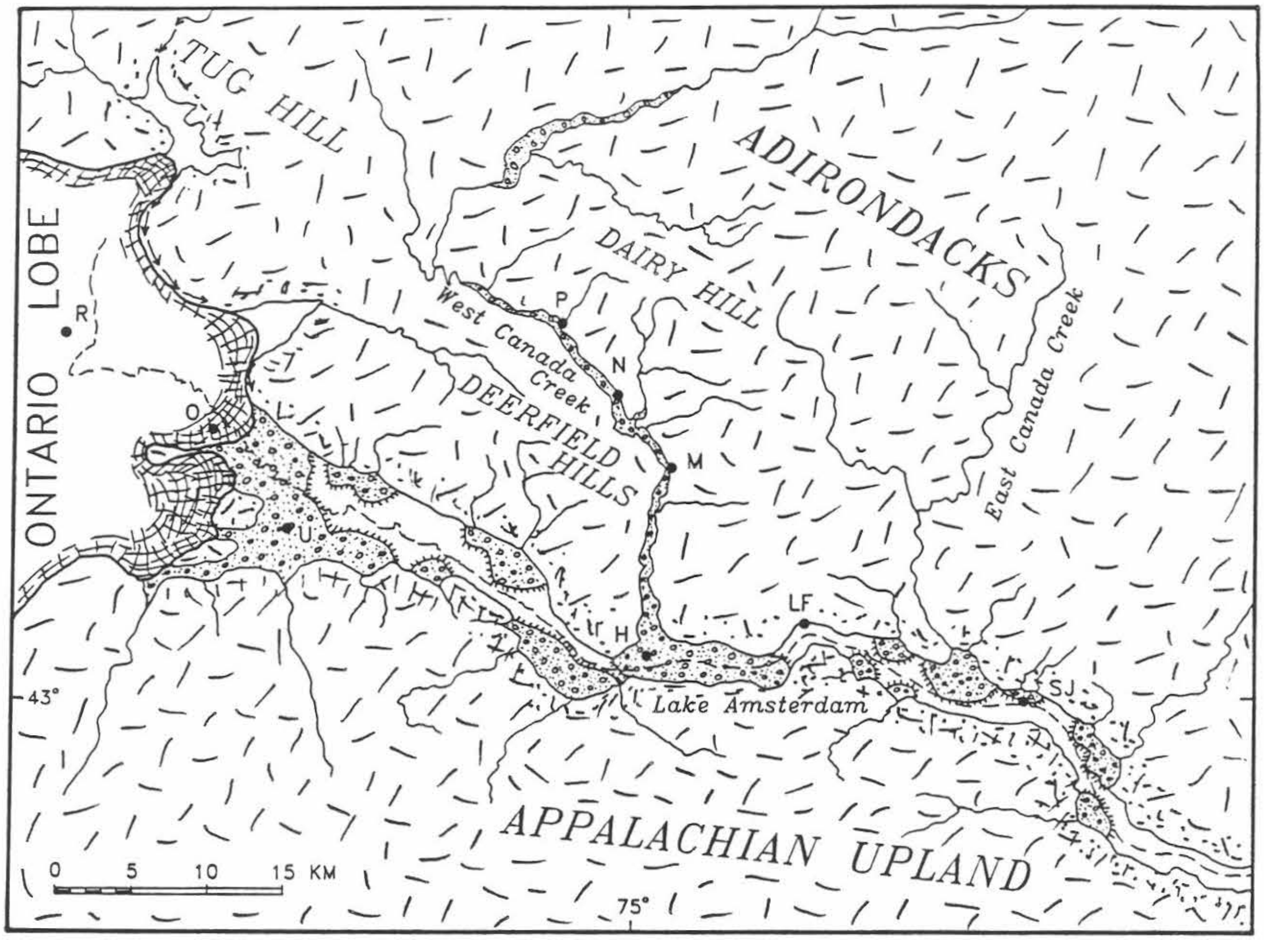
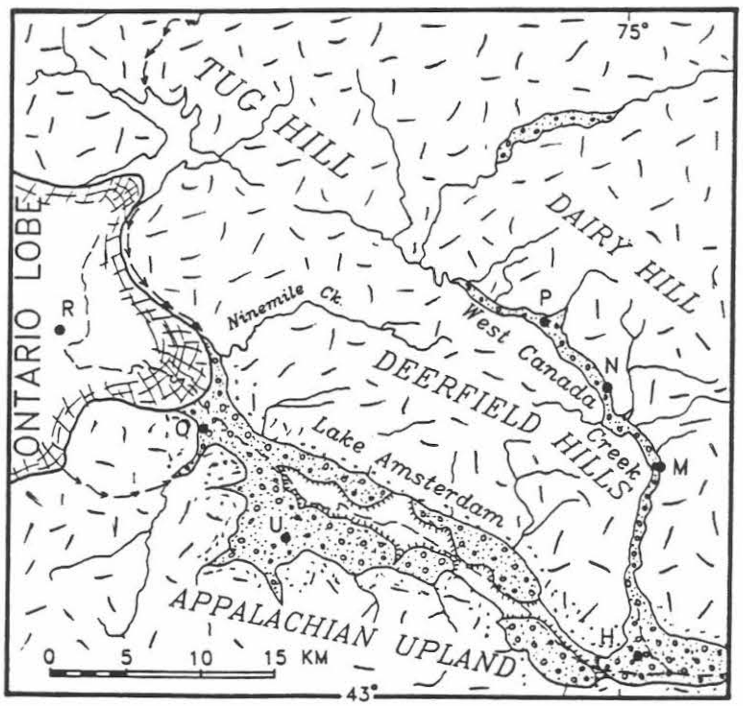


FIGURE 15. Recession of the Ontario Lobe to Oriskany and development of the Whitesboro Terrace in Lake Amsterdam.

FIGURE 16. The Ninemile Readvance and Lake Amsterdam.



Post-Valley Heads Readvance of the Ontario Lobe

The final Late Wisconsinan readvance of the Ontario Lobe into the Mohawk Valley, the Ninemile Readvance, deposited a thin (0.5-2.0 m) red till sheet (Rome till of Muller and others, 1986) on lacustrine sand as far east as Ninemile Creek (Figs. 10 and 16; Loewy, 1983). Deposits and features near Rome, that were used to define the Stanwix Readvance (Fullerton, 1971, 1980) may be the result of the more extensive ice cover of the Ninemile Readvance. A fluvial terrace at the limit of the Ninemile Readvance, which is inset in the Whitesboro Terrace, is probably graded to Lake Amsterdam. Following Ontario Lobe recession, the initial high phase of Lake Iroquois, named Lake Hyper-Iroquois by Fullerton (1971, 1980), may have been a westward extension of Lake Amsterdam. Lake Amsterdam finally drained as Mohawk Lobe recession unplugged the eastern Mohawk Valley allowing free eastward drainage to occur in the Mohawk Valley. Lake Iroquois established a spillway on the Mohawk-Ontario divide at Rome. Lake Iroquois drainage in the western Mohawk Valley (Iro-Mohawk River) is represented by a series of fluvial gravel terraces which can be traced down valley to Schenectady (LaFleur, 1983).

CONCLUSIONS

Several aspects of Late Wisconsinan glaciation in the western Mohawk Valley have broad applications to other areas of New York. In particular, the western Mohawk Valley has well exposed sections of sediment deposited by glaciers which advanced into deep glaciolacustrine troughs. The deposits provide a rare opportunity to study glacial processes and formulate depositional models for this type of environment. The region is especially important because trough environments in many areas are today occupied by modern lakes where core data is difficult to obtain and one must interpret seismic records. Glacial models developed in the western Mohawk Valley may enhance our ability to interpret seismic records.

It is also important to recognize possible glaciological controls imparted by (1) deep lacustrine water (up to 350 m) fronting calving ice margins, (2) the advance of ice lobes into troughs that have varying dimensions, and (3) changing water levels due to lake impoundment or breakout created by oscillating ice lobes. Water depth and trough dimensions in the Mohawk Valley may account for the rapid flow (surging?) and recession of ice lobes, the sudden termination of readvances, and the non synchronous nature of Ontario and Mohawk Lobe readvances on a scale of hundreds of years. Glaciological controls by these parameters have been relatively well studied in glaciomarine settings, but their lacustrine counterparts deserve much more attention because they are likely to be important to understanding the glacial history of lacustrine ice margins across much of the northeastern United States and Great Lakes region.

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ROAD LOG

Please have one person in each vehicle follow the road log to avoid getting lost and separated. Figure references are to those in the guidebook article, <u>except Figure 17 which is from Ridge and others, 1990 and will be handed out separately at the beginning of the trip</u>. All route descriptions and place names are given as shown on NYS Dept. of Transportation 7.5-minute topo maps which are culturally updated versions of USGS topos.

Be prepared for muddy and steep outcrops. A small pick or shovel and a knife for cutting clay are recommended. Two small stream crossings will be necessary depending on weather conditions.

<u>Assembly Point and Departure</u>: Colgate University, Hamilton, NY **8:00 A.M. sharp!**

The road log begins at the intersection of Rt.51 North and Rt.20 in the town of East Winfield or Birmingham Corners. To reach Rt.20 head north from Hamilton on Rt.12B. Take Rt.20 east to East Winfield.

Mileage	Route Description
0.0 (0.0)	From Rt.20 in East Winfield head north on Rt.51 toward Cedarville.

1.8 (1.8) Crossroads in Chepachet. Swampy lowland to the west is the channel leading south from Cedarville col.

3.0 (1.2) Pull over to the right side of Rt. 51.

STOP 1: **Overview of Cedarville col** (1210 ft) which is on the divide between the Unadilla River (Susquehanna Basin) and Steele Creek (Mohawk Basin) drainages. Cedarville col served as the outlet for Lake Cedarville, a high level Valley Heads lake in the Mohawk Valley (Figs. 5 and 10-12).

3.35 (0.35) Continue north on Rt.51. Turn left (west) in Cedarville to stay on Rt.51.

3.5 (0.15) Turn right to stay on Rt.51 which heads northeast into the valley of Steele Ck. Steele Ck. descends into Ilion Gorge which exposes much of the late Ordovician through Silurian stratigraphy of the western Mohawk Valley. Red shale exposed in the gorge is the Vernon Shale which is the source of the red color in glacial units of the region. A large inwash delta at the mouth of Steele Ck. in the Mohawk Valley indicates that Ilion Gorge was largely cut during the late Pleistocene existence of Lake Amsterdam.

10.5 (7.0) Rt.51 is joined by Bell Hill Rd. from the south as it enters Ilion. Follow Rt.51 into downtown Ilion.

12.1 (1.6) Traffic light in Ilion. Continue straight on Rt. 51.

12.5 (0.4) Rt.51 crosses the NYS Barge Canal and ends. Take Rt.5 East toward Herkimer. The trip will head through Herkimer where it will pick up Rt.28 North. Because of traffic lights in Herkimer the caravan will be reassembled on Rt.28 North after it splits away from Rt.5.

14.7 (2.2) First traffic light in Herkimer. Continue on Rt.5 East and watch for signs for Rt.28 North.

15.5 (0.8) Turn left (north) on the east side of Herkimer to follow Rt.28 north. Rt.28 follows the West Canada Creek valley.

18.6 (3.1) Town of Kast Bridge. In the next mile, bluff section 828 (Fig. 17) will be visible across W. Canada Ck. The flat top of the bluff to the east is the surface of the Holland Patent Diamicton and silt and clay of the post-Holland Patent beds (Figs. 2 and 13).

22.8 (4.2) Entering town of Middleville, KOA campground ahead on right. Across W.Canada Ck. from campground is bluff section 811 (Fig. 17).

23.8 (1.0) Rt.28 makes an abrupt right turn in Middleville. After turning right, make an immediate left onto Fishing Rock Rd.

24.2 (0.35) Pull over to the side of Fishing Rock Rd.

STOP 2: **Fishing Rock Rd. bluff**, section 324 (Fig. 17). This section has Precambrian gneiss at road level with two striation directions overlain by thin, discontinuous till in the West Canada Diamicton. The diamicton is overlain by sand and gravel and mass flow diamicton beds of an esker/subaqueous fan complex in the base of the upper Newport Beds. Fan deposition at this location may have been triggered by the lowering of Lake Newport at the close of pre-Valley Heads glaciation (Figs. 2 and 9).

24.55 (0.35) Retrace route back to Rt.28 and make a left onto Rt.28.

24.65 (0.1) After crossing W. Canada Ck. into the center of Middleville, turn left at the traffic light, following Rt.28 toward Newport.

27.0 (2.35) West Canada High School. The school sits on fluvial terraces graded to Lake Amsterdam or its immediate predecessor in the Mohawk Valley.

27.3 (0.3) Rt.28 crosses over White Creek.

27.6 (0.3) Rt.28 follows a bend in W. Canada Ck. Bluff section 451 (Fig. 17) is the embankment to the right.

28.0 (0.4) Pull off of Rt.28 on right taking advantage of the old highway pavement. Walk 0.1 mile north along side of Rt.28 to horse pasture and cross cutoff meander loop to bluff along east side of valley.

<u>STOP 3</u>: **Newport bluff,** section 447 (Fig. 17). This bluff exposes the top of the pre-Valley Heads section (Middleville Fm.) which is overlain by the Valley Heads section (Poland Fm.). The two units are separated by the Shed Brook Discontinuity at this exposure (Fig. 2).

29.1 (1.1) Continue north on Rt.28 after arriving in the center of Newport.

29.5 (0.4) In the north end of Newport, turn right onto Gage Rd. The octagonal limestone house at this corner is the workshop site of Linus Yale, inventor of the Yale Lock.

31.0 (1.5) Gage Rd. climbs to the top of a hill composed of the entire glacial section of the W. Canada Valley (Fig. 2). Turn right onto White Creek Rd.

31.2 (0.2) Turn left onto the Newport-Gray Rd. which crosses over White Ck. and enters the Factory Brook valley. About 0.2 miles beyond the White Ck. bridge pull over to the right side of the road by a cow pasture gate.

<u>STOP 4</u>: **Factory Brook bluffs,** sections 225-226 (Fig. 17). The bluff sections across Factory Bk. contain almost the entire pre-Valley Heads section (Fig. 2). The top of till in the West Canada Diamicton at this exposure shows subglacial grooving that records Mohawk Lobe ice flow to the northwest. The top of pre-Valley Heads deposits (upper Newport Beds) are subglacially deformed beneath Valley Heads till in the Norway Diamicton, an Ontario Lobe deposit from the Hinckley-St. Johnsville Readvance (Figs. 10 and 12).

33.3 (1.9) Retrace route back to Gage Rd. and Newport. Take Rt.28 north in Newport toward Poland.

36.8 (3.5) Follow Rt.28 north to the center of Poland. Continue north on Rt.28.

37.3 (0.5) Pull over to right at gravel pit entrance at north end of Poland. Follow dirt road into pit.

STOP 5: **Poland clay pit**, section 338 (Fig. 17). This section exposes lacustrine sediments in the lower part of the Valley Heads Poland Fm. Two important transitions occur in this section. A sudden change from lacustrine sand to dark gray silt and clay varves in the pre- and post-Hawthorne beds marks rising water levels during the initial impoundment of Lake Cedarville. A transition higher in the section from dark gray varves, with only gray ice-rafted pellets (post-Hawthorne beds), to pinkish gray varves, with red and gray ice-rafted pellets (pre-Norway beds), represents the transition from Mohawk to Ontario provenance that occurred between the Salisbury and Hinckley-St. Johnsville readvances. The pre-Norway beds are subglacially deformed beneath till in the Norway Diamicton.

37.8 (0.5) Retrace route back to center of Poland and turn left onto Rt.8 North.

39.5 (1.7) In the center of Cold Brook, continue north on Rt. 8.

42.1 (2.6) Rt.8 North enters the breakout channel of Lake Miller as it passes by Hurricane Rd. Continue north on Rt.8.

43.2 (1.1) Turn right off of Rt.8 onto Hall Rd.

43.7 (0.5) After Hall Rd. climbs to top of hill, turn left onto Burt Rd. and pull over to right.

STOP 6: **Lake Miller delta** at the maximum position of the Hinckley-St. Johnsville Readvance (Valley Heads, Ontario Lobe, Figs. 10 and 12). Burt Rd. crosses the distal part of the delta which is graded to a water level of 1410 ft that drained across a spillway into Spruce Creek. Map analysis of delta and spillway elevations in the upper West Canada Valley from this stage of Lake Miller indicates that isostatic tilting in the region was about 4-5 ft/mile. When the Ontario Lobe began to recede, Lake Miller drained southward across the delta by way of the breakout channel along Rt. 8 which served as the spillway for a second stage of Lake Miller at about 1390 ft.

44.5 (0.8) Follow Burt Rd. north to Rt.8. Take Rt.8 North.

45.85 (1.35) Continue north on Rt.8 as it crosses Black Creek.

47.85 (2.0) Turn left off of Rt.8 onto Ash Creek Rd. toward Ohio. Ash Creek Rd. will become Pardeeville-Ohio Rd.

50.7 (2.85) Pass through Ohio on Pardeeville-Ohio Rd. and cross over Reese Rd. The next several miles of the trip will cross areas of the Ohio sand plain which represents deposition of deltaic and lake-bottom sand in Lake Miller. Sand in this area sits on clay and silt of the Newport Beds (pre-Valley Heads).

51.8 (1.1) Turn right off of Pardeeville-Ohio Rd. onto Smith Rd.

52.7 (0.9) Turn left off of Smith Rd. onto Dow Rd.

53.6 (0.9) Intersection where Dow Rd. turns sharply to the left and becomes Hemstreet Rd. THIS AREA IS PRIVATE PROPERTY! Continue straight off of Dow Rd. on dirt path toward the Hinckley Reservoir. Access to the reservoir will depend on water levels.

STOP 7: **South shore bluffs of the Hinckley Reservoir.** The bluffs expose the lower through upper Newport Beds, deposited in Lake Newport during the time of the West Canada Readvance which was confined to the lower W. Canada Valley (Figs. 7-9). Readvance activity of the Mohawk Lobe is marked by provenance changes in varves at the Hinckley bluffs. Varves at the base of the section are sandy, have an Adirondack source, and overlie a very sandy and bouldery Adirondack diamicton. The sandy varves give way to clayey dark gray varves of Mohawk Lobe provenance. Mohawk Lobe varves are then overlain by Adirondack varves marking the recession of the Mohawk Lobe and dropping water levels in Lake Newport. The Hinckley bluffs also have spectacular slump and load structures.

54.5 (0.9) Return to Hemstreet Rd. and follow it south to a T with Hill Rd. Turn right (west) onto Hill Rd. and follow it toward Grant. Hill Rd. will turn into Stormy Rd.

56.5 (2.0) After crossing over Black Ck. in Grant, turn right onto Southside Rd. and follow it to the Hinckley Reservoir dam and Rt.365.

60.6 (4.1) Turn left onto Rt.365 toward Barneveld after crossing W. Canada Ck.

61.0 (0.4) Town of Hinckley. The bluff tops across W. Canada Ck. to the south are ice-contact deltas built along the receding margin of the Ontario Lobe (Hinckley-St. Johnsville Readvance, Figs. 10 and 12). The deltas are graded to the second stage of Lake Miller which drained through the channel seen along Rt.8 near Stop 6.

62.5 (1.5) Continue west on Rt.365 north of the town of Prospect. Prospect sits on the surface of a delta built into Lake Prospect (Franzi, 1984) which formed for a short time during the drainage of Lake Cedarville in the Mohawk and W. Canada valleys.

64.7 (2.2) Immediately after crossing under the Conrail railroad bridge turn left onto Parker Hollow Rd. which follows Cincinnati Ck.

65.1 (0.4) Pull over to the right side of Parker Hollow Rd.

STOP 8: **Bluffs on Cincinnati Ck.** expose later Valley Heads deposits in the upper W. Canada Valley. The base of the section is esker/subaqueous sand and gravel overlain by varves with red ice-rafted pellets (post-Norway beds). Till in the Holland Patent Diamicton overlies the post-Norway beds and it is overlain by an ice-contact delta of Lake Gravesville (Figs. 10 and 13). This locality is just inside the eastern limit of the Barneveld-Little Falls Readv. of the Ontario Lobe.

66.4 (0.7) Continue southwest on Parker Hollow Rd., under the Rt.12-28 highway bridge to a stop sign in Barneveld. Continue straight across the intersection onto Rt.365 West.

70.3 (3.9) Center of Holland Patent, continue west on Rt.365. Watch for signs for Floyd.

75.2 (4.9) Center of Floyd. Turn left (south) off of Rt.365 onto Koenig Rd. toward Oriskany.

77.4 (2.2) At end of Koenig Rd. turn right (west) onto River Rd.

77.55 (0.15) Pull off to right onto dirt road at sand pit.

STOP 9: **River Rd. sand pit.** This section is representative of exposures in the broad undulating plain between Rome to the west and Ninemile Ck. to the east. It provides evidence for the final readvance (post-Valley Heads) of the Ontario Lobe into the Mohawk Valley (Ninemile Readvance, Figs. 10 and 16). The section has 0.5-2.0 m of red till (Rome till of Muller and others, 1986) overlying about 10-15 m of lacustrine sand that was deposited in Lake Amsterdam. The sand is subtly deformed at the base of the till which does not occur east of Ninemile Ck.

78.3 (0.6) Retrace route east and continue on River Rd. to bridge over Ninemile Ck. Continue east on River Rd.

78.8 (0.5) After crossing small unnamed stream valley, branch off to south away from River Rd.

79.1 (0.3) You will encounter a complex intersection with a series of stop signs. DO NOT get on Rt.49. Stay to the left and follow signs to Oriskany.

79.7 (0.6) Cross over the Mohawk River bridge.

80.4 (0.7) After crossing the Mohawk River flood plain, turn left onto Rt.69 East in Oriskany.

81.1 (0.7) Pass by landfill entrance on right. Hill top to right is the Whitesboro Terrace. Continue east on Rt.69.

82.2 (1.1) At entrance to Burrows Hauling Co. follow dirt road to back of large sand and gravel pit.

STOP 10: The pit shows the interior of the **Whitesboro Terrace** which is an esker-fed, ice-contact feature formed at a recessional position of the Ontario Lobe during the last gasp of Valley Heads glaciation (Figs. 10 and 15). The terrace appears to be an ice-contact delta because it has 5-10 m of fluvial gravel capping lacustrine sand along what resembles a deltaic topset/foreset contact. However, the contact does not represent a horizontal surface and is tilted regionally down to the east-northeast. The gravel was deposited on a fluvially scoured surface that probably represents trimming of the sand (delta or subaqueous fan?) in response to falling lake levels in the Mohawk Valley. The gravel cap represents the first Ontario Lobe, ice-contact deposit graded to a water level less than 500 ft in the western Mohawk Valley (Lake Amsterdam). Further east in the western Mohawk Valley Lake Amsterdam is only recorded by inwash deltas.

END OF MILEAGE LOG: At this point we will try to return to Hamilton as quickly as possible following the directions below:

- Continue southeast on Rt. 69 for 2 miles to where it turns into Rt.5A. Follow Rt.5A southwest for 3 miles to Rt.5. Turn right (west) onto Rt.5 for 0.15 miles. Turn left (southwest) onto Rt.5B for 1.4 miles. Take Rt.12B southwest for 14 miles through Clinton and Oriskany Falls to Rt.20. Take combined Rts. 20 and 12B west for 3 miles through Madison to where Rt.12B splits to the south. Take Rt.12B south for 6 miles to Hamilton.

GEOLOGY OF THE MOHAWK AND BLACK RIVER VALLEYS - A FIELD TRIP FOR EARTH SCIENCE TEACHERS

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Rocks exposed in the Mohawk and Black River Valleys consist of a sequence of Late Precambrian metamorphic rocks unconformably overlain by a sedimentary succession ranging from Cambrian through Silurian in age. A veneer of unconsolidated Pleistocene sediments mantles the bedrock. Changes in plate tectonic setting with time are reflected both directly and indirectly in the changing character of the rock record, and Pleistocene deposits in the region provide clues to Late Wisconsinan deglaciation. The purpose of this field trip is 3-fold: 1) to examine the character of the bedrock, 2) to see how changes in plate tectonic setting are mirrored by changes in the rock record, and 3) to study the record of Late Pleistocene glacial and peri-glacial environments of central New York. The introductory text provides background on the bedrock geology, plate tectonic setting, and Pleistocene history of the region.

Precambrian Basement Rocks

Late Precambrian metamorphic rocks underlie all sedimentary rocks in central New York State and are exposed at the surface east of the Black River Valley and north of the Mohawk River Valley. The Black River Valley, in fact, lies essentially along the unconformity, separating gently southwest-dipping Paleozoic sedimentary rocks from Precambrian basement of the Adirondack dome to the east. These rocks consist of a complex sequence of sedimentary and igneous rocks intensely deformed and metamorphosed during an event known as the Grenville Orogeny. We know enough about the geology of Precambrian rocks in New York State to begin to put together a picture of events that unfolded during the Late Precambrian. There are enormous gaps in our knowledge, however, and we are forced to speculate at nearly every stage of our reconstruction.

What was the environment like during accumulation of the sedimentary and volcanic rocks? The presence of stromatolites, evaporites, quartz sandstones, and limestones suggests a shallow sea over continental crust. The volcanic rocks, impure sediments, and possible hot springs metal deposits may indicate rifting, perhaps during the early stages of development of a passive margin or a back-arc basin. Radiometric dates tell us that the oldest of these rocks were probably deposited about 1300 million years ago. Until we know more about the original sequence of rock units, we will probably not be able to say much more.

What happened during the Grenville Orogeny? The crust experienced a major shortening event — folding and ductile shearing combined to deform and thicken the crust, deeply burying rocks that had at one time lain at the surface of the Earth. These rocks recrystallized at high temperatures and pressures. Early in the sequence of events, anorthositic magmas seeped up from the mantle, perhaps derived by fractional crystallization of basaltic magmas trapped at the base of the crust. Heat transferred from the hot anorthosite partially melted the surrounding crust, and silicic magmas of varying compositions rose upward. Even when solidified, the anorthosites were less dense than the surrounding country rocks. They rose slowly upward through the deforming mass of crust, the smaller masses forming mushroom-shaped domes, and the largest masses spreading laterally into great sheets with several "roots". The presence of highly foliated margins and unfoliated cores suggests that the anorthosites moved upward by ductile shear along the margins of the bodies. Folding and ductile shear continued throughout the region, refolding the country rock and deforming nearly all of the igneous rocks after they had solidified.

It is difficult to say exactly when the Grenville Orogeny began. If we believe that the anorthosite magma was intruded during the early phases of deformation, the radiometric dates on the anorthosite ranging from 1300 to 1100 million years tell us that the Orogeny was under way at least 1100 million years ago, perhaps earlier. Deformation and metamorphism appear to have peaked between 1020 and 1100 million years ago. By about 900 million years ago, the rocks had cooled enough that most of the sensitive radiometric clocks had been set.

Now, we come to the "why" of the Grenville Orogeny. In proposing a model, we need to keep a number of things in mind: 1) the presence of what may have been a rift basin or passive margin sequence, 2) major crustal shortening during the Grenville, directed from east to west or southeast to northwest, and 3) the presence of crust that was twice as thick as normal continental crust at the end of the Grenville Orogeny.

A number of people have proposed that a continent-continent collision akin to the modern collision of India with Asia is a reasonable model for the Grenville Orogeny. This is a logical conclusion to draw, because, at present, we know of no way other than a major accretionary event to double the normal thickness of continental crust. Such a collision would have severely compressed the crust, causing major overthrusting, repeated folding, and ductile shearing as large slabs of crust were driven northwestward. Figure 1 illustrates a plausible sequence of events for the Grenville Orogeny.

When we look at more modern examples, we see that welding of large terranes such as those involved in the Grenville Province is typically a complicated event, typically involving a number of smaller terranes. So much has been eroded from the Grenville Province that we have no way of guessing how complicated the collision might really have been. How many terranes collided with the margin is unclear. In fact, it isn't even clear where a suture or sutures might lie. The lack of an obvious candidate for a suture zone is one of the main arguments used against a plate tectonic interpretation for the Grenville Orogeny. Those in favor of a plate tectonic interpretation simply place the suture out of sight, buried beneath younger rocks to the east.

What happened after the Grenville Orogeny? Over the course of nearly 400 million years, erosion and tectonic unroofing stripped enormous volumes of rock off the great mountain range, until nearly 30km of material had been removed.

Paleozoic Sedimentary Rocks

Paleozoic sedimentary rocks in Central New York State range in age from the very latest Precambrian through the end of the Silurian, a time span of about 200 million years. No other region in the State preserves such a complete sedimentary record from this period of time. Although rocks of the Taconic region are similar in age, they are complexly deformed and more difficult to interpret. This time period is an important one in the geologic evolution of New York State, because it includes rifting of the eroded supercontinent of Grenvillia, opening and widening of the Iapetus Ocean, closing of the western Iapetus, and collision of the Taconic Island Arc Terrane with the eastern margin of Laurentia (the "North American" segment of rifted Grenvillia) during the Taconic Orogeny. We should be able to find important clues to these events in the rocks of this region. This region is also important to our understanding of the evolution of life during the Silurian Period — the record in New York State is nearly complete and is the best known Silurian record anywhere in eastern North America.

The sediments in this region are remarkably varied. They range from poorly-sorted feldspar-rich sandstones to clean quartz sands, bedded iron ores, layers of salt, fossiliferous limestones, and unfossiliferous black shales. We find such a variety of sediment types that we might begin to suspect quite a variety of sedimentary settings as well. What were these sedimentary settings like? How did they change with time? What do the changes tell us about the geologic evolution of New York State?

Rather than trying to digest the entire section at once, we will divide the sedimentary record into 3 packages, the first representing deposition during the Late Precambrian through Early Ordovician, the second representing the Middle and Late Ordovician, and the third the Silurian. Each time period corresponds to a distinctly different regional tectonic setting — opening of the Iapetus, accretion of the Taconic Island Arc Terrane during the Taconic Orogeny, and aftermath of the Taconic Orogeny. One of the intriguing questions we will want to answer is how an undeformed sedimentary sequence can record significant evidence of major regional tectonic events such as these.

Latest Precambrian Through Early Ordovician. Rocks from this interval of time are represented by scattered patches of poorly-sorted feldspar-rich sandstones of latest Precambrian age overlain by regionally-extensive shallow water marine sandstones and dolostones of Late Cambrian and Early Ordovician age.

What kind of picture emerges when we tie together inferences about the rocks and their sedimentary settings? Between 600 and 650 million years ago, the crust was stretched, and major normal faults trending north-northeast broke the crust. These faults are very different from the ductile shear zones of Grenville age. Because they formed at shallow levels in the crust, they are marked by shattered rocks known as breccias. Basaltic dikes were also intruded at this time. The rift valleys must have looked much like the modern East African and Rio Grande Rifts. Rare deposits of poorly sorted, locally-derived feldspathic sands (the Nicholville arkoses) date from the very end of the Precambrian and suggest that fault-bounded basins were probably scattered throughout the Adirondack region at this time. During the Late Cambrian, a shallow sea advanced across the region from east to west, spreading a blanket of beach sand (the Potsdam Sandstone) on eroded Precambrian basement and on scattered remnants of Late Precambrian rift basin sediments. As the shoreline transgressed westward and sand supply diminished in the Early Ordovician, carbonate sediments were deposited in widespread nearshore dolomitic layers (the Little Falls Formation, which contains Herkimer diamonds, and the Theresa Formation) and thicker offshore limestone reefs and banks to the east. Throughout the Late Cambrian and Early Ordovician, the east coast of Laurentia would have looked much like an unvegetated version of the Florida Coast, with a wide ocean stretching off to the east. Early in the Middle Ordovician, however, this quiet shelf environment was temporarily interrupted by subaerial erosion severe enough to remove part or all of the sedimentary record. In the Black River Valley, all of the Potsdam Sandstone and Theresa/Little Falls Formations were eroded during this event.

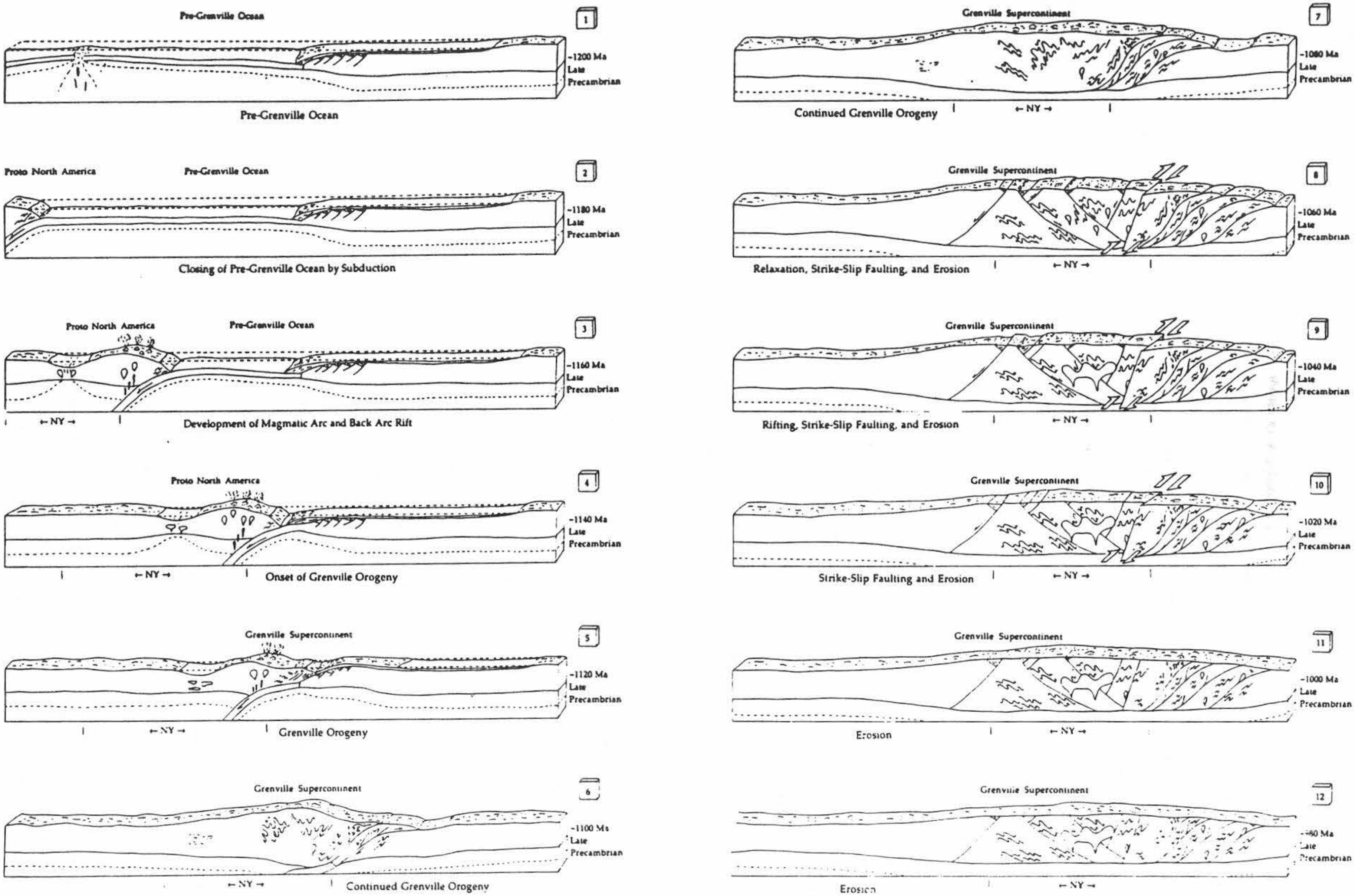


Figure 1. Possible plate tectonic model for Grenville orogenic events.

Why did these events happen? Our current plate tectonic model suggests that the supercontinent of Grenvillia rifted into a number of fragments during the very Late Precambrian (figure 2). The rift valleys in New York State reflected stretching and thinning of Laurentia as rifting proceeded. As the Iapetus Ocean grew, Laurentia moved away from the spreading center. The east-sloping continental margin slowly subsided, and a shallow ocean gradually transgressed westward. Formation of the Iapetus Ocean rather suddenly increased the volume of the Cambrian world rift system. Because spreading centers are topographically high, the effect of adding to the world rift system is much the same as dropping rocks into a bathtub. Displaced water likely spread over the continents in a world-wide rise in sea level. Both subsidence and sea-level rise were probably involved in marine transgression across New York State.

From the Late Cambrian through the Early Ordovician, the passive margin of Laurentia slowly accumulated sediments, with sands coming from the continent to the west. Unbeknownst to the unsuspecting trilobites, however, the Taconic Island Arc was slowly advancing toward Laurentia (figure 2). The Knox Unconformity, reflected in the absence of Late Cambrian and Early Ordovician sediments in the Black River Valley, may have been the herald of the inevitable collision. As the collision began, the Laurentian continental shelf may have buckled up above sea level in a great bulge, exposing the shelf to extensive erosion.

The Middle Through Late Ordovician. The general character of sediments deposited in the region changed dramatically as the Middle and Late Ordovician unfolded. Thin, widespread carbonate sediments were succeeded by interlayered carbonates and shales. Tremendous thicknesses of black shale then blanketed the carbonates. Finally, sands, silts, and shales were deposited across the region.

When we summarize our inferences about the Middle and Late Ordovician, we can see immediately that the picture is more complicated than it was for the Early Ordovician. The Middle Ordovician opened with gradual subsidence of an east-sloping carbonate shelf (the Black River Group – see the stratigraphic column in Plate I). Volcanic ash and muds derived from the east gradually appeared in the record (the Trenton Group) and tell us that open ocean no longer lay to the east. Then, the carbonate shelf buckled and foundered in the east. Earthquakes appear to have shaken loose great slurries of sediments that slid off the shelf to the west and into the basin to the east (turbidites of the eastern Trenton Group and Utica Shale).

The "hinge" of the basin moved westward as the water deepened to as much as 500m, and major carbonate deposition died out altogether. The eastern shoreline of the basin lay in the vicinity of eastern New York, the western shoreline well to the west of the State. As the basin slowly filled with sediment derived from the east, deep water shales (the Utica and Whetstone Gulf Shales) gave way to shallow water siltstones, shoreline sands (the Pulaski Shale and Sandstone and the Oswego Sandstone), and subaerial sediments of the Queenston Delta (the Queenston Shale) (Plate I).

Why did these events happen? These events span the time of the Taconic Orogeny. Although the sedimentary rocks of Central New York State are essentially undeformed, the changes in sedimentary environments reflect major tectonic events taking place farther east. Our current plate tectonic model suggests that collision of the Taconic Island Arc with Laurentia began in earnest in the Middle Ordovician (figure 2). As the arc rode up over the eastern edge of the continental shelf, the margin of Laurentia foundered. At first, subsidence was slow, and deposition of shallow-water carbonates kept pace with the subsidence. Volcanic ash wafted in from the Taconic Island Arc to the east. As the relentless advance of the arc telescoped the continental shelf, the shallow marine basin in New York State buckled and subsided rapidly along great fault blocks. Sediments eroded from the rising Taconic terrane and arc to the east poured westward into the deepening basin, and carbonate sedimentation was extinguished altogether. We do not know whether the flood of Late Ordovician sediments forming the Queenston Delta reflects renewed uplift to the east or simply extensive erosion accompanying a world-wide drop in sea-level related to a major glacial event.

Accretion of the Taconic Island Arc Terrane produced dramatic results in this region. As collision buckled the margin of the continent, water deepened from at most a few tens of meters to over 500 meters. That's almost twice as deep as the North Sea!

The Silurian. Silurian rocks are exposed along the southern margin of the Mohawk River Valley, where they dip gently south beneath the great mass of Devonian rocks exposed on the Appalachian Plateau. Silurian rocks lie above an unconformity developed on the Ordovician Queenston Shale and are exposed at the surface along a wide band extending east from Niagara Falls and narrowing to nothing east of Canajoharie.

The general picture that emerges for this short interval of geologic time is one of terrestrial and nearshore sedimentation in a complexly changing set of environments. Overall, carbonate deposition increased in importance over deposition of sandstones, siltstones, and shales as the Silurian progressed. Although the dramatic changes of the Ordovician are missing, the changes in both sediments and fossils as a result of irregular fluctuations in shoreline and marine circulation patterns tell us an intriguing story.

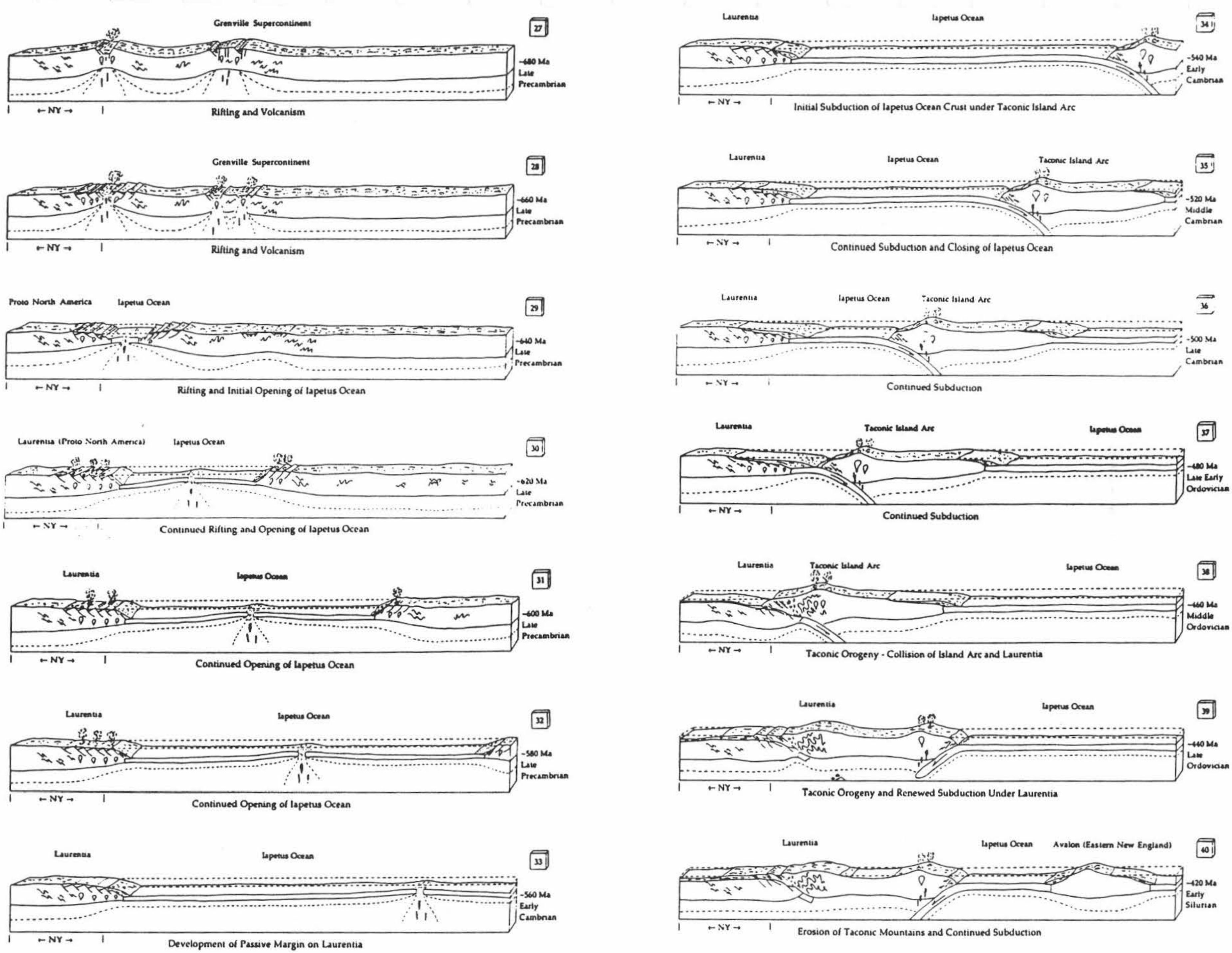


Figure 2. Plate tectonic model for development of an Early Paleozoic passive margin and the Taconic Orogeny.

From the end of the Ordovician through the earliest part of the Silurian, the entire region must have lain above sea level, because we have no rock record from that time. In that respect, New York State is little different from most other places in the world. The unconformity itself, though, is typically very difficult to locate, especially where Lower Silurian deltaic deposits overlie very similar Ordovician rocks of the Queenston Delta.

Shallow seas washed the region throughout much of the Silurian. During the Early Silurian, an ocean briefly advanced eastward as far as Medina, then shrank westward until it once again lay past the border of the State. Several million years later, the ocean readvanced eastward as far as Oneida and Utica. Gravels (the Oneida Conglomerate) (Plate I), sands, silts, and muds (the Clinton Group) accumulated on deltas, beaches, and tidal flats early in the Silurian. Unusual marine water conditions resulted in the formation of Clinton-type iron ores at the time of deposition of some of the Clinton Group sediments. Clastic sediments eventually gave way to carbonates deposited on extensive shallow water shelves (the Lockport Group). Nowhere was the water very deep. Near the end of the Silurian, evaporation and poor circulation produced large tracts of very salty lagoons and tidal flats, in which eurypterids played and layers of salt accumulated (the Vernon Shale and the Salina Group) (Plate I).

What was the tectonic setting of the region during the Silurian? The Taconic Orogeny was over by the end of the Ordovician. By the time the Early Silurian had arrived, the Taconic Island Arc Terrane was not only securely welded to Laurentia but quite thoroughly eroded as well. Major tectonic activity occurred far to the east along the plate boundary, where west-dipping subduction slowly consumed crust of the central Iapetus Ocean (figure 2). New York State itself was a quiet foreland basin, much like the shallow ocean lying between Australia and New Guinea. The origin of the very gentle upwarping and downwarping reflected in shoreline migrations throughout the Silurian is not particularly well understood and may be related to convergence along the plate boundary to the east.

The Fossil Record. Early Paleozoic seas were dominated by invertebrate organisms - brachiopods, clams, worms, snails, trilobites, corals, bryozoans, nautiloids, graptolites, echinoderms, tentaculitids, and ostracodes. Evolution of individual species is beyond the scope of this summary, but a number of landmark evolutionary events deserve comment.

Middle Ordovician seas supported a host of new creatures. A new invertebrate group, the bryozoans, appeared and became important as colonial rock builders. Another conspicuous new arrival, the coral-like stromatoporoids, constructed extensive mound-like reefs. True corals likewise made their initial appearance — the oldest known coral reef in the world is found on Isle La Motte in Lake Champlain.

The Silurian Rochester Shale preserves a remarkably diverse fauna of over 200 species, including 84 of bryozoans alone. There were hosts of ostracodes and stalked echinoderms, and even larger numbers of brachiopods. Although trilobites were on the decline, the surviving families were still important. The tentaculitids, a steadily increasing group of tiny, conical ringed shells, became important. Corals, snails, clams, and nautiloids, though less abundant, also continued to evolve and disperse. The seas were well-populated, and competition for food and survival was keen. It is hardly surprising that air-breathing arthropods began to evolve at this time and eventually colonized the land.

Masked in the generally sparse record of the Silurian is the history of a great evolutionary advance — the lineage of the earliest vertebrates, the fishes. Although fishes are known in the Lower Ordovician and may have appeared even as early as the Cambrian, the Silurian is characterized by appearance of a number of more modern types. During the ensuing Devonian Period, a great variety of fishes populated the seas, but armor-skinned fishes were preserved as fossils in the Silurian Vernon Shale.

Land plants had not made an appearance even as late as the end of the Silurian. When we try to visualize the region at this time, we must picture a bare landscape devoid of trees and grasses — a bleak scene with no vegetation to protect the surface from the onslaught of wind and water.

The Pleistocene

During the Pleistocene, New York State was covered by a succession of continental ice sheets. Little evidence of early ice sheets remains, but the last and most extensive glaciation left a strong stamp on landforms seen in the region today. The Laurentide ice sheet covered northeastern North America during Wisconsinan time and had its center in the Laurentian Mountains of Quebec and the uplands of eastern Quebec and Labrador. The ice sheet was nearly equal in size to the present Antarctic ice sheet and covered up to 12.3 million square kilometers. At the maximum extent about 54,000 to 63,000 years before present (ybp), the Wisconsinan ice sheet covered all of New York State with the exception of the Salamanca Reentrant (figure 3a).

The Laurentide ice sheet flowed across New York in four major "ice streams" or lobes. The lobes formed in response to differential flow within the ice sheet and were probably affected by underlying topography. In north-central New York

State, the Ontario lobe and the Mohawk, Oneida and Black river sub-lobes have been identified through grooves and striations, and by the deposits they left behind. The Wisconsinan ice lobes obliterated earlier proglacial river systems and accentuated the depths and widths of the earlier valleys. Isostatic subsidence reduced gradients and lowered elevations by as much as 120 to 130m (Flint, 1971).

Recession began to take place as the climate warmed and melting of the ice sheet exceeded the southern flow. Recession was probably marked by many still-stands and readvances of varying durations. Today, the locations of these former ice margins are marked by looping morainal belts draped across the hills and valleys. Geologists have identified several stages in retreat of the ice as the margin migrated northward across the State. The first two well-defined stops are called the Binghamton and the Valley Heads stages. The Binghamton Stage has been identified at about 14,000ybp, and the Valley Heads Stage at 12,000 to 13,800ybp. The Valley Heads stage formed a very distinctive line of moraines that can be traced from Cooperstown through Oriskany Falls, Munnsville, to Tully and westward across the State. At this time, major meltwater drainage was primarily southward through the Allegheny, Chemung, Susquehanna, and Hudson Rivers (figures 3b and 3c).

During the 2000-year interval between the Valley Heads Stage and the formation of Glacial Lake Iroquois, ice retreat and readvance formed a very complex series of erosional and depositional features. Regrettably few of these features have been dated radiometrically. As the ice sheet stagnated, sediment-laden meltwater poured into depressions on, in, under, or adjacent to the ice. Meltwater also scoured deep channels as it worked its way toward lower elevations.

As the Ontario lobe retreated up the western Mohawk Valley, and the Hudson lobe retreated toward present-day Albany, meltwater was impounded between the two lobes. This formed glacial lake Herkimer, which ranged from a maximum level of 439m (1440') to a low level of about 305m (1000') (figure 3d). These levels are given in terms of present elevations above mean sea level and reflect some isostatic rebound. Lake Herkimer drained to the south down the Unadilla and Chenango Rivers through a col or low pass at Cedarville and into the Susquehanna River (Ridge *et al.*, 1984). Glacial Lake Newberry was formed by water trapped between the ice margin to the north and the Valley Heads Moraines to the south at an elevation of about 366m (1200'). Lake Newberry formed a series of interconnected finger lakes that filled the through-valleys and were joined by meltwater channels that cut the north-south trending drainage divides (Hand, 1978).

As the ice thinned and retreated, lower drainage channels opened and allowed the ice-marginal lakes to drop to lower levels. Lake Hall to the west dropped to the 274m (900') level and Lakes Amsterdam and Schoharie to the east formed at the 189m (620') and the 207m (620') levels (LaFleur, 1979) (figures 3f and 3g).

As the ice margin continued its northerly retreat from the edge of the Appalachian Plateau, drainage was largely to the east along the southern margin of the ice. Along the southern edge of the ice margin, meltwater formed another lake that was bounded on the south by the higher elevations of the plateau. This large lake, named Lake Iroquois, drained to the east through the Mohawk Valley via an outlet at Rome at an elevation of 137m (450')(figures 3g and 4b).

Excellent examples of Lake Iroquois sediments are preserved just north of Canastota, where varved clays, marl, silt, and peat (in ascending order) record both deposition in and drainage of this large lake. Drainage from Lake Iroquois proceeded eastward through the former Lake Amsterdam channel and into Lake Albany at about 100m (300'). When the ice dam just south of Albany opened, these two lakes drained southward down the Hudson River (LaFleur, 1977). Eastward drainage at Rome continued until northward retreat of the ice opened the St. Lawrence valley to northeastward drainage at a lower level.

The Black River Lobe filled the Black River Valley at the times of Lakes Herkimer, Amsterdam and Iroquois. Meltwater was trapped between the retreating ice tongue and the southeastern edge of the Tug Hill Plateau, forming a series of ice-marginal lakes in this region as well. Lakes Forestport and Port Leyden filled the southern end of the valley and drained to the south. Meltwater pouring from Lake Port Leyden into Lake Iroquois roared down Lansing Kill and scoured the Boonville Gorge out of soft and easily-eroded Utica Shale. A large delta formed as the end of the gorge on the northern side of Lake Iroquois, giving modern day Lake Delta its name.

During retreat of the ice, the land immediately south of the ice margin was predominantly covered with tundra vegetation at higher elevations and grassland in lower regions. As the ice margin crept northward, grasses and forests of spruce and pine developed on the bare glacial deposits. Vegetation apparently took hold fairly rapidly after the ice cleared an area (Flint, 1971). Pollen studies, occurrence of wood fragments, and deposits of pine needles in glacial lake sediments provide evidence of the wide variety of plants in the region. Elk, mastodont, beaver, and Stone Age humans were among the mammals living just south of the ice margin.

After drainage of the proglacial lakes, barren lake bottoms and shorelines lay exposed to the elements. Although vegetation quickly covered much of the area, especially the fertile lake beds, the sandy beaches were exposed to wind erosion. Prevailing westerly winds picked up large amounts of sand on the eastern edges of former lakes and moved it eastward. Just east of the city of Rome, a series of large dunes are preserved in the Rome Sand Plains. The Sand Plains are formed of cross-

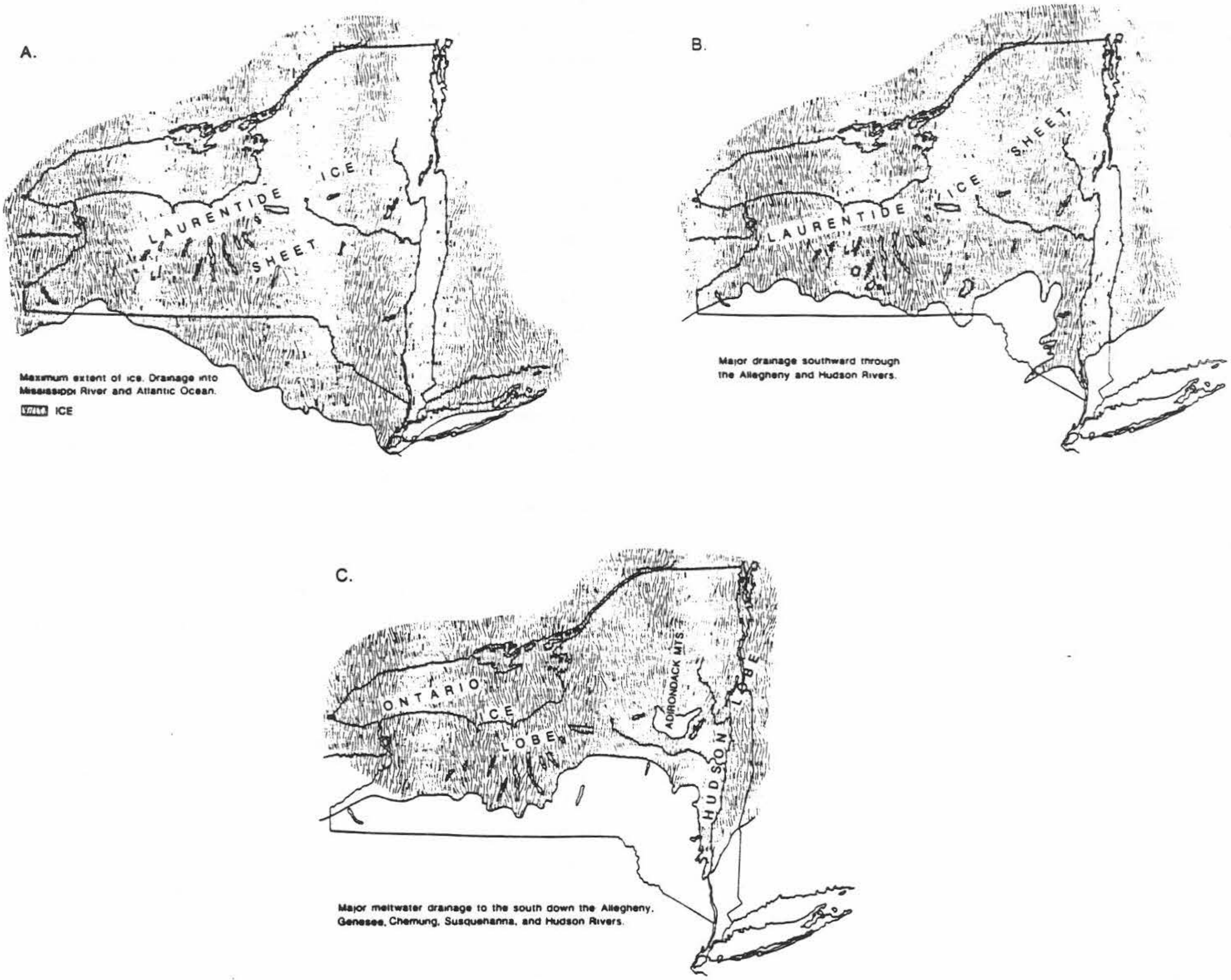
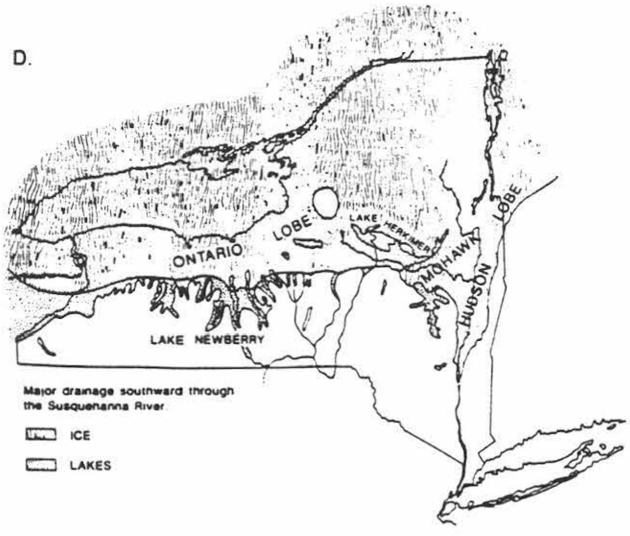


Figure 3. These maps illustrate selected stages during retreat of Wisconsinan ice in New York State. Inferred locations of ice margins and lakes are modifications of maps done by Fairchild (1909), with modifications based on work done by Ridge *et al.* (1984). The diagrams have been further modified by the New York State Geological Survey and appear on pages 178 and 179 of Isachsen *et al.*(1991).

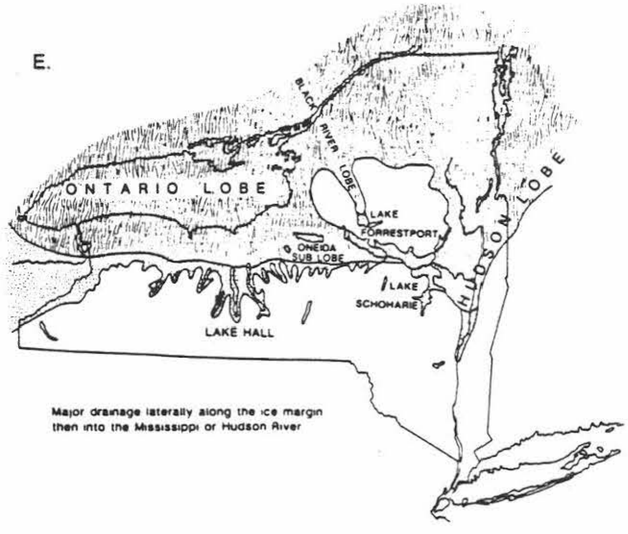
During deglaciation, ice retreat was neither continuous nor regular. Sporadic and localized readvances, still-stands, and retreats caused major changes in paleogeography by closing or opening an important drainage channel or outlet. The maps are presented to give the reader a general overview of the complex series of events that probably occurred in Central New York.

D.



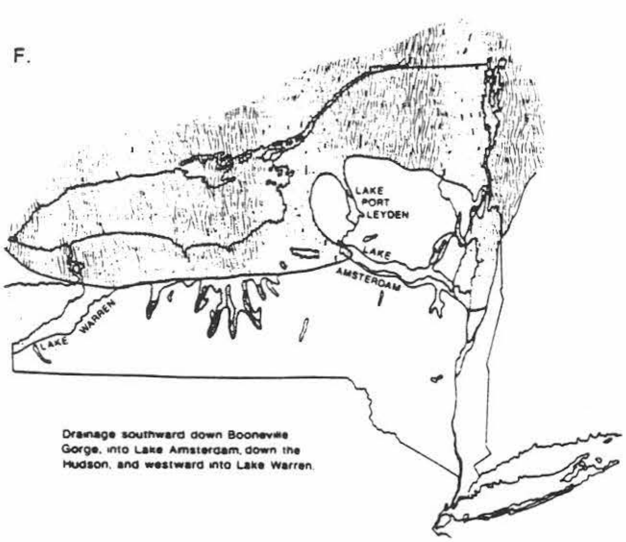
Major drainage southward through the Susquehanna River.

E.



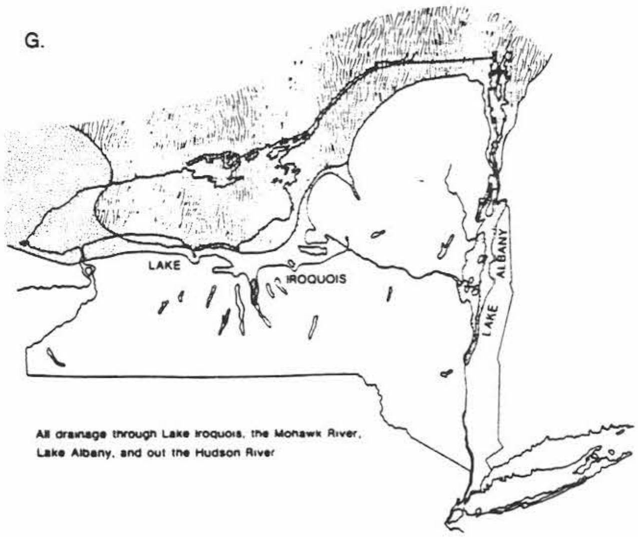
Major drainage laterally along the ice margin then into the Mississippi or Hudson River

F.



Drainage southward down Booneville Gorge, into Lake Amsterdam, down the Hudson, and westward into Lake Warren.

G.



All drainage through Lake Iroquois, the Mohawk River, Lake Albany, and out the Hudson River

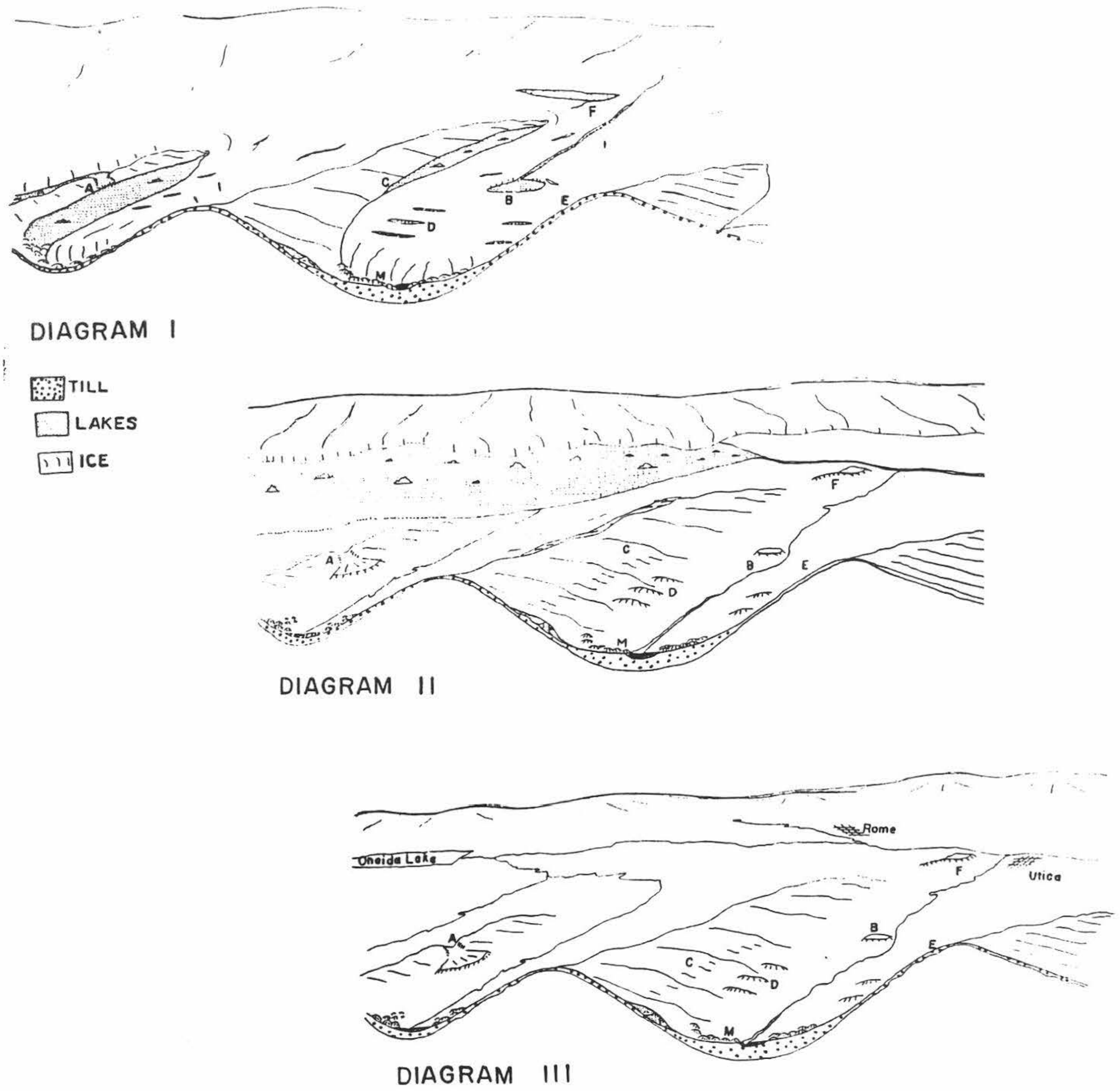


Figure 4. Idealized sketches showing an interpretation of deglaciation of the western Mohawk Valley. Sketches are drawn from a hypothetical aerial vantage point above Oriskany Falls, New York. **Diagram I** shows the looping moraines associated with the Valley Heads Readvance (M). With stagnation and thinning of the ice, meltwater laden with sediment collected along the edges of the ice, forming small lakes (B, C). Just south of Vernon Center, a metlwater channel formed on the west seide of Eaton Hill (A). The channel was cut into bedrock at first, and then flowed north along the ice margin until it was blocked or diverted by ice and cut east across the ridge, forming a large delta slightly west of route 26. Meltwater also formed a large delta in the Oriskany Valley south of Clinton between route 12B and Dugway Road (B). Along the walls of the valley, kame terraces formed from the remains of lateral moraines and sediment left against the ice by meltwater. As the ice melted, away, these deposits slumped and were faulted and broken (C). Crevasses on the ice also filled with sediment, and en echelon crevasse fillings occur in Oriskany Creek Valley along route 12B south of Deansboro (D). Ground moraine was deposited in layers ranging from a few centimeters to tens of meters thick (E). In the Mohawk Valley, an esker and delta formed where metlwater flowed off the stagnant ice into glacial Lake Amsterdam (F). **Diagram II** shows a generalized view of the same area at the time of glacial Lake Iroquois (also see figure 3). The ice margin lay on the northern side of the Mohawk Valley, forming the northern shore of the lake, with drainage eastward from Rome down the Mohawk River at an elevation of 148m (450'). **Diagram III** shows the area in diagrams I and II as it appears today.

bedded, well-sorted, fine grained sand held in place by pine forest. These very distinctive land forms have been set aside and preserved by the Nature Conservancy and the New York State Department of Environmental Conservation.

FIELD TRIP LOG

Many of the outcrops around New York State display beautiful and rare geologic features, and these outcrops should be conserved. No geologist should remove specimens from an outcrop "just for the sake of having a piece." Samples should be collected with an eye to a specific use and should be labelled and documented. Even researchers should think twice about whether removal of an exquisite sample is truly necessary. Talus should be collected whenever possible, in order not to damage exposures for future workers.

PLEASE ENCOURAGE YOUR STUDENTS TO PRACTICE OUTCROP CONSERVATION. Please never let them succumb to indiscriminate hammering. In fact, none of the field trip stops in this guidebook require hammers, so we would encourage you to bring only 1 hammer into the field – your own. Should you wish to make a documented collection for your school while you are on this trip, please let us know, and we'll assist you.

The road log begins at the intersection of routes 412 and 12B in the village of Clinton. All distances in the left margin are in miles. The stops in this road log lie on the following $7^1/_2$' quads: Boonville, Brantingham Lake, Clinton, Croghan, Glenfield, Lowville, Oriskany Falls, Port Leyden, Trenton South, Utica East, and Utica West. Geologic relationships are well portrayed on both the New York State Geologic Map (Rickard and Fisher, 1971) and on the New York State Geological Highway Map (Rogers *et al.*, 1990).

0.0 Go south on route 12B.
2.7 Kame delta on tree-covered hill to the east.
3.8 Hamlet of Deansboro; intersection of routes 315 and 12B.
5.7 Kame terraces on west side of road (see description under STOP 1).
6.2 Crevasse fillings on west side of road (see description under STOP 2).
8.5 Eastern Rock Products Quarry; good outcrops of Devonian Helderberg Group.
9.3 Junction route 26, Oriskany Falls; turn right at flashing light and proceed north on College Street.
9.4 Turn right in upper driveway of former Oriskany Falls School. This is STOP 1.

STOP 1: VALLEY HEADS MORAINE, ORISKANY FALLS

Looking south from this vantage point, one can see topography very different from that to the north in Oriskany Creek valley. At Oriskany Falls, a flat-floored valley gives way to irregular hills and hummocky topography. These hills and depressions are known as kame and kettle topography and are part of the Valley Heads Moraine. These deposits at Oriskany Falls are part of a large belt of recessional moraines marking a still-stand or readvance of the Wisconsinan glacier in New York State. The Valley Heads Moraine marks the southernmost extent of the ice sheet to be visited on this field trip. The Olean and Binghamton stages are also marked by similar deposits farther to the south, but the Valley Heads Moraine is the most distinctive belt of morainal deposits in central New York.

The hummocky terrain at Oriskany Falls formed by differential melting of sediment-laden ice at the glacier terminus. Ice advance from the north and ablation (primarily due to melting) must have reached equilibrium, and, for a time, the ice front neither advanced nor retreated. Judging from the size of the Valley Heads Moraine, this still-stand may have lasted for as much as 100 years.

The moraine loops across the valley, marking the lobate shape of the glacier. Some of the morainal deposits were created as dirty ice melted and released entrained sediment. This process is similar to the process that forms the piles of sediment one finds along roadsides or driveways when snowbanks melt in the spring. Unsorted, unlayered sediment marks the areas where snowbanks were piled in the winter. By analogy, piles of ablation till mark the terminus of the ice sheet and are composed of generally unsorted and unstratified sediment. Stagnant ice blocks trapped in the sediment later melt out, leaving depressions known as kettles (figure 5).

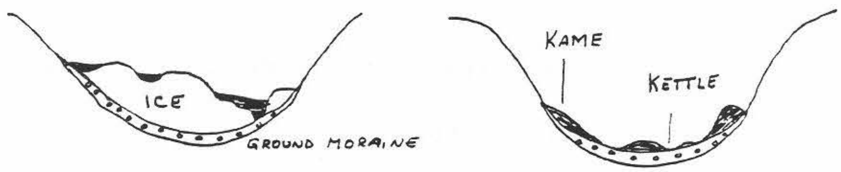


Figure 5. Formation of kame and kettle topography.

Many moraines contain considerable quantities of stratified drift in addition to unstratified ablation till. These stratified sediments are deposited in depressions on the moraine surface by running water — primarily meltwater from the wasting ice sheet. Within individual layers, sediments are typically moderately well-sorted, although grain size ranges widely from layer to layer. Many of the sediments are fluvio-glacial, and cut and fill structures and cross-bedding are common. While buried ice blocks remain intact, stratified drift occupies depressions in the morainal surface. Once buried ice blocks melt and produce kettles, the topography inverts, producing hills of stratified drift known as kames. The inversion process commonly disrupts stratification in the kame deposits. This hummocky topography is known as a kame and kettle moraine. Most kame and kettle moraines are also dissected by meltwaters that spread the eroded material down-valley in outwash plains. Were we to drive south of Oriskany Falls to route 20, we would be able to see the broad flat outwash valleys developed south of the Valley Heads Moraine. Scattered kettles occur on the outwash plain, and one of the best sits in the front yard of the Madison School on route 20.

In the valley to the north of Oriskany Falls, one can see crevasse fillings (at road log mile point 6.2). These kames of stratified drift were formed as sediment accumulated in large cracks or crevasses in the ice. When the ice melted, these sediments were left as en echelon ridges parallel to one another but at an angle to the valley walls. In addition, one can also see examples of kame terraces along the valley walls to the north (road log mile point 5.7). These formed along the sides of the stagnant ice, as sediment-rich meltwater and rainwater poured off the ice and hills separating the ice tongues. As with all kame deposits, these are composed of stratified drift. However, many of the kames show disrupted stratification, produced as ice melted and removed support from the sediment, causing slumping and faulting.

Return to College Street, and continue north. College Street becomes Skyline Drive.

16.2 Large glacial erratic on the west side of the road.

16.8 Good view to the west over Sconondoa Valley and Oneida Lake. Meltwater channel and delta complex can be seen south of Vernon Center in Sconondoa Valley.

17.8 Stop sign at intersection of Skyline Drive and College Hill Road. Turn right.

20.5 Flashing light at intersection of College Hill Road and route 233. Continue straight.

21.4 Traffic light at Clinton village square; follow 12B through Clinton.

23.7 Intersection of routes 5 and 12B; continue on 12B.

25.8 Intersection of routes 12, 5, and 12B. Go north on route 12; 4-lane highway starts.

31.5 Turn right into Riverside Mall, and proceed to the north end of the parking lot behind the Bradlee's store. STOP 2 is on the slope and field north of the parking lot.

STOP 2 - RIVERSIDE MALL

Here, we will see a very different type of Pleistocene deposit than the one we saw at Oriskany Falls. Sediments of the Valley Heads Moraine are ice contact deposits; the sediments at Riverside Mall are pro-glacial lake sediments, deposited in glacial Lake Iroquois or Amsterdam. The slope north of the parking lot is underlain by dark-colored, very fine-grained, finely laminated sediments. At the top of the slope, sediments are conspicuously coarser, composed of fine to medium sands with some pebbly layers.

The fine-grained sediments are varved clays, or rhythmites, composed of layers of gray clay alternating with organic-rich silty layers. Overall fine grain size indicates a low-energy environment, away from lake inlets and deltaic complexes. The layers represent seasonal accumulations, and each couplet (varve) represents one year of sediment accumulation. The silty layers form during rapid sediment influx and probably represent summer accumulation, when the ice is melting and releasing sediment to the lake. The clays accumulate during the winter, when the ice is delivering little sediment to the system and pro-glacial lakes are iced over. Winter accumulations consist only of very finely-divided material that settles very slowly out of the water column. Such material accumulates year-round but is swamped during the summer by coarser detritus fed into the system. Summer layers here are thinner than the winter layers, probably because of the short duration of the melting season.

Clays are notoriously unstable when wet—witness the periodic devastation caused by clay-lubricated mudslides in California. With the right amount of water, clays and silts are very susceptible to liquefaction when shaken. Vibration causes particles to lose frictional contact with one another, and the mass becomes temporarily liquified and has no shear strength. Once vibration stops, the mass regains its shear strength. A close look at the mall parking lot will reveal many humps and swales, and even more numerous patches. The parking lot is underlain by varved clays and silts. When moisture content is right, vibration from vehicles induces some liquefaction, and the parking lot surface deforms in response to flow of material beneath it. A good drainage plan would have prevented much of the problem.

As you walk northward up the slope from the parking lot, you will notice a distinct change in sediment from varved clays to fine-grained sands. These sands were deposited in an environment of higher energy than that in which the varved clays were deposited.

Sediments at this stop exhibit a nice vertical facies change, indicating a change in lake depth with time. In a typical lake, the coarsest sediment settles in shallow, high energy water nearest the shore and lake inlets. As coarse sediment settles out, only finer material passes into deeper water, and only clays and organic material accumulate in the deeper portions of the lake. The change upslope at Riverside Mall from varved clays to sands indicates shallowing of water with time, either by progradation of a delta or by general lowering of lake level during ice retreat.

In some of the glacial lakes (especially Lake Iroquois), carbonate-producing animals and plants flourished, and layers of marl occur between the coarse sand zones and the silt zones. Marl is a calcareous sediment; in this area, marls are typically rich in shells of tiny gastropods and pelecypods and remains of a calcareous plant known as chara.

Return to route 12 and proceed north. Route 12 climbs up Deerfield Hill and traverses a series of Pleistocene beach terraces and deltas.

42.7 Junction of routes 12 and 28; stay on route 12.
51.4 Well-developed Pleistocene ground moraine on west side of road.
54.4 Junction of routes 12 and 28; stay on route 12.
56.6 Nice field of Pleistocene erratics to west of road.
59.9 Well-sorted Pleistocene fluvio-glacial sands to west of road.
61.2 The large, elongate hill east of the road is Park Hill and consists of dissected Pleistocene outwash. Several sand and gravel operations exploit the deposit.
61.4 Intersection of routes 12 and 12D; stay on route 12.
62.4 Traversing bottom of glacial Lake Port Leyden.
69.0 Crossroads in downtown Port Leyden; continue straight.
74.2 Glacially-polished bedrock knobs east and west of the road.
76.8 The horizon to the east across the Black River Valley has a very uniform elevation. That straight horizon corresponds to a well-developed, flat topographic surface at an elevation of about 1200-1250' that shows up very well on the topographic maps of the Brantingham and Port Leyden quads. These flat surfaces are delta tops, formed as sediment was deposited into glacial Lake Port Leyden during the last part of the Pleistocene. Glacial Lake Port Leyden must have been at least as deep as the tops of the deltas; it was no small lake. A look at the topographic maps will show many closed depressions on the delta tops; these were presumably formed as grounded ice, partially covered with deltaic sediments, melted and caused subsidence. The delta slopes have been somewhat dissected since. Glacially-sculpted Deerlick Road lies in the Black River Valley east of the road; it appears to be a roche moutonée.
77.3 Good exposure of glacially-polished bedrock east of the road.
81.5 Route 12 crosses Roaring Brook.
85.7 Junction of routes 26 and 12; continue north on 12.
86.3 Junction of routes 12 and 812. Turn right on 812 at the light.
86.9 Stop sign. Turn left on route 812 and proceed north.
89.8 Floodplain of the Black River, which floods with great regularity.
91.6 Downtown New Bremen.
95.3 Pleistocene boulder fields on west side of road.
95.7 STOP 3, at corner of Brewery Road and route 12, south of Croghan.

STOP 3: CROGHAN ROADCUT (and that's no baloney!)

The rocks in this spectacular roadcut are part of the suite of metasedimentary and metaigneous rocks that make up the bulk of the Adirondacks. These rocks were deformed and metamorphosed during the Grenville Orogeny (1020-1100 Ma), at depths of about 25km below the surface, temperatures of 700-720°C, and pressures of about 7 kilobars. They are quite thoroughly cooked. They are also deformed, and all lithologies in this outcrop, including the ones of igneous parentage, show compositional layering and/or planar alignments of minerals (foliations) acquired during deformation.

There are 3 major lithologies present in this roadcut 1) a mildly-foliated, coarse-grained pink granitic gneiss consisting of pink K-feldspar, plagioclase, quartz, and variable amounts of hornblende with accessory magnetite, biotite, apatite, and zircon, 2) a foliated, very coarse-grained pink augen ("eye") gneiss, with enormous pink feldspar grains up to 2-6cm long set in a matrix of plagioclase, hornblende, and quartz, and 3) well-foliated, banded gneisses composed of black, pyroxene-biotite and hornblende-biotite gneisses interlayered with pink to gray granitic and augen gneisses. The dark-colored gneisses are composed of biotite, plagioclase, and hornblende or pyroxene, with accessory apatite, zircon, and magnetite. The light-colored interlayers are similar in composition to the granitic gneisses in the outcrop.

In addition to these 3 major lithologies, this outcrop contains minor amounts of amphibolite, biotite-oligoclase augen gneiss, quartzofeldspathic hornblende gneiss, and biotite-plagioclase schist all interlayered with granitic gneisses and granitic augen gneisses.

The foliation is defined differently in each of the 3 main lithologies. In the granitic gneiss, the foliation is defined both by the parallel alignment of dispersed biotite and by alternating layers of different grain size (but similar composition). In the augen gneiss, the foliation is defined by large, flattened and streamlined K-feldspar grains outlined by ribbons of quartz. The foliation defined by the flattened grains is enhanced by an irregular compositional layering defined by lenses of pink K-feldspar (commonly strung out into layers) alternating with stringers of quartz, plagioclase, and hornblende. In the banded gneiss, foliation is defined primarily by alternating layers of different compositions, one light-colored and rich in quartz and feldspar, the other dark-colored and rich in mafic minerals. Within the dark-colored gneiss, foliation is enhanced by the parallel alignment of biotite. None of the 3 major lithologies has enough mica in it for the foliation to also be a good rock cleavage direction. As a result, all of the major lithologies are blocky and tough.

Although these rocks are metamorphic, the outcrop displays many classic relict igneous features. The granitic gneisses were evidently originally granitic magmas that thoroughly and intimately invaded a country rock of dark-colored gneisses. The dark-colored gneisses, then, might well be xenoliths, remnants of the country rock caught up during intrusion of granitic magmas. The great variety of grain size amongst the granitic gneisses, and the complicated multiple cross-cutting relationships attest to a period of multiple intrusions. Evidence in this outcrop also suggests that deformation was going on as the magmas were being intruded (*i.e.*, that the granitic magmas were syntectonic). Some folds in the dark-colored gneisses are truncated by the granitic gneisses, implying intrusion after folding. On the other hand, some granitic gneisses interlayered with the dark-colored gneisses are folded right along with the dark-colored gneiss. In fact, some of the extremely flattened augen in the augen gneisses are folded. Both suggest folding *after* granite emplacement. Folding after and folding before granite emplacement implies protracted and syntectonic intrusion. Because all units are foliated, deformation must have outlasted the intrusive episodes.

The other relict igneous features in this outcrop are the augen in the augen gneisses. The augen gneiss may also be referred to as a megacrystic gneiss, in allusion to the enormous size of the deformed feldspar crystals that make up the augen. The metacrysts or augen are very likely relict phenocrysts formed during early slow cooling of a granitic magma with a two-stage cooling history.

As suggested in the previous paragraph, the protolith (pre-metamorphic rock type) for the granitic gneiss was probably a granite. Metamorphism did very little to its original mineral composition, because temperatures during metamorphism were nearly as high as the temperature of a granitic melt at those depths. In other words, the minerals in the granite were perfectly happy under the conditions of metamorphism. Even though the overall mineral content did not change appreciably during metamorphism, individual mineral grains did become unstable and recrystallize in new orientations in response to deformation in the rock. In this manner, the granite acquired a foliation of recrystallized and aligned biotite and became a granitic gneiss. The foliation defined by interlayers of different grain sizes is likely due in part to inheritance of primary textural differences and in part to deformation.

The protolith for the augen gneiss was probably a porphyritic granite. Its metamorphic minerals are not substantially different from those in the protolith, but deformation and recrystallization produced a well-developed foliation in the rock. If you look carefully at the augen gneisses in this outcrop, you will see that there is a range in texture from blocky pink feldspars set in a poorly-foliated matrix to stringers of pink feldspar set in a well-foliated matrix. These differences in texture undoubtedly result from differences in intensity of deformation. As the amount of shearing in a layer increases, the feldspar tabs flatten. Strain is highest along the margins of the grains, particularly at the corners; the edges and corners of the feldspars begin to recrystallize into networks of tiny, stable feldspar grains (figure 6). These tiny grains are then no longer part of the larger feldspar tab - they have been, in effect, transferred to the matrix. Plastic flattening of the grain, then, is accompanied by some grain size reduction and a change of grain shape from blocky to eye-shaped. Some of the large feldspar grains are also fractured, if plastic deformation in the grain can't keep up with deformation in the rock. Changes are also happening in the matrix minerals. Platy and prismatic minerals such as biotite and hornblende recrystallize parallel to the flattening plane of the feldspar tabs, creating a foliation wrapping around the developing augen. Coarse quartz grains become flattened, ribbon-shaped grains oriented parallel to the foliation and wrapping around the augen. Many of these deformed quartz grains undergo the same kind of dynamic recrystallization that affects the margins of the feldspars. New, tiny grains of quartz form from the old strained grains. Quartz is very susceptible to the dynamic recrystallization process and undergoes grain size reduction much more readily than feldspars. Therefore, when you look at the augen gneiss, realize that some of the disparity in grain size between augen and matrix was produced during deformation. In fact, many augen gneisses of the world began life as uniformly coarse-grained rocks; not all augen gneisses had porphyritic parents.

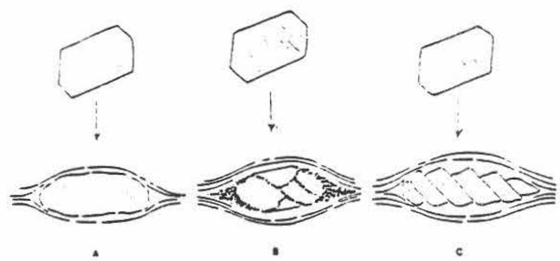


Figure 6. Formation of augen by crystal plastic processes (a), dynamic recrystallization (b), and brittle fracture (c). These processes need not operate separately.

Determining the protolith for the dark-colored gneisses is somewhat more problematic, and both igneous and sedimentary protoliths are possible. The dark-colored gneisses are intimately interlayered with light-colored gneisses and augen gneisses to form packages of conspicuously banded gneiss. Questions concerning the origin of the banding cannot be answered unequivocally. The fact that many of the light-colored layers can be traced to larger masses of granitic gneisses suggests that at least some of the banding may be intrusive, the result of intimate interfingering of dark gneiss and granitic magma (a lit-par-lit gneiss). Layers that cannot be traced to a granitic mass may be either light-colored layers in the original country rock, or granitic material derived very locally by partial melting of the dark-colored gneiss and creation of migmatites. Considering that metamorphic temperatures were very close to the minimum melting temperatures of many types of metasediments, the ultimate source of the large volumes of granitic magma in the Croghan area was very likely the metamorphic complex itself. Local partial melting of the dark gneisses may have taken place as well.

After viewing the Precambrian geology, let's jump ahead about a billion years to see what effect the Wisconsinan ice sheet had on this locality. The top of the outcrop is a smoothed and polished surface, produced as glacial ice with entrained sediment sand-papered and rounded the hard Precambrian rock. Some differential erosion can be seen along zones of dark gneiss. The grooves and shallow potholes suggest that the outcrop may not only have been glacially scoured but also eroded by running water. Marginal or subglacial meltwaters, particularly related to a stagnant ice mass, are good candidates.

Proceed eastward over the top of the outcrop to a low exposure of Precambrian bedrock about 30m away. This outcrop displays distinctive scratches and grooves (glacial striae) trending S10-20°W, parallel to the last ice movement direction in the area. Looking south from this outcrop, you will see a hill with a sand quarry. This hill is part of a series of hills extending the entire length of the Lowville quadrangle. The line of hills trends north-south, with a slight dog-leg near this stop. The deposits that underlie these elongate hills are largely unconsolidated sediments dating from the late Pleistocene. As discussed at stops 1 and 2, Pleistocene deposits in this area include ground moraine (both ablation and lodgement till), morainal ridges (typically composed of till overlain by patchy stratified drift), kame terraces and crevasse fillings (stratified drift), eskers (stratified drift), outwash (fluvial deposits), and pro-glacial lake deposits (beach and deltaic deposits, varved clays, etc.). Each of these deposits has distinctive characteristics, and examination of the deposits in this quarry should allow us at least to rule out some of the possibilities. This deposit was originally described by Buddington (1934) as a moraine. However, close examination shows structures consistent with sediment deposited by running water, rather than directly by melting ice. If these deposits are ice contact sediments, deposition was dominated by fluvioglacial processes.

Sediments in the quarry are conspicuously stratified. Scour and fill structures are common. Cross bedding occurs at many different scales, but many layers exhibit planar laminae. A majority of the deposit consists of sand-sized grains, although there are some layers of very fine grains as well as some layers of pebbly sand. Individual layers are typically moderately well-sorted to well-sorted. The characteristics of sorting and stratification in these sediments indicate deposition from a moving current; direct ice deposition does not produce stratified and sorted sediments. Therefore, these sediments are not tills.

Cut and fill structures, cross-bedding, and some pebble imbrication are consistent with fluvial deposition. At this stop, stratified sediment lies directly on water-scoured Precambrian bedrock; there is no intervening morainal till. It appears, then, that at least this portion of the deposit does not represent a veneer of stratified drift over ridges of morainal debris.

Some of the horizons in the quarry consist of thinly laminated find sand and silt. Cross-bedding is absent, and individual layers are laterally extensive and easily-traceable. These may represent sediments deposited in closed depressions such as small ponds. Some of these fine-grained layers contain cobbles that may well be dropstones.

The fact that these deposits are part of an elongate but not sinuous ridge system suggests that these deposits are not part of an esker, a pro-glacial lake, or an outwash plain (although dissected outwash is a possibility). A stratified ice contact deposit is the most logical choice - perhaps an ice-marginal deposit or a very large crevasse filling in stagnant ice. The fact that slopes on the west sides of the hills are steeper than those on the east sides suggests that the main ice contact side of the deposit may have been on the west.

Turn around, and return south on route 812 towards Lowville.

97.4 Spectacular view to the west of the Tug Hill Plateau, underlain by Middle and Upper Ordovician sedimentary rocks.

99.8 New Bremen bridge.

101.0 An even better view of the Tug Hill Plateau.

103.3 STOP 4. Pull off the road across from the Dadville Quarry.

STOP 4: DADVILLE QUARRY

After repeated searches, neither of us has ever found an exposure of the Precambrian/Paleozoic unconformity (Plate I) in the Black River Valley, although such exposures have been reported in the literature (Miller, 1910). The unconformity *is* spectacularly exposed in roadcuts near Kingston, Ontario, but that's a bit far away for this field trip. The outcrops at the Dadville Quarry constrain the position of the unconformity better than almost anywhere in the Black River Valley, and we'll have to settle for that.

Precambrian granitic gneisses crop out on the north side of the road in front of the Highway Department Building. Standing on the Precambrian, one can look east and see many knobs of Precambrian rock scattered throughout the Black River Valley and nearly buried by Pleistocene sediments. Directly southeast across the road, Middle Ordovician limestones of the Black River Group are exposed in an active quarry. The trace of the unconformity must lie just below the quarry floor, swinging around the base of the hill and across the road west of where we are standing. From the base of the hill, the unconformity dips very gently (only a few degrees) to the southwest.

Why can't we see the unconformity in the quarry? Surely they must have quarried through to the Precambrian in places, you think. In point of fact, they haven't quarried to the Precambrian, and a look at the stratigraphic column in Plate I will tell us why. The quarry owners are interested in limestones, and the Black River Group has good limestones in all but the Pamelia Formation at the bottom of the Group. As soon as they reached the Pamelia, they stopped quarrying.

A look at the topographic map would show that the position of the unconformity is marked here by a distinct break in topography. As we continue to drive south, you might find it interesting to check the topographic map for other breaks in slope, and examine the countryside and outcrops to see whether those breaks in slope mark the Precambrian/Paleozoic unconformity or not.

104.0 Continue south on route 812. The houses on the left are perched on a bench in the Paleozoic that marks the contact between the Black River Group and the Trenton Group, both of Ordovician age. That bench and contact can be traced south on the topo map from stop 4. The next major bench to the west lies at a stratigraphic horizon within the lower portion of the Denley Limestone.

104.4 Turn right on E. State Street (route 812).

105.0 Junction routes 12 and 26; turn left, and go south on routes 12 and 26.

105.6 Bear left at the Y, and proceed south on route 12.

109.6 Crossing Roaring Brook. Precambrian basement rock is exposed in the field to the east of the bridge, and Paleozoic sediments crop out immediately west of the bridge in the stream channel. The unconformity must lie somewhere between the 2 exposures, although the actual contact is not exposed at the surface.

111.7 Rounded knobs of Precambrian gneisses dot the field to the east of the road; Paleozoic sediments begin to the west of the road at the first topographic rise.

Town of Glenfield to the east. The topo map shows how well the topography reflects the geology. The base of the first sharp slope west of Glenfield lies at the Precambrian/Paleozoic unconformity. The first bench marks the top of the Black River Group. The second major bench marks a horizon within the lower Denley Limestone. The third bench marks the top of the Denley, and the 4th bench marks the top of the Steuben Limestone. The base of the major slope at the Whetstone Gulf Campground marks the contact between the Steuben Limestone and the Utica Shale. The last bench lies at the contact between the Utica Shale and the Lorraine Group (equivalent to the Frankfort Shale).

111.9 Pleistocene glacial deposits mask most of the rocks near the unconformity.

117.1 Route 12 lies on Precambrian bedrock; an outcrop of the Pamelia Formation (lower Black River Group) lies approximately 50m west up Snugsboro Road.

118.8 Outcrops of the lower Black River Group line the lower part of Turin Road to the west.

119.3 Between the last mileage point and this one, the road has crossed the unconformity; outcrops of the Lowville Formation (Black River Group) lie immediately west of the road.

119.6 Junction with 12D; continue south on route 12.

119.9 From here south, the railroad tracks approximately follow the unconformity.

120.4 Outcrop of Precambrian granitic gneiss east of road; glacial deposits blanket much of the Precambrian in this area.

122.3 Crossroads at Port Leyden. Continue south on route 12.

124.1 Precambrian diopside-quartz-phlogopite gneiss on the east side of the road. The protolith for these metasedimentary units was probably a dirty carbonate, such as a sandy dolomstone.

125.5 Historic marker for the Black River Canal. The Canal was completed in 1855 and linked Carthage and Rome. In a 35-mile stretch through the steepest terrain, workers constructed 109 locks. The locks were built with limestones quarried from the Black River Group, as were many of the dove-gray stone buildings in the North Country.

126.5 STOPS 5 AND 6. Park well off the road north of the bridge. Watch your socks.

STOP 5: SUGAR RIVER, EAST OF ROUTE 12

THIS IS PRIVATE PROPERTY. You *MUST* obtain permission from Barrett Paving Company (the quarry owners) if you wish to examine the rocks along the Sugar River. The owners have been very generous in the past in allowing groups to visit the exquisite features on their property. However, repeated unauthorized entry by inconsiderate geologists to properties at many other places around the country has rightfully angered many property owners. As a result, many classic localities are now completely off limits, even to those who properly seek permission. Don't take a chance on spoiling the Sugar River spot for everyone else. Be *sure* to get permission. Encourage others to do the same.

ALSO - BEWARE THE UNUSUALLY LUSH AND PREDATORY POISON IVY.

These beautiful gray limestones are part of the Watertown Formation at the top of the Black River Group (Plate I) and were deposited as a subtital[1] sequence on the North American passive continental margin during the Early Middle Ordovician. At several stops on this field trip, we will be examining carbonate sequences. Limestones and dolostones, composed of calcite and dolomite respectively, exhibit a wide range of textures and structures, but all share the common characteristic of being *biogenically-derived*. Very few post-Precambrian carbonates are pure, non-biogenic chemical precipitates. In addition, one should keep the distinction in mind between detrital clastic rocks such as sandstones and shales, whose fragments are derived from weathering of a *distant* source, and carbonates, whose fragments are derived very locally by organic precipitation of calcite. Neither animals nor plants precipitate dolomite, and virtually all dolomite in the sedimentary record results from replacement of calcite by its magnesium-rich counterpart, dolomite.

In order to help unravel the environment of deposition of a carbonate rock, it is useful to know in what form the calcite or dolomite occurs in a rock. Calcite can occur in a carbonate rock as skeletal debris, carbonate mud, sparry calcite, intraclasts, and pellets; dolomite can occur as early dolomite or secondary dolomite.

- *Skeletal debris* consists of the remains of organisms, sometimes well-preserved in life position and not transported very far. Skeletal debris commonly consists of broken and disarticulated fossils and may be as fine as silt-sized particles.
- When communication by currents or biologic activity reduces fossil fragments to mud-sized particles, the material is called *carbonate mud*. Lithified carbonate mud is referred to as "micrite" – microcrystalline calcite.
- *Sparry calcite* is coarsely crystalline calcite precipitated as a secondary mineral in pores and burrows. Some

[1] *subtital*: almost always below low tide; *intertidal*: almost always between high and low tide; *supratidal*: almost always above high tide.

skeletal carbonates have a micritic matrix, others have a sparry matrix, depending upon whether the environment winnowed out very fine carbonate particles at the time of deposition.

- *Intraclasts* are non-skeletal fragments >.2mm in size and associated with erosional intervals. They represent fragments of previously-deposited and partly-consolidated sediment (very locally-derived, of course).
- Many *pellets* are probably intraclasts <.2mm in size, although some may well be fecal pellets.
- *Early dolomite* commonly occurs in thin laminae and forms by migration of magnesium-rich waters upward through the sediments, followed by penecontemporaneous replacement of calcite by dolomite. Early dolomite is almost always found in supratidal settings.
- *Secondary dolomite* forms well after deposition of the sediment and is highly variable, filling vugs, replacing calcite, or cementing clastic rocks.

A given carbonate rock may be composed of any combination of the ingredients listed above. Which components are combined and how they are combined governs the appearance of any individual carbonate rock. The word "limestone" describes anything from a hash of boulder-sized fossil fragments to a featureless, ultrafine-grained micrite. In self-defense, sedimentologists have derived a whole raft of esoteric terms to convey more than simple "limestone". The following definitions should help you to wade through most field guides:

calcirudites: carbonate equivalent of conglomeratic grain size
calcarenites: carbonate equivalent of sandstone grain size
calcisiltites: carbonate equivalent of siltstone grain size
calcilutites: carbonate equivalent of mudstone grain size
micrite: limestone formed from a carbonate mud
biomicrite: limestone formed from skeletal debris & carbonate mud
biosparite: limestone formed from skeletal debris & sparry calcite

Grain size is generally indicative of the energy of the environment, although one must bear in mind that large *fossils* do not by themselves indicate a high energy environment, because the animals *lived* in the environment and were not transported there.

If exposed to periodic exposure and dessication, carbonates may show mudcracks and local erosion surfaces. If deposited by currents, carbonate sediments may be cross-stratified. If burrowing organisms lived and fed in the carbonate substrate, the sediment may show burrows (horizontal or vertical). Active bioturbation (i.e., churning of the sediment by burrowing organisms) may completely destroy any fine-scale depositional structures such as bedding or lamination.

The Watertown Formation is a very dark gray limestone with fossils and fossil fragments floating in a very fine-grained matrix. The matrix is micritic, and "clumpiness" of the micrite has prompted Walker (1973) to suggest that the micrite may have been pelletal. Bioclasts (both unabraded fossils and abraded fossil fragments) generally make up 10-30% of the rock, although percentages range as low as 4% and as high as 53% (Walker, 1973). The more fossiliferous lithologies would be properly called biomicrites. In all cases, bioclasts "float" in a matrix of micrite. Black chert nodules are common throughout the unit.

Stratification is virtually absent in the Watertown Formation, because the carbonate muds were completely churned by burrowing organisms who ate their way through the soft sediment. Much of the pelletal texture in the micrite may have originated as fecal pellets excreted as these organisms reworked the sediment. Networks of horizontal burrows thoroughly mottle this limestone and show up well on weathered horizontal surfaces. Mudcracks are absent, as are intraclasts and erosional intervals. Sparry calcite is confined to burrows.

Fossils in the Watertown Formation are illustrated in figure 7 and include:

rugose (horn) corals (*Lambeophyllum*)
tabulate corals (*Foerstephyllum*)
stromatoporoids (*Stromatocerium*)
echinoderm fragments
straight-shelled nautiloid cephalopods (*Actinoceras* & *Endoceras*)
bryozoans (*Eridotrypa*)
calcareous algae (*Hedstroemia*)
rare brachiopods (*Dalmanella* & *Strophomena*)

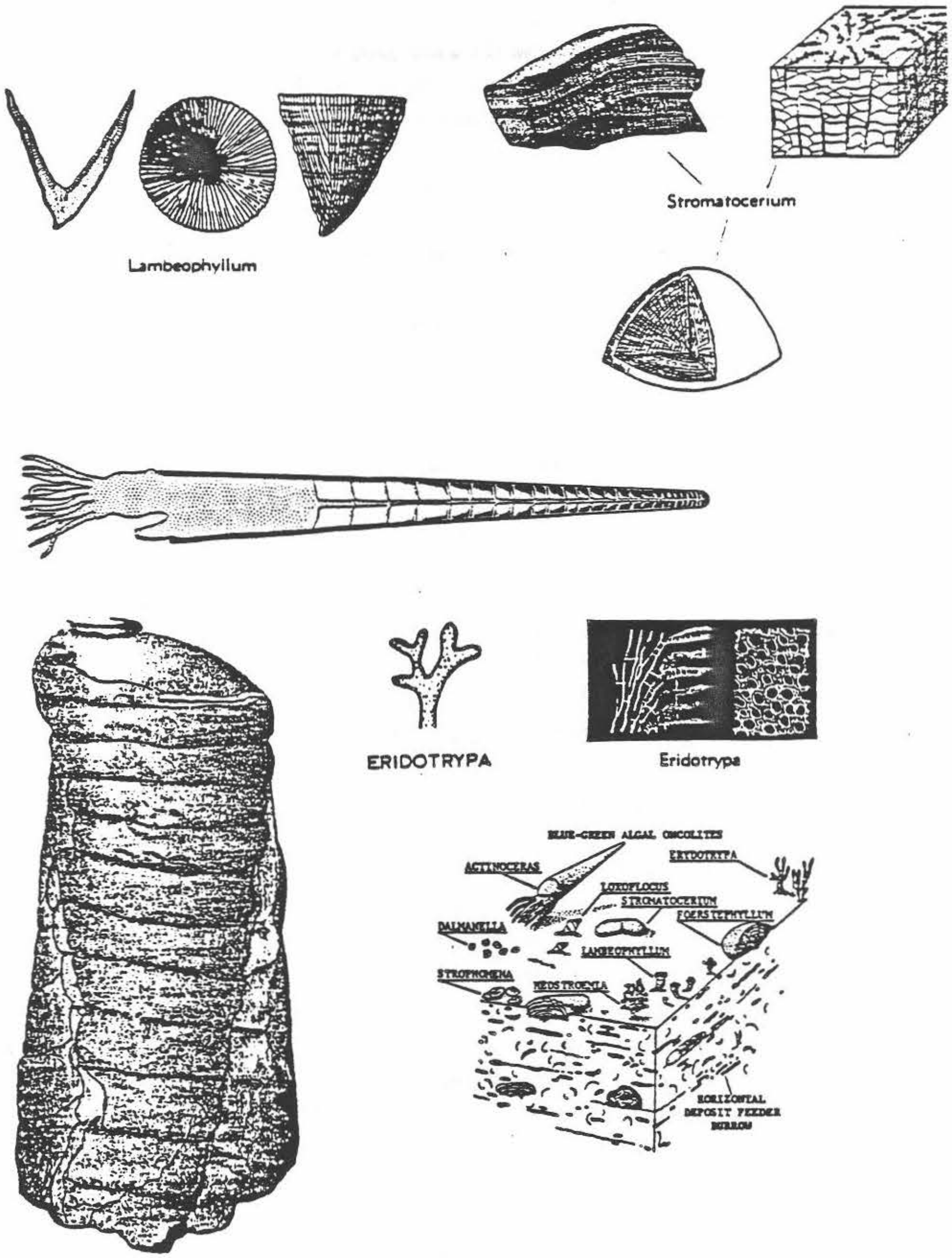


Figure 7. Fossils from the Watertown Formation. Diagrams from Cameron et al., 1972; Moore et al., 1952; Tasch, 1980; Titus, 1977; Walker and Laporte, 1980.

rare coiled gastropods (*Loxoplocus*)
horizontal burrows of deposit feeders

How high was the energy of the environment in which the Watertown was deposited? The fact that the rock is micrite-dominated suggests that the environment was a quiet, low-energy one, generally incapable of transporting anything but fine to very fine grains. How, then, do we explain the abundant sand-sized bioclasts (which can be felt by rubbing a hand over a weathered surface - they feel like sand scattered in a nice coat of varnish)? Walker (1973) has suggested that most of the sand-sized bioclasts are porous echinoderm fragments. Porosity would reduce their densities and allow them to "behave" as smaller particles would in a current. Okay, but then how do we explain the very large fossil fragments, like the 3-meter-long nautiloids? Large fossil fragments can easily accumulate in a very fine-grained matrix if the animals lived on the bottom or died in the water column above and sank to the bottom. Currents are not required to move them to their sites of accumulation. One would expect that these large pieces might not be significantly abraded, although they may well be broken if someone had dined on the carcasses before they were buried. Lastly, how do we explain the presence of large chert nodules? It turns out that chert nodules are a secondary phenomenon and are formed in a sediment after deposition. Solutions rich in silica soak through the sediment and replace portions of the rock with silica, creating chert nodules. In a carbonate rock, it's not immediately obvious where the silica should come from. In the Black River Group, silica may have come from leached beds of volcanic ash or from solution of fragments of siliceous sponges. Volcanic ash is a good candidate, because thin volcanic ash layers blown from eruptions in the approaching Taconic Island Arc occur sporadically throughout the upper Black River Group.

What clues have emerged regarding the environment of deposition of the Watertown Formation? Fine grain size of the matrix points to a dominantly low-energy environment. Lack of mudcracks, intraclasts, and erosional intervals suggests that the bottom was always below low tide (a subtidal environment). Lack of deep burrowers supports this, suggesting that organisms did not need to burrow deeply to escape a hostile environment. The fauna consists of animals related to those in the Recent that require normal marine salinities (Walker, 1973) and prefer not to be bothered by daily tides. Modern calcareous algae require water depths of less than 80m in clear water in order to photosynthesize, and Walker (1973) suggests that carbonate mud might have made the water turbid enough during deposition of the Watertown that 10m might have been the maximum depth for photosynthesis. If we put this all together, we come up with an offshore shallow marine environment with good circulation and rare turbulence. The area would have normally been below wave base, but rare storms likely stirred up the bottom. The water would also likely have been warm, because central New York State was located about 20° south of the equator during the Ordovician. A pleasant place to live.

The limestones along the Sugar River display spectacular solution features, particularly in the massively-bedded Watertown Formation. One is first struck by the smooth, rounded, doughy-looking outcrops, and then by solution enlargement of joints, producing huge blocks of limestone along the streambed. Farther downstream, a magnificent train of potholes lines the stream channel, many containing pothole stones. Solution is as important as abrasion in creating these potholes. The most spectacular solution features, however, are the swallow holes. During a dry summer or fall, the Sugar River vanishes completely into swallow holes and never reaches the confluence with the Black River. However, during high water, a decrease in volume is difficult to see. Some of the lost water emerges as substantial springs along the face of the quarry. A look at the quarry face will suggest to you that the subsurface flow takes place through solution-enlarged joints, many along bedding surfaces, rather than through gigantic subterranean tunnels. DO NOT WALK TO THE EDGE OF THE QUARRY FACE - MUCH OF IT IS OVERHUNG AND UNSAFE. If you walk a short distance northeast, you will find the outlet for most of the swallowed Sugar River water — the Little Sugar River heads in a spectacular blind valley just easy of the quarry. The blind valley shows up well on the topographic map.

Glacial erratics dot the Black River Valley, and there is a *very* large granitic gneiss erratic smack in the middle of the Sugar River channel. The Sugar River probably makes headway in moving it about once every 200 years or so. (By the by, why aren't there any glacial striae on the bedrock at the Sugar River??)

STOP 6 - LOWER TRENTON GROUP, SUGAR RIVER

The fine-grained, fossiliferous limestones and interlayered shales in this outcrop are part of the lower portion of the Trenton Group (Plate I) and were deposited in shallow Middle Ordovician seas at a time when the Taconic Island Arc Terrane had just begun to collide with the continental margin of North America.

The lower 2/3 of the outcrop belongs to the Napanee Limestone, the upper 1/3 to the Kings Falls Limestone. Both are part of the Trenton Group. Geologists who have studied the Trenton have chosen to divide the sequence into a total of 6 formations, only 2 of which are exposed here, and it is interesting to consider why they have done so and on what basis. Rock sequences are divided into formations within which rock types show some degree of internal homogeneity or sequence that distinguishes them from adjacent sequences. Some genetic connotation is implied, and one generally views the conditions or range of conditions of formation for a particular formation to be different from those in adjacent units. If they weren't, they'd all be part of the same formation. Formations must also be thick enough to be mappable at a reasonable scale. Some formation boundaries are obvious. For instance, there is a conspicuous difference between the Napanee Limestone and the Watertown Formation, which we just visited. Anyone would put them into different formations. However, we did not see the upper Watertown or lower Napanee. If they grade into one another, you might not place the formation boundary at the same place I would, and there's nothing wrong with that. Other formational boundaries are not so obvious. Take the one in this outcrop, for instance. The formation boundary between the Napanee below and the Kings Falls above separates thinly interbedded limestones and shales from thickly interbedded limestones and shales. One's first reaction is that there's not that much difference between the 2 units and that it all ought to be one formation. But, when we examine the details, we find that the limestone in the Napanee is dominantly very fine grained and sparsely fossiliferous, while the limestone in the Kings Falls is coarse and very fossiliferous. The subtle change in thickness of beds also marks a change in the environment of deposition, and it *does* make sense to place a formation boundary between the two.

The limestones in the Napanee are primarily sparsely fossiliferous micrites composed dominantly of carbonate mud (some pelletal) with infrequent, discontinuous skeletal laminae. There are some highly fossiliferous micrites (biomicrites), with many fossils and fossil fragments set in a carbonate mud matrix, and rare biopelmicrites, a fossil and pellet hash cemented by secondary calcite. The limestones are interlayered with black, sparsely-fossiliferous calcareous shales. In general, the coarser, fossiliferous limestones increase in abundance upward (Cameron *et al.*, 1972). The coarser layers show some ripples and cross-stratification, but none of the layers contains mudcracks. The limestones display some vertical burrows, but they are not thoroughly burrow-mottled, as the Watertown Formation is.

In contrast, limestones in the Kings Falls are primarily shelly limestones and fossil hashes; matrix material is dominantly sparry rather than micritic. These limestones are referred to as biosparites or shelly calcarenties. Layers of thinly bedded calcareous shale alternate with layers of limestone, as they do in the Napanee, but limestone layers are considerably thicker in the Kings Falls than in the Napanee. Limestones in the Kings Falls show abundant high energy features, including cross-stratification, ripples, erosional surfaces, intraclasts, and an absence of micritic material (Titus and Cameron, 1976). These limestones accumulated as a collection of fossil fragments in an environment that winnowed out carbonate mud; spaces between fossil fragments were later filled with secondary, coarsely crystalline (sparry) calcite. A close look at a talus block from one of the Kings Falls layers will show you reflections from calcite cleavage surfaces 1mm or more across - these are calcite grains of the secondary sparry matrix.

The fauna is brachiopod-dominated — about 1/3 of the species present are brachiopods (Titus, 1977). Fossils in the Napanee Limestone are illustrated in figure 8 and include:

brachiopods (*Triplesia, Dalmanella, Sowerbyella, Rafinesquina, Strophomena, Paucicrura, Hesperorthis*)
broyzoa (*Prasopora, Stictopora, Amplexopora*)
crinold fragments
trilobites (*Flexicalymene, Isotelus, Ceraurus*)
snails - particularly in the lower Kings Falls (*Sinuites, Liospira, Subulites, Hormotoma, Loxoplocus, Phragmolites*)
ostracodes
coral - in the lower Kings Falls (*Lambeophyllum*)

The fauna in these 2 formations is dominated by low filter feeders (such as brachiopods and bryozoa), with some grazers (such as trilobites, gastropods, and ostracodes).

What clues have emerged regarding the environment of deposition of the Napanee and Kings Falls Limestones? The absence of mudcracks and other evidence of dessication in both formations, taken together with the abundant remains of marine organisms (few of whom could have managed a twice-daily low tide), suggest a subtidal environment. The dominance of sparry matrix in the Kings Falls and micritic matrix in the Napanee indicates that the Kings Falls was deposited in a high energy environment that winnowed carbonate mud, while the Napanee was deposited in a quieter environment that enabled carbonate mud to accumulate. However, presence of shaly laminae in the Kings Falls suggests that quiet water prevailed from time to time. The presence of cross-stratification, ripples, and erosional intervals, particularly in the Kings Falls, indicates that the environment was frequently above wave base. This conclusion is supported by the presence of fossil hash (coquina) layers, particularly in the Kings Falls. These layers are composed of disarticulated and broken skeletal materials, transported to the site of deposition by currents. Taken together, the evidence indicates deposition in shallow water, in a high subtidal

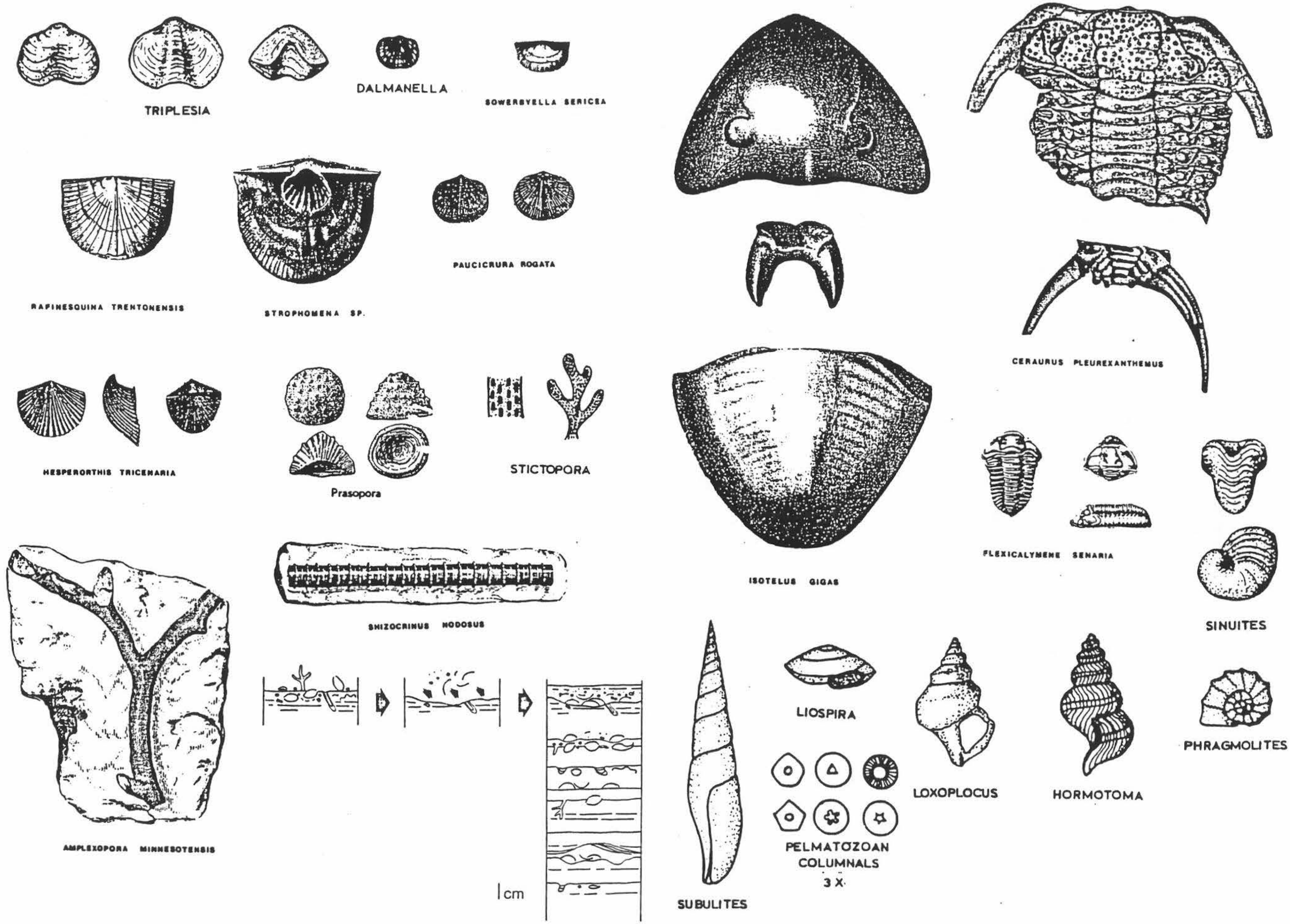


Figure 8. Fossils from the Lower Trenton Group. Diagrams from Cameron er al., 1972; Moore et al., 1952; Titus, 1977; Tasch, 1980.

environment. The Napanee was probably deposited in a shallow, shelf-lagoon, while the coarse, sparry Kings Falls was deposited in a waveswept offshore shoal (Cameron *et al.*, 1972; Titus and Cameron, 1976). Both were deposited in water shallower (or at least more agitated) than that characteristic of the Watertown Formation, which we just visited.

The repetitious interlayering of limestone and shale is an interesting puzzle. D.W. Larson (pers. comm., 1984) suggests that the interlayering is a storm-generated phenomenon, with the following scenario. Clay mixed with some carbonate may well have been the steady-state, everyday, slowly-accumulated sediment. The product would have been a calcareous clay mud deposited in a quiet-water environment incapable of transporting the silt-sized to coarse sand or pebble-sized carbonate grains and fossil fragments characteristic of the limestone layers. During storms, energy in the environment would have increased dramatically, bringing coarser carbonate detritus in from adjacent, higher-energy environments, and redistributing carbonate material previously deposited locally. The energy during storms would have been enough to winnow clays at the sediment/water interface, and carbonates of silt size or greater would have accumulated relatively rapidly. If you look carefully, you might even be able to convince yourself that some of the limestone layers in the Napanee are graded, with coarse fossil fragments decreasing in both abundance and size upward from the base of a layer.

In this scenario, the limestone layers represent relatively rapid, storm-related deposition. The rare, thin skeletal laminae within and at the top of limestone layers presumable contain fossils in or near life position and represent a community of organisms developed on the bottom after a storm. Storm-sequence deposition is illustrated in figure 8.

As one goes east in the Trenton Group, sediments record features of deeper water deposition. Here, too, limestones alternate with shales, but the limestones are very different in character. They show features that suggest deposition by turbidity currents in quite deep water. Earthquakes may well have shaken carbonate sediments lying on the shallow shelf to the west, mobilizing them into a great slurry that slid eastward just above the bottom. As the turbidity current slowed, layers of lime sands and muds settled out of the water, draped over the clay muds normally deposited in the environment.

Studies of features such as this tell us that the Trenton Group was deposited in sedimentary environments that ranged from low-energy lagoons to wave scoured barrier shoals, shallow and deep shelves, and a slope connecting the shelf to the west with a basin to the east. One shoreline appears to have lain well to the west of New York State. A shallow shelf extended as far east as the Black River Valley. East of there, water depths varied a surprising amount. Deep water changed to shallow water and back to deep water again several times between Utica and Albany. We think that indicates the presence of a number of fault blocks - horsts and grabens - in the transition zone from the western shelf to the eastern basin (Cisne *et al.*, 1982; Kidd, 1991). The eastern shoreline of the basin lay approximately at the eastern border of New York State at the beginning of Trenton time.

Return to the vehicles, and go ***north*** on route 12.

127.9 Port Leyden
130.1 Turn right and follow the spur to route 12D south.
130.3 Blinking light at base of hill. Continue westward on 12D.
133.2 Turn right at intersection with route 26 north.
140.1 Whetstone Gulf Park entrance. STOP 7.

STOP 7 WHETSTONE GULF

Throughout the region, the limestones and interlayered shales of the Trenton Group that we saw at Stops 5 and 6 are overlain by hundreds of meters of black Utica Shale (Plate I). Over the past few thousand years, Whetstone Creek has carved a steep walled gorge into the very fine-grained, soft black shales and siltstones of the Utica Shale and overlying Whetstone Gulf siltstones and shales. The sediments are much finer grained than any others we will see today and were deposited in an environment little stirred by significant currents. The black color derives from abundant unoxidized organic matter, suggesting that the sediments were deposited in a putrid, anoxic environment incapable of oxidizing dead material as it settled quietly to the bottom of the sea. The Utica Shale in central New York State is considerably thicker (over 10X thicker!) than the earlier sediments we have examined, indicating that significant subsidence must have taken place in order to make room for such a thick accumulation of sediment. The muds accumulated in a basin estimated to have been as much as 500 meters deep, produced as the edge of the Laurentian continent foundered as it collided with the Taconic Island Arc.

The change in fossil assemblages is as striking as the change in rock types. The brachiopods, corals, bryozoans and trilobites that are so common in the Trenton Group and characteristic of well-aerated, warm, shallow waters are missing from the overlying shales. Instead, we find scattered remains of graptolites and trilobites preserved in a rock full of unoxidized organic material. Poor circulation in the deep basin produced an oxygen- and food-deficient environment that was highly charged with iron and poisoned by hydrogen sulfide. The water depths were so foul that no organisms lived there. All of the fossils are of organisms that were swimmers or floaters. Their remains evidently settled to the poisonous bottom when they died, and nothing was there to eat them.

Limestones of the Upper Trenton Group in the western part of the region were being deposited at the same time as the black shales in the Champlain Valley and central Mohawk Valley. If we were to look at a number of sections, we would see that the black shale environment gradually spread westward, until all of the Trenton limestones were blanketed by black muds derived by erosion of a high sediment source in the east.

As the Ordovician drew to a close, the sediments changed yet again. The Utica Shale and Whetstone Gulf shales and siltstones are overlain by Pulaski and Oswego Formations (Plate I). It is these more resistant units that hold up the Tug Hill Plateau above Whetstone Gulf. There are many indications that water became progressively shallower in the Late Ordovician basin. Younger sediments are coarser in grain size, reflecting higher energy environments. The graptolite and trilobite-bearing Whetstone Gulf Formation is overlain by the Pulaski Sandstone, with abundant shallow-water marine clams and brachiopods. These are in turn overlain by the poorly-fossiliferous coarse beach sands of the Oswego Sandstone. The last units deposited during the Ordovician are not even marine - the basin had been completely filled in. The distinctive red shales siltstones, and sandstones of the Queenston Shale, remants of which cap the Tug Hill Plateau, show all the features of an enormous deltaic deposit that spread westward across the State.

While walking along the stream channel, notice that the shale and siltstone layers are strongly jointed. Also keep a sharp eye open for fossils in the talus and along the stream bed. Many cephalopods, brachiopods, trilobites, and graptolites can be found. Some of the cephalopods have been replaced by pyrite and look like tiny rolls of gold coins.

Return to the vehicles, and leave Whetstone Gulf State Park. Return to Route 26, and head south.

147.8 Turn left at intersection with route 12D North
149.9 Flashing light at base of hill.
150.1 Turn right toward 12S.
152.3 Turn left onto Route 12 South.
156.7 Barrett Paving Co. Quarries. Continue south on route 12.
155.5 Good view of Pleistocene delta tops to the northeast and southeast across the Black River Valley. If you look carefully, you can see sand and gravel quarries in the deltaic deposits. Notice how well these delta tops show up on the topo map.
156.8 Turn right on Schuyler Street (at sign to Boonville business district).
157.0 Turn left on route 46 at the stoplight.
157.1 Intersection of routes 294 and 46; continue south on route 46.
157.7 Valley floor near head of Boonville Gorge.
158.7 Sand quarry in Pleistocene meltwater deposits.
159.2 Channel narrows into Gorge, where Pleistocene meltwaters were channeled into Lansing Kill. This was the outlet for glacial Lake Port Leyden, and the topographic feature is known as a meltwater channel. Boonville Gorge was cut by meltwaters relatively quickly during the Pleistocene and has been little modified since. The topo map on the next page shows the abandoned channel north of Lansing Kill and the knickpoint where the channel drops into Boonville Gorge.
160.3 View southward down steep-walled meltwater channel; Lansing Kill is now a small stream in a large, steep-sided valley.
160.9 Outcrop of Middle Ordovician Utica Shale (the uppermost formation in the Trenton Group); the Pleistocene meltwater channel in Boonville Gorge was easily incised into the soft shales of the Utica.
163.3 Entrance to Pixley Falls State Park.
163.5 More Utica Shale.
167.5 For about 1km, undercutting along meanders in Lansing Kill is well-developed.
170.3 Valley widens; erratic on floor to west of road.
170.4 Delta Lake State Park entrance.
171.1 Good view of Delta Lake.
179.9 Traversing floor of glacial Lake Iroquois (before it dropped to the 450' elevation level).
181.4 Junction of routes 26 and 46. Continue south on route 46.
181.8 Traversing glacial Lake Iroquois floor at Fort Stanwix.
181.9 Traffic light; turn left on East Dominic Street.

182.2 Turn right at traffic light, and proceed south on Mill Street.
182.7 Cross Barge Canal.
182.8 At light, turn left on Martin Street.
183.3 Continue straight on route 233.
183.6 Junction of routes 233 and 69; continue south on route 233. The road passes upward from the floor of glacial Lake Iroquois (at 450' level).
189.4 Crossroads and light at Westmoreland; continue south on route 233.
192.2 Intersection of routes 5 and 233; continue south on route 233.
192.7 Kames or dissected esker in fields to east of road.
193.4 Turn left on Norton Avenue.
193.9 Kame or esker remnant at Christmas Knob to north of the road. This feature shows up best during the winter when the leaves are off the trees.
194.5 Turn left onto Kirkland Avenue near the Clinton Arena.
194.6 Take first right, and proceed east on McBride Avenue.
194.7 Crossing filled Chenango Canal.
194.8 Turn left onto Utica Street (route 12B), and proceed north.
195.2 Turn right onto Brimfield Street.
195.4 STOP 8.

STOP 8: CLINTON IRON ORES

The sediments of the Clinton Group (Plate I) are a sequence of extremely varied clastic rocks deposited in nearshore marine environments during the Middle Silurian. Sources for the detritus lay to the east in the eroding highlands of the accreted Taconic Island Arc Terrane. At stop 7, we will have a look at spoil piles dumped during extraction of the Westmoreland Hematite, one of the 2 iron ore horizons in the Clinton Group.

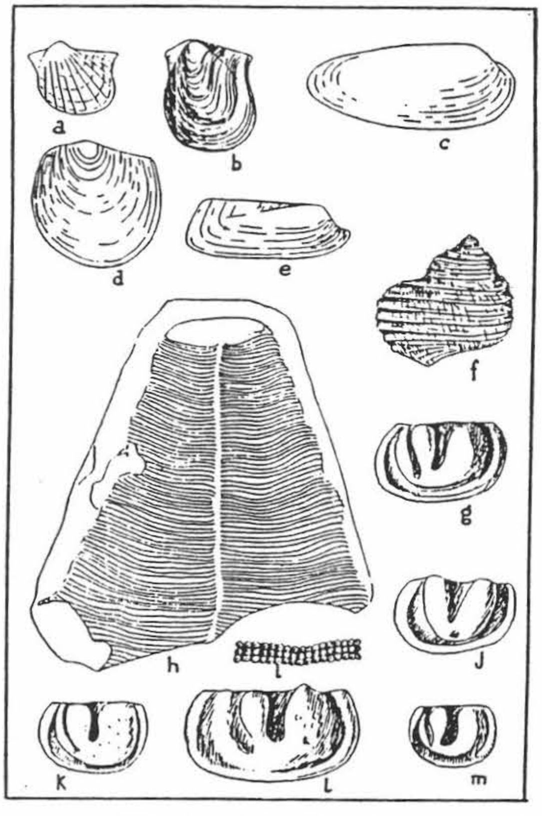
The samples in the spoil piles are dominated by brownish-red calcareous oolitic hematite ore from the Westmoreland Hematite. The ooliths (pron. *oh-oh-liths*) are approximately 1mm in diameter and range in shape from spheroids to oblate spheroids. The ooliths are composed of concentric layers of hematite and chamosite (an iron-silicate mineral) deposited around a nucleus, commonly a well-rounded quartz grain (Dale, 1953; Muskatt, 1972). Sliced in half, the ooliths resemble an old-fashioned fireball candy. Ooliths form when minerals precipitate chemically from seawater around a sand grain or fossil fragment nucleus. As currents roll the ooliths around on the bottom, they acquire layer upon layer of chemical precipitate, in snowball fashion. Calcite ooliths are currently forming on wave-agitated portions of the Bahama banks. In the Westmoreland Hematite, ooliths accumulated hematite instead of calcite. The ooliths are accompanied by rare fossil fragments and are set in a matrix of hematite, sparry calcite, and dolomite (Muskatt, 1972). In some layers, hydration has converted hematite to limonite or goethite, giving the layers an orange, rather than red-brown, color.

In addition to oolitic hematite, specimens in the spoil pile contain layers of gray-green siltstone (no ooliths), fossiliferous red shale (with some ooliths), fossiliferous and burrowed green shale, and sparsely oolitic layers with an abundance of rounded intraclasts and quartz fragments up to 1cm across. Layers are lense-shaped and somewhat discontinuous. Some specimens show well-developed trough cross-stratification, and many of the oolitic hematites show hints of cross-stratification. Megaripples with a wavelength of about .5m occur along the path east of the spoil pile.

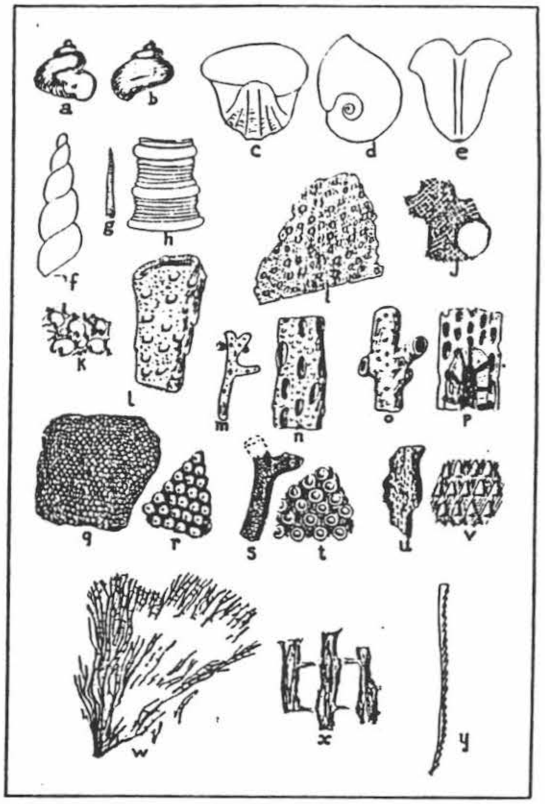
Fossils include brachiopods, trilobites, cephalopods, ostracodes, crinoids, conularids, and traces of many burrowing organisms. Fossils from the Clinton Group are illustrated in figure 9.

What clues emerge concerning the environment of deposition of these units? Presence of cross-stratification, megaripples, coarse detrital particles, intraclasts, erosional intervals, and ooliths all point to a well-agitated environment for the rocks in which these features occur. Finer-grained lithologies demand a quieter environment, and, predictably, these are the lithologies with most of the animal remains and trace fossils. An area above wave base and perhaps occasionally above low tide level is consistent with both sediment types and fauna. Tidal flats, channels, and lagoons interfingering with one another seem to be a reasonable scenario for these rocks.

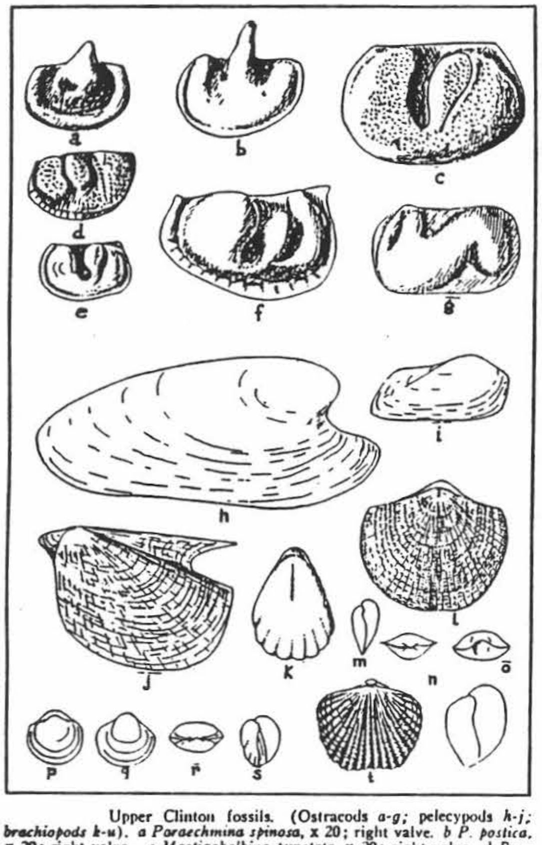
The hematite in these rocks is primary — it accumulated in the sediment right along with the clastic material. The most curious aspect of this, however, is the fact that the iron must have been in solution in the seawater in order for hematite to have precipitated and accumulated as ooliths. This may not seem like must of a problem; after all, seawater is a real soup of dissolved material. The problem is that, under oxidizing marine conditions, iron is extremely insoluble. (Salt water may rust your car, but, once it's rusted, sea water won't dissolve iron oxides off the rusty spots on your car.) Modern marine waters and most river waters contain very little dissolved iron, and there is no evidence to suggest that Middle Silurian seas were very must different overall.



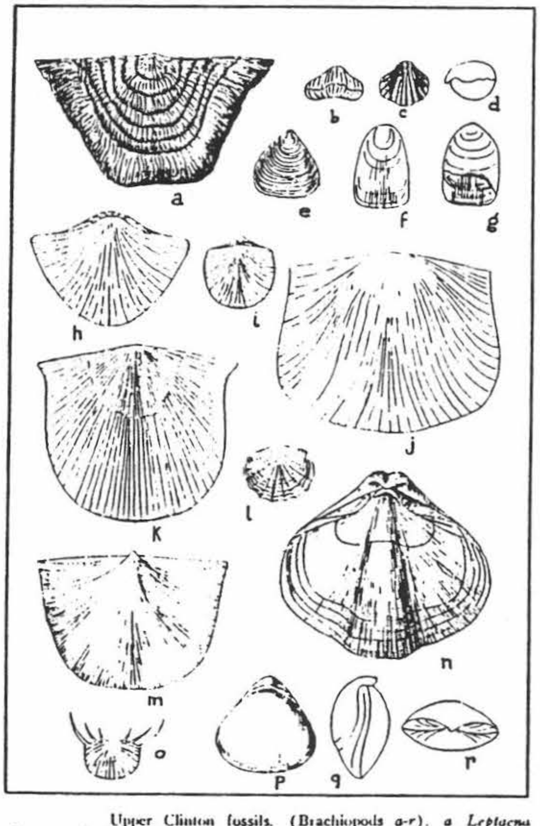
Middle Clinton fossils. (Pelecypods *a-e*; gastropod *f*; conularids *h, i*; ostracods *g, j-m*). *a Pterinea emacerata.* *b. Leptodesma rhomboidalis.* *c Ctenodonta machaeriformis.* *d Cyrtodonta alata.* *e Orthonota curta.* *f Cyclonema varicosum.* *g Mastigobolbina lata*; squeeze of right valve; x 8. *h, i Conularia niagarensis*; with surface detail x 5. *j Zygobolbina conradi*; squeeze of right valve, x 8. *k Mastigobolbina clarkei*; squeeze of right valve, x 8. *l M. vanuxemi*; natural cast of interior of right valve, x 8. *m M. lata*; squeeze of right valve, x 8. (Drawings by V. Caldwell)



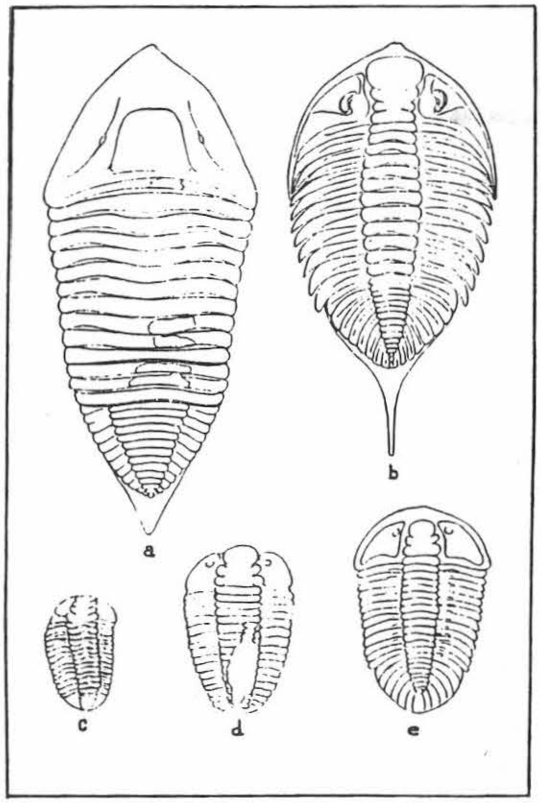
Upper Clinton fossils. (Gastropods, *a, b*; pteropod, *g, h*; bryozoans, *i-v*; graptolites, *w-y*). *a, b Strophostylus cancellatus.* *c, d, e Bucania bellerophon.* *f Hormotoma subulata.* *g, h Tentaculites niagarensis* with enlargement. x 5. *i, j Clathropora frondosa* with enlargement, x 5. *k, l Lioclema asperum* with enlargement, x 5. *m, n Acanthoclema asperum* with enlargement, x 5. *o, p Eridotrypa solida* with enlargement, x 5. *q, r Rhinopora verrucosa*, with enlargement, x 5. *s, t Stictotrypa punctipora* with enlargement, x 5. *u, v Fistulipora crustula* with enlargement, x 5. *w, x Dictyonema sculariforme* with enlargement, x 5. *y Monograptus clintonensis.*



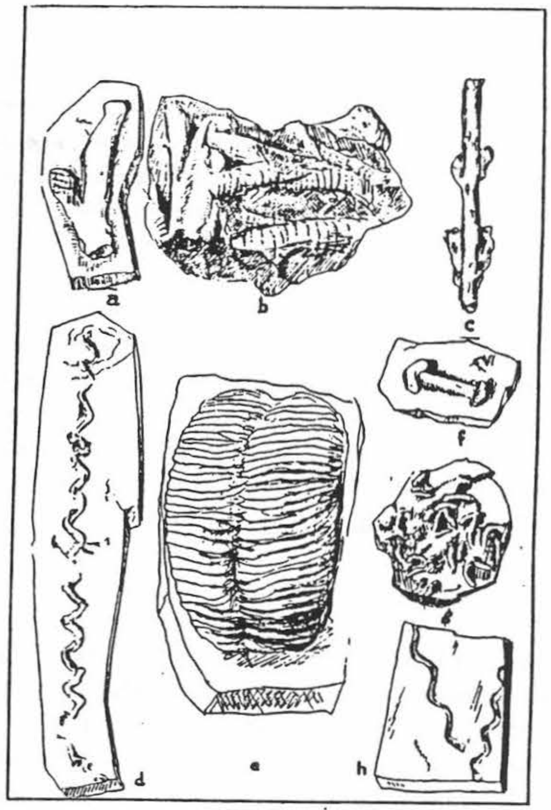
Upper Clinton fossils. (Ostracods *a-g*; pelecypods *h-j*; brachiopods *k-u*). *a Paraechmina spinosa*, x 20; right valve. *b P. postica*, x 20; right valve. *c Mastigobolbina punctata*, x 20; right valve. *d Beyrichia veronica*; testiferous right valve, x 20. *e. Mastigobolbina trilobita*, x 12; testiferous left valve, male. *f Beyrichia lukemontensis* var. *horsti*, x 16; left valve, male. *g Disygopleura proutyi*, x 20; left valve, male. *h Modiolopsis valida.* *i M.subcarinata.* *j Pterinea emacaeata.* *k Camarotoechia acinus.* *l, m, n, o Atrypa reticularis.* *p, q, r, s Nucleospira pisiformis.* *t, u Atrypa nodostriata.* (Drawn by V. Caldwell)



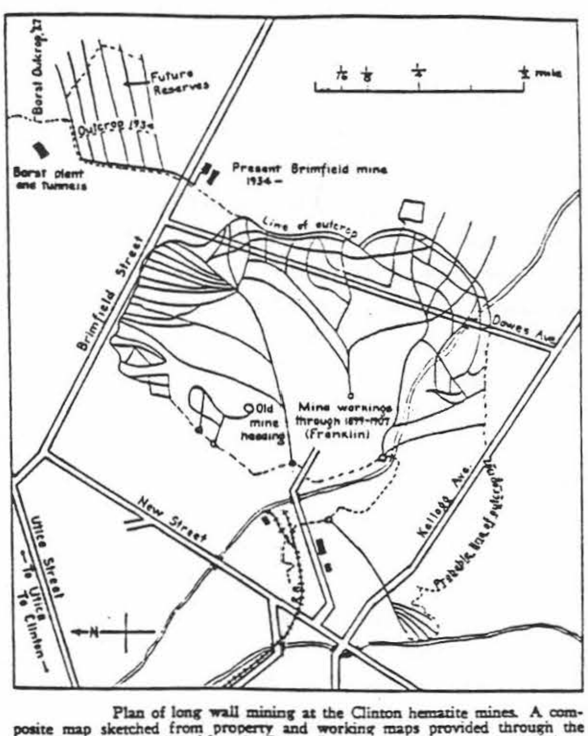
Upper Clinton fossils. (Brachiopods *a-r*). *a Leptaena rhomboidalis.* *b, c, d Camarotoechia neglecta.* *e Lingula perovata.* *f Lingula clintoni.* *g L. oblata.* *h Plectambonites transversalis.* *i Dalmanella elegantula.* *j Strophonella patenta.* *k Schuchertella subplana.* *l Coelospira hemispherica.* *m Rafinesquina obscura.* *n "Spirifer" radiatus.* *o Chonetes cornutus.* *p, q, r Whitfieldella intermedia.* (Drawings by V. Caldwell.)



Clinton trilobites. *a Homalonotus delphinocephalus*, x ⅓. *b Dalmanites limulurus*. *c Liocalymene clintoni*. *d Calymene conradi?* *e C. niagarensis*.



Clinton fossils. (Sea weeds *a, b, c, e, g;* worm burrow *f;* trail of gastropod *d, h.*). *a Buthotrephis impudica*. *b Arthrophycus alleghaniensis*. *c* Stem of sea weed? *d* Trail of gastropod. *e Rusophycus biloba*. *f* Worm burrow. *g* Stems of marine plants? *h* Trail of gastropod, Herkimer sandstone. (Drawn from specimens by V. Caldwell)



Plan of long wall mining at the Clinton hematite mines. A composite map sketched from property and working maps provided through the courtesy of the Clinton Metallic Paint Co. Stripping operations carried on entirely prior to 1897 are omitted, but followed the 700-foot contour in general; openings being made between New Hartford in the Sauquoit valley and at Clinton, Kirkland, Lairdsville and Hecla Works.

Figure 9. Fossils from the Clinton Group and map of hematite mine workings in Clinton, New York (Dale, 1953).

So, what can make iron soluble? Iron dissolves in very acidic waters and very reducing waters; neutralize or oxidize such water, and the iron with precipitate out. Many workers have suggested that sedimentary iron ores such as the hematitic ores of the Westmoreland formed where hot or cold submarine springs debouched acidic and/or reducing water laden with dissolved iron. Upon mixing with normal marine water, iron oxides and silicates precipitated as ooliths or replaced calcite in fossil material (as in the Kirkland Hematite or "red flux") (Stanton, 1972).

Sedimentary ironstones like the Westmoreland and Kirkland ores are referred to all over the world as "Clinton-type ores", names for the Clinton Group and Clinton, New York. Clinton-type ores have been important sources of iron in the past, supporting such famous steel towns as Birmingham, Aabama. Ores in the Clinton, New York area were exploited as early as 1797, and active mining of ore for pig iron continued until the end of World War I (Dale, 1953). As the huge easily-extractable ores of the Lake Superior Iron Ranges were developed during World War I, mine owners found it less and less profitable to extract ore from the 1-2m-thick hematite layers in Clinton. After World War I, iron ore was extracted for paint and brick pigment only. The operations at stop 7 were run by the Clinton Metallic Paint Company, and workings were entirely underground. Operations began on Brimfield Street in 1928 and shut down in 1963. The map in figure 9 shows the areas of Clinton undermined by iron ore extraction. Many claim that these underground workings have adversely affected groundwater flow in this area.

195.5 Continue east on Brimfield Street.
196.1 Turn right on Dawes Avenue, and proceed south.
196.3 Bridge over Dawes Creek. Herkimer Sandstone (Clinton Group) crops out east of the bridge. Megaripples of two different orientations grace the bedding surfaces along the creek.
197.1 Turn right on Kellogg Street, and proceed west.
197.3 Intersection of Kellogg Street and route 12B. Go straight, and proceed through Clinton on route 12B.
201.9 Intersection of routes 12B and 412. Turn left, and proceed south on route 12B.
204.5 Hamlet of Deansboro.
214.1 Eastern Rock Products Quarry and STOP 9. Park in the parking area east of the road. At the end of this stop, we will return to Clinton via route 12B.

STOP 9: EASTERN ROCK PRODUCTS QUARRY, ORISKANY FALLS

At this stop, we will be interested primarily in glacial features exposed at the rim of the quarry. However, we will have a look at the bedrock if time permits.

Portions of 5 Devonian formations are exposed in the quarry. The lower 35m of gray limestones belong to the Helderberg Group, including 12-15m of the Manlius Formation, 15m of the Coeymans Formation, and 2m of the Kalkberg Formation. 3m of white Oriskany Sandstone forms the prominent band high on the quarry face above the gray Helderberg. 20m of gray Onondaga Limestones at the top of the quarry face is overlain by glacial tills of variable thickness. All of the bedrock units were deposited in shallow epicontinental seas during the Early Devonian, after substantial erosion of the accreted Taconic Island Arc Terrane, and prior to accretion of the Avalon Terrane. In a trip eastward from the ocean here at Oriskany Falls during the Early Devonian, one would have encountered a low, broad landmass in what is now New England, more ocean (this time a real ocean basin with oceanic crust), and, finally, the small continent of Avalon.

The Helderberg limestones are a varied lot. Differences in lithology reflect differences in environment of deposition. Carbonate mud matrix in some reflects relatively low energy environments, while sparry matrix in others suggests high energy. Mudcracks show dessication in some, but not all units. Some of the limestones are cross-stratified and were deposited above wave base; others show evidence for being affected by waves only during storms.

The Manlius shows features consistent with environments oscillating between supratidal, intertidal, and shallow subtidal lagoon environments. Walker and Laporte (1970) have suggested that good analogies may be drawn between these limestones and their environments of deposition and those in the early Middle Ordovician Black River Group. The organisms in the communities are, of course, different species, but they occupy similar nitches in strikingly similar rocks.

Water deepened episodically but progressively during deposition of the Helderberg. The Coeymans Formation was deposited on a discontinuous barrier shoal in environments that ranged from high to low energy, and the Kalkberg was deposited in a shallow, open marine environment dominantly below wave base (Laporte, 1969; Walker and Laporte, 1970).

The Oriskany Sandstone lies disconformably on the Helderberg Limestone and represents a shoreline sand sequence deposited at the base of the last transgressive sequence in the Early Devonian. These sandstones are coarse-grained and contain remains of enormous brachiopods. Evidently, the energy in the environment was either too high to preserve more delicate fossils, or else was too high to allow less robust organisms to live happily. In southwest New York State, the Oriskany

Sandstone is hundreds of meters below the surface and serves as a good reservoir rock for petroleum.

The lower Onondaga Limestone is chertier and coarser-grained than much of the Helderberg Group, but individual pieces on the quarry floor are difficult to distinguish from the gray limestones of the Helderberg. Environments during deposition of the Onondaga were primarily those on wave-affected, shallowly-submerged lagoonal shelves (Lindemann, 1979). Reefs are common and are known to contain natural gas. The Onondaga has been used extensively for building stone, and, obviously, is quarried for aggregate.

At the top of the Devonian sediments in this quarry, we find a very distinctive erosion surface that shows conspicuous evidence of glaciation. If you examine the limestone carefully, you will be able to find examples of striae, crescentic gouges, and chatter marks. The pebbles, cobbles, and boulders that were stuck in the underside of the ice or pressed between ice and rock acted like sandpaper, forming the scratches, grooves, and gouges. Friction between the pebbles and the bedrock caused the pebbles to jump like chalk on a blackboard, forming chatter marks. Direction of striae indicate that this area was overridden both by south-flowing ice from the Laurentide ice sheet and by southeastward-flowing ice from the Ontario lobe.

Unsorted, unstratified glacial till overlies the Devonian rocks and makes up the overburden in this quarry. The till is removed and used by local municipalities as fill. In the unstratified till, you can see the direct contact of lodgement till (ground moraine) with the underlying bedrock. In some of the overburden, areas of stratification occur, where meltwater poured off the ridge to the northeast onto or along the ice sheet that filled the valley.

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PLATE I

Composite Stratigraphic Section for Central New York State - Precambrian through Early Devonian

			lithologic description	fossils	inferred sedimentary environ.	inferred tectonic environment	references	
Mid. Devonian	Hamilton Gp.	*Marcellus Fm. (Union Springs Mbr.)* *4-5m thick*	***Black Shales*** black, thin-bedded, thinly laminated, calcareous shale with black , thin-bedded, fine-grained limestone or limestone concretions; no burrows	sparsely fossiliferous; planktic organisms; cephalopods, hitchhiking bivalves, terrestrial wood	deep, anoxic basin conditions, quiet water; no large animals dwelled on the foul, mucky bottom; source of clastics to the east; analagous to Utica Shale environment.	foreland basin immediately cratonward of accreted Taconic Island Arc; accretion of Avalon Terrane east of the Taconic Arc Terrane (Acadian Orogeny) caused uplift in the Taconic Arc region and subsidence in the foreland basin; in the deepening basin, limestone deposition was extinguished and replaced by clastic sedimentation; Trenton/Utica re-run time. No good modern analog.	Rickard, 1975 Brower et al, 1975 Grasso, 1978 Grasso & Wolf, 1977 Selleck et al, 1977	Hamilton, NY area
Early Devonian		*Onondaga Limestone* *20m thick*	***Limestones*** middle and upper: very fine grained limestones with some interlayered clastic mud derived from the north during deposition of the middle Onondaga. lower: light gray, coarse grained, crinoidal limestone with well-developed coral bioherms; chert nodules. gas-bearing reefs first discovered in 1967.	rugose & tabulate corals brachiopods bryozoans gastropods trilobites	shallowly submerged lagoonal shelves with extensive tracts of carbonate reef-building organisms; lower member in shallow wave-affected waters; middle and upper members deposited below wave base; deposited in west-transgressing sea; slow submergence.	foreland basin immediately cratonward of accreted Taconic Island Arc Terrane; tectonic activity limited to intermittent uplift and subsidence of limited magnitude, presumably caused by plate margin activity east of the Taconic Arc Terrane. As the accreted and inactive Taconic Arc Terrane was progressively eroded, sediments changed from dominantly clastic (Silurian) to dominantly carbonate (Devonian). Modern analog: Arafura Sea between Australia and New Guinea.	Lindemann, 1979 Lindemann & Simonds, 1977 Rickard, 1975	Clinton/Oriskany Falls area
		disconformity						
		Oriskany Sandstone *3m thick*	***Calcareous Quartz Sandstone*** yellowish-brown, thick bedded, medium to coarse grained calcareous quartz sandstones.	large, robust brachiopods	shoreline deposit; preserved in discontinuous lenses in central New York State		Rickard, 1975	
		disconformity						
	Helderberg Group	*Kalkberg Fm.* *2m thick*	***Limestones*** blue, generally massive to irregularly bedded, med to fine grained limestones with fossil fragments set in a carbonate mud rather than sparry matrix; horizontal-burrowed; some black chert beds; no cross stratification or mudcracks; bentonites.	bryozoans brachiopods ostracodes trilobites	shallow water, stable, open marine environment with good circulation on an extensive shelf seaward & eastward of shoal environment of the Kalkberg; below wave base; not far enough east to receive terrigenous detritus fromTaconic Arc Terrane.		Anderson et al, 1978 Dale, 1953 Fisher, 1980 Laporte, 1967 Laporte, 1969 Rickard, 1975	
		Coeymans Fm. (Deansboro Mbr) *15m thick*	***Limestones*** blue, irregular to massive bedded, med grained fossiliferous limestone with fragments of brachiopods and crinoids set in a matrix of dominantly sparry calcite rather than carbonate mud; erosion srufaces with intraclasts; abundant burrow-mottling, both vertical and horizontal; cross stratification.	crinoids and brachiopods tabulate corals bryozoans ostracodes	discontinuous barrier/shoal of shallowly submerged crinoidal mounds and meadows separating open ocean to the east (Kalkberg environment) from protected lagoon (Manlius environment); both high and low energy.			
		Manlius Fm. *14m thick*	***Limestones and Argillaceous Limestones*** the following are complexly interbedded: a) blue and drab, thinly laminated, fine grained dolomitic limestones; bird's-eyes; mudcracks; no burrows. b) dark blue, thin and even bedded, fine grained limestone interlayered with med to coarse grained fossiliferous limestone; some mudcracks and flat pebble conglomerates; scattered vertical burrows. c) blue thin to med bedded, fine grained limestones interlayered with gray, massive and irregularly bedded crinoidal and "reefy" limestones; no mudcracks; scattered vertical burrows.	a) fossils scarce; ostracodes, algal laminae b) types few but individuals abundant; ostracodes, brachiopods, stromatolites c) abundant diverse fauna requiring submergence; ostracodes, rugose corals, brachiopods, stromatoporoids	supratidal (a), intertidal (b), and shallow, protected subtidal broad shelf lagoon (c); little circulation; little tidal effect (storm waves only); first in westward transgressive sequence showing intermittent (punctuated), rather than continuous, submergence.			
		Rondout Fm. *15m thick*	***Argillaceous Dolostones*** argillaceous and shaly dolostones interlayered with dolostones	bryozoans, tabulates, rugose corals	shallow water; dolomite probably secondary		Fisher, 1980 Rickard, 1975	
Late Silurian	Salina Gp.	*Bertie Formation* *16m thick*	***Dolostones and Shaly Dolostones*** gray to brown, thin to thick bedded, finely laminated dolostone; brown, thin bedded, finely laminated shaly dolostone; some bedded gypsum; mudcracks, small erosional channels; burrows; halite hoppers; flat-pebble conglomerates.	attack of the eurypterids ostracodes	intertidal to low supratidal, restricted, hypersaline environment; much like the Syracuse Formation.		Ciurca, 1978 Dale, 1953 Fisher, 1957 Fisher, 1980 Rickard, 1975 Treesh, 1972	Clintn./Utica area

compiled and prepared by Barbara Tewksbury, Hamilton College, 1992

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			lithologic description	fossils	inferred sedimentary environ.	inferred tectonic environment	references	
Late Silurian	Salina Group	*Camillus Shale* *70-100m thick*	***Dolomitic Shales and Dolostones*** red and olive dolomitic shales with some interbedded 1) rippled and mudcracked dolostone, 2) gypsum, and 3) quartz sandstone with well-rounded grains; dolomite content increases upward.	no fossils	?marginal marine with some aeolian sand?	ditto previous page	Ciurca, 1978 Dale, 1953 Fisher, 1957 Fisher, 1980 Rickard, 1975 Treesh, 1972	Clinton/Utica area
Late Silurian	Salina Group	*Syracuse Fm.* *30m thick*	***Dolostone, Shales, and Evaporites*** light gray to gray green, thin to thick bedded, thinly laminated dolostones with halite crystal casts and gypsum nodules replaced by calcite; abundant mudcracks, ripples, and flat-pebble conglomerates. Correlative with subsurface salt beds.	algal mounds ostracodes pelecypods eurypterids	intertidal to low supratidal, restricted, hypersaline environment; salt beds intercalated with peritidal dolomites; modern analog - Persian Gulf sabkhas			
Late Silurian	Salina Group	*Vernon Shale* *50-100m thick*	***Shales*** bright red, poorly fissile, unfossiliferous, mudcracked shale with green reduction spots and cracks. Local beds of 1) green, poorly fissile shale, 2) med dark gray gypsiferous and fossiliferous shale, 3) greenish-black eurypterid-bearing dolomitic shale, 4) sandstone, 5) gypsum.	in entirety, fossils very rare; in hypersaline marine members, fossils include eurypterids, brachiopods, ostracodes, pelecypods, cephalopods, & gastropods.	delta silt of river flowing into restricted lagoons or playa-type lakes (littoral & deltaic env.); temporary marine encroachment brought hypersaline conditions and a majority of the Vernon fossils; beginning of transgressive, hypersaline sequence accomplished by basin subsidence. Modern analog: Gulf of California (for sed., but not tect., env.)			
		disconformity						
Middle Silurian	Lockport Gp.	*Ilion Shale* *25m thick*	***Shales and Mudstones with Thin Dolostones*** green-gray to gray-black mudstones and very fissile shales with thin interlayers of dark gray, fine grained argillaceous dolostones containing stromatolites and edge-wise conglomerates; abundant ripples and mudcracks. Stromatolites are vuggy and contain sphalerite, dolomite, calcite, and quartz.	fossils rare, but lingulid brachiopods and stromatolites are the most common	shallow water, tidal flats in the eastern reaches of shallow embayment in Lockport Sea; salinity normal due to proximity to shore and influx of fresh water; rest of Lockport Sea restricted & became hyersaline; eastern source for Clinton Gp. clastics petered out by Lockport time and supplied only fine silt & clay.		Dale, 1953 Fisher, 1980 Rickard, 1975 Zenger, 1965	
Middle Silurian	Clinton Group	*Herkimer Fm.* *(Joslin Hill Mbr)* *25m thick*	***Sandstone, Siltstones, Sandy Shales and Dolomitic Sandstones*** variably interbedded sequence of 1) dark gray, thin to thickly laminated, generally unfossiliferous silty shale; 2) gray to brownish-gray, fine to med grained dolomitic sandstone; 3) gray calcareous siltstone; 4) hematitic sandstone; and 5) phosphatic layers. % sandstone in sequence increases to east. Small and large scale ripples, cross stratification, channels, mudcracks, rounded fossil fragments, no carbonate mud.	abundant trilobite trace fossils; crinoidal in upper part	high energy, near-shore environment; tidally-influenced; grades eastward into beach facies of Jordanville Member east of Joslin Hill		Dale, 1953 Fisher, 1980 Muskatt, 1972 Rickard, 1975 Zenger, 1971	
Middle Silurian	Clinton Group	*Kirkland Dolostone* *1.5m thick*	***Fossiliferous Hematitic Dolostones*** grayish red to moderately red, coarse grained to conglomeratic fossil fragmental, calcareous, slightly sandy dolostone; irregularly and discontinuously bedded; fossils replaced by hematite; 10-40% hematite; not oolitic; thin interbeds of green shale; known locally as the "red flux".	brachiopods bryozoans crinoids coelenterates	lensoid unit; shallow marine; lagoonal environment?			
Middle Silurian	Clinton Group	*Willowvale Shale* *10m thick*	***Shales with Sandy Dolostones*** upper: gray silty shale with interlayered light gray to dark gray fossiliferous sandy dolomitic limestone and sandy dolostone. Lower: greenish shales and shaly mudstones with some interbedded siltstones.	most fossiliferous formation in Clinton Gp; abundant brachiopods, pelecypods, & crinoids; upper units show broken & rounded fossil fragments; cephalopods	shallow, subtidal environment ranging from quiet to agitated; may be lagoon/offshore bar sequence or lagoon/tidal flat and channel sequence.			
Middle Silurian	Clinton Group	*Westmoreland Hematite* *1m thick*	***Calcareous Oolitic Hematite*** red to dull brown, med to coarse grained, calcareous oolitic hematite ore with some sandy layers; oolites composed of layers of hematite and chamosite around quartz grains and fossil fragments.	trilobites ostracodes brachiopods	shallow, agitated water, intertidal to shallow subtidal.			

PLATE I

			lithologic description	fossils	inferred sedimentary environ.	inferred tectonic environment	references	
Middle Silurian	Clinton Gp.	*Sauquoit Formation* *40m thick*	***Shales with Siltstones and Sandstones*** green, fissile shales interbedded with med gray to greenish-gray siltstones, gray shales, sandy shales, and gray to red, fine to coarse grained, cross-bedded quartz sandstone; ferruginous beds are rare; shale content decreases eastward. Ripples and mudcracks abundant; phosphate nodules locally.	few fossils in the sandstones; mudstones & siltstones contain pelecypods, brachiopods, & ostracodes; gastropods & trilobites are rare	tidal flats and channels with shoreline east of Utica; interfingers east of Joslin Hill with non-marine Otsquago redbeds.	ditto previous page	ditto previous page	Clinton/Utica area
Middle Silurian	Clinton Gp.	*Oneida Conglomerate* *3-10m thick*	***Pebbly Quartz Sandstones and Conglomerates*** light gray, thickly bedded, quartz-cemented conglomerate with interbedded light gray to rusty, finely cross-laminated, fine to coarse grained quartz sandstone	few fossils, some trace fossils and broken fragments of inarticulate brachiopods.	shallow water, high energy marine environment, probably near shoreline; some fluvial component; sediment sources to the east.			Clinton/Utica area
disconformity								
Late Ordovician		*Queenston Shale* *200m thick*	***Red Shales with Interbedded Siltstone & Sandstone*** distinctive reddish shale with subordinate interlayers of siltstone and sandstone.	unfossiliferous	deltaic environment and terrestrial environments; sediment source to the east.	deltaic deposits blanketing the last of the sediments deposited in the foreland basin of the Taconic Orogeny. These deposits are known as the Queenston Delta.		Black River Valley Area
Late Ordovician	Lorraine Group	*Oswego Sandstone* *35m thick*	***Massive Sandstone with Rare Shale Beds*** gray, fine-grained, massive, unfossiliferous quartz sandstone with a few thin black shale interbeds; some cross bedding in the sandstones.	unfossiliferous	shallow water, high energy marine environment, probably a nearshore or beach environment; sediment source to the east.	slowly-filling foreland basin immediately cratonward of accreted Taconic Arc Terrane. Modern analog: Arafura Sea between Australia & New Guinea	Bretsky, 1970 Bretsky & Thomas, 1978 Dale, 1953 Fisher, 1977 Fisher, 1980 Miller, 1910	Black River Valley Area
Late Ordovician	Lorraine Group	*Pulaski Shale & Sandstone* *200m thick*	***Sandstone, Siltstone, and Shale*** gray, fine-grained sandstone beds alternating with black to dark gray shale, siltstone, and occasional thin beds of impure limestone; ripple marks, mudcracks, burrows.	highly fossiliferous crinoids, trilobites brachiopods gastropods, bryozoans	tidal to subtidal environment; sediment source to the east			Black River Valley Area
Late Ordovician	Lorraine Group	*Whetstone Gulf Shale* *65m thick*	***Black Shales with Thin Sandstone Layers*** black to dark gray shales with occasional thin beds of fine-grained sandstone.	not very fossiliferous trilobites graptolites	shallow basin, water depth below wave base, but basin no longer anoxic as it had been during Utica time; sediment source to the east.			Black River Valley Area
Late Ordovician	Trenton Group	*Utica Shale* *230m thick*	***Black Shales*** black and gray, fissile to massive, graptolite-bearing shales intercalated with lenses of black massive calcareous mudstone. A monotonous sequence.	graptolites trilobites (pyritized) cephalopods (pyritized)	deep anoxic basin conditions over vast area and long period of time; source of clastics to the east.	shelf deepening into basin on plate being subducted beneath Taconic Arc Terrane. Taconic Arc Terrane collides with continental margin of ancestral North America (Laurentia) during the Taconic Orogeny; attempted subduction of Laurentian continental crust results in thrusting of continental slope and rise sediments back on the Laurentian continental shelf (emplacement of the Taconic thrust slices in eastern NYS) and development of a rapidly deepening trough whose axis migrates westward from eastern to central NYS, accumulating first a transgressive limestone/shale sequence, then a deep basin euxinic shale; bentonites record activity in the Taconic Arc Terrane shortly before collision. Source of clastics is the advancing Taconic Arc Terrane. Modern analog: Australian shelf/Timor Trough in the Southwest Pacific; island of Taiwan.	Anderson et al., 1978 Cameron, 1972 Cameron et al., 1972 Cisne et al., 1982 Dales, 1953 Fisher, 1977 Fisher, 1980 Larson, 1984 Titus, 1977 Titus & Cameron, 1976	Black River Valley Area
Middle Ordovician	Trenton Group	*Steuben Limestone* *8m thick*	***Limestones*** dark gray, heavy-ledged, med to coarse grained massive crinoidal limestone with little interbedded shale; cross laminations and burrows are common; forms scarp above Denley.	abundant crinoids & brachiopods; gastropods, trilobites, & rugose corals less common	subtidal, higher energy environment than Denley; shallower shelf; at, near, and above wave base.			Black River Valley Area
Middle Ordovician	Trenton Group	*Denley Limestone* *70m thick*	***Limestones and Calcareous Shales*** variable sequence of dark gray to blue-gray fine grained to very fine grained limestones and argillaceous limestones interlayered with dark gray, thinly laminated calcareous shales; some coquinal limestones; horizontal burrow networks; bentonites; overall finer grain size than earlier Trenton. Lower Denley contains slump breccias at Trenton Falls.	brachiopods bryozoans trilobites cephalopods crinoids	subtidal, relatively deep shelf, with a silty substrate; storm-influenced sedimentation			Black River Valley Area
Middle Ordovician	Trenton Group	*Sugar River Limestone* *16m thick*	***Limestones and Calcareous Shales*** thinly interlayered 1) dark gray to black, thin to med bedded, fine to med grained, non-shelly, highly fossiliferous limestones and 2) dark gray, thinly laminated calcareous shales; carbonate mud matrix content higher than Kings Falls. No coquina beds; bentonites; some burrows.	diverse fauna bryozoans crinoids trilobites brachiopods	subtidal; quiet shelf environment; relatively deep; increase in carbonate mud content over Kings Falls Unit.			Black River Valley Area

			lithologic description	fossils	inferred sedimentary environ.	inferred tectonic environment	references	
Middle Ordovician	Trenton Group	*Kings Falls Limestone 20m thick*	***Limestones and Calcareous Shales*** upper: interlayered 1) dark gray, med to thick bedded, coarse grained, non-shelly fossiliferous limestone and 2) dark gray, thinly bedded calcareous shale; limestones have lower % sparry calcite and higher % carbonate mud in the matrix than the lower unit. lower: coarsely interlayered 1) dark gray, med to thick bedded, coarse grained very fossiliferous shelly limestone and coquina and 2) dark gray, thinly bedded calcareous shale; high energy features; limestones have high % sparry calcite and low % of carbonate mud matrix.	middle & upper bryozoan-dominated but includes trilobites, brachiopods, gastropods, & crinoids; lower is also brachiopod-dominated but contains some corals	middle and upper part were offshore, shallow shelf, quieter and deeper than shoal of lower part; lower part was high subtidal to low subtidal, waveswept offshore shoal.	ditto previous page	ditto previous page	
Middle Ordovician	Trenton Group	*Napanee Limestone 6m thick*	***Limestones and Calcareous Shales*** thinly interlayered 1) dark gray, very fine grained sparsely fossiliferous limestone with infrequent, discontinuous laminae of fossil fragments and 2) dark gray, thinly laminated, fossiliferous calcareous shale. vertical burrows; no mudcracks, bentonites; med grained, fossiliferous limestones increase in abundance upward.	brachiopod-dominated low diversity bryozoans gastropods trilobites	very high to high subtidal; shallow shelf-lagoon; normal salinity.			
		disconformity						
Middle Ordovician	Black River Group	*Watertown Fm. 3m thick*	***Limestones*** dark gray, thick to very thick, lumpy discontinuously bedded, fine grained limestone with fossil fragments floating in the matrix; thoroughly horizontal burrow-mottled; black chert nodules common; bentonites; more biogenic reworking than other Black River Group units and contains more fossils.	nautiloids calcareous algae stromatoporoids tabulate & rugose corals horizontal burrowers	subtidal, level bottom; water depth probably about 10m; reflects maximum transgression.	passive (rifted) margin shelf; local shoreline to the east against a low island or islands of Precambrian on the shelf; Taconic Arc Terrane not yet impinging on continental margin; bentonites record proximity of the arc, however; slow subsidence of the shelf is reflected in transgressive sequence. Modern analog: Bahamian carbonate banks	Cameron & Kamal, 1977 Fisher, 1977 Fisher, 1980 Walker, 1973	Black River Valley Area
Middle Ordovician	Black River Group	*Lowville Fm. 18m thick*	***Limestones with Some Dolostones, Incl. Bentonites*** the following are complexly interbedded: a) pale to med gray, thin bedded, wavy laminated fine to coarse limestone; vertical burrows, mudcracks, scour structures. b) dark gray, med to thick bedded, thinly laminated fine to very fine grained stylolitic limestone and med gray thick bedded, thinly laminated coarse fossiliferous limestone. c) medium dark gray, thin to med lumpy bedded, coarse bioclastic limestone d) lithology similar to upper Pamelia (see below)	trilobites (a,b) deep burrowers (a) shallow burrowers (b,c) ostracodes (a, b, d) tabulate corals (c) gastropods (c) bryozoans (c) pelecypods (b)	middle and upper: oscillating restricted intertidal mudflats (a), protected subtidal lagoons and channels (b), and aerated shoals seaward of the lagoons (c). lower: supratidal dolomitic mudflats (d). Oscillation of environments was caused by intermittent subsidence.			
Middle Ordovician	Black River Group	*Pamelia Fm. 6m thick*	***Sandstones and Sandy Dolostones*** upper: pale gray to buff, thin to med bedded, wavy to thinly laminated, fine to med grained dolostone; small-scale mudcracks; bird's eyes. lower: tan, thin to med bedded, med to coarse grained dolomitic sandstone	ostracodes some trilobites vertical burrowers	supratidal dolomitic mudflats and supratidal regolith on the Precambrian; transgressive to the east; paleoshoreline in the Black River Valley ran approx. north-south			
		nonconformity						
Precambrian		*various lithologies (no formal stratigraphic units in the Black River Valley)*	biotite and hornblende granitic gneisses; granitic augen gneisses; biotite-plagioclase-quartz gneisses; pyroxene-biotite gneisses; migmatites; quartz-feldspar gneisses with garnet and/or sillimanite; amphibolite; syenitic and charnockitic gneisses; alaskitic gneisses. all lithologies are metamorphic and deformed, including those with igneous protoliths.	rare stromatolites in marbles near Balmat, NY	conditions of metamorphism in the Black River Valley area: 700-720°C 7.5 kilobars ~25km depth	convergent margin, continent/continent collision; Grenville sediments (perhaps deposited on a rifted margin or in an arc-related basin) were deformed and metamorphosed as a terrane or terranes of unknown size and extent collided with the Grenville continent. Deformation and metamorphism was complex and multiphase, stretching over a time range from roughly 1250Ma to about 1000Ma. Deformation and metamorphism related to this event are referred to as the Grenville Orogeny. The suture zone may lie southeast of exposed Grenville rocks. modern analog: Himalayan Range/Tibetan Plateau collision was followed by Late Precambrian rifting. modern analog: Red Sea area	Buddington, 1934 Fisher, 1980 Whitney et al., 1989	

SEDIMENTATION-EROSION PATTERNS ALONG THE SOUTHEASTERN SHORELINE OF LAKE ONTARIO

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Introduction

The southeastern shoreline of Lake Ontario (Fig. 1) is dynamic and evolving rapidly in response to wave conditions, rainfall patterns, groundwater seepage, and short- and long-term variations in lake levels. We have been studying these changes in a systematic way for three years, trying to document the nature of these processess as they interact to produce depositional-erosional effects. The field trip is designed to acquaint you with the results of our work and to show those of you who are in academia the potential this area has for field instruction about glacial and coastal deposits and processes. Furthermore, we hope that some of you may opt to direct your research efforts to some of the many unsolved problems that we will be discussing during the trip.

Relevant Background Geology

The Oswego Sandstone, an Upper Ordovician (440-445 MYBP) unit, is the dominant bedrock of the southeastern shore of Lake Ontario. This red-to-grey-colored unit is composed of clastic sediment that accumulated in fluvial, deltaic, and shallow marine environments. As you would expect, there are extensive outcrops around the coastal town of Oswego. The erosion rate of coastal sections underlain by outcrops of the Oswego Sandstone is lower than for regions that are not, because of the resistant character of the rock. The two field sites that we will visit do not have outcrops of the Oswego Sandstone, although the glacial till of the areas contains many clasts of the unit.

The stratigraphic units that are most relevant to the modern-day coastal processes are a variety of unconsolidated to semi-consolidated glacial deposits of Late Wisconsin age that are about 12,000 to 20,000 years old. Multiple cycles of ice advance and retreat scoured the Ontario Basin and laid down an uneven cover of glacial material, which along the southern edge of Lake Ontario consists of extensive drumlin fields, kame deposits, ground moraines, and proglacial lake-bed sediments (Kaiser, 1962). Reconstructions of ice-flow patterns and deglaciation events in and around the Lake Ontario region are presented by Shaw and Gilbert (1990), Ridky and Bindschadler (1990), Hicock and Dreimanis (1989), Mullins and Hinchey (1989), and Gadd (1980). The latest readvancement of the ice sheet shaped a variety of morainal deposits into streamlined drumlins that trend south-southeast (Dreimanis and Goldthwaite, 1973); many of these have been truncated at the shoreline by wave erosion, exposing their interiors for inspection. The bedded kame deposits formed in ponds, lakes, and around the margin of the retreating ice sheet by the reworking of glacial sediments by meltwater (Solomon, 1976).

About 12,000 yrs BP, glacial Lake Iroquois formed south of the retreating ice sheet with an outlet to the east near Rome, N. Y. (Muller and Prest, 1985). The complex drainage history of this lake, including a reconstruction of water-level variations, has been inferred from a study of sediment cores and shoreline outcrops of Lake Iroquois deposits (Anderson and Lewis, 1985; Anderson and Lewis, 1982; Sutton and others, 1972; Karrow and others, 1961). Some 10,000 to 11, 000 yrs BP, a eustatic rise of sea level flooded the isostatically-depressed crust of northern New York, resulting in a marine incursion, the Champlain Sea, into the St.Lawrence Valley lowlands and its environs. Studies of bottom sediments from Lake Ontario have not yet provided compelling evidence for the incursion of seawater into the Ontario Basin (Muller and Prest, 1985; Clark and Karrow, 1984; Schroeder and Bada, 1978). In any event, the present rate of crustal rebound on the northern side of the lake exceeds that on the south, causing a lake transgression of the southern shore (Clark and Personage, 1970). This differential isostatic rebound which is tilting the Ontario Basin to the south has important implications for present-day erosional processes of the southeastern shoreline (Drexhage and Calkin, 1981).

A few thousand years ago, Lake Ontario attained its present water level. In the process, the shoreline, including the landscape and its deposits, was transformed . For example, the refraction of waves focused energy on drumlins which were coastal promontories; these glacial hills were eroded and truncated, forming numerous coastal bluffs that continue to dominate much of the landscape of the southeastern shoreline (Fig. 2). The wave-driven longshore drift of gravel and sand, derived from erosion of the bluffs, formed barrier spits and baymouth bars, which separate the inlets, bays, creeks, and wetlands from Lake Ontario. Thus, a very irregular shoreline is being straightened by the erosion of headlands and by the infilling of coastal indentations and lowlands. The Lake Ontario bottom received and continues to receive an influx of fine sediment, largely mud that is winnowed out of of the glacial till deposits (Kemp and Thomas, 1976; Sutton and others, 1974; Thomas and others, 1972).

Previous work has documented the prevalence of coastal erosion by mass wasting and wave activity along the southeastern shore of Lake Ontario (Brennan and Calkin, 1984). See Martin (1901) for an interesting turn-of-the-century perspective on the nature of coastal changes in this region. Also, Kemp and Harper (1976) and McAndrews and Power (1973) provide important insights into the nature of Lake Ontario sedimentation.

Field-Site Visits

We are going to make two stops during the course of this trip. The first will be McIntyres Bluff located in the township of Sterling in Cayuga County; the site is about 4 km to the northeast of Fair Haven State Park (Fig.2). Here we will walk about one kilometer of the shoreline, examining in detail the gullies that are carved into McIntyres Bluff, the cobble beach fronting the bluff, and a baymouth barrier in front of Juniper Pond. The second stop will be near the jetties constructed at the mouth of Little Sodus Bay in the town of North Fair Haven (Fig.2). Here we will study the impact of shoreline-stabilization projects on erosional-depositional patterns of the adjoining shorelines.

Bluff Erosion

Much of the shoreline physiography is dominated by tall bluffs separated by bays, ponds, or wetland marshes that are fronted by barriers (Christensen and others, 1990; McClennen and Pinet, 1990). The bluffs, which have moderate to steep slopes and heights ranging between 10 and 50 m above lake level, are backed up by north-south-trending drumlins that have been truncated by a combination of wave erosion and mass-wasting processes. The drumlin bluffs, which are composed of unconsolidated to semi-consolidated Pleistocene till, are eroding at rapid rates (0.5 to >1.0 m/yr) and provide the main input of sediment to the nearshore zone and barrier island-bay complexes of the area.

Surveys (Drexhage and Calkin, 1981) indicate that the rapid retreat of the bluffs along the lakeshore of Ontario is attributable to a number of factors. These include:

i) bluff height: tall bluffs tend to recede at a faster rate than short bluffs;
ii) bluff steepness: the steeper the slope, the faster the rate of erosion;
iii) bluff orientation: bluffs that face northwest erode significantly faster than bluffs with other orientations because of exposure to waves generated by the dominant northwesterly winds;
iv) till composition: bluffs composed of mud-rich till tend to be unvegetated and erode more rapidly than the vegetated bluffs composed of sand-rich till;
v) beach width: the wider the beach at the base of the bluff, the slower is the rate of erosion;
vi) beach composition: bluffs fronted by gravel and cobble beaches retreat more rapidly than those fronted by sand beaches, because the former tend to be high wave-energy zones;
vii) slope of the beach and nearshore bottom: beach and bottom declivities control the amount of wave energy that is expended against the toe of the bluff;
viii) lake levels: bluffs are most susceptible to undermining by wave attack during high stands of the lake level.

We would add several other factors to the list based on our own work. Groundwater seepage may exert considerable control on the durability of the bluff to withstand erosion, particularly if there are sand-rich and clay-rich horizons in the till of the bluff. Sites where groundwater seepage occurs are revealed by near-horizontal bands of moist till that appear dark in color relative to the drier and lighter-colored zones (O'Neill, 1985). Seepage lubricates surfaces that may act as potential glide planes for slumps. Also, if the outflow of water from a seepage point is substantial, rills and gullies with deep relief are carved into the cliff's surface. Bluff surfaces that are smooth and planar undergo a slower rate of inland retreat than those that are deeply incised by an network of drainage gullies. The reason for this, one of our primary reasearch goals, is discussed at length below and will be one of the main focuses of the visit to our field sites.

Because so many natural factors directly and indirectly control the rate of bluff recession along the Lake Ontario shoreline, it is difficult to isolate the primary causative factors for any specific bluff, or for that matter for any stretch of shoreline. Furthermore, human activity -- developing property near bluffs and constructing of shore-protective structures -- has influenced, sometime profoundly, the nature, degree, and rate of coastal erosion. However, it is possible to say that in principle, a high, steep, gullied bluff, oriented to the northwest and composed of mud-rich till, and fronted by a narrow cobble beach that slopes steeply into the nearshore zone, likely will be eroding at an alarming rate, well in excess of one meter per year.

Based on our frequent trips to the lake shore under all types of weather conditions, we have observed a variety of erosional processes that are denudating the drumlin cliffs (Table 1). Our work has been directed at studying deeply gullied bluffs; we have not yet examined denudational processses of smooth, planar bluffs, which undoubtedly have a different style of cliff recession. Not surprisingly, gullied bluffs are most active during periods of heavy rainfall that lasts for several days. At such times, surface runoff becomes channelized into the rills and gullies that are incised into the cliff. Mud, sand, and gravel are flushed down the thalweg of the gullies and are deposited near the base of the cliff as small alluvial fans that radiate out from the mouth of the gully onto the upper beach. If the rain persists so that infiltration into the ground occurs, the till loses its cohesiveness. This leads to the generation of mud flows. These viscous slurries, a mixture of water and mud, have enough cohesive strength that they prevent the settlement of even large particles (> 50 cm) out of suspension. The mud flows are not restricted to the gullies, but occur everywhere on the bluff slope, wherever the ground is saturated with water and particularly where the gradients are moderate (about 20 to 60 degrees). These mud flows collectively transport a large quantity of unsorted sediment to the gullies and eventually to the upper beach where they accumulate at the base of the cliff as a wedge-shaped deposit of unstratified, poorly-sorted sediment with particle sizes ranging from clays to boulders. In effect, they resemble a glacial till, and are, in fact, difficult to distinguish from glacial debris on the basis of textural characteristics.

The mud-flow fill in the gullies is reworked by water supplied by groundwater seepage and/or surface runoff, and by rainshowers. This flowing water cuts small channels into the gully fill as it winnows out mud and entrains sand. The stream also cuts a channel through the wedge-shaped mud-flow deposits that collect at the base of the bluff. All this eroded material accumulates on the upper beach in the form of small, thin alluvial fans. These small fan deposits are quite temporary features, as waves quickly erode them when they break on the upper beach.

Slumping occurs on different scales, as masses of till, owing to shear failure, become separated from the face of the bluff and slide downward along glide planes. Fresh slump scars are a common sighting during most visits to the field sites. Large and small masses of till covered with topsoil and vegetation, including grasses, bushes and trees, occur in the middle and lower reaches of many of the gullies, and represent material that slumped off the very top of the bluff. The slumping process is enhanced substantially by periodic wave notching of the cliff base, a process which undermines the lower sections of the bluff and causes their collapse.

Under dry conditions, the bluff face is quite stable and, hence, inactive. Occasionally a boulder or pebble is dislodged and falls downslope. Wind deflation winnows and scours out clay, silt, and sand. Also, minor gravity slides have been observed where a section of the bluff has become oversteepened and collapsed, dislodging material downslope. During the winter season when storms are frequent and intense, the bluff face is remarkably stable, because of the extensive build-up of ice, as thick as a few meters, on the beach that fronts the cliff. This mass of ice serves as a natural protective revetment, and the storm waves expend their energy breaking on the ice rather than eroding the base of the cliff. Also, the bluff is frozen and covered with snow, both of which tend to insulate the cliff face from erosion. During winter or early spring thaws, mud flows transport large quantities of sediment onto the ice surface; these deposits remain perched high above the lake level until the ice melts later in the spring, and waves degrade them.

Field Measurements and Observations of Bluff Erosion

In an effort to surmise the nature, rate, and regularity of bluff retreat along southeastern Lake Ontario, we installed an array of nine steel rods on small planar sections of McIntyres Bluff (Fig.2) and a bluff near Brown Road in Wolcott, Wayne County (located about 10 km west of the jetties at the mouth of Little Sodus Bay, our second field stop). These rods were emplaced perpendicular to the slope along the bluffs' upper, middle, and lower sections, in order to establish whether the degree of erosion varies with elevation at a specific site. We attempted to select a region of each bluff that was not extensively gullied. The degree of cliff denudation was estimated by visiting the sites regularly and measuring the height of the exposed porion of the rods.

Our results are tabulated in Figure 3. Note that most of our rods (13 out of 18 installed) currently have been eroded out of the cliff or have been buried under sediment. Also, all but one of the rods emplaced near the base of the bluffs disappeared within two months of their installation as a result of burial beneath a pile of sediment debris that accumulated in this zone. However, we have sufficient data for the upper and middle regions of the bluffs to make some reasonable statements about temporal and spatial variations in the rate of cliff erosion at these two sites.

The most obvious features of our data are the changes that occur in a stepwise fashion over time (Fig.3). The bluff faces are reasonably stable during most of the year; little measurable change was noted for the seven-month stretch extending from May to November during the two-year monitoring period. Erosion seems to occur mainly during the winter or early spring These data suggest that snowmelt, thawing of ground frost, and spring precipitation, which induced seepage, surface runoff, slumping, and mud flows, were the principal factors that caused cliff recession during the measurement period. Waves, which attack the cliff at its base, cutting a notch and undermining the bluff from below, did not seem to have been a factor during this time. Bluffs are most susceptible to wave attack when lake levels are high which occurs during the early and middle summer in Lake Ontario, a time when the bluffs, based on our erosion-rod data, were inactive and stable. Another way to state this is that bluff erosion tended to occur when lake levels were low to moderately high, a time when cliff-base erosion by waves is less likely to occur. However, we know from previous observations that wave notching of the cliffs periodically contributes significantly to gravity slides and slumping. What happened fortuitously is that significant wave cutting of the bluff base did not occur at the two field sites during the monitoring period. Therefore, the erosion that we measured reflects the effects of precipitation, runoff, and seepage. Because the data are limited to two small areas at two bluff sites and to two years, these conclusions are necessarily limited and tentative. But they are a beginning at understanding details of short-term events in the long history of bluff recession.

In an attempt to identify and possibly quantify erosional and depositional processes in major gullies that are carved into the bluff face, we initiated a five-month-long profiling study of Sitts Bluff located in Cayuga County, about 2 km northeast of The Pond at Fair Haven National Park (Frederick and others, 1991). Five large gullies were profiled at three-to-four-week intervals, and a series of erosional rods were emplaced in the gullies proper and in the side slopes of the adjoining ridges. Grain-size analyses of the till indicated that it is a sand-rich deposit, comprised in terms of weight percent dominantly of sand (30 - 50 %), comparable amounts of silt and gravel (each 20 - 35 %), and a minor admixture of clay (1 - 5 %).

A summary compilation of some of our profiling data for two gullies are presented in Figure 4. They show clearly that the thalweg of both gullies periodically was lowered (erosion) and raised (deposition) as a function of time. Although this has yet to be confirmed, we suspect, based on field observations, that aggradation in the gully occurred during very wet periods when mud flows off the gullies' side slopes supplied large quantities of sediment to the gully. Subsequently, during periods of normal rainfall, channelized runoff cut into these deposits, scouring mainly mud and sand, and lowering the floor of the gully. Scouring by channelized water flow is indicated by boulder-lag deposits in the thalweg of the gully.

During the short observation period, the gullies served mainly as chutes for the dispersal of sediment shed from adjoining ridge slopes and drainage rills. Sometimes the floors of the gullies were filled with sediment, at other times they were emptied of sediment. Note that over the five-month survey there was no net change in the level of the floor in gully 1 and a net aggradation of about 60 to 80 cm in gully 2. What is perplexing, however, is that phases of erosion and deposition are not synchronous in these two gullies even though they adjoin one another (Fig.4)! The only obvious difference between the two sites is their size and orientation. Gully 2 drains directly north and is protected from the prevailing northwesterly winds by a 75-m-high ridge; gully 1 drains northwest and receives inflow from a larger tributary network carved into the bluff face than gully 2. Whether these factors somehow are responsible for the out-of-phase relationship remains problematical.

Baymouth Barrier and Beach Processes

Material eroded from and deposited at the base of the drumlin bluffs eventually is reworked by wave activity. The waves fractionate the sediment into various size fractions; the gravel and sand components are transported to the east by longshore drift across the embayed areas between drumlin bluffs where they become incorporated into the deposits of baymouth barriers and spits. This section examines the coastal processes and their resultant deposits that characterize segments of two barriers that we chose for profiling studies.

The northwest-facing baymouth barrier that encloses Juniper Pond and the surrounding wetlands to the west of McIntyres Bluff (Fig.2) is about 25-m across and relatively high, rising more than two meters above the September lake level. In profile, it resembles a broad-based triangle with a slightly steeper pondside than lakeside (Fig.5). Its crest is heavily vegetated with brush and trees and the beach is composed of medium to fine sand with a significant admixture of gravel, pebbles and cobbles..

During our three-year surveying period, the barrier remained remarkably stable despite its exposure to the prevailing winds and waves, and inclement weather. This is revealed by superimposing three topographic profiles of the same barrier site, each measured a year apart (Fig. 5). There are no substantial changes in the overall shape of the barrier, suggesting that this landform has attained a stable, steady-state configuration, at least in the short term, for the inputs and outputs of sediment, and wave-energy conditions.

Seasonal patterns of erosion and deposition are evident on the beach side of the barrier (Fig.6). The summer beach profiles are characterized best as rolling or "lumpy", and reflect minor aggradation and shifting of sand under low-to-moderate wave energy conditions. During the fall season, two to three distinct berms tend to accrete to the beach face, each typically having a slightly oblique orientation with respect to the lake's waterline. The uppermost berm represents deposition under storm conditions, a time when wave approach is from the north or northeast, rather than the more typical west or northwest quadrants. The lowest berm is the more active of the set, as it is reworked regularly by the prevailing fair-weather waves. By winter, ice builds up on the beach and the nearshore zone from swash and surf spray. The ice cover attains a maximum thickness in excess of two meters, and extends almost to the crestline of the barrier. As is the case for the bluffs, the ice which is grounded solidly to the barrier acts as an effective bulwark and protects the beach from the damaging effects of winter storms. Large storm breakers collapse against the wall of ice and fling pebbles and even cobbles up onto the ice's surface; these large particles often become covered by ice with the subsequent freezing of spray. The exposed sand at the crestline is immobilized as well by freezing spray which temporarily cements the grains into a resistant "sandstone" and "conglomerate". The lower trunks of trees and their overhanging branches get coated with ice also; some become top heavy because of the ice load and fall over, uprooting sediment. By spring, ice break up, which we have not witnessed, but has been reported by others, can scrape and bulldoze the beach, and drop pebbles and cobbles onto the sand as the beach ice melts. At this time, a reasonably prominent berm tends to form on the lower beach, which gets smeared out across the beach face by summer, creating the aforementioned lumpy microtopography. This cyclical response of the beach to the weather, we assume, is representative of tall barriers along the southeastern shore of Lake Ontario.

By contrast, low-lying barriers respond quite differantly to wave conditions, particularly those associated with storms or strong winds. A case in point is our other profiling site on a short baymouth barrier that closes off a creek and wetland area that drain the lowlying area between two prominent drumlins to the west of Brown Road (Fig.2). This barrier is narrow, about 16-m across, and rises no higher than 1.25 m above the September lake level. A comparison of three profiles taken a year apart from this site reveals its susceptibility to washover processes (Fig.7). No significant changes occurred to the barrier between September 1989 and September 1990, except for the deposition of about 10 cm of sand on the crest and upper beach face. However, a year later the entire barrier shifted its position landward by almost three meters, as a consequence of washover processes, whereby sediment on the lakeside of the barrier was transferred to its backside. The migration of the barrier shoreward was not gradual, but occurred over a very short period of time, likely during the course of a single storm event. This is suggested by a comparison of the the April and May 1991 profiles (Fig.8). Note that the "rollover" process, the transfer of sand and gravel from the front to the backside of the barrier, was completed by early May. We surmise that a combination of factors is responsible for the sudden landward displacement of the barrier. Late spring is the time of year when the lake level is rising in response to rainfall and runoff, and high river discharge. The effect of these factors is clearly evident in our May profile of the barrier which shows unusually high lake and pond levels at this time, such that merely six meters of the barrier remained emergent, rising near our profiling site no higher than 75 cm above the lake level. A storm surge created by strong onshore winds could easily have raised the water level by that much, such that storm waves would have broached the barrier, carrying sand and gravel to the backside of the barrier. The storm event responsible for overwashing the barrier probably had little rainfall associated with it. This inference is based on the fact that the metal rods set into McIntyres bluff show no sign of erosion during April 1991 (Fig.3) which they would have if there had been substantial amounts of rainfall.

During the same month-long interval (April to May, 1991), the tall barrier near McIntyres bluff showed the greatest lakeward progradation of the surveying period (Fig.8). A broad storm berm was accreted to the middle part of the beach face, widening it by four meters. If what we measured at our survey site is representative of the system, a tremendous volume of sand was plastered against the barrier at this time. Because the amount of accretion during this month-long interval is so anomalous when compared to all of our other measurements during the three-year monitering period, we assume that the storm responsible for the landward migration of the Brown Road barrier caused this depositional event as well. However, unlike the rollover effect at the Brown Road barrier,which resulted in the permanent shift of that barrier, the widening of the beach near McIntyres Bluff was temporary and did not represent a net gain of sand to the barrier system. By mid-August, McIntyres barrier had regained its former triangular profile that seems to be the stable configuration for the system (Fig.5). We should note, however, that a small washover fan occurred recently (spring 1992) on the McIntyres Bluff barrier, about 15 m to the east of our profiling transect.

Our data, which are limited, indicate that low-lying barriers are susceptible to washover processes and landward drift at times when a storm coincides with high lake levels. These conditions are more likely to occur in the spring or early summer, because lake levels are naturally higher at this time than during the remainder of the year. Moreover, it appears that washovers are uncommon during the winter season despite the high frequency of storms, because of the wide expanse of ice build-up against the shore. This ice mass absorbs the impact of waves and minimizes sediment entrainment despite the high-energy conditons. If washovers do occur during the winter, it is likely that little sediment is transported to the backside of the barrier because of the ice cover and the frozen nature of the exposed beach sand. The tentative results of our survey to date suggest that low-lying barriers along the southeastern shore of Lake Ontario are undergoing active landward migration by washover effects. This process is episodic and requires special hydrologic (high lake levels) and storm conditions. The barrier we studied off Brown Road shifted pondward by almost three meters probably during the course of a single storm. This is the only significant landward retreat of a barrier that we noted during our three year survey.

Coastal Sediment Fractionation Model

Based on our observervartions and measurements over a three-year period, we have constructed a qualitative model that purports to describe the fractionation of sediment along the southeastern coastline of Lake Ontario. Our model applies to the shoreline zone extending to the west and east of Fair Haven State Park. This coastal sector is dominated by coastal bluffs (70-80 %) separated by lowlands and wetlands, most of which are fronted with a baymouth barrier or spit (Christensen and others, 1990). Few natural inlets are evident along this shoreline segment. Hence, our model does not incorporate the complicating sedimentological effects associated with water and current discharge through an inlet. Occasionally an inlet is cut through a barrier, but it typically is short-lived because of the longshore drift of sand and gravel across the opening. Given these general physiographic conditions, we believe that the shoreline can be divided into a system of reasonably discrete coastal compartments, each consisting of three principal elements: a sediment source, a sediment dispersal network, and sediment sinks (Fig.9). Each element is described separately below.

Sediment source: For the vast majority of the shoreline under study, the active sedimentary cover is derived from a drumlin which serves as a point source for the compartment. Much of the shoreline is dotted with cliffs which represent drumlins that have been truncated by erosion. The till in these drumlins is the source of virtually all of the sediment currently being reworked by waves and coastal currents. The drumlin till is unsorted and is comprised of a heterogeneous mixture of clay, silt, sand, gravel, cobbles and boulders. This glacial material, once released from the drumlin, is what is eventually dispersed downcurrent by longshore and offshore currents.

Sediment dispersal: The network for sediment dispersal is complicated and involves two distinct pathways: one out of the source area (the drumlin bluff) to the nearshore zone, the other parallel to the shore. The dispersal mechanisms are distinct at each site of the compartment, with gravitationally-induced transport dominant in the source area and wave-generated transport prevailing in the nearshore zone. The removal of till from the drumlin bluff is accomplished by mass slumping and by water-induced transport -- surface runoff and mud flows -- down rills and gullies. In both cases, till is removed from the drumlin proper and placed at the toe of the cliff where it is eventually reworked by waves. Slumping leads to the sliding of cohesive packages of till to the cliff base. The confined flow of sediment through gullies, in contrast, leads to the formation of unsorted mud-flow deposits and stratified alluvial fans at the mouths of the gullies; if the gully system carved into the bluff is extensive, the mud-flow sediments coalesce into a wedge-shaped deposit on the upper beach. Waves then begin eating away at these deposits, eroding the fan sediment and notching the cliff or slump masses, and, by so doing, undermine the slope and induce additional gravitational sliding of till.

Waves then fractionate the sediment into various size classes, each of which has different flow paths through the coastal compartment (Fig. 9). Particles coarser than gravel tend to collect as a lag deposit on the beach that fronts the bluff. Apparently, the bluff-backed shore is a high-energy zone, likely due to wave refraction over a bathymetric swell caused by the incompletely eroded base of the drumlins and the deeper embayed areas of the lowlands to either side of the drumlins. The domimant longshore currents carry material to the east, creating a barrier or less commonly a barrier spit. The updrift end of the barrier near the bluff is composed of angular coarse gravel and pebbles; the sediment of the barrier grades into finer gravel and sand, and becomes better sorted and rounded with distance away from the bluff source. The mud fraction is put in suspension just offshore of the gravel/cobble beach that fronts the drumlin cliff, and is dispersed to the east alongshore as a discolored band of turbid water in the nearshore zone.

Sediment sinks: The boulder- and-cobble-sized particles tend to collect at the base of the drumlin bluffs. These large particles typically become well rounded and are commonly imbricated.

As the bluff face retreats in response to erosion, the boulder and cobble deposits are drowned by the advancing shoreline, and with time probably are covereded by finer sediment as water deepens. The gravel and sand fraction that is fluxed along the spit by longshore drift is molded to the beach face and nearshore zone (Fig. 9). If the barrier or spit is low-lying, then some fraction is transported across the island by washover processes where it is deposited either on the backside of the barrier or in the pond or bay proper. If the barrier is topographically high so that washovers are less likely, then the sand must be dispersed offshore in some manner that is not yet documented or is fluxed into the next compartment that begins at the base of the adjoining drumlin cliff. The suspended mud in the nearshore band is dispersed lakeward where it settles to the lake bottom.

Human Impact on the Shoreline

Many changes in depositional-erosional patterns along the present-day shoreline are the direct result of human intervention. A case in point is the construction of jetties at the entrance to Little Sodus Bay to the north of Fair Haven (Fig. 10). These structures have influenced profoundly the disposition of sediment on both the updrift and downdrift sides of the inlet. We will stop at a public beach located to the west of the inlet and walk the entire length of the barrier, trying to surmise what exact changes have been brought about to this coastal system as a result of stabilization and dredging of the bay inlet. Also, looking eastward from the western jetty, we will get a superb view of three prominent drumlin bluffs (Fig. 2), the farthest being McIntyres Bluff, our first stop of the trip. Note that the nearest bluff is vegetated, and that the other two are not. Any speculations why this is the case?

Rather than detailing the post-construction history of the shoreline adjoining the jetties, we thought it would be more valuable to make observations and attempt as a group to reconstruct what effects stabilization of the jetty has had on this coastal system. Following the discussion, we will provide each of you with a handout summarizing the sedimentation history of Little Sodus Bay and its environs. We will include information about the time of construction and the physical nature of the jetties and breakwater, the configuration of the shoreline and nearshore lake bottom prior to jetty construction, and the sedimentation-erosional patterns after jetty construction.

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Table 1 Erosional Processes along Coastal Lake Ontario

Bluffs:
- winnowing and scouring by rain, sheetwash, and surface runoff
- groundwater percolation and seepage
- mud flows
- slumping
- gravity slides
- rock falls
- wave notching
- wind deflation
- freeze-thaw creep
- animal and human activity

Barriers:
- wave sorting, winnowing, and suspension
- longshore currents
- scour at stream outlets
- stream scour along the backside of the barrier
- storm washover
- inlet formation
- creep induced by vehicular traffic
- human construction and maintenace dredging

Beach Face:
- wave sorting, winnowing, and suspension
- scouring and rafting of sediment by beach ice
- longshore currents
- creep induced by vehicular traffic
- enhanced scour from backwash unable to penetrate frozen beach face

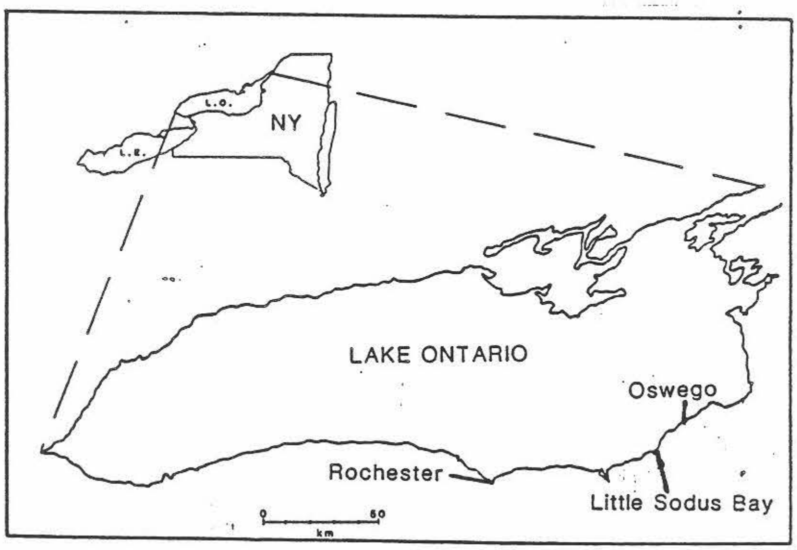


Figure 1. General location map.

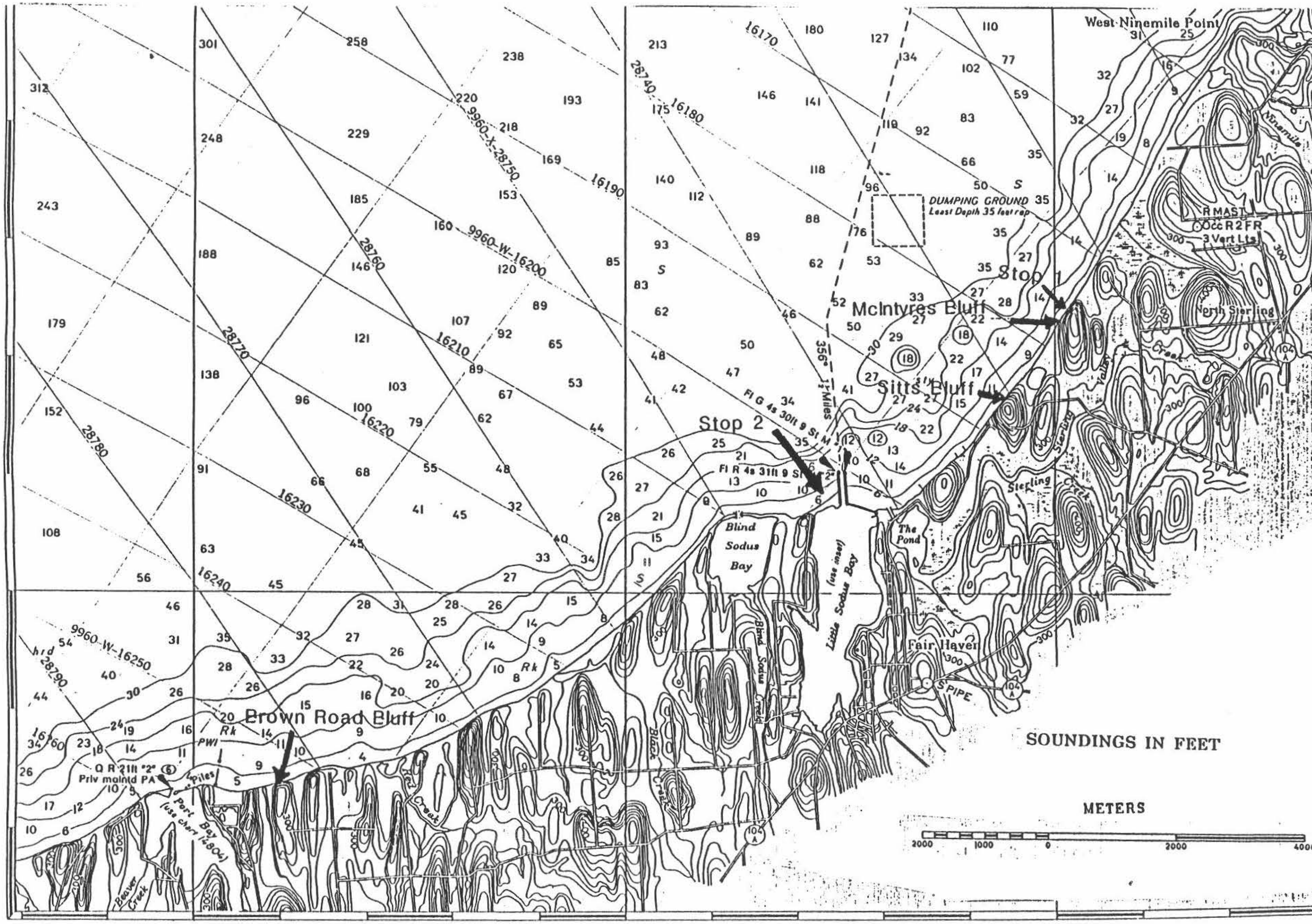


Figure 2. Southeastern shoreline of Lake Ontario showing location of the two field-trip stops.

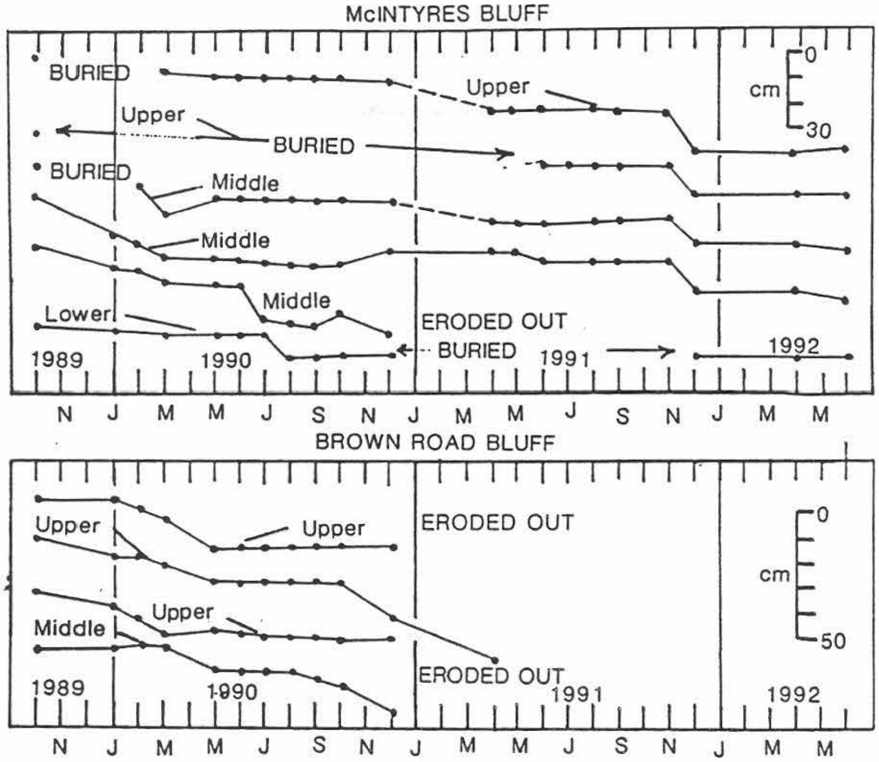


Figure 3. A plot of cumulative erosion around steel rods that were driven into the face of two bluffs.

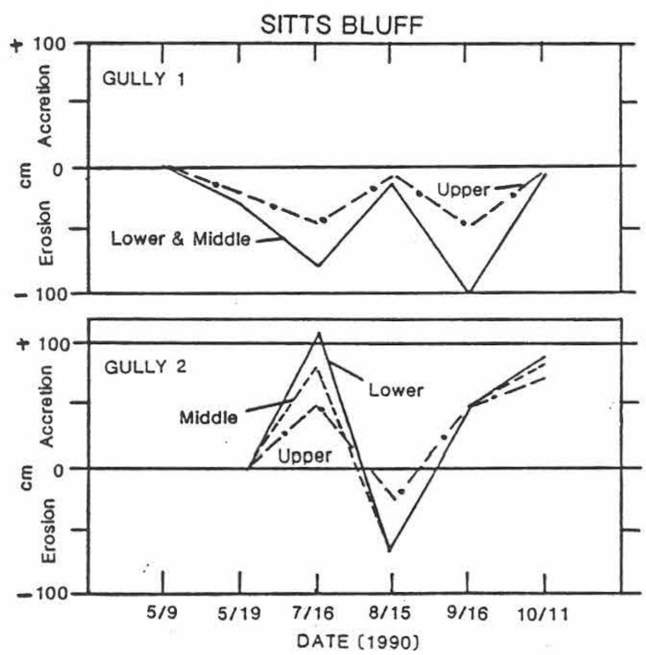


Figure 4. A time plot of variations in the floor depth of two gullies of Sitts Bluff. Gully 1 showed no net change, gully 2 a net accretion of 60 to 80 cm.

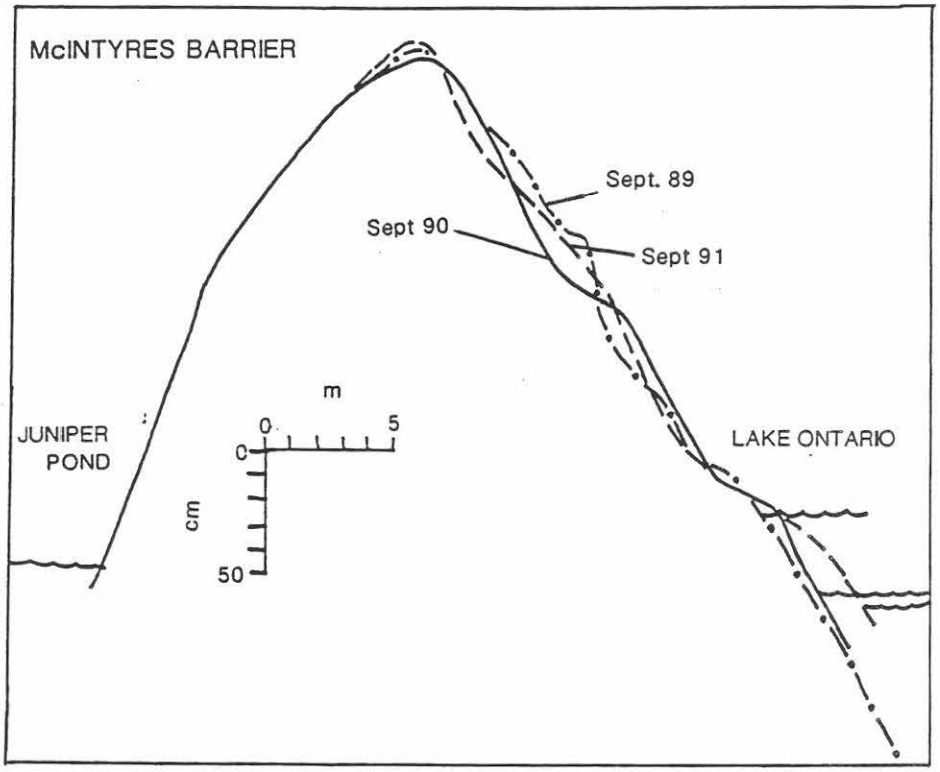


Figure 5. Beach profiles (vertical exaggeration = 10x) of McIntyres barrier.

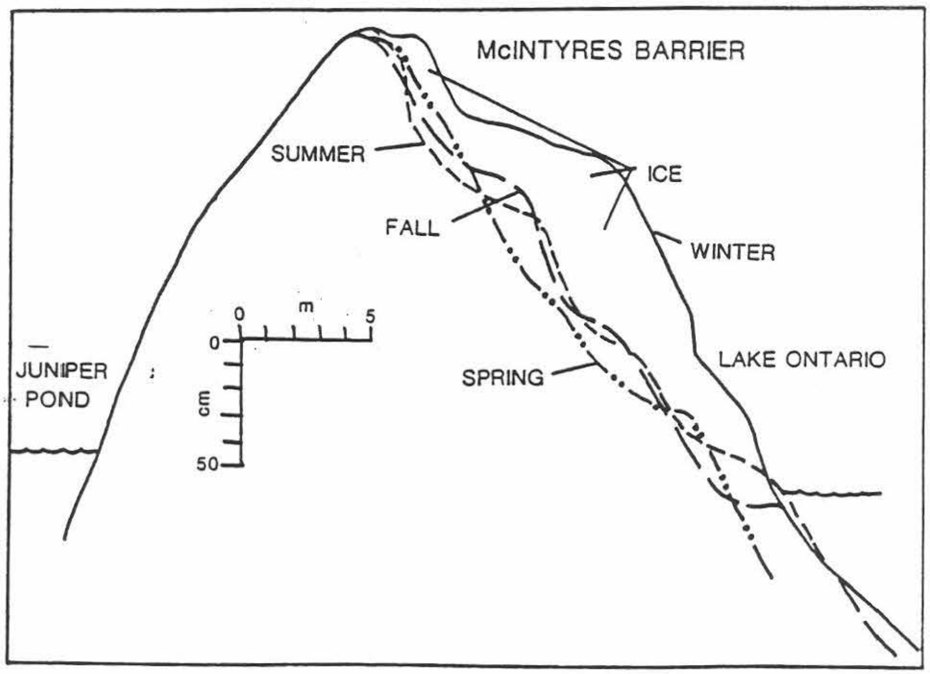


Figure 6. Seasonal variations in the beach profiles (vertical exaggeration = 10x) of McIntyres barrier.

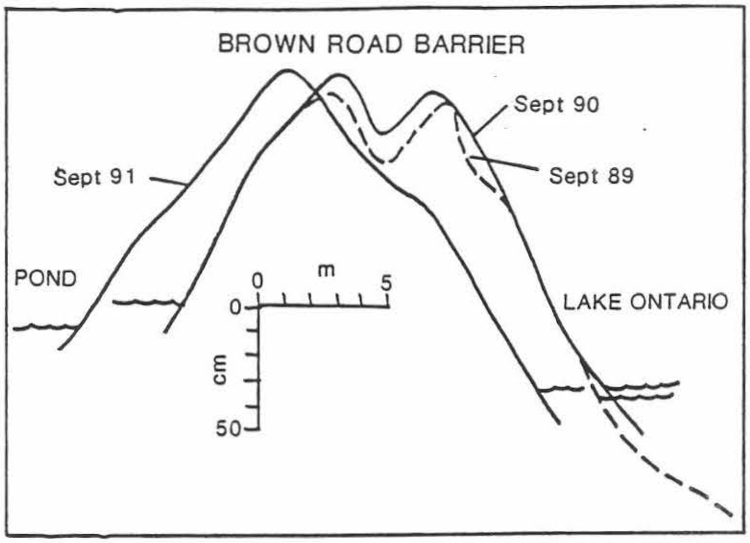


Figure 7. Beach profiles (vertical exaggeration = 10x) of Brown Road barrier.

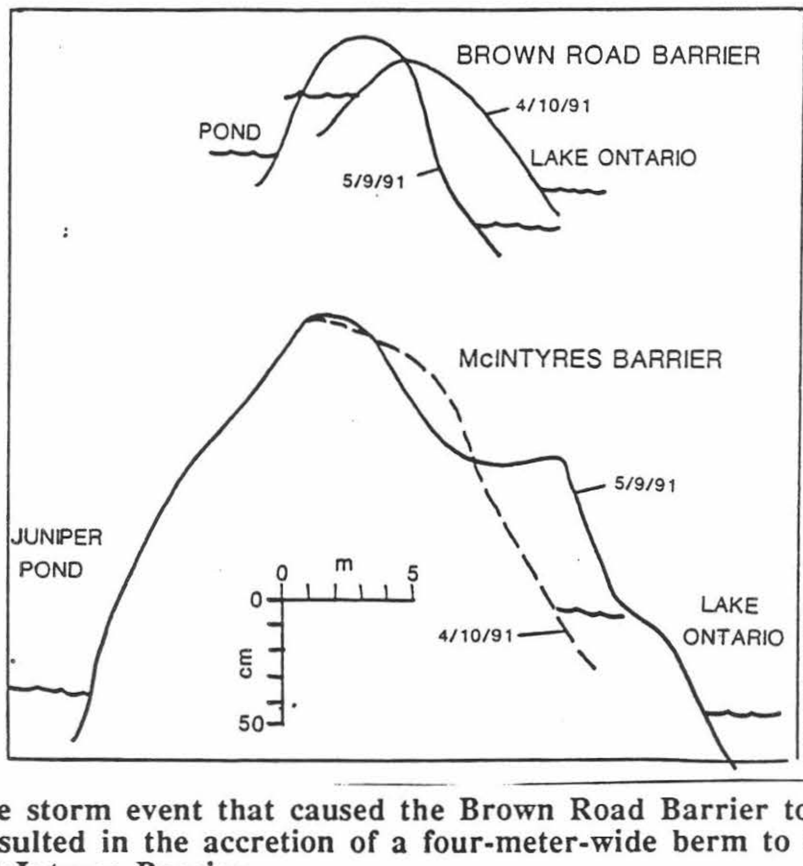


Figure 8. The storm event that caused the Brown Road Barrier to shift landward resulted in the accretion of a four-meter-wide berm to the beach of McIntyres Barrier.

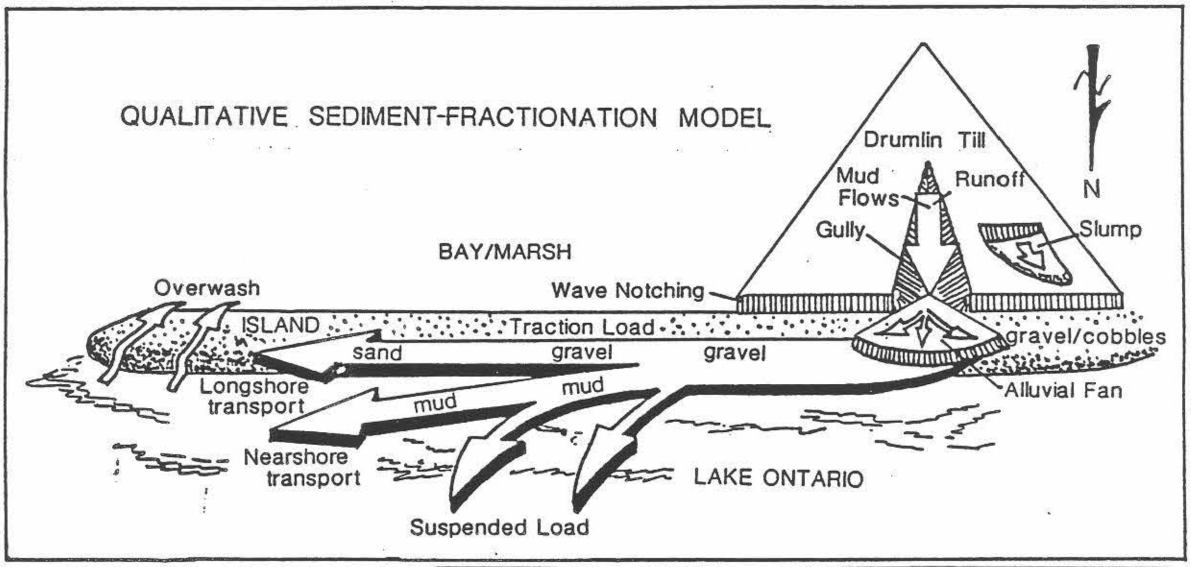


Figure 9. A model of a coastal compartment along the southeastern shoreline of Lake Ontario.

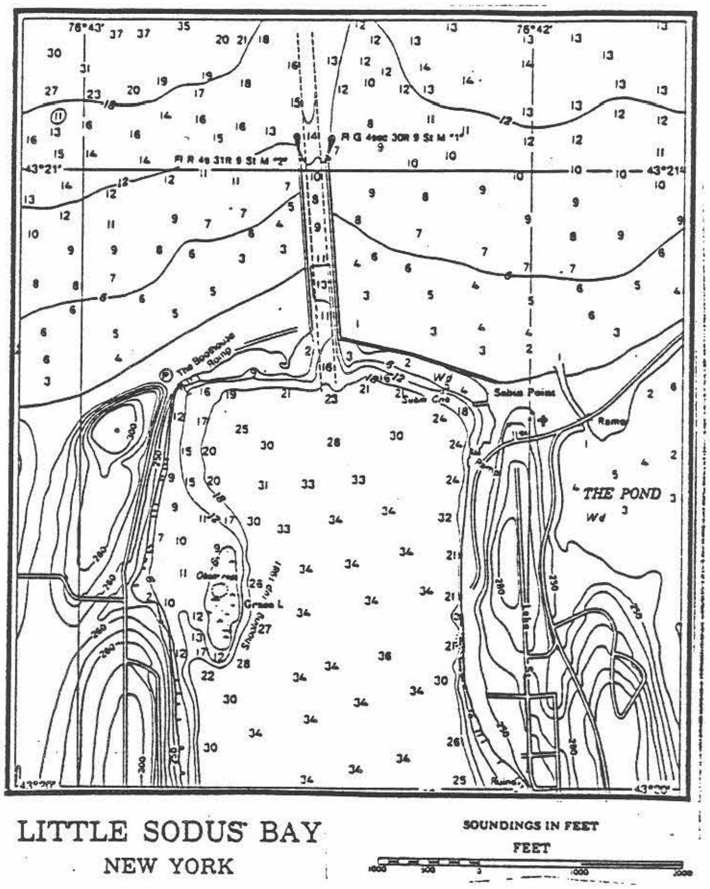


Figure 10. Chart of the jettied inlet to Little Sodus Bay.

TRIP B-1: QUANTITATIVE PALEOECOLOGY OF HAMILTON GROUP LOCALITIES IN CENTRAL NEW YORK

Trip Leaders: Willis B. Newman, James C. Brower, and Cathryn R. Newton, Syracuse University, Syracuse, NY 13244-1070.

INTRODUCTION

The Middle Devonian Hamilton Group has in recent years attracted much interest, including studies published in two volumes edited by Brett (1986) and Landing and Brett (1991). Transgressive and regressive cycles are a prominent feature of the marine sediments (e.g., Brett and Baird, 1985; Brett, Baird and Miller, 1986). The thin cycles in western New York consist of shale, calcareous shale, and limestone and are roughly symmetrical (Miller, 1986; Savarese, Gray and Brett, 1986). Approaching the source area in eastern New York, the cycles become thicker and are dominated by a regressive phase of mostly fine and coarse clastics.

This paper reviews the paleoecology of six typical regressive cycles in central New York based on quantitative multivariate techniques. The work is preliminary and more detailed studies are in preparation. Although Hamilton communities and paleoecology have been the subject of a legion of papers, most of these are qualitative. Examples can be found in the Hamilton volumes mentioned above. Papers emphasizing quantitative methods in the study of Hamilton communities include Brower (1987), Brower and Kile (1988), Brower, Thomson and Kile (1988), and Brower and Nye (1991). Other works using quantitative techniques, mostly cluster analysis, are Miller (1986), Savarese, Gray and Brett (1986) and McCollum (1991). Our analytical strategy operates in a series of steps. In the first stage, communities and gradients are characterized by means of various clustering and ordination algorithms. Both types of techniques are necessary because they are subject to different forms of distortion and they show different structures in the data set. The communities and gradients are tested for statistical significance through a variety of approaches. Both communities and gradients are then analyzed in terms of parameters such as diversity, equitability, and trophic structure. Finally, the causal factors that generated the sequence of communities and gradients are determined.

STATISTICAL METHODS

The data were analyzed with multivariate statistics. Several lucid textbooks dealing with this subject in ecology include Clifford and Stephenson (1975), Gauch (1982), Legendre and Legendre (1983), Pielou (1984) and Ludwig and Reynolds (1988). The basic analytical strategy follows that of Brower (1987), Brower, Thomson and Kile (1988), and Brower and Nye (1991). First, clustering and ordination are used to identify and characterize the communities and gradients present in the data. These are tested subsequently to ascertain their statistical significance. In the final step, plotting the communities and gradients against geographic and stratigraphic position leads to the construction of a paleoecological model.

The data treated represent the percentages of species in samples. This results in communities that reflect a combination of fidelity and dominance. Some analyses are based on the original percents or proportions of the species in the samples. In other cases, the percentages were transformed to $\log_e$ of (percent + 1). The type of data selected depends on which follows the most linear distribution.

Both ordination and clustering methods are necessary for the recognition of faunal communities and gradients because they are subject to different types of distortion. Ordination methods preserve the main patterns in the data, but the relationships between similar species and samples are often shown incorrectly. Clustering retains the distances between similar items, but at the cost of large-scale deformation. Computational details of the algorithms are available in the text-books listed above. Clusters were calculated by the unweighted-pair-group-method (UPGM) on matrices of various types of correlation and distance coefficients. In our study, ordination and cluster techniques yield conclusions that are similar and complementary.

Several methods were selected to test the significance of the clustering and correlations. Calculated correlations can easily be compared with a population or parametric value of zero (e.g. Legendre and Legendre, 1983). Cumulative frequency plots of correlation coefficients show patterns of correlations within clusters and within the entire data set. If the linkages on a dendrogram formed purely by chance, one would expect that the correlations between species or samples within the clusters would equal those for the same items between or among clusters. This hypothesis can be treated with numerous nonparametric statistics such as the Mann-Whitney U-test.

The groups of samples recognized in the clusters and ordinations should be tested for statistical significance. We elected to employ significance tests for clusters rather than discriminant analysis because they are largely independent of the number of variables (taxa in this case) and they are less sensitive to statistical assumptions. We will briefly discuss the cluster significance hypothesis tested for the communities (see Sneath, 1977, 1979 for details). The conventional hypothesis for univariate or multivariate means ascertains whether or not the differences between them are statistically significant subject to a predetermined risk level. This is obviously trivial for groups which have been previously identified by cluster analysis; one almost invariably rejects the null hypotheses even if the two groups overlap greatly. It is much more informative to query whether the samples were drawn from two populations which overlap more or less than some specified amount. We regard communities as consistent associations of organisms which may intergrade. The communities studied here have mostly changed in response to paleoecological gradients caused by regressions. Consequently, a rectangular distribution of the data provides a reasonable null hypothesis (Sneath, 1977). The null hypothesis specifies that the two clusters are drawn from overlapping rectangular distributions. The alternate hypothesis is that the distributions of the two clusters are disjunct (see Sneath, 1979, fig. 2 for critical values).

STOP 1, ROADCUT ON SWAMP ROAD

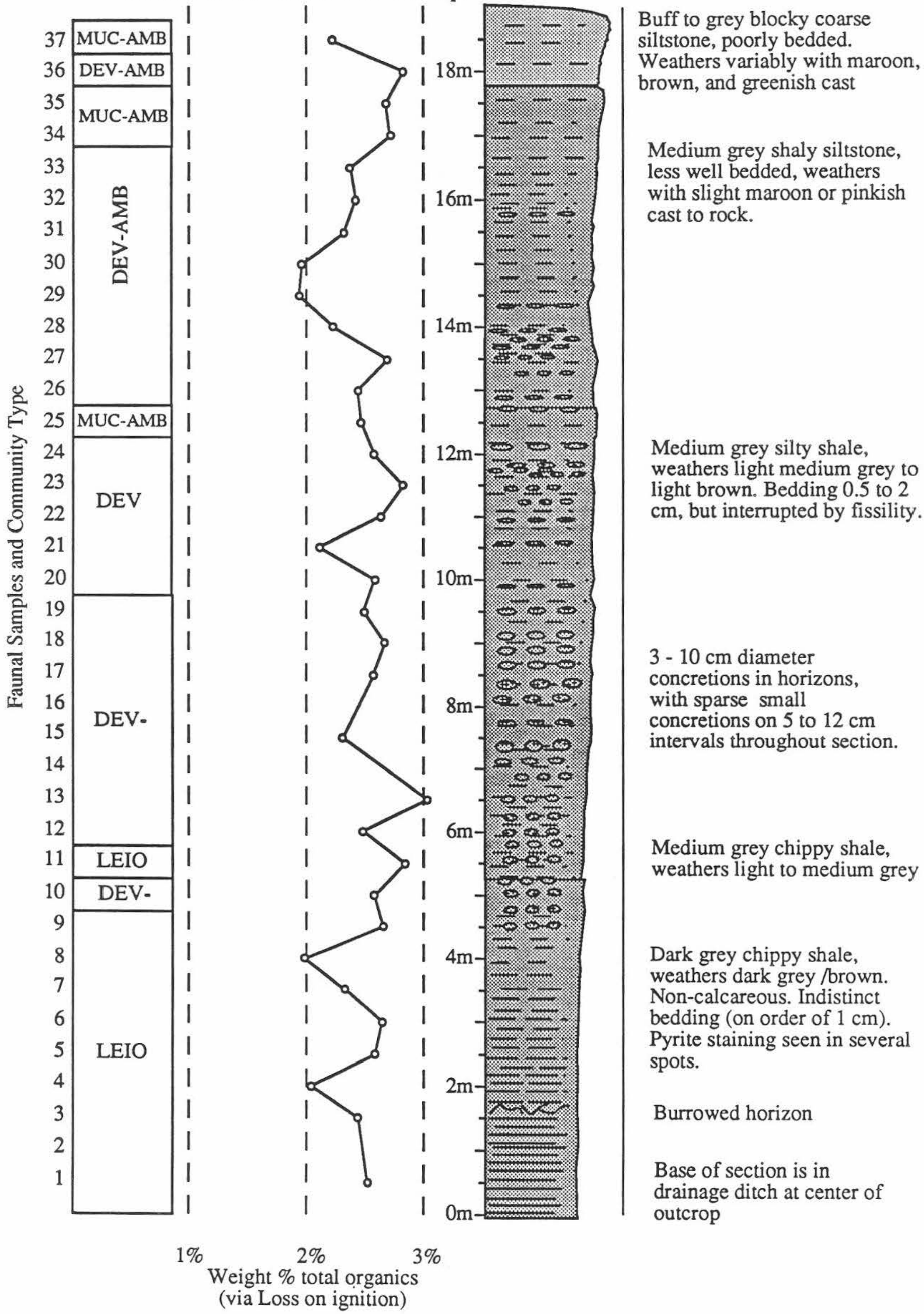
Introduction

This outcrop exposes about 20 m of the upper part of the Pecksport Member of the Marcellus Formation (Grasso, Brett, and Baird, 1985). The dark grey shales at the base grade into fine grey to brown coarse siltstones at the top (Fig. 1). Moving upwards, there is a gradual increase in species richness, adult size of the different species, proportion of epifaunal taxa, and a decrease in the abundance of deeper water taxa like Leiorhynchus and Pterochaenia.

The exceptional preservation of calcite and aragonite shells at this locality has been the focus of many studies, for example Bailey (1983) on bivalves, Rollins, Eldredge and Spiller (1971) on gastropods and monoplacophorans, Cameron (1967) on a uniquely preserved annelid worm, and Carter and Tevesz (1978) on the shell microstructure of various groups.

On initial qualitative examination, we recognized a Leiorhynchus-nuculid dominated community at the base, grading into a Mucrospirifer dominated assemblage in the middle which was capped by a large clam fauna at the top. Selleck and Linsley (1989) reported that "the fauna is

Figure 1: Stratigraphic section for Swamp Road, Pecksport Member, Marcellus Formation, Hamilton Group.



dominated by the brachiopods Spinocyrtia in the upper sandier facies, and Mucrospirifer in the middle siltier layers, along with the bivalves Ptychopteria flabellum, Gosseletia triquetra and a variety of nuculids, gastropods including Bembexia sulcomarginata and Palaeozygopleura hamiltoniae plus a variety of orthoconic cephalopods", which is also a fair qualitative statement. The preservation is deceiving. Many of the taxa possess well preserved intact or nearly intact shells composed of black calcite; these weather out and attract much attention as one searches for fossils.

More detailed sampling and quantitative analysis produces different results. The study examined 37 samples, spaced at 0.5 m intervals, consisting of 300 specimens each for a total of 11,100 individual fossils and found a very different picture. Mucrospirifer mucronatus ranges from 0 to 30 percent in any sample and does not dominate a single sample. Gosseletia triquetra yielded only 26 specimens and comprised less than 2 percent of any sample in which it was present. Spinocyrtia granulosa ranged from 0 to 4 percent of the samples. Leiorhynchus multicostum does dominate many of the lower samples found in dark shales, but Devonochonetes scitulus and Ambocoelia umbonata are more common in the higher samples.

Cluster analysis

The dendrogram for the samples reveals two large clusters, each of which can be divided into several subclusters (Fig. 2a). The clusters are generally tightly structured with most of the links being located at correlations of 0.9 and above. The large cluster on the left side of the diagram represents the deeper and quieter water samples found in the shales in the lower 10 m of the outcrop. These rocks are dominated by the pedicle-attached Leiorhynchus multicostum and the recliner Devonochonetes scitulus in conjunction with significant amounts of Pterochaenia fragilis, Ambocoelia umbonata, and the infaunal deposit feeding bivalves Nuculoidea sp., Nuculites oblongatus, N. triqueter, and Palaeoneilo sp. The pedicle attached brachiopod Leiorhynchus is the most common species in the subcluster including sample numbers 1 to 9 on the cluster diagram, and this assemblage will be termed the Leiorhynchus community. In the other subcluster with samples 10 through 18, Devonochonetes outnumbers Leiorhynchus, and Mucrospirifer becomes more abundant whereas Pterochaenia begins to disappear. This subcluster probably constitutes a relatively low diversity segment of the Devonochonetes community, which is one of the most pervasive (and internally variable) faunal assemblages within the Hamilton Group.

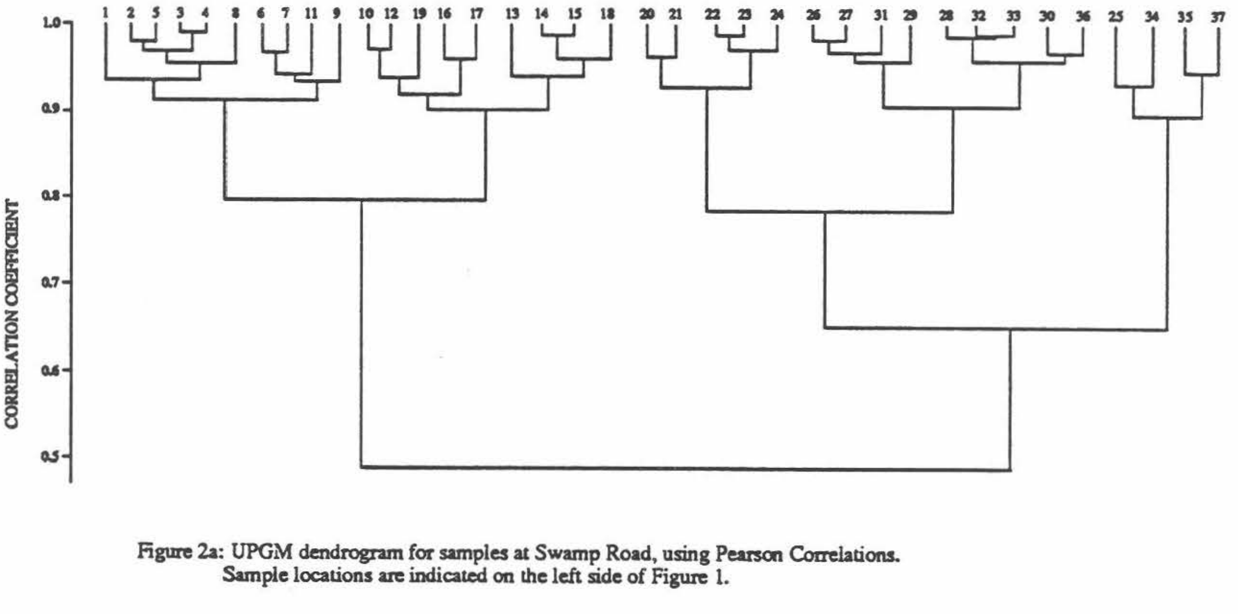


Figure 2a: UPGM dendrogram for samples at Swamp Road, using Pearson Correlations.
Sample locations are indicated on the left side of Figure 1.

The large cluster on the right of the dendrogram, ranging from samples 20 to 37, is consistently high in Devonochonetes; significant amounts of Ambocoelia and/or Mucrospirifer are also always present. The deeper water taxa, Leiorhynchus and Pterochaenia, are not quantitatively important. For the present, we will also refer to this community as that of Devonochonetes. The variance within this community is extremely high, and we may subdivide it after further study. The typical lithology is silty shale. Returning to the dendrogram, the oldest subset, samples 20 to 24, still retains a significant component of infaunal bivalves which were present in the shales below; in addition a number of the semi-infaunal brachiopod Spinulicosta spinulicosta have been found. The next subcluster with samples 26 to 36 suggests a more agitated environment. Devonochonetes and Ambocoelia are co-dominant and there are increases in the brachiopods Mucrospirifer, Tropidoleptus carinatus, Protoleptostrophia perplana, and Spinocyrtia granulosa, the semi-infaunal bivalves Modiomorpha sp. and Grammysia bisulcata, and the epibyssate bivalve Gosseletia triquetra. The last subcluster, samples 25 to 37, is separated by the dominance of Ambocoelia and Mucrospirifer, along with much variation in the numbers of Cornellites flabellum, Leiopteria sp. and Goniophora hamiltonensis. This assemblage is consistently found in the coarsest sediments which were probably deposited in the shallowest water environment within this cycle.

A dendrogram was also computed for the taxa (Fig 2b). The structure is loose and only about 10 percent of the links take place at correlations over 0.7. Despite the rather nebulous nature of the clusters, the species found in the coarser rocks are generally separated from those in the finer sediments.

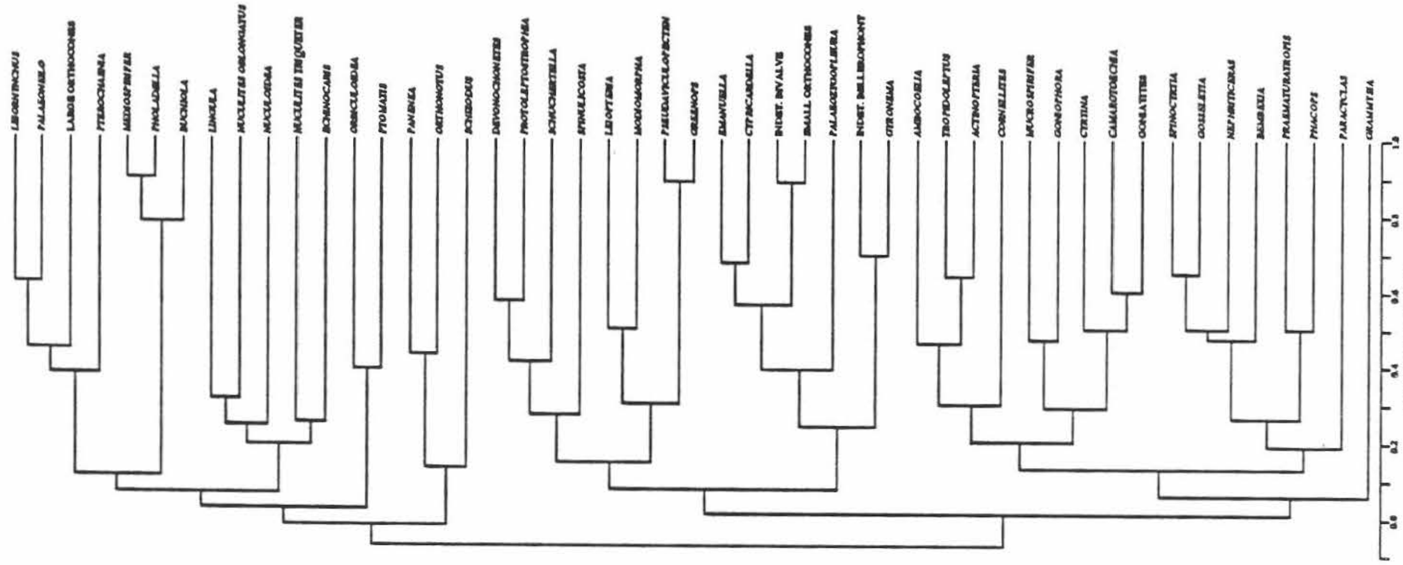


Figure 2b: UPGM dendrogram for species at Swamp Road, using Pearson Correlations.

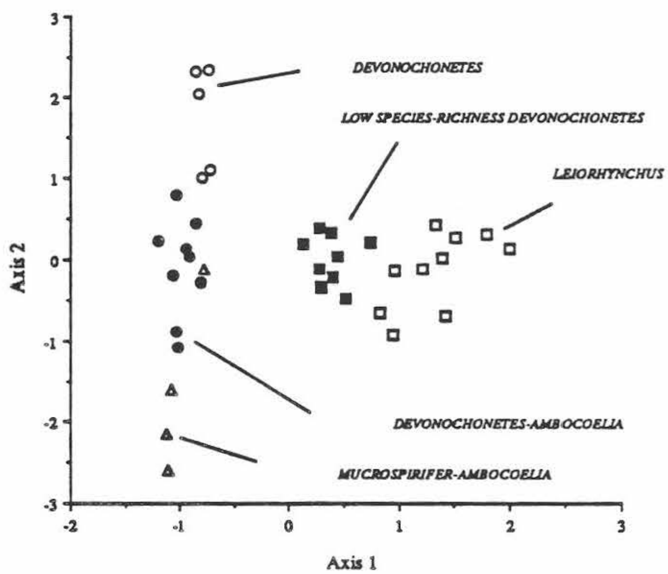
Correspondence analysis

Figure 3 contains the first two axes of the correspondence analysis for the samples, and they explain 48 percent of the information in the data set. Axis 1 arrays the Leiorhynchus community, the low species-richness segment of the Devonochonetes community, and the more typical samples of the Devonochonetes community from right to left. This is apparently associated with the major theme of change from shales to siltstones, which reflects decreasing depth and increasing amounts of agitation. The second axis dissects the three subclusters within the Devonochonetes community. Listed from top to bottom, these subclusters are those with the infaunal bivalves, Devonochonetes and Ambocoelia as co-dominants, and Mucrospirifer and Ambocoelia as the most abundant taxa. These are also grouped in order of decreasing depth.

Cluster significance tests

Cluster significance tests were computed for the two large clusters discussed previously. The results denote that the two clusters represent samples that were drawn from populations that overlap in terms of rectangular distributions at the 0.05 probability level. This was not expected inasmuch as the two clusters are markedly distinct on the dendrogram and the correspondence analysis plot. In fact, there is no observed overlap between the samples assigned to the two clusters. Significance tests were not calculated on the subclusters because each of them only contains a few samples.

Figure 3: Correspondence Analysis Plot for Swamp Road.



Sequence of communities

The stratigraphic section is illustrated in Figure 1 and a schematic sequence of communities is given in Figures 4a&b. The general pattern of decreasing depth from bottom to top of the section is clearly reflected in the lithological and faunal changes as well as the scores for Correspondence Axis 1. These scores decrease from bottom to top of the outcrop along with decreasing depth and increasing agitation. The Leiorhynchus community in the dark grey shale at the base is followed by

Figure 4a: Community progression using Correspondence Axis 1.

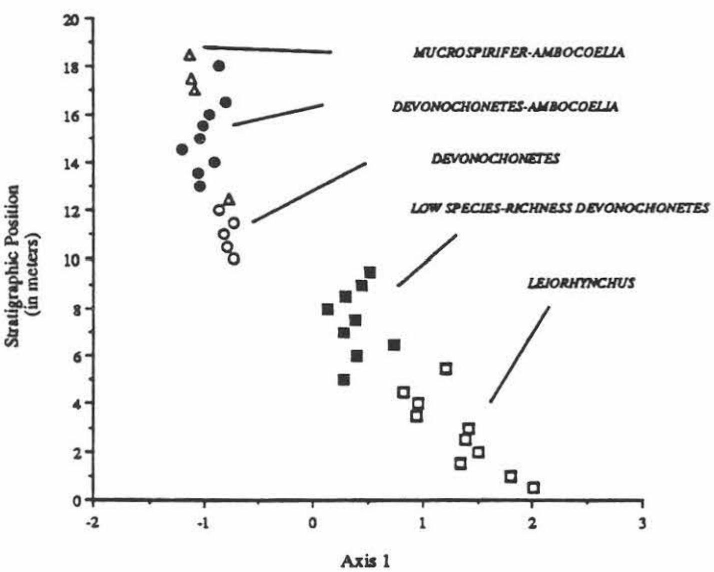


Figure 4b: Sequence of communities at Swamp Road.

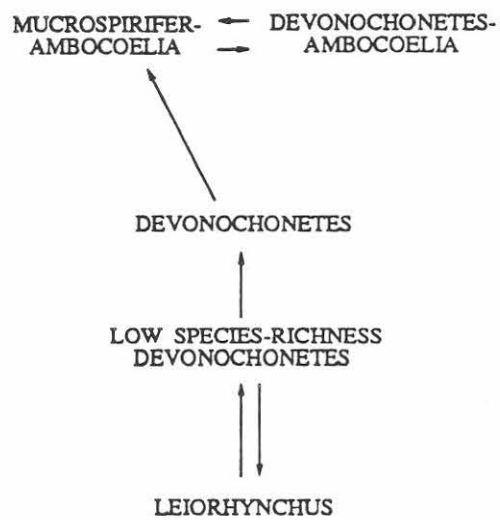
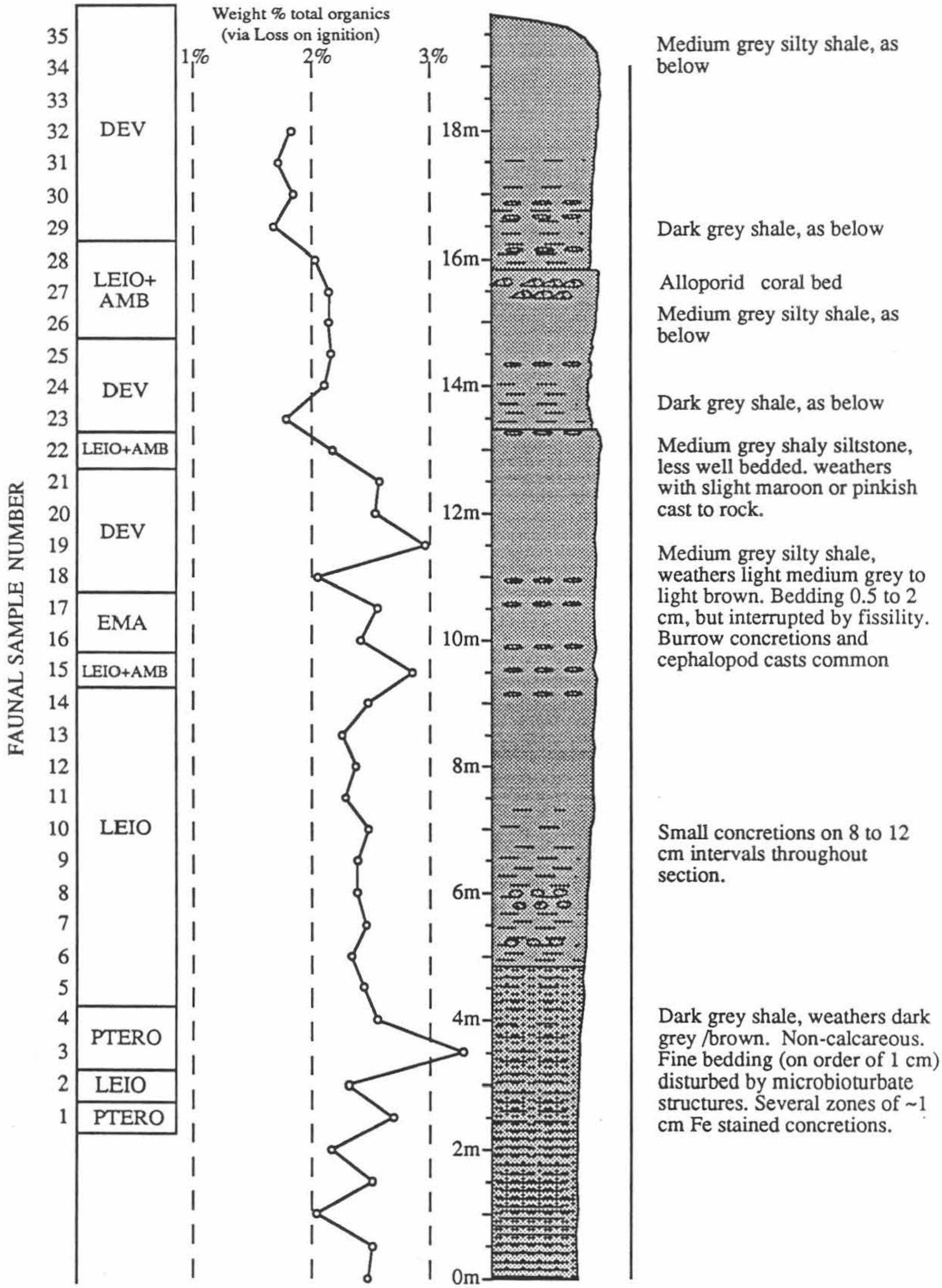


Figure 5: Stratigraphic section with Loss on ignition and faunal data for Cheese Factory Gulf, Upper Cardiff Member, Marcellus Formation.



the low diversity segment of the Devonochonetes community after a short transitional interval where the two assemblages alternate. The Devonochonetes fauna with the infaunal bivalves is next in the sequence. The uppermost part of the section is characterized by silty shale and siltstone. Here, the community composition fluctuates between the Mucrospirifer-Ambocoelia and Devonochonetes-Ambocoelia assemblages.

STOP 2, ROCKSLIDE AND GULLY IN CHEESE FACTORY GULF

Introduction

The dark grey and silty shales at this outcrop are assigned to the upper part of the Cardiff Member of the Marcellus Formation (Fig. 5). In general, it represents a more distal equivalent to the upwardly-coarsening sequence seen at Swamp Road. The most interesting aspect about this sequence is that it records the establishment of a stable benthic fauna in an area previously dominated by sparse pelagic taxa such as orthocones and Pterochaenia fragilis. As noted by Brower and Nye (1991), some individuals of Pterochaenia were probably epiplanktonic whereas others were most likely byssally attached to the substrate.

This part of the section is not generally well exposed because the soft shales weather rapidly and are quickly overgrown. The two exposures here allow comparison of weathered and fresh outcrops. The slide section dates from the early 1970's whereas the ravine developed during the summer of 1990.

The study was conducted on 35 samples taken from the ravine at 0.5 meter intervals (Fig. 5). Approximately 10,500 fossils were counted. Five depauperate samples at the base of the ravine were omitted from the analysis because it was not possible to count 300 specimens.

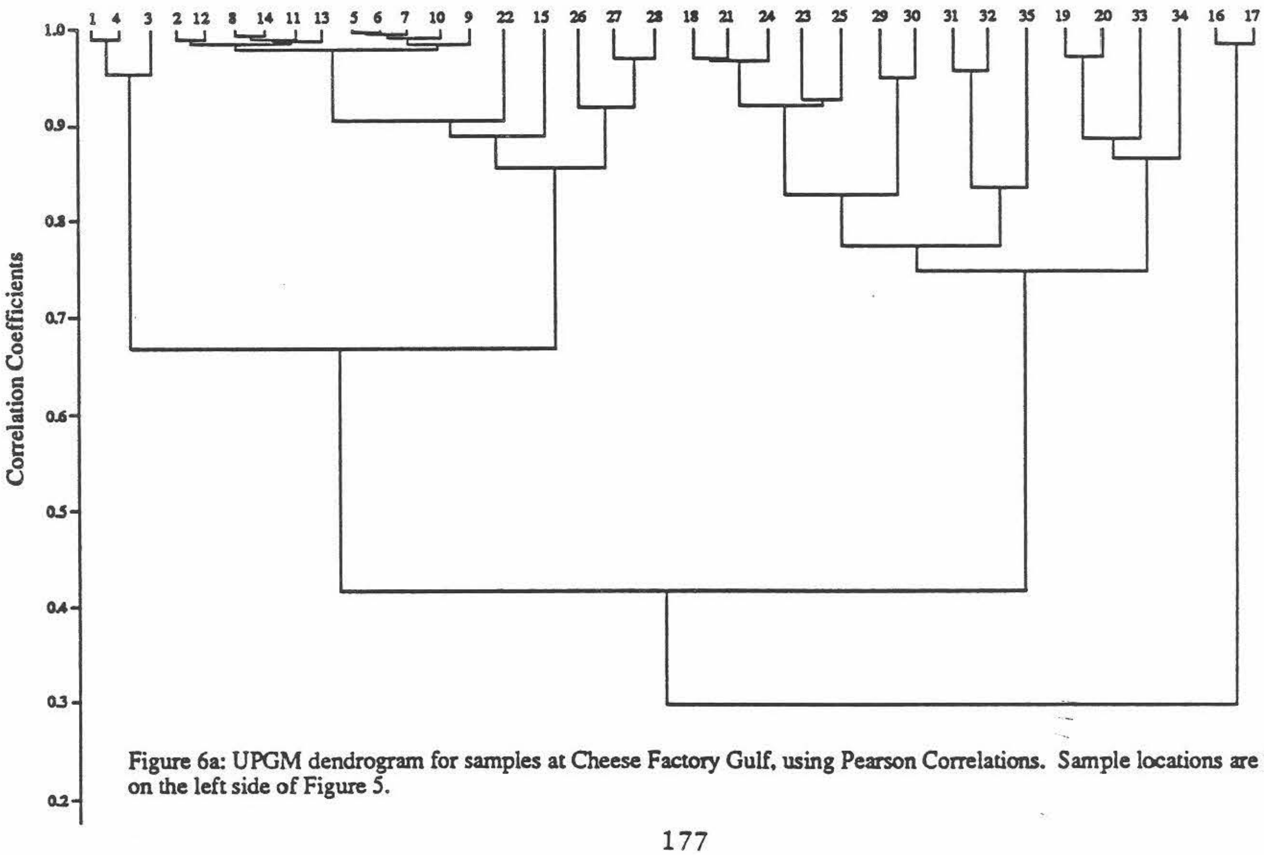


Figure 6a: UPGM dendrogram for samples at Cheese Factory Gulf, using Pearson Correlations. Sample locations are indicated on the left side of Figure 5.

Cluster analysis

Figure 6a illustrates the dendrogram for the samples. The data are tightly structured and four distinct clusters can be seen. All samples within the four clusters join at correlation coefficients that are larger than 0.7. Samples dominated by Pterochaenia fragilis form the group of three samples with numbers 1, 3 and 4 (Pterochaenia community). In the large series of samples ranging from numbers 2 to 28 on the cluster diagram, Leiorhynchus multicostum is always the most common species. Pterochaenia and Emanuella subumbona are also frequent, along with smaller amounts of Spinulicosta spinulicosta and Nuculoidea sp. These samples belong to the Leiorhynchus community. Most samples in this cluster join at high similarities where the correlation coefficients are 0.96 or higher. Samples 22 to 28 group at lower levels ranging from 0.85 to 0.95. These latter samples differ from the more typical samples, numbers 2 through 9, in having more Ambocoelia umbonata and somewhat higher faunal diversity. The large block of samples from number 18 to number 34 represent those where Devonochonetes scitulus, Leiorhynchus and Nuculoidea sp. are all common. In addition significant numbers of Ambocoelia umbonata, Mucrospirifer mucronatus, Nuculites oblongatus and gastropods are encountered. This cluster falls into a relatively low diversity segment of the highly heterogeneous Devonochonetes community. Samples 16 and 17 differ greatly from all others because they contain 50 percent or more of the small brachiopod Emanuella subumbona. This assemblage is termed the Emanuella community.

A cluster diagram was also generated for the species (Fig. 6b). As usual for paleoecological data, this dendrogram is much more loosely structured than that for the samples. For example, only 14 percent of the links are above 0.7.

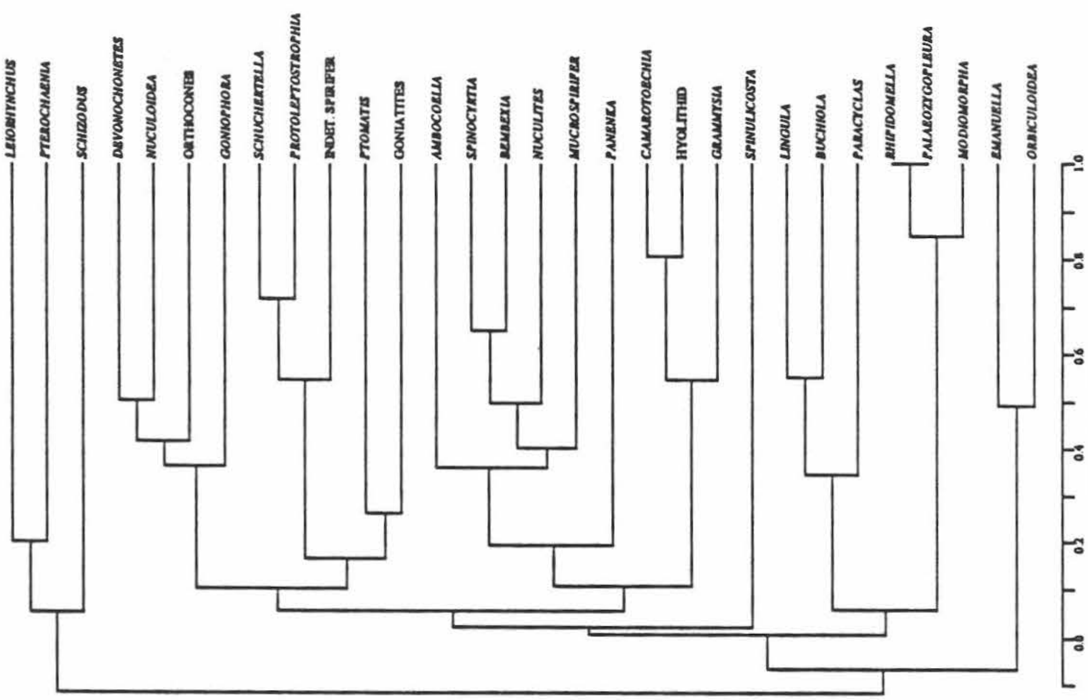


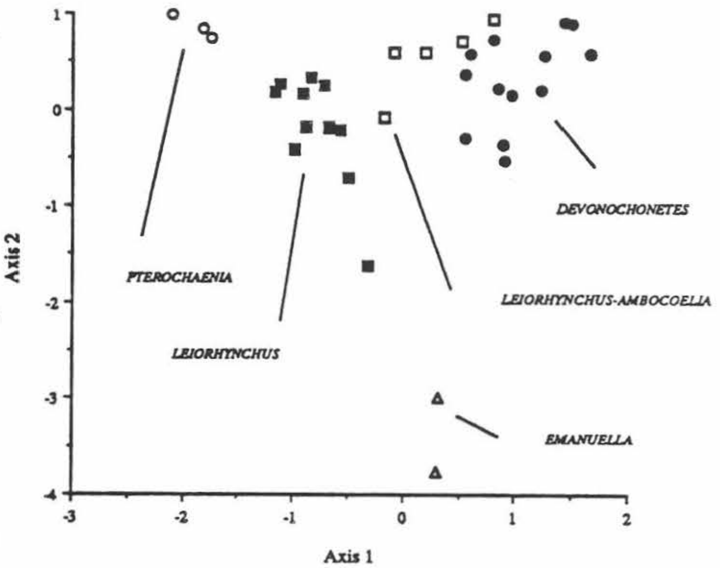
Figure 6b: UPGM dendrogram for species at Cheese Factory Gulf, using Pearson Correlations.

Correspondence analysis

The first two axes for the samples of the correspondence analysis are pictured in Figure 7. Together, these account for 47 percent of the variance in the data and they display the overall relations between the communities. Although, a small amount of overlap can be seen, the Leiorhynchus and

Devonochonetes communities generally occupy separate regions in the middle of the plot. The Pterochaenia community lies closest to and is most similar to that of Leiorhynchus. The two samples with Emanuella are widely separated and only marginally similar to some samples in the Leiorhynchus community. As mentioned later, these relations are consistent with the sequence of the communities.

Figure 7: Correspondence Analysis Plot for Cheese Factory Gulf.



Cluster significance tests

Cluster significance tests were applied to the samples from the Leiorhynchus and Devonochonetes communities. These indicate that these two groups of samples are drawn from overlapping rectangular distributions at the risk level of 0.05. Considering the results of the correspondence analysis, this is not surprising and was expected. Furthermore the significance tests suggest that these two communities may not occupy separate regions in the environmental gradient seen at this outcrop. The other groups were not tested because of the small numbers of samples involved.

Figure 8a: Community progression using Correpondence Axis 1.

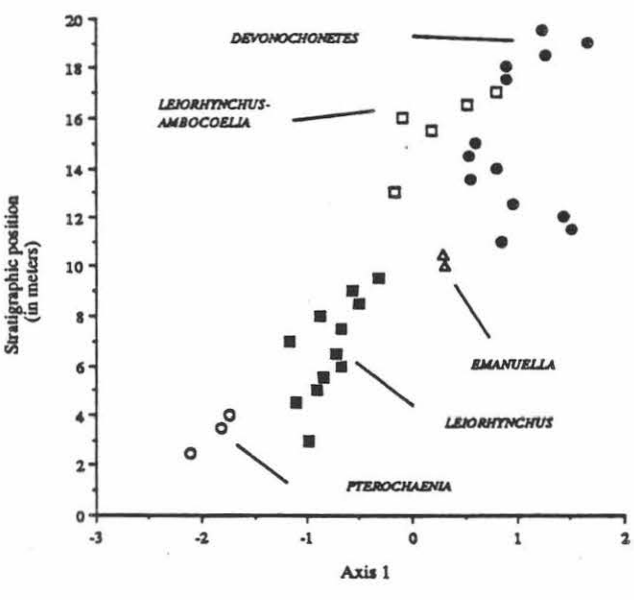
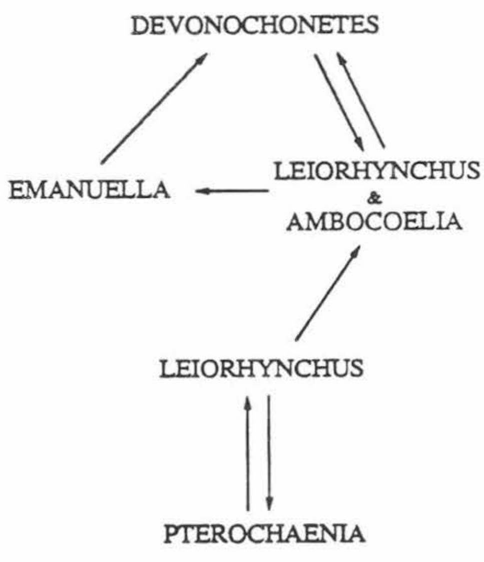


Figure 8b: Sequence of communities at Cheese Factory Gulf.



Sequence of communities

Figure 5 shows the stratigraphic section and vertical sequence of communities which is summarized in Figure 8. Overall, there is a reasonable correlation between stratigraphic level, rock type, organic content, and community composition. The oldest community, Pterochaenia, is generally in the finest sediments with the highest organic content. We interpret these specimens of Pterochaenia as being byssally attached to the seafloor. As water depth decreased, the Pterochaenia community was eventually replaced by a Leiorhynchus community after a short interval of fluctuation between these two assemblages. The Emanuella community follows that of a sample in the Leiorhynchus community with many Ambocoelia. Both Leiorhynchus and Emanuella are pedicle-attached brachiopods although they exhibit few other common features. The environmental changes involved here are not certain and are currently under study. The lithology and organic content of the two samples placed in the Emanuella community resemble those of many samples in the Leiorhynchus community. The Devonochonetes community predominates in the upper half of the section although there are several intervals with the Leiorhynchus community. These two segments of the latter community are characterized by higher diversity, more Ambocoelia, lower organic content, and slightly coarser lithology than the typical members of the Leiorhynchus community lower in the outcrop. Most of the more diverse samples in the Devonochonetes community are found in the siltier sediments with the lowest amounts of organic matter. The most frequent ecological groups in the Devonochonetes community are reclining and pedicle-attached brachiopods. The lithological changes imply that this community lived in the shallowest water. The most obvious environmental change in the section is decreasing water depth. Note that Correspondence Axis 1 provides a reasonable measure of relative depth. Low values are associated with deep water samples and vice versa. However, changes in oxygen content and substrate properties may also have taken place. In general, conditions improved sufficiently to allow colonization by a more diverse benthic fauna.

STOP 3, ROADCUT ON ROUTE 20 NEAR POMPEY CENTER

Introduction

The Route 20 roadcut exposes the Delphi Station Member of the Skaneateles Formation except for the basal and topmost beds. Brower, Thomson and Kile (1988) published a detailed description of this outcrop. Seventy samples were obtained from the section which ranges from shale at the base to fine siltstone and sandstone at the top (Figure 9). The cluster (Figs. 10a&b) and correspondence (Fig. 11) analyses reveal three communities, Nuculoidea, Schuchertella, and Actinopteria, listed in ascending stratigraphic order. The Nuculoidea community can be divided into three subcommunities. Significance tests on the clusters suggest that they are relatively tightly structured. All assemblages found in adjacent beds represent samples that were drawn from disjunct rectangular distributions at the 0.05 or 0.01 risk levels.

All three communities have equitabilities that are 0.78 or higher and are interpreted as biologically accommodated rather than opportunistic. For example, Brower and Nye (1991) report equitabilities ranging from 0.19 to 0.49 for three Hamilton benthic communities that formed under stressed conditions. Proceeding from the Nuculoidea to Schuchertella to the Actinopteria community, the grain size becomes coarser, ranging from mainly shale to mostly siltstone and sandstone, whereas the organic content decreases from 3.01 to 2.11 percent. The number of common taxa per sample rises from about 16 to 23. The lithology and sedimentary structures demonstrate that the Nuculoidea community occupied the deeper and more offshore areas whereas the Actinopteria assemblage lived in the most shallow regions closer to the shoreline.

Nuculoidea community

The abundant organic matter and fine grained sediment indicate generally quiet water

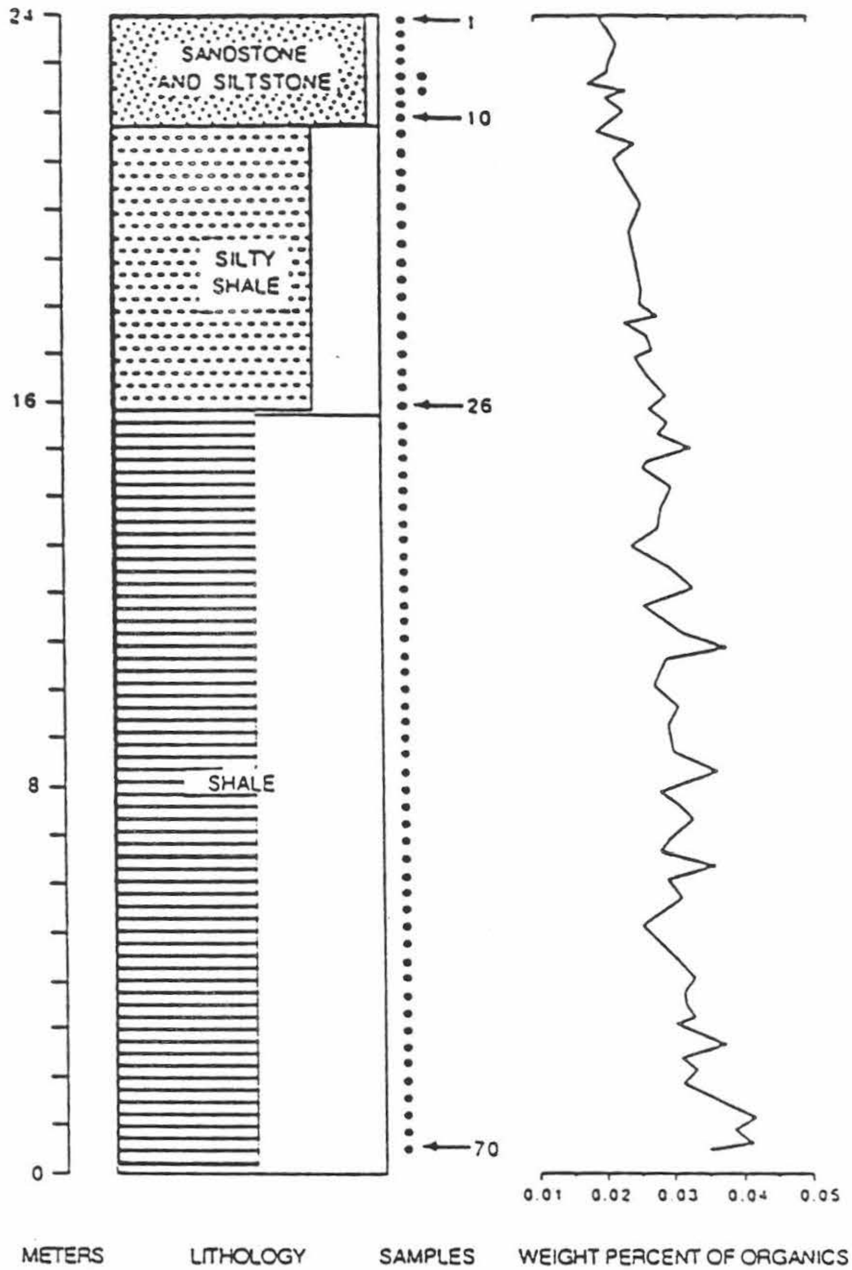


Figure 9: Stratigraphic section for the Delphi Station Member (Skaneateles Formation, Hamilton Group) exposed at Route 20.

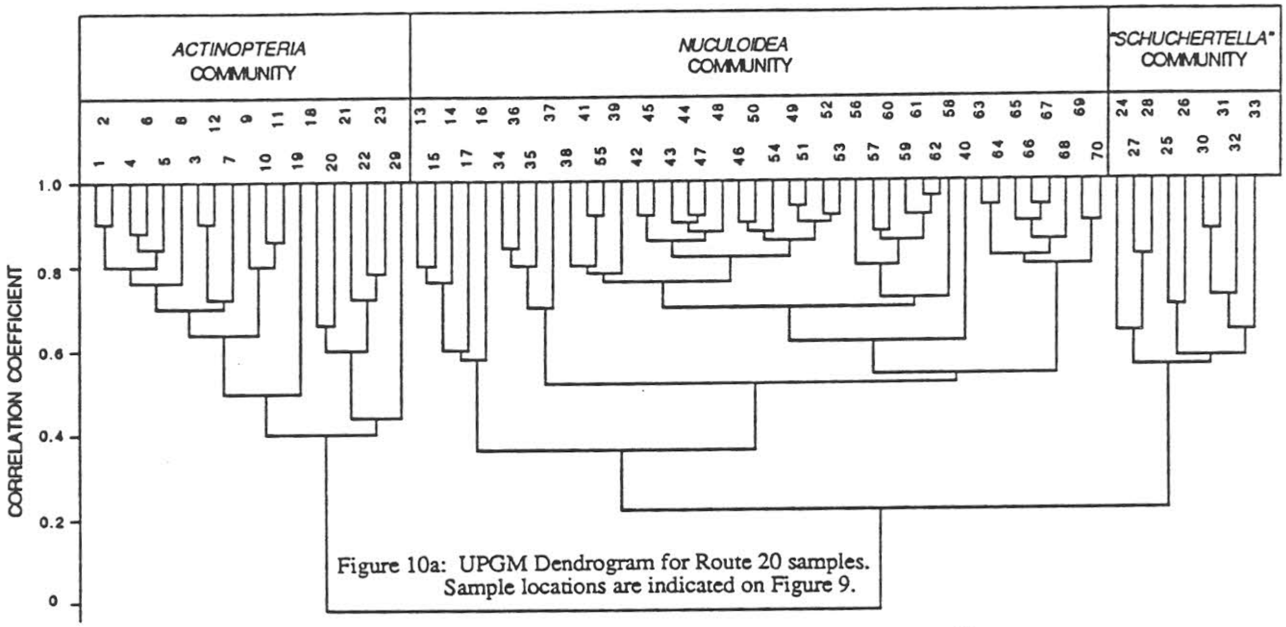


Figure 10a: UPGM Dendrogram for Route 20 samples. Sample locations are indicated on Figure 9.

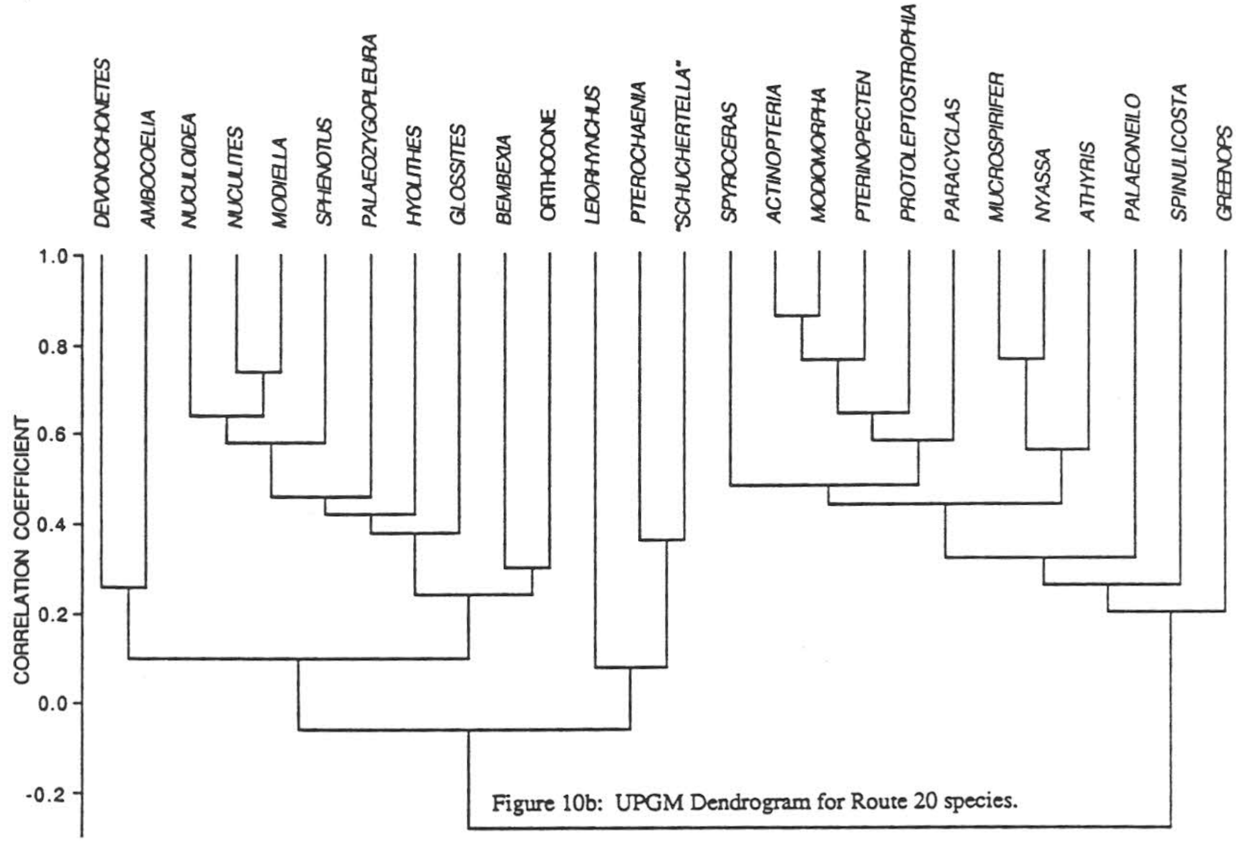


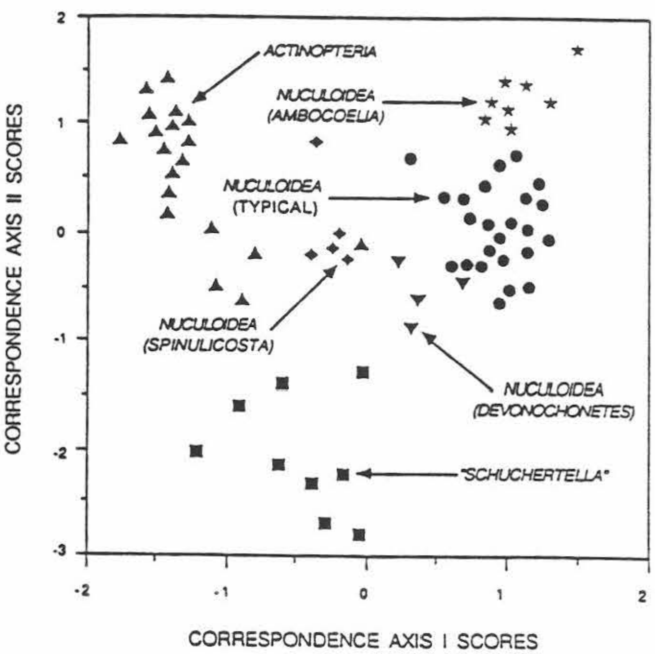
Figure 10b: UPGM Dendrogram for Route 20 species.

conditions, although some gentle agitation was necessary to maintain the numerous filter-feeders present. Most of the lower and several of the upper samples are assigned to this community (Fig. 12). Mobile infaunal deposit feeders make up 29 percent of this suite and form the most abundant ecological group. The two main constituents, Nuculoidea sp. and Nuculites oblongatus, are ecologically different inasmuch as they burrowed at somewhat different depths (see Brower and Nye, 1991 for references used to reconstruct living habits). Twenty percent of reclining filter-feeders, mostly small shells of Devonochonetes scitulus, are included. The three common endobyssate filter-feeders, Glossites, Modiella and Sphenotus, represent almost 19 percent of the community, and all three can be observed on a single bedding plane. All are small, thin-shelled types that were adapted for soft substrates. About 11 percent of pedicle-attached brachiopods can be counted.

Leiorhynchus and Ambocoelia are most common, but they probably did not compete because of differences in size and orientation on the seafloor. The larger leiorhynchids exploited the higher levels. Hyolithes and Greenops account for nearly 10 percent of the individuals, and these crawled along or ploughed through the mud, eating organic detritus, micro-organisms and small soft-bodied creatures. Gastropods comprise five percent, and their feeding habits are conjectural. The most frequent taxa, Bembexia and Palaeozygopleura, are obviously surface dwellers and they typically occur together in dark shales. Recent archaeogastropods of shallow waters are algal grazers, but those living in deep water are deposit feeders. We tentatively classify Bembexia and Palaeozygopleura as herbivores because lithologic and stratigraphic considerations show that these sediments were deposited at moderate depths within the reach of storm waves.

The subcommunities dwelled in somewhat different habitats. The subgroup with Spinulicosta and Actinopteria is the most diverse, observed in the coarsest sediments with the smallest amounts of organic matter, and always in close proximity to the Actinopteria community. This assemblage definitely lies at the shallow end of the Nuculoidea spectrum (Fig. 12). Although reasonably diverse,the Ambocoelia subdivision is dominated by comparatively few species and occurs in the finest sediments. This subcommunity grades into the true Ambocoelia assemblage, which is one of the deep water benthic communities of the Hamilton Group. The Ambocoelia subcommunity was clearly located in the deepest waters utilized by the Nuculoidea community. The ecological differences between typical samples of the Nuculoidea community and those with numerous Devonochonetes and Leiorhynchus are less certain. In terms of lithologic and faunal content, these samples are transitional between the typical Nuculoidea subcommunity and the Schuchertella community, which is consistent with stratigraphic position and the correspondence analysis scores (Fig. 12). These considerations suggest intermediate water depths.

Figure 11: Correspondence Analysis Plot for Route 20.



Schuchertella community

This assemblage is rich in reclining filter-feeders, especially Schuchertella and Devonochonetes, which make up over forty percent of the specimens. Many ecological categories and their constituents are intermediate between those of the Nuculoidea and Actinopteria communities, i.e. infaunal and epifaunal deposit feeders, buried filter-feeders and nektonic predators. Several features suggest that the Schuchertella community may have lived in an unfavorable environment; these are the abundance of small recliners along with the rarity of deeply-buried endobyssate bivalves and pedicle-attached brachiopods. Furthermore, the shallow-buried endobyssates are dominated by Pterochaenia, a small animal which could presumably attach to weak substrates. The common taxa are all characterized by small adult body sizes, and the animals may have been short-lived. The silty shale lithology is not consistent with soft and soupy sediment. The most likely unfavorable parameters include episodic sedimentation and/or some other aspect of the seafloor such as thixotropy. Samples of this community occur between those of the Nuculoidea and Actinopteria assemblages.

Figure 12a: Community progression using Correspondence Axis 1.

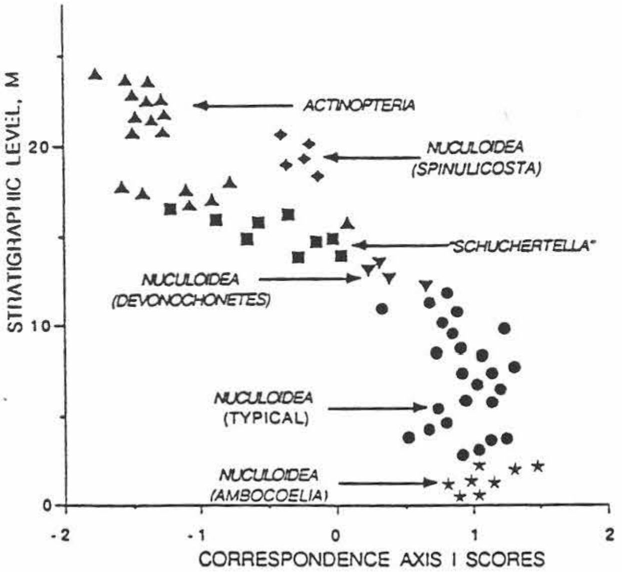
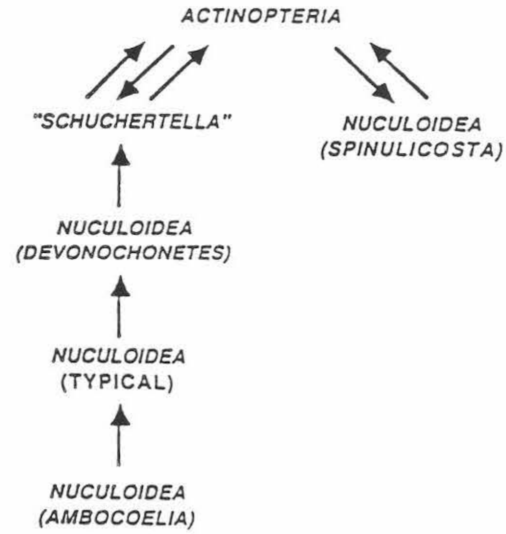


Figure 12b: Sequence of communities at Route 20.



Actinopteria community

The presence of siltstone and sandstone with sedimentary structures like cross-bedding and ripple marks testifies to shallow and agitated waters. Filter-feeders dominate the assemblage, comprising nearly 85 percent of all specimens. Much of the ecological diversity in this suite stems from different adaptations to filter-feeding. The low amounts of organic matter produced a diminished population of deposit feeders compared to the other communities. Most of the stratigraphically higher samples are assigned to this community.

The ubiquitous reclining filter-feeders lead the roster of ecological guilds present by representing 28 percent of the total individuals. The typical recliners are characterized by minimal profiles, which would minimize the danger of erosion by waves and currents. Spinulicosta proves an exception but its spines would have stabilized the shell in the sediment. Over 40 percent of the community is composed of byssate bivalves, ranging from surface-dwelling pectinaceans (Pterinopecten) and partially-buried taxa like Actinopteria to deeply buried animals such as Modiomorpha and Nyassa. Most of these species lived and fed at different levels, and shells were sometimes buried in living position. Aside from Athyris, Camarotoechia and some spirifers, brachiopods with pedicles are rare. The presence of many byssally attached bivalves and some pedunculate brachiopods denotes a firm and stable substrate in contrast to the soft muds of the Nuculoidea community. This is consistent with the clay matrix of these sandstones and siltstones. Nuculoidea and Palaeoneilo are most frequent in the reduced suite of deposit feeders.

Sequence of communities

Figure 12 summarizes the vertical changes by showing plots of the Axis 1 scores of the correspondence analysis versus stratigraphic level and the sequence of assemblages. The scores provide a generalized measure of relative depth and low values coincide with the most shallow samples. Although not completely available for sampling, the lower beds of the member contain the Nuculoidea subcommunity with many Ambocoelia and, thus, represent comparatively deep water. The overall pattern comprises regression or falling sealevel with progressively younger beds. Minor oscillations of depth are superimposed. The deep and shallow points are located at the following levels (Fig. 12): base of section deep, 1.0 m shallow, 2.2 m deep, 3.8 m shallow, 6.3 m deep, 17.8 m shallow, 20.2 m deep, and top of section shallow. At Pratts Falls, about 5.5 km west of the road cut, the uppermost part of the Actinopteria community is abruptly overlain by a thin siltstone or sandstone bed with phosphate pebbles which probably marks the transgression at the beginning of the next sedimentary cycle (see Brett and Baird, 1985; Brett, Baird and Miller, 1986 for discussion of these cycles). The pebble bed is not exposed at the road cut.

Variable sequences of communities are observed in the stratigraphic section. Each arrow denotes a change in fauna from older to younger (Fig. 12). For example, the typical Nuculoidea subcommunity can grade into two of the other Nuculoidea subcommunities. Similarly, the Actinopteria community can follow or precede the Schuchertella or the Spinulicosta brachiopod assemblage of the Nuculoidea community. The Schuchertella community is linked to both the Actinopteria community and the Nuculoidea subcommunity with many brachiopods like Devonochonetes and Leiorhynchus.

Although some assemblages can be followed by a deeper or shallower water environment, the general trend is towards more shallow conditions. It is important to note that the vertical changes are not biologically produced but are caused by physical changes in depth, agitation, substrate type, and possibly dissolved oxygen content.

Although relative depths can be inferred with confidence, absolute figures are less certain. Brett, Baird and Miller (1986) and Vogel, Golubic and Brett (1987) postulate depths of less than 20 m and approximately 50 m for communities that are similar to those of Actinopteria and Nuculoidea herein. Their estimates rest on a variety of faunal and lithologic criteria. The Niger delta most likely forms a reasonable sedimentological analog for the clastic Hamilton in the eastern part of New York (Mazzullo, 1973). The lithofacies suggest similar water depths of 10 to 20 m and 40 m for the Actinopteria and Nuculoidea communities, respectively (Allen, 1970). Assuming a linear relationship between depth and the Axis 1 scores for the correspondence analysis, the following depths were derived on a scale ranging from 10 to 50 m: Actinopteria community 10-31 m but mostly less than 21 m, Schuchertella community 17-31.5 m; subcommunities of the Nuculoidea community, samples with many Spinulicosta 27-30 m, Devonochonetes assemblage 34-40 m, typical type 35.5-47.5 m and beds with many Ambocoelia 42-50 m.

STOP 4, PRIVATE QUARRY ON ROAD NUMBER 4 EAST

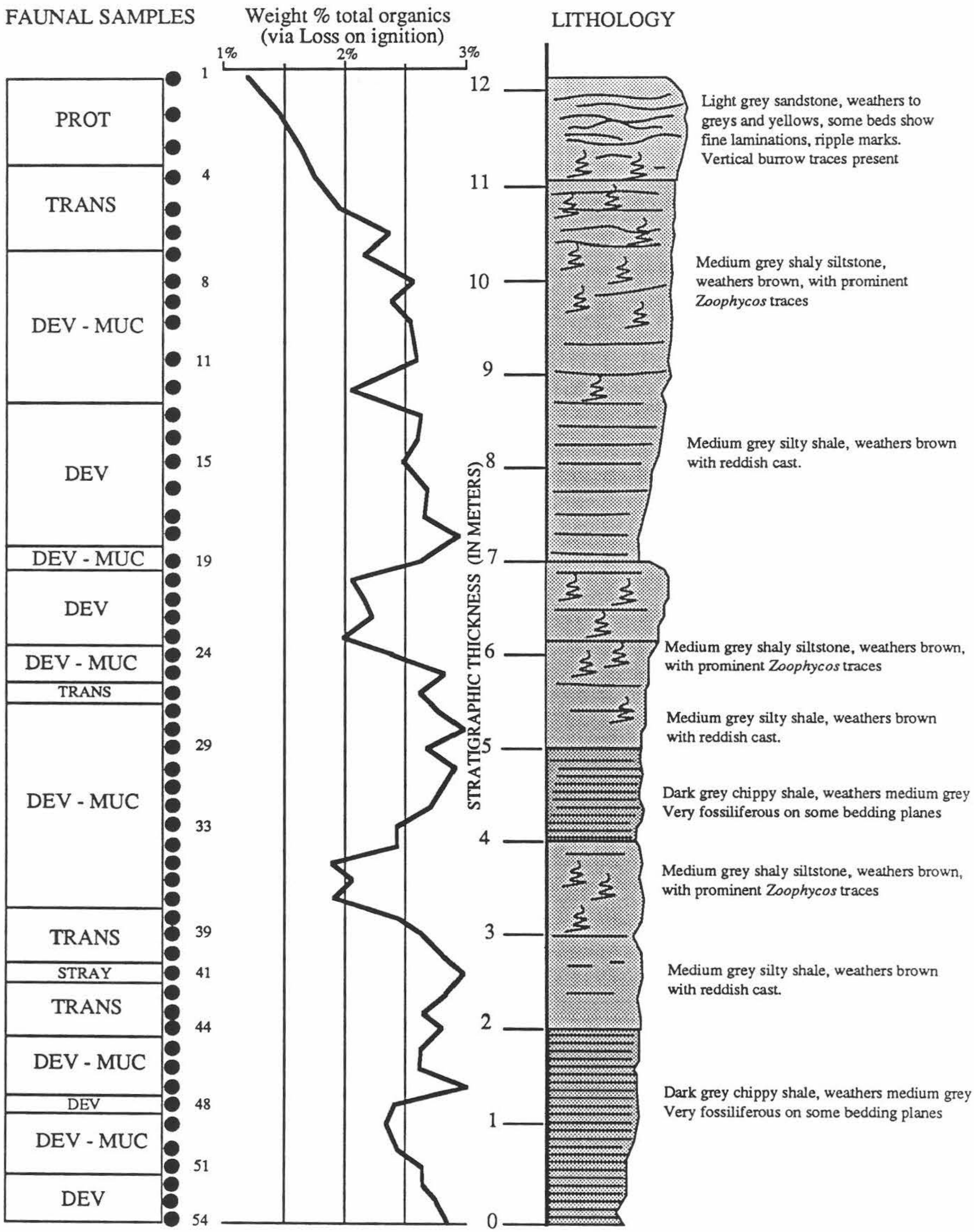
Introduction

About 12.5 m of the lowermost part of the Otisco Member of the Ludlowville Formation is present at this stop (Fig. 13). The lithologies range from shale to fine-grained sandstone and the organic matter content varies from 3.0 percent in some of the shales to 1.2 percent in the uppermost sandstone. Examination of the lithology indicates that three upwardly-coarsening cycles can be seen; the tops of these cycles are at roughly 4 m, 7 m and 12.5 m. Each cycle begins with shale; the two lower cycles are capped by a siltstone that contains abundant Zoophycos caudagalli trace fossils, whereas the upper one ends in sandstone containing a large brachiopod and bivalve fauna.

Three faunal associations were initially apparent in qualitative field studies: 1) A Devonochonetes association characterized by infaunal, deposit-feeding bivalves, gastropods, and abundant Devonochonetes scitulus. 2) A Mucrospirifer association containing a higher diversity fauna of semi-infaunal and epifaunal brachiopods and bivalves. 3) A Protoleptostrophia association dominated by large reclining brachiopods and bivalves. However, this is rather misleading. As in the case of the Swamp Road, the multivariate analyses based on clustering and correspondence analysis reveals a slightly different paleoecological structure.

Large samples were taken — each sample consists of 1000 individuals in order to study the effects of sample size on faunal diversity and proportions. These results suggest that a sample size of 300 individuals is adequate to characterize these parameters for most diverse Hamilton Group communities. The 54 samples are distributed throughout the stratigraphic section (Fig. 13).

Figure 13 : Stratigraphic section for Road Number Four East, Otisco Member, Ludlowville Formation.



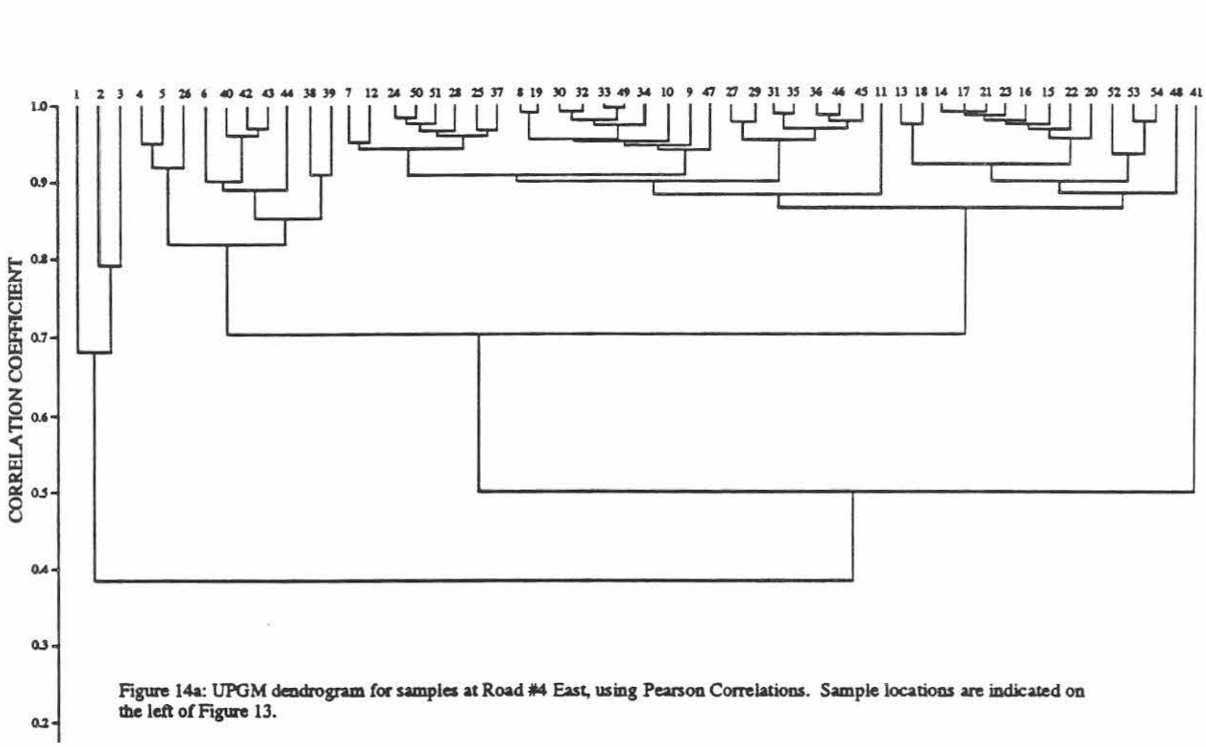


Figure 14a: UPGM dendrogram for samples at Road #4 East, using Pearson Correlations. Sample locations are indicated on the left of Figure 13.

Cluster analysis

The dendrogram for the samples shows one stray sample and three main clusters in which most of the samples join at correlations of 0.8 and above (Fig. 14a). Sample 41 is anomalous because of the high abundance of Tropidoleptus carinatus, Spinulicosta spinulicosta, and Camarotoechia congregata which occur in clusters at this site. The small cluster with samples 1 to 3 represents the Protoleptostrophia fauna which is only found in the sandstone at the top of the outcrop. Common constituents are the brachiopods Mucrospirifer mucronatus, Spinocyrtia granulosa, and Devonochonetes coronatus and the bivalve Actinopteria boydi. Although the cluster is clearly identified on the dendrogram, the links between the three samples are relatively low at 0.68 and 0.79 which is related to wide variation of faunal proportions in the samples.

The large cluster extending from samples 7 to 48 on the dendrogram is highly heterogeneous. The most common species is almost invariably Devonochonetes scitulus. Mucrospirifer and Tropidoleptus are usually also common. This cluster can be divided into two groups. The samples ranging from 13 to 48 in the cluster are dominated by Devonochonetes scitulus with comparatively small numbers of Mucrospirifer and Tropidoleptus. Infaunal deposit-feeding bivalves, Nuculoidea sp., Nuculites oblongatus and Palaeoneilo sp. are abundant and make up roughly eight percent of the assemblage. These samples would be assigned to the Devonochonetes community of most authors. Mucrospirifer and Devonochonetes co-dominate the subcluster composed of samples 7 to 11 on the dendrogram. In addition, Tropidoleptus is more common and the infaunal deposit-feeding bivalves are rare. For the present, we refer to this fauna as the Mucrospirifer-Devonochonetes community.

The block of samples from 4 to 39 on the cluster diagram is transitional between the Mucrospirifer-Devonochonetes community and the Protoleptostrophia fauna in some respects. For example, the amounts of Devonochonetes scitulus and Protoleptostrophia are intermediate between those of the previous two clusters whereas Mucrospirifer is about equally common in all three groups of samples. However, Tropidoleptus is most frequent in the transitional samples.

As usual, the clusters for the species are more complex (Fig 14b). Three of the 26 links (11.5 percent) take place at correlations of 0.95 and above. However, the other joins are all below 0.7. Nevertheless, several categories of taxa can be recognized, namely forms that are typical of fine-grained sediments, those mostly occurring in coarser rocks, rare species, and organisms that are widely or erratically distributed.

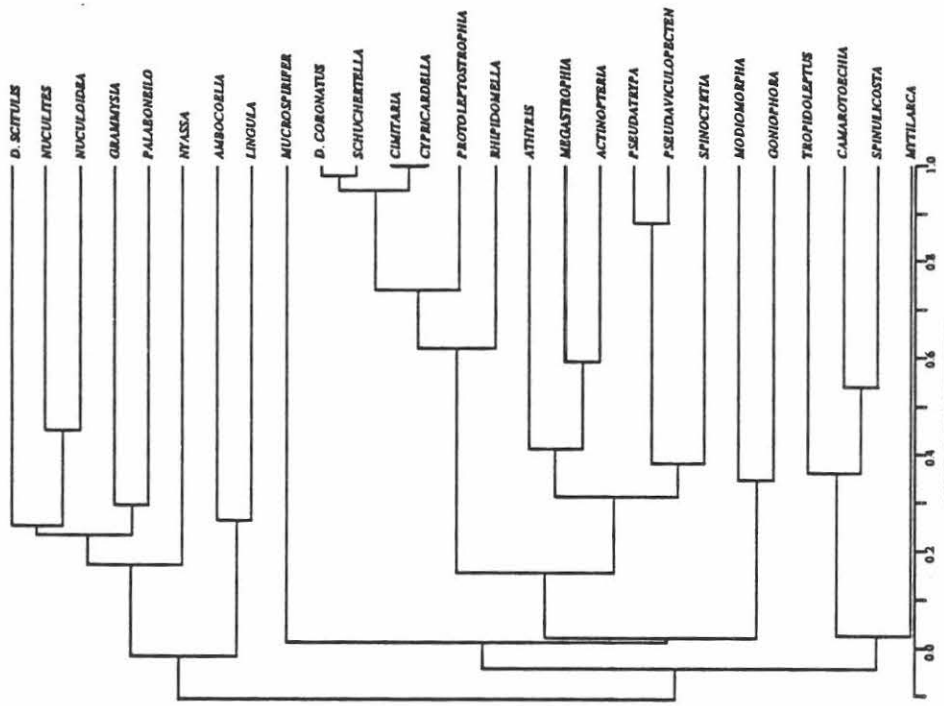
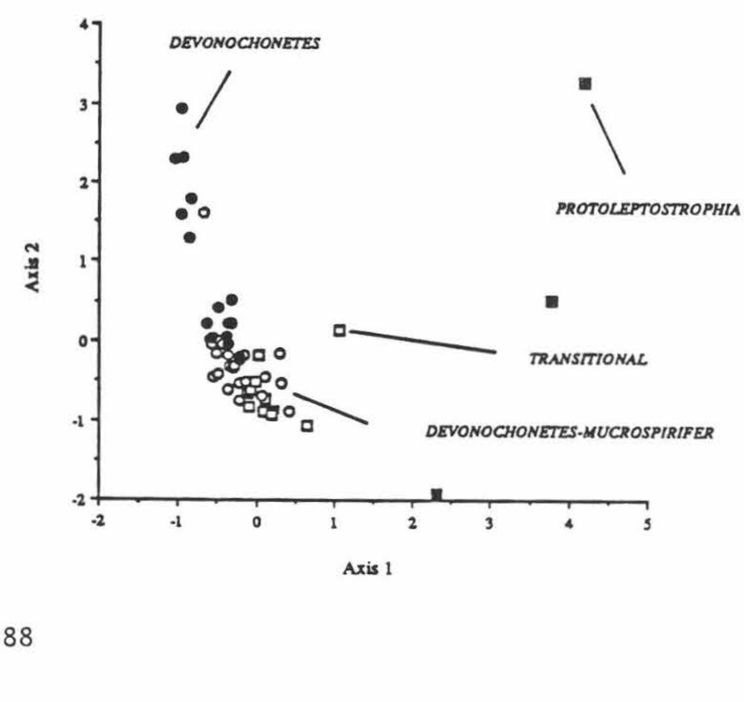


Figure 14b: UPGM dendrogram for species at Road #4 East, using Pearson Correlations.

Correspondence analysis

The first two axes of the correspondence analysis for the samples are graphed in Figure 15, and they explain 36 percent of the variation in the data. The different clusters and communities are mostly concentrated in different regions of the plot with various amounts of overlap being observed between adjacent groups of samples. The three samples with the Protoleptostrophia fauna are strongly segregated from the rest. Although only three samples are involved, a wide region of the plot is occupied because these samples vary considerably in faunal composition. Although most samples with the Devonochonetes

Figure 15: Correspondence Analysis Plot for Road #4 East.



community fall in the upper left part of the plot, these samples exhibit considerable overlap with those of the Devonochonetes - Mucrospirifer community in the lower left. Likewise, much intergradation takes place between the Devonochonetes - Mucrospirifer community and the transitional group. The correspondence analysis plot arranges the communities in the following order: Devonochonetes community, Devonochonetes - Mucrospirifer community, transitional group, Protoleptostrophia fauna. As mentioned later, this order is closely related to the stratigraphic sequence of communities.

Cluster significance tests

The cluster significance tests on the adjacent pairs of communities denote that all clusters comprise samples that represent samples that have overlapping rectangular distributions at the 0.05 alpha level. At least some of the communities or clusters intergrade widely and the parent populations may overlap as much as 50 percent if tested for Gaussian distributions at a probability of 0.05. These Hamilton communities are clearly not discrete.

Sequence of communities

The samples on the stratigraphic section (Fig. 13) are categorized in terms of the communities and the vertical changes of the communities are schematically tabulated in Figure 16. Note that the stray sample was omitted from the latter figure. Numerous alternations are shown by two pairs of communities, namely Devonochonetes community to Devonochonetes - Mucrospirifer community, and Devonochonetes - Mucrospirifer community and transitional cluster. The three samples with the Protoleptostrophia assemblage are underlain by the transitional fauna. The lithologies suggest that the various communities tended to occupy different depth and agitation zones, ranging from Devonochonetes community as deepest and quietest to Devonochonetes - Mucrospirifer community to transitional group to Protoleptostrophia fauna as the most shallow and highly agitated on average. The Devonochonetes community is found in the finest shales with the highest organic matter contents. The Protoleptostrophia fauna is restricted to the fine sandstones with less than 1.7 percent of organics. Other changes probably also took place, for example differences in substrate texture, oxygenation, degrees of bioturbation, and amounts of alternation between relatively rough and quiet water intervals. However, these changes are at least generally correlated with and linked to water

Figure 16a: Community progression using Correspondence Axis 1.

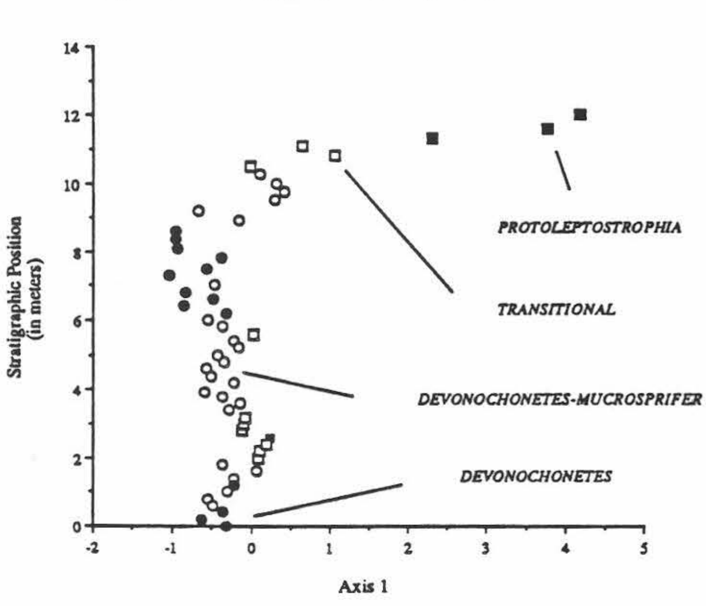
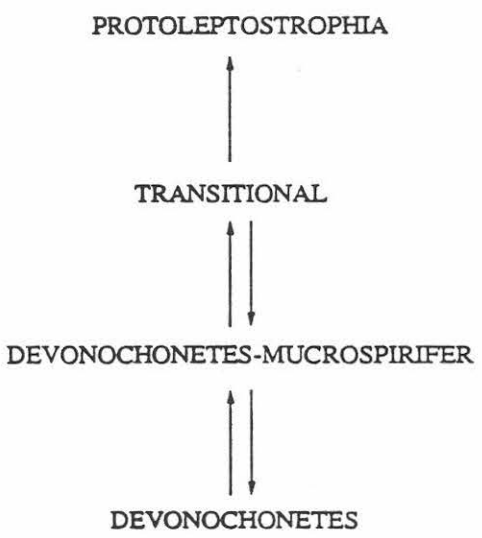


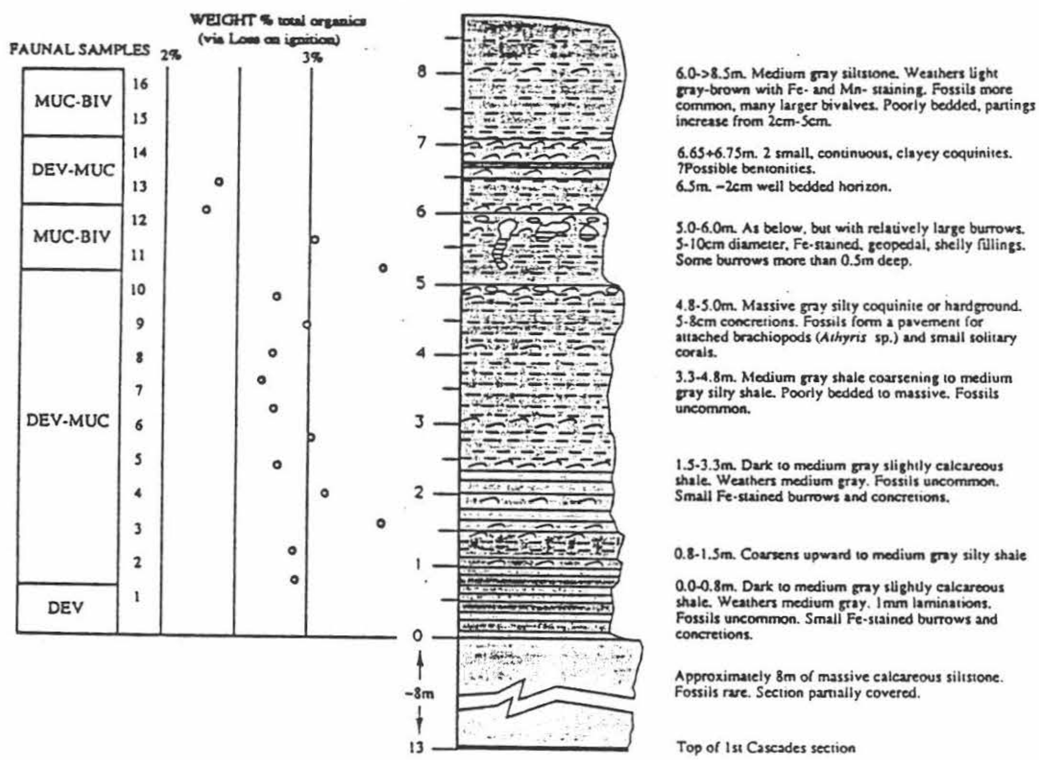
Figure 16b: Sequence of communities at Road #4 East.



depth and agitation. As with the other exposures, the scores for Correspondence Axis 1 appear to provide a surrogate measure for relative depth. The high scores are grouped with the most shallow samples whereas low values are observed for the samples from deeper water (Fig. 16). The major pattern noted for the entire outcrop is generally deep water followed by a marked interval of regression. Here, the asymmetery is quite striking. The correspondence axis scores imply that the Protoleptostrophia fauna lived in much more shallow water than the other communities. Two comparatively symmetrical sequences of regression and transgression are superimposed on the samples from the deeper water. The change of symmetry is notable. The communities are variably developed in different parts of the outcrop. In the two lower regressive and transgressive cycles, the main changes, omitting minor oscillations, are Devonochonetes community to Devonochonetes - Mucrospirifer community to transitional assemblage to Devonochonetes - Mucrospirifer community for the lower cycle and Devonochonetes - Mucrospirifer community to transitional cluster to Devonochonetes - Mucrospirifer community to Devonochonetes community in the second cycle. The full sequence of Devonochonetes community to Devonochonetes - Mucrospirifer community to transitional group to Protoleptostrophia fauna is only present in the uppermost regressive part of the outcrop.

An interesting side note is that our initial qualitative and quantitative analyses on the Road Number 4 outcrop yield communities that are quite similar; here, the style of preservation does not introduce a large amount of bias during collecting. The situation at Swamp Road is greatly different. Our qualitative and quantitative "communities" were not the same because of the collecting bias which leads one to preferentially find the well-preserved calcitic specimens and ignore the comparatively drab molds and casts.

Figure 17a: Stratigraphic section for Upper Cascade, Ivy Point Member of the Ludlowville Formation, Hamilton Group

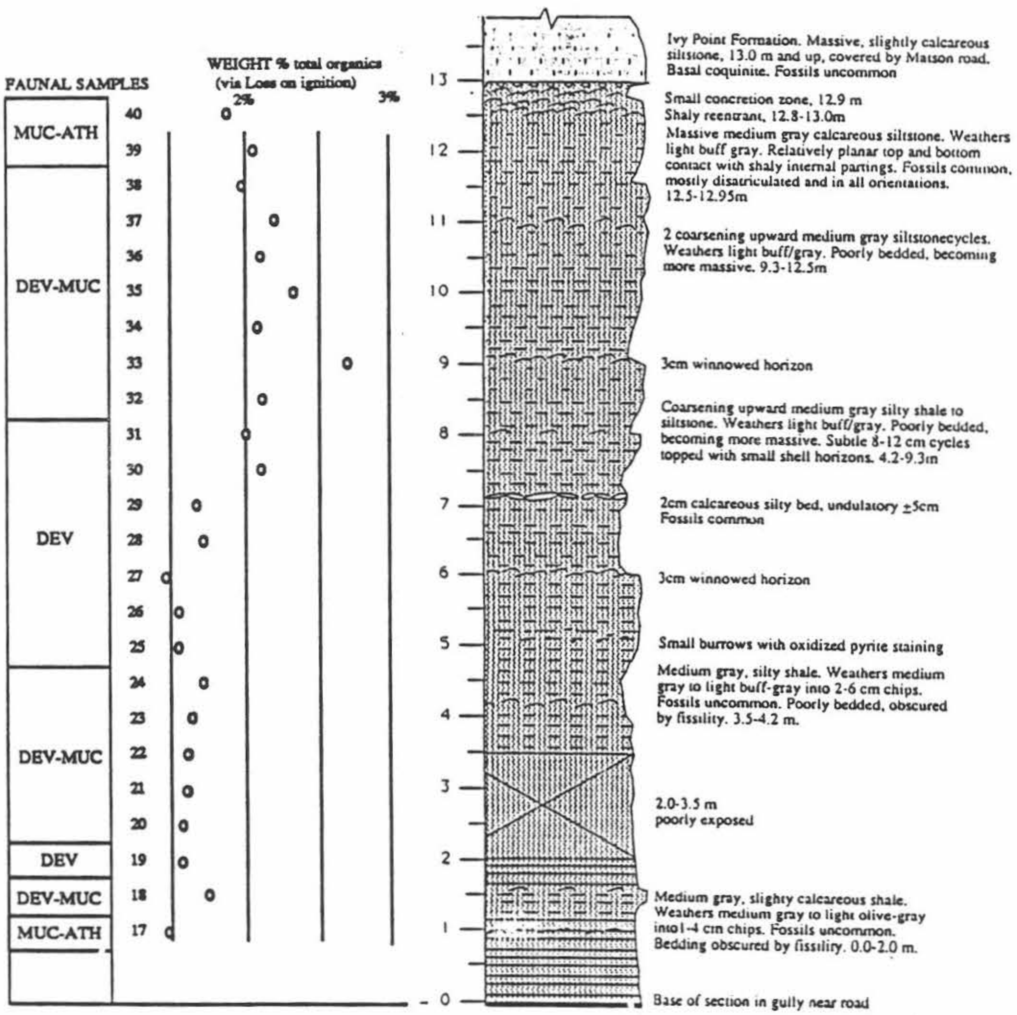


STOP 5, ROADCUT ON ROUTE 38 AT CASCADE

Introduction

Two asymmetrical cycles which coarsen upwards were sampled at this outcrop; an upper one in the Ivy Point Member of the Ludlowville Formation (Fig. 17a) and a lower one in the Otisco Member (Fig. 17b). Brett, Baird and Miller (1986) published a complete stratigraphic section for this roadcut. The two cycles were selected because large areas were available for sampling which contain representative samples of the overall fauna. A variety of lithologies are present which vary from dark grey shale with fine laminations and comparatively high amounts of organic matter to fine grey siltstones where organics are relatively sparse. Three hundred fossils were counted in each sample. Inasmuch as only 24 and 16 samples were collected from the lower and upper cycles, respectively, they were amalgamated to give a larger data set with 40 samples. In contrast to the other outcrops examined, two sources of variation reside in the data, namely changes within the cycles as well as contrasts between them.

Figure 17b: Stratigraphic section for Lower Cascade, Otisco Member of the Ludlowville Formation, Hamilton Group.



Cluster analysis

The dendrogram for the samples from both cycles can be seen in Figure 18a. Samples from both cycles are keyed into the stratigraphic sections in Figure 17a&b. The samples from 1 to 29 of the clustering diagram belong to the Devonochonetes community which is known from the finer shales in the two cycles. Devonochonetes scitulus dominates the assemblage with Mucrospirifer mucronatus and Nuculoidea next in frequency. Significant amounts of Tropidoleptus carinatus and Athyris spiriferoides are also encountered.

The next two clusters include the samples from numbers 2 to 10 and numbers 18 to 38. Devonochonetes scitulus and Mucrospirifer are co-dominant and these samples are placed in the Devonochonetes - Mucrospirifer community. The group ranging from samples 2 to 10 is restricted to the upper cycle whereas the other one hails from the lower cycle. The lower cycle group is

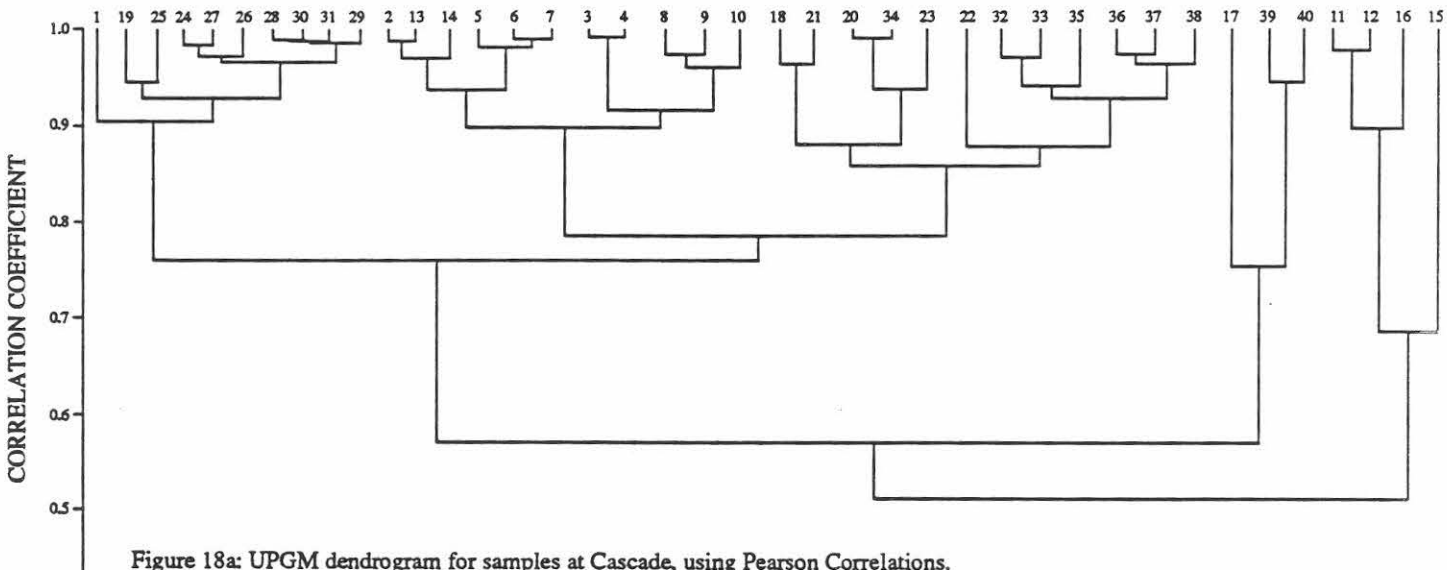


Figure 18a: UPGM dendrogram for samples at Cascade, using Pearson Correlations. Sample locations are indicated on Figure 17.

comparatively rich in Athyris whereas the upper one contains more abundant Tropidoleptus. The lower cycle segment of this community lies in slightly finer sediments than the cluster from the upper cycle. Observe that these groupings relate to differences between rather than within the cycles. The contrasts in grain size also suggest somewhat different depositional regimes.

The three samples in the small cluster with numbers 17, 39 and 40 are from the base and top of the lower cycle. The enclosing rocks are calcareous siltstones where the calcium carbonate content exceeds that of the other rocks in both cycles. The most abundant forms are Mucrospirifer and Athyris with smaller amounts of Devonochonetes scitulus, Rhipidomella vanuxemi, Nuculoidea and Cypricardella. This represents a Mucrospirifer - Athyris community which is much more common in western than in central New York.

The cluster with samples 11 to 15 in Figure 18 is only known in the lower cycle. Mucrospirifer is abundant along with numbers of Spinocyrtia granulosa, Tropidoleptus, Actinopteria and Modiomorpha and is only found in the coarser rocks (siltstones) with little organic matter or calcium carbonate. This common Hamilton community will be termed the Mucrospirifer - bivalve community.

As usual, the dendrogram for the species is rather nebulous (Fig 18b) and about 21 percent of the links take place at correlations of 0.7 and higher. Many of the associations hover around values of nil which indicate no association at all. Despite this, the order of the taxa along the dendrogram generally distinguished the deeper water species in the finer shales from those most common in coarser sediments which were probably deposited in more shallow water.

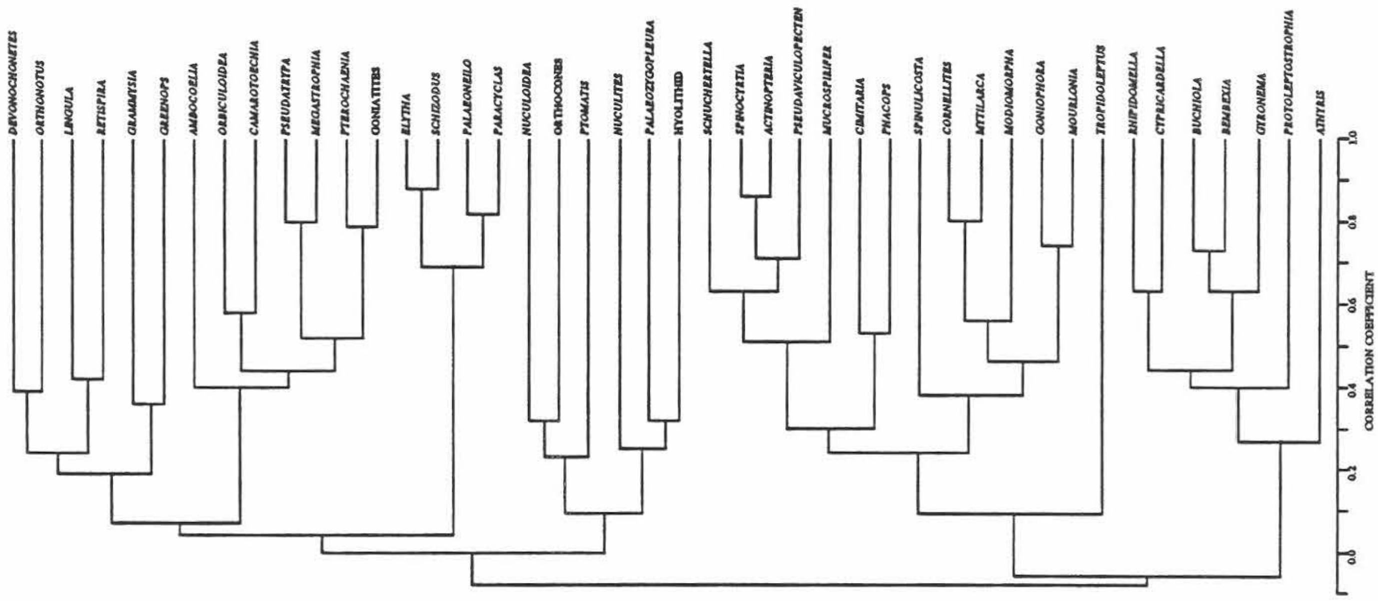
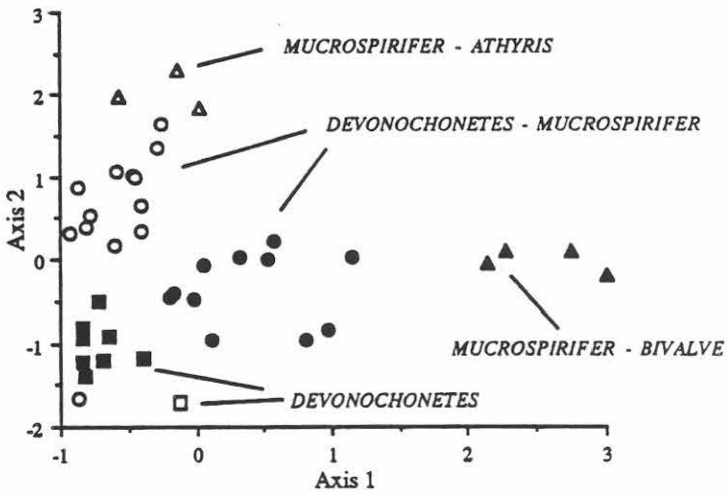


Figure 18b: UPGM dendrogram for species at Cascade, using Pearson Correlations.

Correspondence analysis

The first two correspondence analysis axes explain 42.5 percent of the variance in the data. The communities show little or no overlap on the diagram (Fig. 19). Axis 1 basically polarizes the community in the coarsest rocks, Mucrospirifer - bivalve, from the others which are characteristic of the finer rocks. To some extent, the Devonochonetes - Mucrospirifer community from the upper cycle occupies an intermediate position along Axis 1. The second axis arrays the clusters from the Devonochonetes community with the lowest scores to that of Mucrospirifer - Athyris with the highest values. Although, the results from the correspondence analysis are not yet fully interpreted, the data are consistent with those derived from the cluster analysis.

Figure 19: Correspondence Analysis for Cascade. Black symbols are for samples from Upper Cascade.



Cluster significance tests

Adjacent pairs of communities can be inferred from the plot of the sample scores for the first two correspondence analysis axes in Figure 19 as well as from the vertical sequence of communities discussed next. The observed measures of overlap denote that the following pairs of communities are drawn from parent rectangular distributions that do not intergrade at the 0.05 probability level (note that in all cases, the results are only marginally significant at this alpha level): Devonochonetes versus Devonochonetes - Mucrospirifer from the lower cycle; Devonochonetes versus Devonochonetes - Mucrospirifer from the upper cycle; Devonochonetes - Mucrospirifer from the lower and upper cycles; Devonochonetes - Mucrospirifer from the upper cycle versus Mucrospirifer - bivalve. The only community which overlaps its nearest neighbor at a risk level of 0.05 is the Mucrospirifer - Athyris community and the segment of the Devonochonetes - Mucrospirifer community from the lower cycle. This pattern contrasts with that seen at the Road Number 4 outcrop which is also from the Ludlowville Formation where all adjacent communities tested were clearly from overlapping rectangular distributions at 0.05 probability. Thus, the communities at Cascade seem to be more highly structured than at Road Number 4. These differences will be subjected to further investigation.

Sequence of communities

The lithologies and organic matter contents lead one to conclude that the communities are mostly related to changes in depth. The Mucrospirifer - bivalve community is associated with the coarsest sediments and least amounts of organic matter. Conversely the Devonochonetes community resides in the finer and darker shales with the most organics. The Devonochonetes - Mucrospirifer community and its intermediate lithologies probably occupied a wide range of intermediate depths along the gradient. The exact position of the Mucrospirifer - Athyris community is somewhat uncertain but it most likely falls within the depth range of the Devonochonetes - Mucrospirifer community, although in somewhat more calcareous settings. The Mucrospirifer - bivalve community is associated with the coarsest siltstones with the lowest amounts of organic matter and low calcium carbonate contents.

Figure 20a: Community progression at Upper Cascade using Correspondence Axis 1.

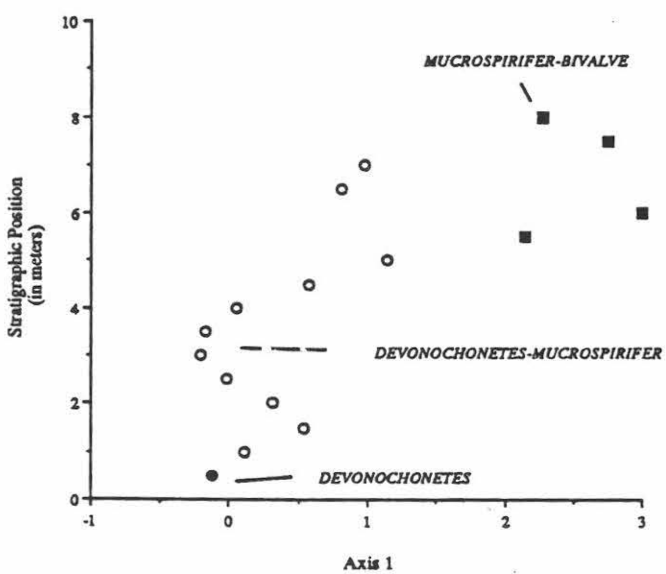


Figure 20b: Sequence of communities at Upper Cascade.

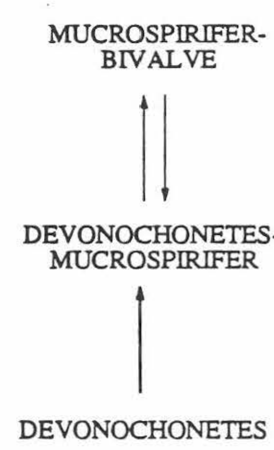


Figure 21a: Community progression at Lower Cascade using Correspondence Axis 1.

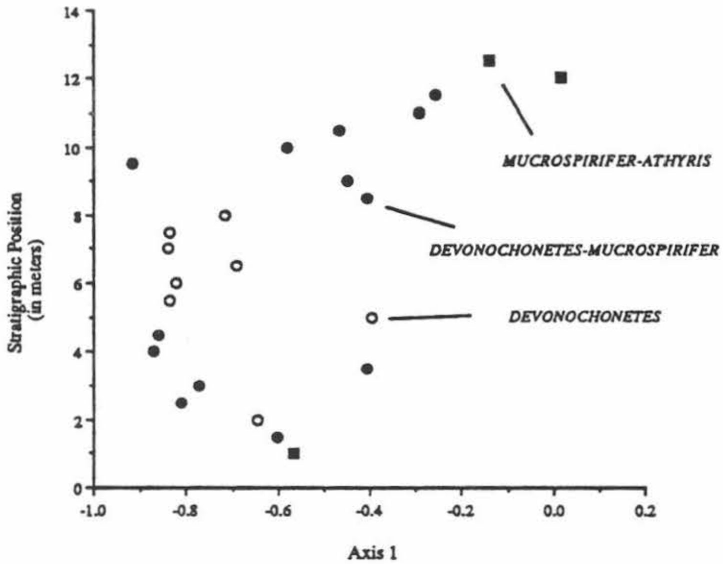
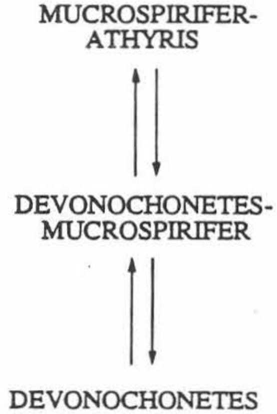


Figure 21b: Sequence of communities at Lower Cascade.



The community sequence and allied data are most clear for the upper cycle (Fig. 20a&b). An initial sample with the Devonochonetes community is followed by numerous samples with the Devonochonetes - Mucrospirifer community. Next are two samples, each with the Mucrospirifer - bivalve assemblage and the Devonochonetes - Mucrospirifer community. The section is capped by the Mucrospirifer - bivalve assemblage. As previously, the Correspondence Axis 1 scores for the samples are thought to be linked to relative depth and the most shallow samples have the highest scores. The plot of the Axis 1 scores against stratigraphic position (Fig. 20a) shows the overall effect of falling sealevel. There are two regressive - transgressive cycles superimposed on the main theme. As at Road # 4, the major changes are located in the upper part of the cycle.

In terms of lithologic and faunal changes the lower cycle is more subtle (Figure 21a&b). Lithologies span the interval between silty shale and siltstone. The calcareous siltstones with the Mucrospirifer - Athyris community are also in this cycle. Community composition is dominated by alternations between the Devonochonetes and the Devonochonetes - Mucrospirifer assemblages but several samples with the Mucrospirifer - Athyris community are encountered at the base and top of the lower cycle. Assuming that the scores for the samples on Correspondence Axis 1 provide a surrogate parameter for relative depth, the older part of the cycle represents deeper water with many minor oscillations of depth whereas the upper part of the cycle records a marked shoaling interval.

SUMMARY AND CONCLUSIONS

During this study, six typical cycles of upwardly coarsening clastic sediments have been examined in the Hamilton Group of central New York. Sedimentary sequences were selected from the Marcellus, Skaneateles, and Ludlowville Formations which represent the lower and upper parts of the group. Closely-spaced and large samples, mostly composed of 300 or 1000 specimens, were taken in order to characterize faunal diversity, proportions and ecological structure. The following conclusions are drawn:

1. Samples of 300 specimens are large enough to characterize the most diverse Hamilton communities in this area.

2. The main pattern in all cycles consists of decreasing depth and correlated differences as evidenced by change in sediment grain size, sedimentary structures, amount of organic matter, and faunal composition.

3. The multivariate techniques of cluster analysis and correspondence analysis are successful in the identification of faunal communities or assemblages. They also point to relations between and within communities. The ordination method of correspondence analysis and the clustering algorithms produce complimentary data, and both types of techniques should be employed in concert. Once the analyses are complete, it will be possible to specify average composition and variances for the various communities.

4. Statistical significance tests at the 0.05 probability level have been performed on all pairs of related communities as ascertained from the correspondence analysis plots and the known vertical sequences of communities. As expected, the multivariate means of all communities are statistically different, and all communities differ in average composition. Significance tests on clusters were also employed on the data to test the amounts of overlap in the populations from which the samples were derived. In two locations, Route 20 and Cascade, most or all of the communities comprise samples from rectangular distributions that do not overlap along the observed faunal gradients. These communities seem to have been tightly structured. At the other outcrops, Swamp Road, Cheese Factory Gulf, and Road #4, the pairs of communities come from populations that are characterized by overlapping rectangular distributions. In fact, the parent populations may exhibit as much as 50 percent of overlap (calculated for Gaussian or normal distributions) although lower percentages are more typical. Widely overlapping communities are deemed to be weakly structured and, perhaps, poorly integrated. The significance tests are still being explored.

5. The scores for the samples along the first correspondence axis extract the main pattern of change and relations between the communities for each outcrop and cycle treated. As mentioned previously, the cycles were probably produced by decreasing depth or shoaling, so these scores yield a generalized measure of relative depth. The exact relation between the first correspondence axis and depth cannot be specified at present, and the relationship could be rank order, linear or nonlinear.

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TRIP B-1: QUANTITATIVE PALEOECOLOGY OF HAMILTON GROUP LOCALITIES IN CENTRAL NEW YORK.

ALL OF THE SECTIONS OBSERVED ON THIS TRIP ARE EITHER STEEP OUTCROPS OR ON RELATIVELY WELL TRAVELED ROADS OR BOTH! PLEASE BE ALERT AND TAKE NECESSARY PRECAUTIONS.

Start	0.0 mi.	Morrisville, NY. Paired Stop Lights on Route 20. Proceed North on Cedar Street which becomes Swamp Road
2.7	2.7	**STOP 1: Swamp Road outcrop** on right, park either in small quarry on right or off the road on the left.
2.7	5.4	Return south to Morrisville and Route 20, turn right (west). Proceed west over hill and dale through
6.7	12.1	Nelson,and
3.4	15.5	Cazenovia. As you drive through town
1.0	16.5	Route 20 turns south (left), then west (right) to get around Cazenovia Lake.
0.2	16.7	Junction of Route 13 and 20
0.7	17.4	Junction of Route 20 and 92 Continue west on Route 20. Proceed down steep hill and cross valley floor. Turn south (left) onto Oran-Delphi road
2.5	19.9	Turn south (left) onto Oran-Delphi road
0.8	20.7	Turn west (right) onto Cheese Factory Gulf road (marked Gulf road)
0.7	21.4	Proceed up this road past sharp right turn and lower Cardiff
0.2	21.6	exposure, past the culvert at bottom of the ravine section to
0.1	21.7	park at the wide spot in the road. **STOP 2: Cheese Factory Gulf** slide section. Walk back to ravine section. Be aware of Poison Ivy on creek banks.
1.4	23.1	Continue uphill on Gulf road (north) to return to Route 20
0.3	23.4	Turn east (right) onto Route 20 for a short distance, and park off the road near the yellow warning signs for the hill, but not too far onto the farmers field. **STOP 3: Route 20 outcrop** is on both sides of the hill below.
1.0	24.4	Turn vehicles around and go west on Route 20 to Pompey Center, turn south (left) onto Pompey Center road.
2.2	26.6	Turn east (left) onto Road #4 East
0.3	26.9	Private **STOP 4: Road #4 quarry** is on the south side of the road at the crest of the hill. The present owner lives across the road.
2.5	29.4	Backtrack to Route 20.
32	62	To continue to the Cascade outcrop, drive approximately 32 miles west on Route 20 through Lafayette, Skaneateles, to Auburn. Turn south onto Route 38 south.
14	76	Continue south on Route 38 approximately 14 miles to Cascade, NY, at the south end of Owasco Lake. **STOP 5: Cascade outcrop** is on the west (right) side of the road along a half-mile stretch at the base of the hill. Park on the edge of the highway.

THE LITTLE FALLS DOLOSTONE (LATE CAMBRIAN): STRATIGRAPHY AND MINERALOGY

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The Late Cambrian Little Falls Dolostone is well known for the occurrence of exceptionally clear, doubly terminated quartz crystals, known as "diamonds", and for the irregular hemispherical masses of algal stromatolites, known as "cryptozoons". The areal relations of the Little Falls Dolostone in the field trip area is shown in Figure 1.

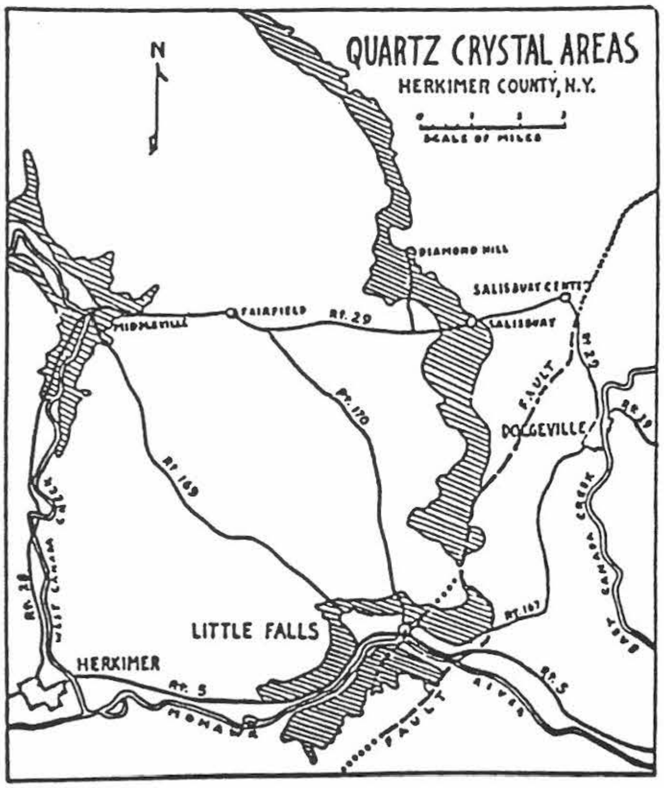

Figure 1.

Areal relations of the Little Falls Dolostone (diagonal lines) in the Little Falls, New York, region. (From Tuttle, 1973, after Cushing, 1905).

Clarke (1903, p. 16) formally designated the Little Falls Dolostone for the "...highly magnesian, sparsely fossiliferous phase of the Calciferous Sandrock in the Mohawk Valley...". The type area was first indicated by Hartnagel (1912, p. 29) as "...the pass in the Mohawk Valley at Little Falls, Herkimer County". Zenger (1976) designated a composite type section (Figure 2), and his definition is reproduced here in its entirety for the convenience of the reader.

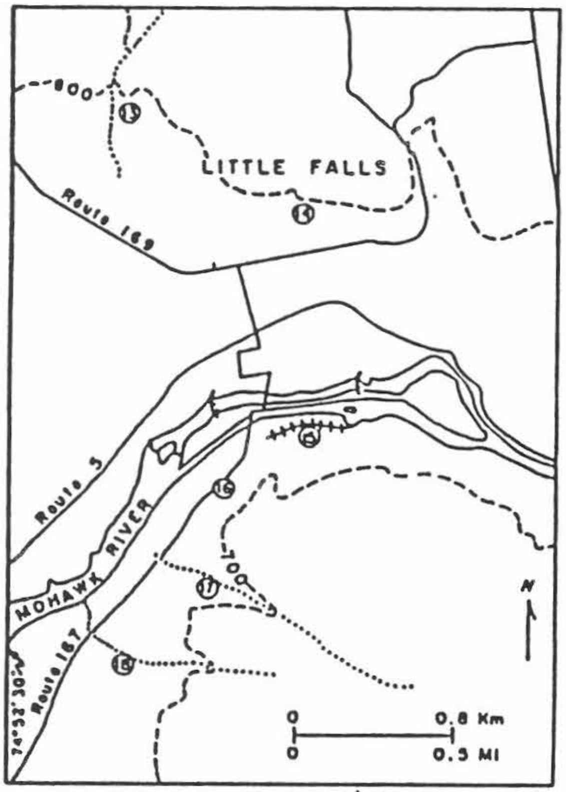

FIG. 2—Sketch map showing location of composite type section at Little Falls. Locality numbers referred to in text. Except for outline of Mohawk River, solid lines are roads; dashed lines are contour lines; dotted lines are streams.

(From Zenger, 1976, p.1571)

DEFINITION OF TYPE SECTION

(From Zenger, 1976, p.1571)

The composite type section consists of the following.

Loc. 13—Section exposed along south-flowing stream south of Little Falls Reservoir and east of Top Notch Road, in and above Buttermilk Falls; W ½ Little Falls 7 ½-minute quadrangle. Upper 90 ft (29 m) of Little Falls and contact with overlying Tribes Hill Formation (Ordovician) exposed at about 750-ft (230 m) elevation.

Loc. 14—Partly covered section exposed along secondary road (leading to Burrell's Mansion) and along hillside north of road beginning at elevation of about 590 ft (180 m) at northern end of Williams Street; W ½ Little Falls 7 ½-minute quadrangle. Approximately 205 ft (62 m) of section with neither lower nor upper contact exposed.

Loc. 15—Exposures at top of south side of railroad cut just south of old Jefferson Street School, on south side of Mohawk River and Barge Canal; top of cut at about 400-ft (122 m) elevation; northernmost part of SW ½ Little Falls 7 ½-minute quadrangle. Exposed is basal part of Little Falls and nonconformable contact with underlying Proterozoic gneiss.

Loc. 16—Section exposed in steep face just south of New York Route 167, 0.5 mi (0.8 km) east of north-flowing tributary at loc. 17 (see below); base of measured section at about 410-ft (125 m) elevation; SW ½ Little Falls 7 ½-minute quadrangle. About 65 ft (20 m) of lower Little Falls exposed.

Loc. 17—Section along northwest-flowing tributary of Mohawk River, directly southwest of steep bluff termed "Rollaway;" measured section begins at 420-ft (129 m) elevation, few hundred feet southeast of Route 167; SW ½ Little Falls 7 ½-minute quadrangle. Nearly 300 ft (92 m) and contact with overlying Tribes Hill exposed.

Loc. 18—Section along another northwest-flowing tributary of Mohawk River, about 1.2 mi (1.9 km) southwest of city hall and 0.5 mi (0.8 km) west of loc. 17; section begins just southeast of New York Route 167 at about 410-ft (125 m) elevation; SW ½ Little Falls 7 ½-minute quadrangle. About 200 ft (61 m) of middle part of formation exposed.

The lithology of the Little Falls Dolostone is quite varied. Although no one lithology is typical of the entire formation, dolostone is predominant, with sandstone and mixed dolostone-sandstone varieties. The beds are commonly gray, medium-to thick-bedded, with vuggy beds containing dolomite, calcite, anthraxolite, and of course the famous quartz crystals.

With the exception of the algal stromatolites, the Little Falls Dolostone very rarely yields fossils. The algal structures are more common in informal unit B, and range in thickness from less than a foot to three feet. Note there are several different stratigraphic levels of occurrence. As for other fossils, Zenger (1976, 1981) reported the linguid brachiopod *Lingulepis*? from unit B at several localities, including the quarry at Middleville, New York.

The stratigraphic and lateral relationships as determined by Zenger (1981) are shown in Figure 3. Zenger (1981) has demonstrated that the environment of deposition was a series of subtidal and peritidal conditions on an inner shelf area.

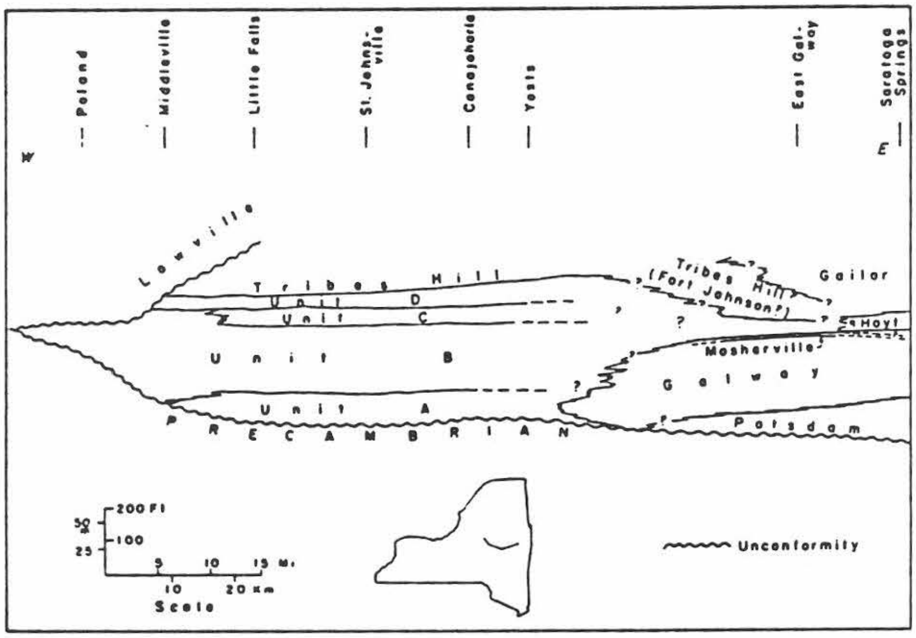

Figure 3. Generalized stratigraphic and lateral relations of the Little Falls Dolostone in east-central and eastern New York State. (From Zenger, 1981, figure 27, p.48).

Interest in the mineralogy of the Little Falls Dolostone, especially in the quartz crystals, goes back a long time. According to Ulrich (1989) the earliest documentation of the quartz crystals is a notice by Silliman (1819). According to Beck (1842) and Tuttle (1973), an article by Hadley (1823) was the first account to emphasize the abundance of the quartz crystals. The variety of external crystal morphologies was illustrated by Beck (1842, p. 261-264) and Ulrich (1989, p. 112-113). A complete list of minerals known from the Little Falls Dolostone is given in Table 1. The most commonly occurring minerals are dolomite, calcite, quartz, and the solidified hydrocarbon, anthraxolite.

The origin of the minerals is complex and as yet unresolved. According to Ulrich (1989, p. 116), the occurrences of the secondary dolomite and quartz crystals, and the anthraxolite, overlap considerably. His suggested paragenesis is shown in Figure 4. Just where the remaining minerals fit into this paragenesis is not known. An earlier paragenesis by Dunn and Fisher (1954) includes aspects of the regional geological history (see Table 2).

TABLE 1. MINERALS OF THE LITTLE FALLS DOLOSTONE

Quartz - Exceptionally clear, equant, doubly terminated. Occurs as rather large (up to 4") crystals in pockets, as smaller (up to 1") in vugs, and as linings of pockets and vugs.

Calcite - Usually yellow to brown in relatively well-formed crystals.

Dolomite- Usually cream to gray to light pink, as well-formed crystals with curved faces. Is the most common mineral.

Pyrite- Usually found as crusts up to 1/4 inch thick. Also reported as solid inclusions in the quartz crystals.

Marcasite-Usually found as solid inclusions in the calcite, mostly as wire-like bladed crystals.

Galena-Reported as very small masses. No crystals.

Sphalerite-Occurs as small (up to 1/4 inch) crystals. Also reported as solid inclusions in the quartz.

Limonite-Occurs as a weathering product of pyrite, marcasite, chalcopyrite, and hematite.

Chalcopyrite-Occurs as small, dark, rusty-looking isolated crystals and as thin, dark, rusty-looking crusts.

Hematite-Reported as solid inclusions in both the dolomite of the matrix and in the quartz crystals.

"Glauconite"-Occurs as blue to blue-green spots and stringers. Analysis by Zenger (1981) shows less iron than expected for this mineral.

Anthraxolite-Occurs as either small (less than one inch) black lustrous masses, or more commonly as a fine powder in some of the vugs. Analysis by Dunn and Fisher (1954) shows this to be a hydrocarbon. Two forms of this hydrocarbon were noted by Keith and Tuttle (1952), and may indicate two episodes of mineralization.

Figure 4. Paragenesis of some minerals of the Little Falls Dolostone, as suggested by Ulrich (1989, p.116).

TABLE 2. SEQUENCE OF EVENTS AS DETERMINED BY DUNN AND FISHER (1954).

1. Precipitation of limestone.
2. Silicification.
3. Dolomitization.
4. Formation of pyrite and sphalerite.
5. Entry of liquid anthraxolite.
6. Folding.
7. Formation of secondary dolomite.
8. Formation of secondary calcite.
9. Loss of gas, and solidification of anthraxolite.
10. Origin of the quartz crystals.
11. Block faulting.

Most research on the minerals has concentrated on the origin of the quartz crystals. Recently, Chamberlain (1988) has suggested that organic hydrocarbon complexes were of substantial importance for the long-term, low-temperature formation of the quartz. This suggestion follows the discovery by Bennett and Siegel (1987) that the solubility of quartz in water can be greatly increased when it is complexed by organic molecules According to Chamberlain (1988), the "Herkimer diamonds" formed while the organic complexes were broken down by bacterial or thermolytic mechanisms. Ulrich (1989) implies that the algal stromatolites may have been the original source of the hydrocarbons.

Unfortunately, little work has been done on the remaining minerals. It would be interesting to known if the temperatures and pressures of formation of the sphalerite are in agreement with the quartz data (approximately 50° C; based on fluid inclusion studies (Roedder, 1979), and inferred burial depths (as much as 7km of overlying sediments; Friedman, 1987)). It would also be interesting to ascertain time of formation of the galena and if it is in agreement with the inferred age of the quartz (presently accepted as Carboniferous).

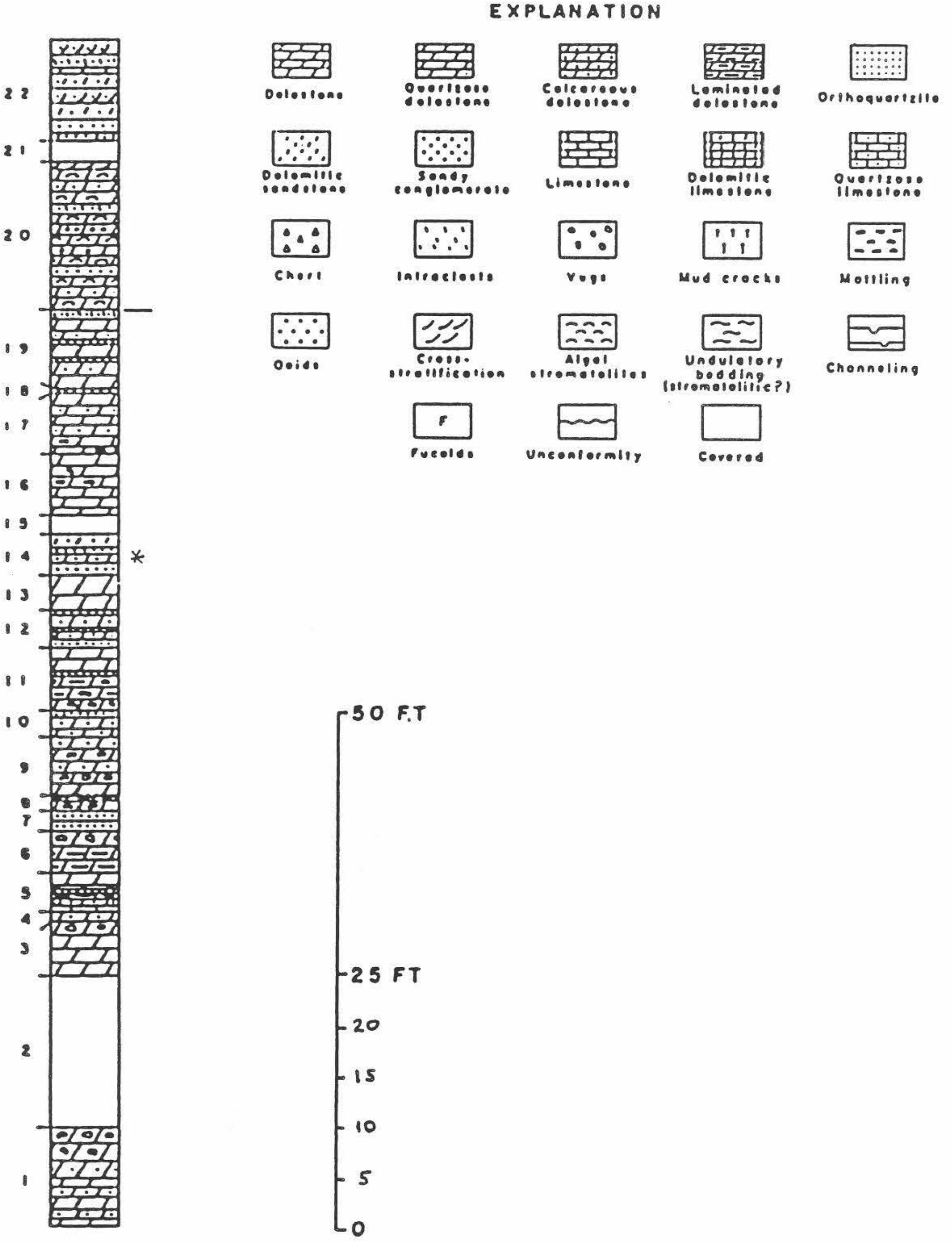

Figure 5. Columnar section for the Eastern Rock Products, Inc., quarry, Middleville, New York. (From Zenger, 1981, Plate 1).

According to Zenger (1976), the Little Falls Dolostone is about 200 feet thick at Middleville, New York. The formation non-conformably overlies the Precambrian gneiss and disconformably underlies the Ordovician Lowville Formation. The measured section (Figure 5) is from Zenger (1981) and is our major field stop. According to Zenger (1981), this section occurs entirely within his informal unit B (see Figure 3). Zenger's (1981) detailed description of this section is included here for the convenience of the reader. His terms for crystal size and crystallization fabrics follow the classification of Friedman (1965).

Unit 22. Sandstone and quartzite, light-gray to light brownish-gray, medium- to coarse-grained, medium-bedded and cross-laminated; intercalated light-gray, decimicron-sized dolostone laminations; some dolomitic sandstone; two feet above base is blackish sandstone in stylolitic contact with underlying dolostone; clasts as well as laminae of dolostone in upper bed; some sandstone very porous; some ripple marks; cross-laminations generally SW, S or SE, very rarely have northerly azimuth. 10 feet.

Unit 21. Dolostones, medium dark-gray to brownish-gray with some vugs, fine-grained (decimicron-sized); stylolitic contact with overlying sandstone sequence. 2 feet.

Unit 20. Sandstone and dolomitic sandstone, grayish-orange to brownish-gray and medium dark-gray, medium- to coarse-grained, commonly cross-laminated; some beds include dolostone clasts, some calcite; alternating stromatolitic dolostone biostromes which are generally abruptly terminated by the sandstone; bedding or parting is of medium thickness; stromatolitic dolostone is fine-grained, brownish-gray to medium dark-gray; about 7 cycles of stromatolitic biostromes capped by snady beds; two of the cycles have oolites above the stromatolites (oolites in moderate brown to moderate yellowish-brown dolostone; many oolites are hollow); matrix of sand and/or oolite in the biostromes with larger heads; relatively few vugs in biostromes - practically none in sandstones; some contacts between sandstone and stromatolitic layers are greenish glauconite(?). 15 feet.

Unit 19. Sandstone, light-gray to olive-gray or medium-gray, medium- to coarse-grained, medium- to thick-bedded, dolomitic, alternating with medium beds of vuggier, medium-gray to brownish-gray, in places laminated, decimicron-sized dolostone; cyclical nature, i.e., three couplets of dolostone overlain by sandstone; sandstones contain rounded dolostone clasts and seams of more dolomitic material; middle sandstone capped by coarser layer which is slightly green (glauconitic ?); upper sandstone has dolostone clasts. 8 feet.

Unit 18. Sandstone, pale yellowish-green, glauconitic, medium-grained; grades upward to sandy dolostone; darker "quartzite" seams within the greenish sandstone. 0.2 feet.

Unit 17. Dolostone, centimicron-sized, saccharoidal with quartzose dolostone intercalations, both being medium dark-gray; within and capping the unit are two medium beds of decimicron-sized, silty dolostone weathering a bit more olive; vugs present but not nume rous. 6 feet.

Unit 16. Dolostone, dark-gray to medium dark-gray with some brownish-gray splotches; thick parting to massive; quartz present but minor; some light greenish, circular to oblong mottles or clots; distinct dark weathering. 6 feet.

Unit 15. Inaccessible. 2 feet.

Unit 14. Sandstone, dolomitic, and quartzite; light-gray to medium light-gray, mostly medium-grained; conspicuous laminations; sharp contact with overlying dark bed; stylolite at contact with underlying dolostone; in lower layers are thin elongate quartzite bodies, forms prominent light band around entire quarry. 4 feet.

Unit 13. Dolostone, medium dark-gray to dark brownish-gray, saccharoidal, medium to thick parting, appearing massive due to weathering, weathers darker than conspicuous sandstone above; mineral filled vugs most abundant in lowest part; convex up laminae - may represent stromatolites (separated). 3.5 feet.

Unit 12. Dolostone, medium dark-gray with brownish-gray splotches, quartzose; intercalated laminae (up to 2" thick) of yellowish-gray, fine- to medium-grained sandstone; lowest half foot has small, slit-like vugs; capping thin sandstone continuous around quarry. 3.5 feet.

Unit 11. Dolostone, medium-gray and brownish-gray, decimicron- to centimicron-sized, slightly quartzose; laminated and mottled; some laminations of light-gray sandstone; thin to thick parting (1" to 1'+); vuggy, especially lower bed; top marked by less resistant zone. 6 feet.

Unit 10. Dolostone, dark medium-gray, centimicron-sized, medium to thick parting, very quartzose; some blebs; vugs practically absent; capped by this light-gray to very light-gray, medium-grained sandstone in white matrix (chert ?). 3 feet.

Unit 9. Dolostone, centimicron-sized, medium- to thick-bedded, slightly quartzose, brownish-gray to medium dark-gray, saccharoidal; very vuggy with secondary calcite and dolomite rhombs and quartz crystals; some pockets and stringers of dolomite rhombs in light, dense matrix, some of which is glauconitic (?). 6 feet.

Unit 8. Dolostone, decimicron-sized, medium-bedded, medium dark-gray; numerous vugs with predominantly quartz crystals and dolomite rhombs; unit exposed near base of main face. 1.7 feet.

Unit 7. Sandstone, fine- to medium-grained, light-gray to yellowish-gray, even surfaced; darker (cherty) pods within more quartzitic material. 2 feet.

Unit 6. Dolostone, medium-gray to dark-gray, decimicron- to centimicron-sized, thick-bedded, laminated, vuggy, particularly in upper bed; some brownish-gray, sandy pods in lower part. 4 feet.

Unit 5. Dolostone, medium dark-gray to brownish-gray, centimicron-sized, slightly quartzose; sandstone laminations; vugs contain common secondary minerals and anthraxolite; some finer-grained beds of laminated, decimicron-sized, silty dolostone; upper fine-grained beds contains *Lingulepis*? 3.5 ft.

Unit 4. Sandstone, medium-grained, dolomitic and calcareous (?); irregular, medium parting; no vugs. 1 foot.

Unit 3. Dolostone, dark-gray to dark brownish-gray, centimicron-sized, medium-bedded; upper 2 feet rather vuggy and contains abundant calcite. 5.5 feet.

Unit 2. Concealed. 15 to 20 feet.

Unit 1. Dolostone and quartzose dolostone, thin- to medium-bedded, medium light-gray to greenish-gray, fine centimicron-sized; upper part thicker bedded, light brownish-gray in splotches and layers alternating with dark-gray dolostone, low in quartz and more vuggy; dark color due to disseminated anthraxolite (?), exposed along West Canada Creek by Eastern Rock Products, Inc. office. 10 feet.

Total thickness, 120 +/- feet.

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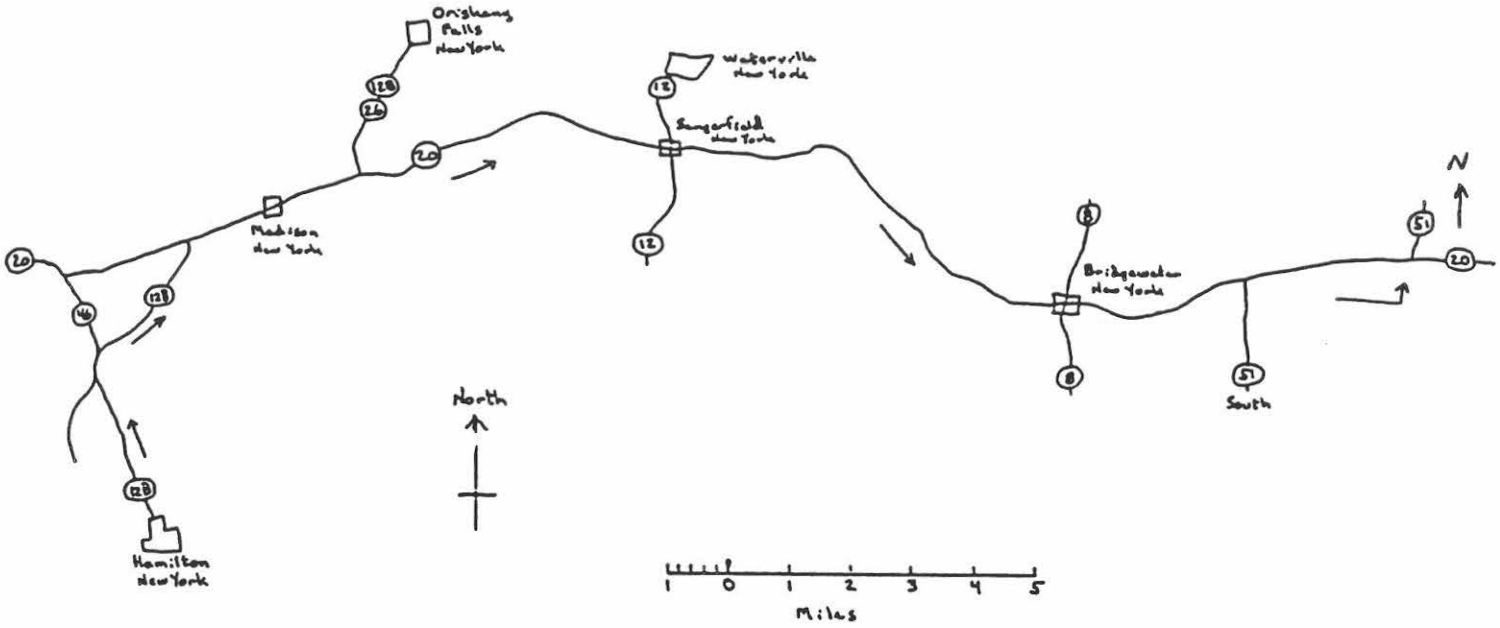

Figure 6. Generalized map for directions to start of road log.

ROAD LOG

Although this field trip departs from Colgate University, the road log starts at the junction of New York Routes 51 and 20. Proceed north on Route 12B to Route 20. Turn right, heading east on Route 20 for approximately 21 miles to the junction of Route 20 with Route 51 North. Turn left, heading north on Route 51. Refer to figure 6 for the general route directions to the start of the road log.

Cumulative mileage	Incremental mileage	Route description and remarks
0.0	0.0	Junction of Routes 51 and 20. Proceed north on Route 51 (this will be a left turn from Route 20 as we head east).
3.2	3.2	Junction of Route 51 with Jordanville Road in Cedarville, New York. Turn left (heading west) and continue on Route 51 for one-tenth of a mile.
3.3	0.1	Turn right (heading north) on Route 51 North. This section of road, known as Ilion Gorge, is somewhat narrow and quite winding. The road follows Steele Creek.
4.8	1.5	Junction of Route 51 with Holcomb Gulf Road (on the left). Continue north on Route 51.
5.05	0.25	Outcrop (on the right) of gray Syracuse Fm. (Upper Silurian).
5.6	0.55	Outcrop of the red Vernon Shale (Upper Silurian). See Treesh, 1972, for more details. These exposures of the Vernon Shale continue for the next 0.3 mile.
5.9	0.3	Junction of Route 51 with Jerusalem Hill Road on the left. This section of Jerusalem Hill Road is closed. See Treesh (1972) for more details on the units exposed along this closed road. Continue north on Route 51.
7.6	1.7	Short stop. Pull off CAREFULLY to the very wide shoulder on the left. Two items of interest. 1) is the small 15' thick outcrop of Otsquago Sandstone (Middle Silurian, Clinton Group; see Muskatt, 1972, for more details). 2) is the damage to trees caused by a tornado in the late 1980's (1989?). Continue north on Route 51, CAREFULLY.

8.05	0.35	Bridge over Steele Creek. On the left, in the stream, are thick layers of the Oneida Conclomerate (Middle Silurian, Clinton Group; see Muskatt, 1972, for more details).
8.3	0.25	Historical marker on the right, to mark the place where Eliphalet Remington 2nd made the first Remington gun.
8.4	0.1	Exposures of Frankfort Shale (Upper Middle Ordovician). These road cut and stream bank exposures on the left and right, continue for 1.4 miles.
10.1	1.7	Junction of Route 51 and Spinnerville Gulf Road (on the right). Continue north on Route 51.
10.3	0.2	Sign (on right) marking the village limits of Ilion. At this point, Route 51 becomes Otsego Street. Continue north on Route 51.
10.8	0.5	Route 51 (Otsego Street) bears to the right at the Y-intersection. Continue north on Route 51.
11.1	0.3	Traffic signal. Continue north (straight) on Route 51.
11.6	0.5	Junction of Route 51 with Route 5S (East). Route 5S in Ilion is Clark Street. Continue north on Route 51. (After crossing Clark Street, Route 51 becomes Central Avenue).
12.1	0.5	After going under the underpass, and crossing the Mohawk River, make a right turn onto the ramp for Route 5 heading for Herkimer, New York.
12.7	0.6	Exposure of glacial fluvial deposit in a sand and gravel pit.
14.1	1.4	Junction of Route 5 with Route 28. At this point, Route 5 becomes Main Street, and Route 28 South becomes Caroline Street. This is the first of eight traffic signals in downtown Herkimer. Continue east on Route 5 to the seventh traffic signal (junction of Route 5 and Route 28 North).
14.8	0.7	Junction of Route 5 and Route 28 North. Turn left (heading north on Route 28) and proceed about 8 miles to the village of Middleville.
15.5	0.7	Traffic signal at the junction of Route 28 and E. German Street. Continue north on Route 28.
15.8	0.3	P&C supermarket on the left. Behind the houses on the right is the levee system for flood control of the West Canada Creek.

17.5	1.7	Small exposure of the Dolgeville Fm (Middle Ordovician) on the left.
17.9	0.4	Sign on the right for Kast Bridge. Continue north on Route 28.
18.2	0.3	On the left for 0.4 mile is an abandoned stream meander of West Canada Creek. The stream channel was diverted by the DOT. Note also the braided nature of West Canada Creek as it tries to adjust to the changes.
18.6	0.4	On the right in the hillside, note the undercutting of the glacial material by a meander of West Canada Creek.
22.1	3.5	On the left is one of two commercial collecting localities for the quartz crystals known as "Herkimer Diamonds". Continue north on Route 28.
22.4	0.3	This is the second commercial site for "Herkimer Diamonds". Note also the exposure of the Little Falls Dolostone (Upper Cambrian) on the left. Continue north on Route 28.

Optional side trip

0.0 (22.9)	0.0 (0.5).	Sign on the right for the junction of Routes 28, 29, and 169. Turn left onto Fishing Rock Road, NOT Summit Road.
0.4 (23.3)	0.4 (0.4)	On the left is a large exposure of a glacial kame deposit. Also of note are the glacial striations on an exposure Precambrian syenite gneiss.
0.5 (23.4)	0.1 (0.1)	On the left is another exposure of the Precambrian syenite gneiss. These two exposures are part of an uplifted block of Precambrian bedrock. It is similar to the material exposed on Moss Island (see Muskatt, 1978, for more details). Note also the nonconformable contact of over 1 billion years. Turn around and return to Route 28.
1.0 (23.9)	0.5 (0.5)	Junction of Fishing Rock Road with Route 28 North. Turn left (heading north) onto Route 28.

End of road log for optional stop. Cumulative mileage in parentheses are those if you took the optional side trip.

23.0 (24.0)	0.1	Bridge over West Canada Creek.
23.05 (24.05)	0.05	Junction of Routes 28, 29, and 169. Turn left (heading west) onto Route 28 North.
24.2 (25.2)	1.15	Entrance on the left for the Eastern Rock Products, Inc. Quarry. Park here. We will be crossing the road (on foot) to enter the quarry. There are no further field stops, and you may leave at your leisure. PLEASE, stay away from the water-filled areas, and DO NOT CLIMB THE QUARRY WALLS.

To reach the New York State Thruway, return to Herkimer, New York, via Route 28 South. When you get to Route 5 in Herkimer, turn right, and follow the signs for the Thruway.

LATE PLEISTOCENE MELTWATER DRAINAGE THROUGH CENTRAL NEW YORK

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Introduction

When the retreating margin of Wisconsinan ice last stood in central New York, large quantities of meltwater poured eastward across uplands separating north-south valleys occupied by finger lakes. In some places, these ancient meltwater routes are inconspicuous; more commonly, they are impressive gorges produced by vertical incision or waterfall migration. On this trip we will examine channels, waterfalls, dry finger lakes, and remnants of deltas produced during this phase of deglaciation.

The importance of meltwater in the deglacial history of central New York has been recognized for many years. Fairchild (1909; 1932) mapped channel complexes from Batavia to Oneida, relating them to discharge from proglacial lakes north of the Valley Heads moraine (which today dams the southern ends of Seneca, Cayuga, and other Finger Lakes), and to "Great Lakes" flow from the Lake Erie basin and beyond. Sissons (1960) described meltwater scour features and sedimentary deposits between Syracuse and Oneida, and demonstrated that englacial, subglacial, and supraglacial flow were locally important. (These types of flow undoubtedly occurred in our area, but ice-marginal drainage was paramount.) Krall (1966) studied fluvioglacial drainage features in the region between Skaneateles and Syracuse. Other discussions bearing on the Syracuse channels have appeared in guidebooks of the New York State Geological Association (Muller, 1964; Grasso, 1970; Hand and Muller, 1972; Hand, 1978).

Despite the remarkable record of meltwater events displayed in central New York, complete historical reconstruction may never be possible. Lack of continuity is a problem; channels traversing upland regions are interrupted by north-south "through" valleys that held lakes (or ice) when the channels were active. Moreover, many erosional features are products of multiple episodes of ice advance and retreat, each accompanied by meltwater drainage that may or may not have been similar to drainage during other episodes. Constructional features (deltas) are temporally more coherent, in that they all must be products of the final phase of ice withdrawal. However, as drainage patterns evolved in response to changing lake levels or the opening of new escape routes, deltas have been destroyed or modified to the point where their significance is not always clear. Finally, the ever-changing configuration of the ice margin (which often provided one bank for the meltwater stream) is not well constrained.

The point of this trip is to observe a Whitman's Sampler of the types of evidence that remain. Hopefully, ambiguities generated by the inadequate record will be compensated by the fun of weighing alternatives and identifying fatal flaws in our colleagues' (and your leader's) interpretations.

Regional Setting

The features we will see are located along the northern margin of the Appalachian (Allegheny) Plateau, which rises southward from Syracuse and is bordered on the north by the Ontario Lowland. Ancient north-south valleys within the Plateau provided favorably oriented avenues for Pleistocene ice tongues in advance of the main glacier front. These north-south valleys became widened and deepened, and modified into the U-shaped troughs that today are termed "through" valleys. Otisco, Onondaga and Butternut valleys are examples, as are the valleys occupied by finger lakes, farther west.

During a pause in ice retreat (Port Bruce Stade), the Valley Heads moraine developed as a massive drift barrier plugging these many "through" valleys. ("Valley Heads" derives from the fact that the moraine now serves as a drainage divide between northward and southward drainage in trough-like valleys that otherwise are through-going.) Ice withdrawal from the Valley Heads position created local "finger" lakes that initially drained southward, over the moraine and into the Susquehanna basin. In time, these lakes developed communication with their siblings to the east or west through spillways or open water.

As the ice continued to recede from the Appalachian plateau's northward-sloping margin, eastward drainage became possible and the lakes' southward outlets were abandoned. During much of this time, Great Lakes water from the Erie and Huron basins (Lake Warren) augmented meltwater from across central and western New York to provide discharge that must have been comparable to that of the Niagara or St. Lawrence today. As these torrents spilled eastward from one "finger" lake to the next, they created the spectacular channels and deltas that we will examine today.

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ROAD LOG

USGS 7 1/2 minute topographic quadrangles relevant to this field trip:

OTISCO VALLEY
MARCELLUS
SOUTH ONONDAGA
JAMESVILLE
SYRACUSE WEST
SYRACUSE EAST

Cumulative Miles	**Interval Miles**	**Route Description**
		■ From Hamilton, north on NY 12B to US 20. West on US 20 to I-81 at LaFayette.
		Route crosses Cowaselon, Chittenango, Limestone and Butternut Valleys, and intervening upland spurs.
		Field trip mileage log begins at

LaFayette (US 20) interchange with I-81 (Interchange #15).

0.0	0.0	■ Left, entering I-81 southbound.
2.5	2.5	■ Right, into Rest Area. **STOP 1**.

View across Tully valley segment of Onondaga valley. This is the southward-directed "stem" of the Y-shaped trough ("through valley") system that will be the focus of this trip; Cedarvale trough can be seen extending northwestward from the three-way junction, but the northeastward continuation of Onondaga valley (toward Syracuse) is not visible from here.

As Wisconsinan ice retreated from the Valley Heads moraine (Tully moraine segment, four miles south of here), proglacial Lake Cardiff was impounded between the moraine and the receding ice front. Initial drainage was southward across the Tully moraine (1200') to the Susquehanna River. Later escape was westward (through Joshua and Navarino Channels) into adjacent lakes controlled by slightly lower segments of Valley Heads moraine.

The focus of this trip, however, will be on still later phases of drainage, corresponding to ice-front positions within about 4 miles of downtown Syracuse (and lake levels below about 900'), when meltwater from as far west as Lake Huron poured through central New York. It was this flow that scoured the Split Rock channels, Pumpkin Hollow, "Syracuse" channels (the several outlets crossing the upland between Onondaga and Butternut valleys), and other prominent canyons across uplands farther east and west.

■ Continue south on I-81.

5.2	2.7	*View of Tully moraine.*
6.6	1.4	■ Exit I-81 at Tully (Interchange #14).
6.7	0.1	■ Left at "T" (end of off-ramp).
6.8	0.1	■ Right (W) on NY 80.

The road follows the crest of the Tully moraine. This is predominantly a kame moraine, composed chiefly of outwash. Irregular *topography is due to kettles and (to a lesser extent) channeling.*

9.2 2.4 ■ Right on Woodmancy Road.

9.4 0.2 ■ Hidden Falls Road. **STOP 2.**

Excellent view of the north (proximal) face of Tully moraine. This steep face, which was banked against the ice, contrasts with the gently sloping (kettled) topography south of the moraine crest. This is the plug that held a lake in Onondaga valley.

Crest of the moraine stands at 1200'; 2 miles north of here, the valley floor is 600' lower. Several hundred feet of unconsolidated valley fill (till, outwash, lake clays, and postglacial alluvium) occur even in the deeper parts of the valley.

Rounded gravels in the roadside exposures are typical of the kame materials making up the bulk of the moraine. Clasts of local bedrock predominate: Lower Devonian limestones and dolostones are especially abundant; brown siltstones from the Hamilton Group are somewhat underrepresented, because they are less durable. Somewhat farther-traveled are red Medina sandstone and white Potsdam(?) sandstone. Gneisses come from the Canadian shield.

■ Turn around and return to NY 80.

9.6 0.2 ■ Right (NW) on NY 80.

17.8 8.2 ■ Left (W) on US 20.

Shortly after turning onto US 20, observe delta remnants in Cedarvale trough. The flat tops of these remnants represent various lake levels.

21.0 3.2 *Road descends into and crosses Navarino channel. The lake in Cedarvale/Onondaga valley discharged for a while through this channel, which held lake level at just over 1000'. To the southwest (downstream), Navarino channel terminates in a small delta on the east wall of Otisco valley.*

22.4 1.4 ■ Right on Slate Hill Road.

25.9	3.5	■ Right (N) on NY 17.
26.1	0.2	■ Right, into Marcellus Town Dump. **STOP 3.**

This is the entrance to Pumpkin Hollow, easily the largest and most impressive meltwater cross channel in central New York. Carved in Marcellus (black) Shale, Pumpkin Hollow is over 300 ft deep, 500 to 1500 ft wide at the floor, and 3 mi long. It carried Great Lakes drainage from Otisco valley to the head of Cedarvale trough. (Virtually all of this flow had previously entered Otisco valley through Guppy Gulph.)

We have here clear evidence that retreat of Pleistocene ice from the Valley Heads position was interrupted by at least one brief readvance. Three miles north of Pumpkin Hollow begins a group of cross channels and scourways collectively called the "Split Rock series." Many of these scourways are one-sided, implying that the other bank was the ice itself, and that the axis of flow shifted laterally (northward) as the ice retreated. Like Pumpkin Hollow, the Split Rock series carried substantial throughput from farther west, probably including outflow from an early phase of Lake Warren.

The highest of the Split Rock series stands at about 900', whereas Pumpkin Hollow is excavated to 700'. Clearly, Pumpkin Hollow could not have been open when the Split Rock series was active, because drainage would simply have escaped via the lower (Pumpkin Hollow) route. Moreover, incision of Pumpkin Hollow <u>after ice had withdrawn far enough to permit development of the Split Rock series</u> requires <u>readvance to ≥900'</u> to initiate cutting of Pumpkin Hollow.

If, as suggested by Calkin and Muller (1992, pers. comm.), the Valley Heads moraine represents the Port Bruce Stade (≈14,000 Ka), the Split Rock series is reasonably assigned to ice retreat during the Mackinaw Interstade, and readvance (activating Pumpkin Hollow) to the Port Huron Stade (≈13 Ka).

It is not known how far the ice front retreated during the Mackinaw Interstade. It could well have cleared the northern margin of *the Appalachian Plateau, permitting an interval of "free drainage" (Erie channel?), as envisioned by Fairchild. The limit of Port Huron readvance is probably defined by the end moraine (eastern equivalent of Auburn/Geneva/Canandaigua moraines) about 2 mi south of Pumpkin Hollow.*

Although the Split Rock series is usually envisioned as the product of ice recession, it is probably equally valid to view it in the context of ice readvance. The flow is then seen as being pushed progressively southward and to higher elevations, until (somewhat above 900') it was either stopped or redirected to Pumpkin Hollow.

Disappearing Lake, in the west end of Pumpkin Hollow, occupies a shallow basin in the top of the Onondaga Limestone. If time permits, we may consider the origin of this depression and some of the lake's interesting behavior.

■ Return to NY 17.

26.3 0.2 ■ Right (N) on NY 17.

26.4 0.1 ■ Right (E) on Pleasant Valley Road.

Driving through Pumpkin Hollow. After about 2 mi, the valley suddenly widens as the north slope recedes. Krall (1966) has shown that Cedarvale trough once continued northwest (i.e., into the north wall of the valley), but became filled with till and was not subsequently reexcavated. A well along Seneca Tpk (NY 175) a mile northwest of here (el. 1030') penetrated 200+ ft of drift without encountering bedrock.

31.5 5.1 ■ Right (S) on Cedarvale Road.

31.7 0.2 ■ Left (SE) on Cedarvale Road. (Straight is Amber Road.)

33.6 1.9 ■ Left (E) on Tanner Road.

Road climbs onto remnants of deltas deposited by flow that entered the lake in Cedarvale trough after being discharged from Pumpkin Hollow. Delta sands and gravels are underlain by pink lake clays that can be seen here in the roadside ditch, the gully along old dirt road leading to gravel pit on right, and in a bank on left side of road.

34.6 1.0 ■ Right (S) on Makyes Road ("Onondaga Hill Road" on older maps).

Crossing delta surfaces at 720', 660', and 600'. All of these deposits relate to flow from Pumpkin Hollow. (Some higher levels, above 780', appear to be scoured till surfaces and small kame deltas from local drainage.) Early (high) deltas were eroded and their materials (in part) redistributed into lower deltas as the lake level fell. Note well preserved distributary channels.

35.6 1.0 ■ Right (W) on Cedarvale Road.

35.7 0.1 ■ Right (N) at Onondaga Fire Station.

35.8 0.1 ■ Enter W.F. Saunders & Sons' gravel pit. (PERMISSION REQUIRED.)

36.0 0.2 ■ W.F. Saunders & Sons' gravel pit. **STOP 4**.

Sand and gravel are being extracted from the Gilbert-style delta whose topset beds define the 720' and 780' levels. Angle-of-repose foresets dip down-valley (ESE) in a single spectacular set whose thickness reaches 70' or more.

Lake level for the 780' delta was probably controlled by the 770' threshold of the approach to Clark Reservation channel. The 720' delta probably formed after the lake outlet had shifted to the (700') threshold of Nottingham Road channel.

Delta deposits in Cedarvale trough must have formed during the final retreat of ice. Had they formed earlier, they would have been vulnerable to destruction or drastic alteration during subsequent ice advance. Also, it was not until final deglaciation that flow through Pumpkin Hollow became established (recall discussion at Stop 3).

The clast composition of these gravels (pebbles/cobbles of limestone, dolostone, red Medina sandstone, gneiss) indicates derivation overwhelmingly from glacial drift. Pumpkin Hollow, which provided the flow that built the delta, is cut in Marcellus (black) Shale, but there are few, if any, clasts of black shale in this pit. This suggests that Pumpkin Hollow *first formed in pre-Port Bruce time, became filled with drift during a subsequent glacial advance, and remained filled until final deglaciation (retreat from the Port Huron position). Excavation of the modern gorge would have begun with removal of this drift and its incorporation into the Cedarvale deltas.*

The alternative possibility, that pebbles of Marcellus Shale would not survive transport, is rejected because such clasts are very abundant in the extensive 620' delta deposits even farther from source. These shale-rich deposits prove that some deepening of Pumpkin Hollow occurred through bedrock erosion during deposition of the youngest (and lowest) deltas. Even the low delta gravels consist chiefly of "bright" (recycled drift) clasts, most of which were probably reworked from older deltas.

■ Return to Cedarvale Road.

36.3 0.3 ■ Left (E) on Cedarvale Road.

36.4 0.1 ■ Continue straight (E) on NY 80.

Driving on 620' delta terrace, which displays many well preserved distributary channels. The delta prograded southeastward, down Cedarvale trough to the 3-way junction of the "Y"-shaped valley system, then northeastward toward Syracuse. Further progradation into the Tully valley "stem" was impossible because Tully valley provided no exit for the water. In this direction, the delta terminates in a primary depositional front.

The outlet for the Onondaga Trough Lake at the time of the 620' delta is uncertain. Grasso (1970) assigned it to an intermediate level in the incision of Rock Cut channel (initial threshold ≈700; later lowered to 550'). Though I accept Rock Cut as the probable exit route, there seems no reason for a persistent "hang-up" near 600'. Indeed, the floor of Rock Cut seems to have been lowered suddenly from 700' to 550' (Hand and Muller, 1972). In light of this, I would suggest that Rock Cut and High Bridge channels were clear and over-deepened when the 620' deltas formed, and that tailwater control was established farther east by the 580' floor of Pools Brook channel or by the prevailing level of

Lake Iroquois into which Pools Brook channel emptied.

38.4 2.0 ■ Right (E) on unnamed road, immediately after entering Onondaga Nation Territory.

39.3 0.9 ■ Left (N) at "Stop" sign in Indian Village.

41.7 2.4 ■ Left (N) on NY 11.

42.1 0.4 ■ Right (E) on Rockwell Road, which becomes Sentinel Heights Road.

43.1 1.0 ■ Left (N) on Graham Road (first left after crossing I-81).

View of northern part of Onondaga valley. Gravel operation on far side of valley is in the lowest (520') terrace. Escape of water at this time must have been directly controlled by the level of Lake Iroquois, possibly by way of Erie channel (<410').

43.8 0.7 ■ Left (N) on LaFayette Road.

44.0 0.2 *Crossing western (inlet) end of Smoky Hollow. This is the southernmost of the "Syracuse" channels cut into the upland between Onondaga valley and Butternut valley. Smoky Hollow is excavated in Marcellus Shale. Incision began at 900'. The channel floor was at 790' at time of abandonment. Hence, while Smoky Hollow was its outlet, the lake in Onondaga valley at this point varied in depth from about 480' to 370'.*

45.3 1.3 ■ Right (S, becoming E) on NY 173 (Seneca Tpk).

47.2 0.9 ■ Left into Clark Reservation State Park.

47.4 0.2 ■ Parking lot in Clark Reservation State Park. **LUNCH and STOP 5**.

The abandoned waterfall at the head of Clark Reservation channel illustrates an efficient and characteristic mode of cutting through the Onondaga Limestone: by headward migration of a waterfall. (Channels in the Marcellus Shale more likely were created by vertical downcutting, i.e., incision.) The bedrock threshold here (770') probably controlled lake level in Cedarvale trough while the delta at Stop 4

(Saunders' gravel pit) was forming.

Though not as high as Niagara (120' vs. 160'), Clark Reservation falls is virtually identical to Horseshoe Falls of the Niagara in terms of both planform and size.

At this Stop we will consider the activation sequence of the "Syracuse" channels. All have dimensions requiring "Great Lakes" (Lake Warren) drainage. All were active in late Wisconsinan time. Most (all?) have had complex histories, involving more than one period of excavation.

Restricted (channelized) flow across the Onondaga/Butternut spur began with Smoky Hollow. Scour began at the 900' level, and the channel was abandoned when its floor (at the west-end entry) had been lowered to 790'. Reworked drift in delta deposits in Butternut valley shows that final excavation involved removal of fill from a channel predating the last glacial override. Origin of the cut-off "meander" loop is not well understood, but relict drift deposits in this loop are further evidence of complex history. The scour and delta surfaces between 800' and 900' in Cedarvale trough probably relate to a gradually falling lake controlled by the Smoky Hollow outlet.

Ice retreat soon opened a more northerly escape route through Clark Reservation, establishing an Onondaga-valley lake level controlled by the bedrock threshold above the falls (770'). Cedarvale deltas tied to this threshold include the one seen at Stop 4. Drainage evidently followed the retreating ice margin northward, scouring to bedrock, until Nottingham Road channel became active. The sill leading to the abandoned waterfall at the head of Nottingham channel stands at 700'.

Intermediate points of stabilization were the two abandoned waterfalls in the south wall of Rock Cut (760', 730'). These falls show that the <u>eastern</u> half of Rock Cut was open at this time. However, as discussed by Hand and Muller (1972), subsequent activation of Nottingham channel would not have been possible unless the <u>western</u> half of Rock Cut remained plugged, and a (drift? ice?) barrier separated the approach to Nottingham from the two Rock Cut plunge basins. Catastrophic *failure of this narrow barrier was invoked by Hand and Muller to clear Rock Cut and High Bridge channels to their present levels, creating prominent deltas at the eastern ends of these channels and abruptly lowering the lake level in Onondaga valley by 100'.*

Delta terraces in Cedarvale trough seem compatible with this scenario. They suggest that after a period of gradual decline from about 790' to 700' (corresponding to gradual lowering of bedrock thresholds between Clark Reservation and Nottingham), there was an abrupt drop to 660'-600'. None of the "Syracuse" channels could have provided suitable control for this latter (very extensive) set of deltas. However, if Rock Cut and High Bridge channels were cleared as proposed by Hand and Muller, they would have provided an over-deepened (slow-flowing) exit route to a level-control point still farther east, namely Pools Brook channel (590') or young Lake Iroquois beyond.

Other questions surround Meadowbrook channel, whose floor elevation (western end) is 550', essentially matching that of Rock Cut. Original cutting of Meadowbrook undoubtedly occurred a long time ago, but its graded floor and mild truncation of drumlins indicate that it carried at least some flow during final deglaciation. How could flow have been forced through Meadowbrook if Rock Cut had already been opened to the same level? (Indeed, a somewhat lower level, since evidently some material has been removed from the floor of Meadowbrook channel.) Hand (1978) invoked an otherwise undocumented oscillation of the ice front to permit clearing Meadowbrook prior to opening the western half of Rock Cut. Subsequent blockage of Meadowbrook by ice was then called upon to reactivate Nottingham Road channel and set the stage for catastrophic flushing of the western part of Rock Cut. A simpler and perhaps better alternative envisions a "base" discharge adequate to have some erosive capability even when divided between two major channels. (Channels east of Limestone valley would have been abandoned by this time, and flow from Rock Cut and Meadowbrook would have escaped directly out the northern end of Butternut valley.)

■ Return to NY 173.

47.6	0.2	■ Right on NY 173.
49.5	1.9	■ Continue straight on Brighton Avenue. (NY 173 turns left.)
49.8	0.3	■ Right on Rock Cut Road, following signs to I-481.
50.5	0.7	■ Enter I-481 eastbound. **STOP 6**.

Observe Rock Cut channel: 100-150' deep, 1000' wide (floor), 2 mi long, threshold elevation 550'. Entrance ramp and the Onondaga Resource Recovery facility are directly below where Great Lakes discharge must have flowed in its approach to Nottingham channel. At that time, this (western) part of Rock Cut must have remained full of drift or ice. The postulated barrier that prevented Nottingham discharge from diverting to the (cleared) eastern end of Rock Cut was located just beyond (east of) the Resource Recovery facility. Next come the two abandoned waterfalls spilling over the south wall. These provided the flow that cleared out Rock Cut from here on east, prior to activation of Nottingham channel. A trailer park occupies one plunge basin; the other (upstream) became partially filled by a boulder bar when the barrier failed, and is not easily seen from the highway.

Gravel quarrying operations at the east end of Rock Cut have removed much of the delta attributed to the brief episode of catastrophic discharge attending barrier failure.

52.6	1.1	■ Exit I-81 (Jamesville Exit).
53.0	0.4	■ Right (S) on NY 91.
53.1	0.1	■ Left (E) on Woodchuck Hill Road.

You are climbing onto a remnant of the delta deposits dumped into Butternut valley during catastrophic flow from Rock Cut. Butternut Creek subsequently isolated this delta remnant and created the steep slope.

On the right side of the road, property of Dewitt Sportsmen's Club preserves original *configuration of the top of the deposit. Note the adverse slope of the broad distributary-like scour. To construct this "delta" (expansion bar?), flow in Rock Cut must have exceeded 65 ft!*

53.7 0.6 ■ Left on Will-O-Wind Drive.

53.75 0.05 ■ Left at "T" (Will-O-Wind Drive). **<u>STOP 7</u>**.

This is the top of the delta/expansion bar attributed to the Rock Cut flood. In the early 1970's, a large highway cut (now vegetated and uninstructive) exposed the full thickness (50' or more) of the NE-dipping foresets within this delta/bar.

The main point of driving through this development is to admire the 6-foot boulders used as lawn decorations. These particular boulders (chiefly Onondaga Limestone) were swept out of Rock Cut and stranded here on the top of the delta. Most, however, were swept to the lip of the delta, from where they rolled down the delta front to form a layer of 2-6' boulders several feet thick beneath the main foresets. The size of these boulders and the need to push them up the backside of this "delta" helped convince Hand and Muller of the reality of the Rock Cut dam burst.

53.8 0.05 ■ Right on Cedar Heights Drive.

54.3 0.5 ■ Left (E) on Woodchuck Hill Road.

■■■■■

Detailed Road Log ends here. To return to Hamilton,

Continue E on Woodchuck Hill Road to NY 92 (≈2 1/4 mi);

Right (SE) on NY 92, through Manlius, to US 20 (≈12 mi);

Left (E) on US 20 to NY 12B (≈15 mi).

Right (S) on NY 12B to Hamilton (≈5 mi).

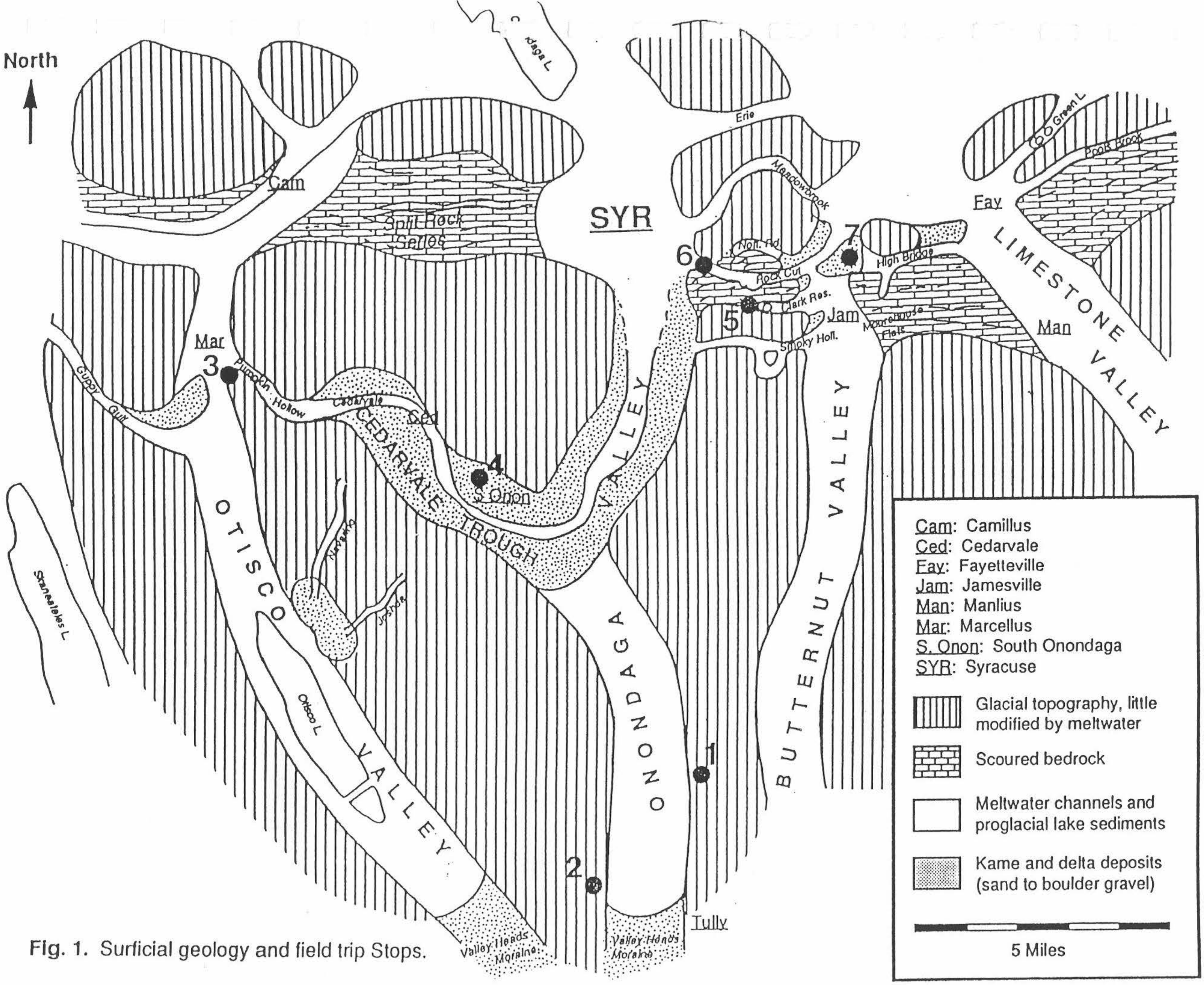

Fig. 1. Surficial geology and field trip Stops.

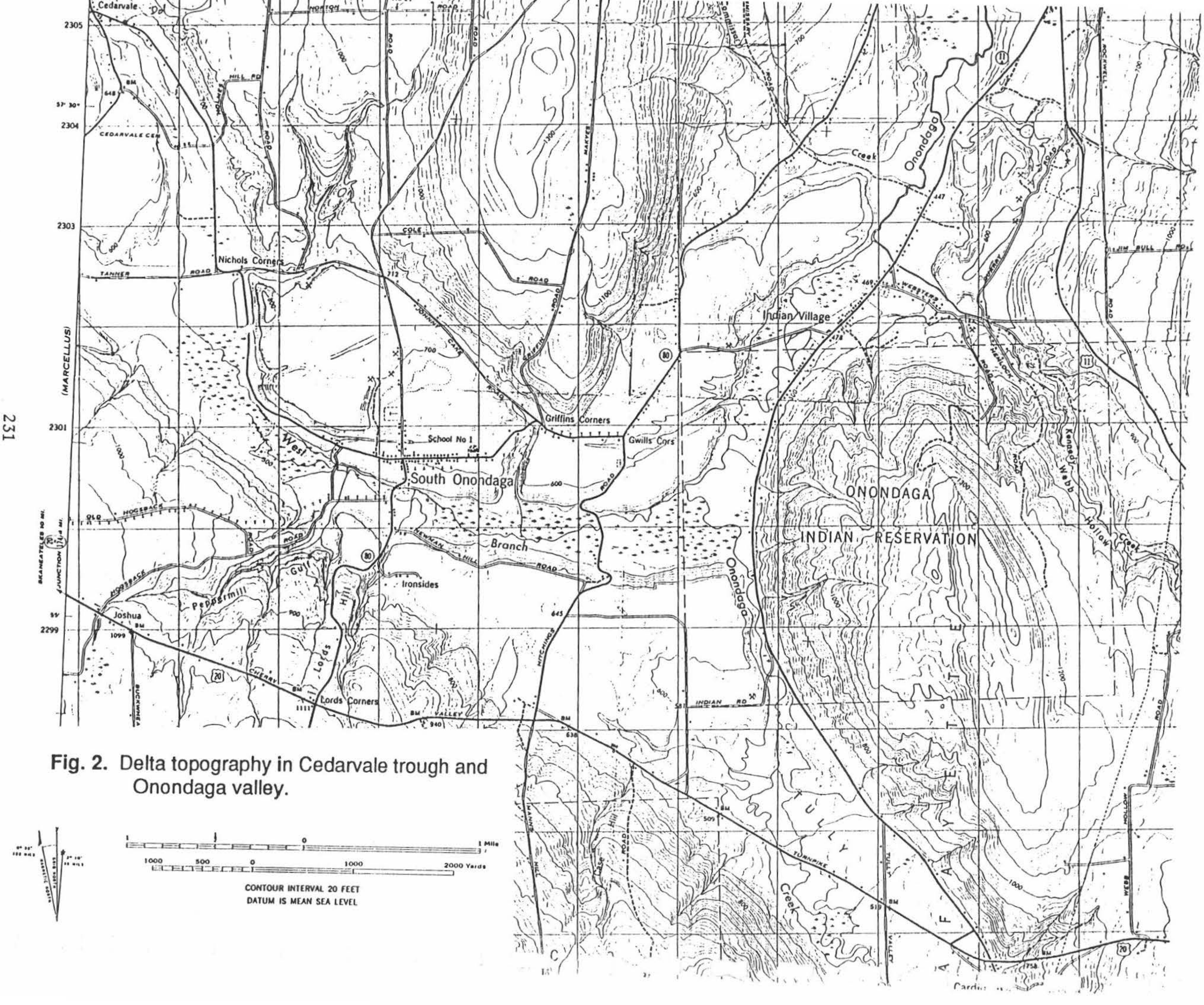

Fig. 2. Delta topography in Cedarvale trough and Onondaga valley.

Fig. 3. Topographic map of "Syracuse" channels.

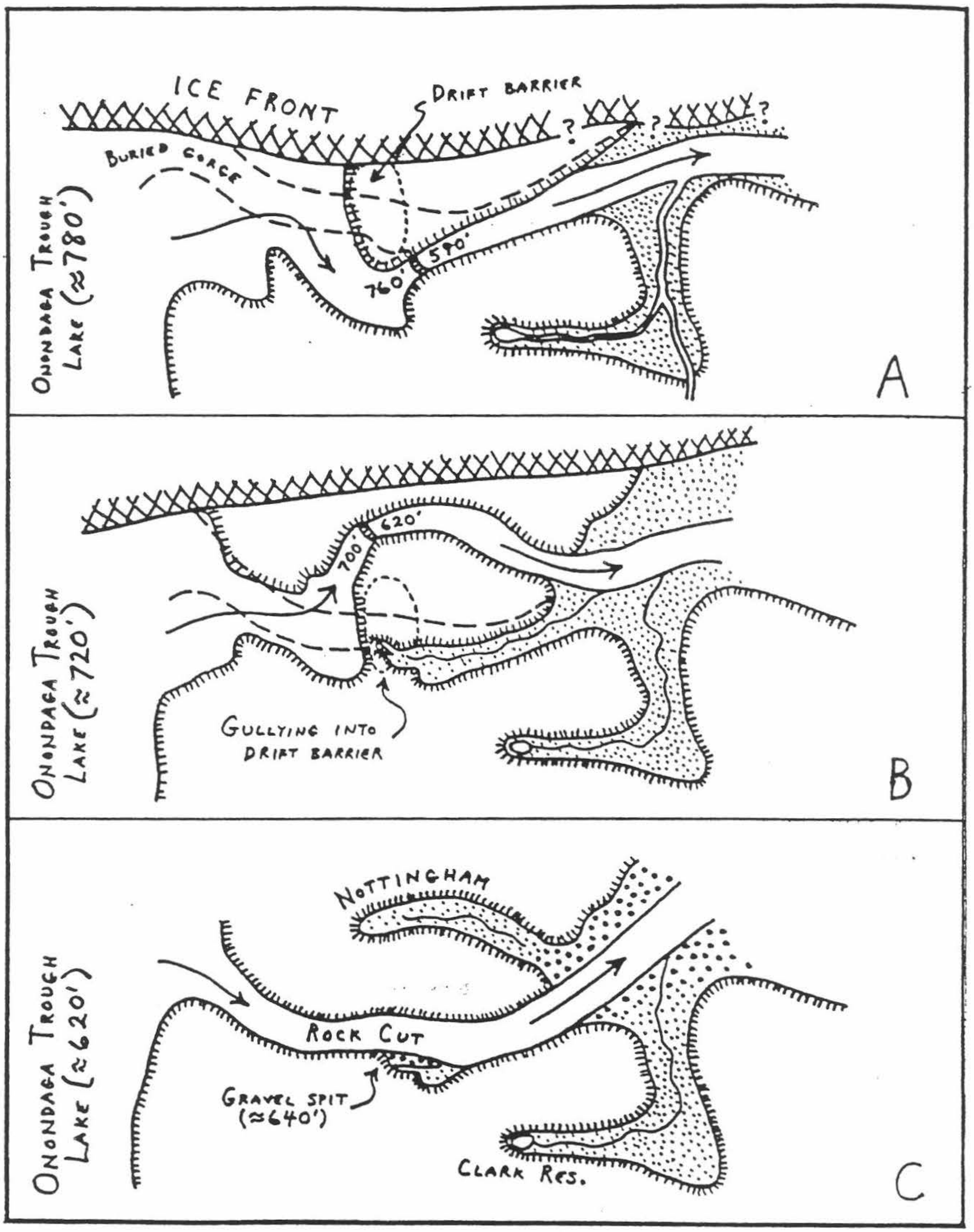

Fig. 4. Evolution of drainage routes leading to (re-)excavation of Rock Cut and Nottingham Road channels. (Slightly modified from Hand and Muller, 1972.)

PATTERNS OF PHYLETIC EVOLUTION IN THE TRENTON GROUP

By

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INTRODUCTION

The Trenton Group has been the subject of 60 years of continuous biostratigraphic study since Marshall Kay began his work during the early 1930's. The unit displays a remarkably rich and diverse fossil assemblage which, along with its facies patterns, is now quite well known. The Trenton Group is thus very well suited for research directed at understanding patterns of evolution. This ongoing research program is currently focused on recognizing patterns of intraspecific clinal variation and relating these to facies change in order to develop comprehensive records of evolutionary events. Two case studies are now complete. Evolution within the crinoid genus *Ectenocrinus* has been described (Titus, 1989). A similar paper on the brachiopod genus Sowerbyella is pending (Titus, 1992). This field guide will serve as an appendix to both of those papers.

PATTERNS OF EVOLUTION

The two broadest categories of evolutionary patterns are cladogenesis and anagenesis. Cladogenesis occurs when a population is divided and, from the two isolates, separate species are derived. This is also called allopatric speciation. Cladogenesis is multiplicative, as two or more taxa are descended from a single ancestral form. Anagenesis, or phyletic evolution, is not. It consists of evolutionary change within an undivided lineage. It is driven by natural selection and thus conforms to Darwin's original view of evolution. Virtually all recent workers agree that both patterns do occur and much of the current debate centers over their relative frequencies and macroevolutionary significance.

The Trenton Group is well suited for the study of anagenesis. The unit is well exposed across an outcrop belt extending from Canajoharie to Ontario (Figs. 1, 2). Deposition of the unit spanned a period of at least 8 million years without any significant breaks in the record. There is a rich and diverse fossil fauna which is very well preserved. These assemblages are found within a diverse facies mosaic (Fig. 3). The author has spent 20 years studying the Trenton and has documented its facies patterns and the biostratigraphy of over 200 of its species (Titus, 1986, 1988). This provides a wealth

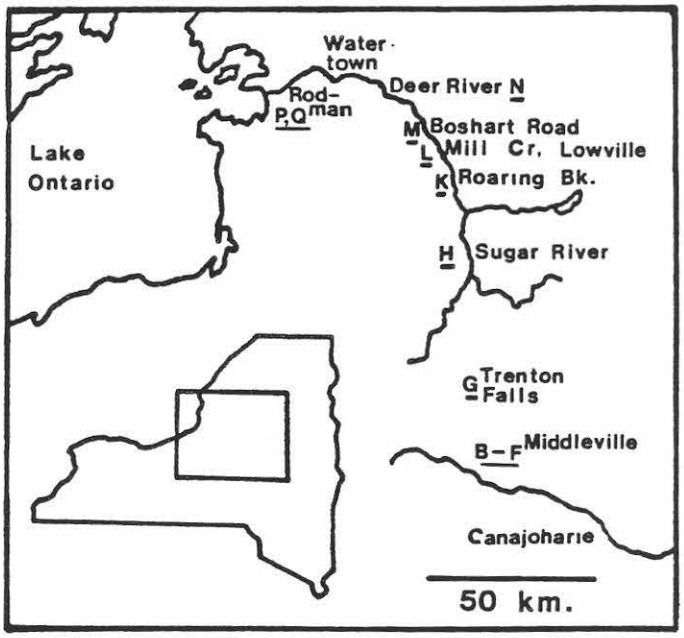


Figure 1. Map of the major locations of the Trenton Group.

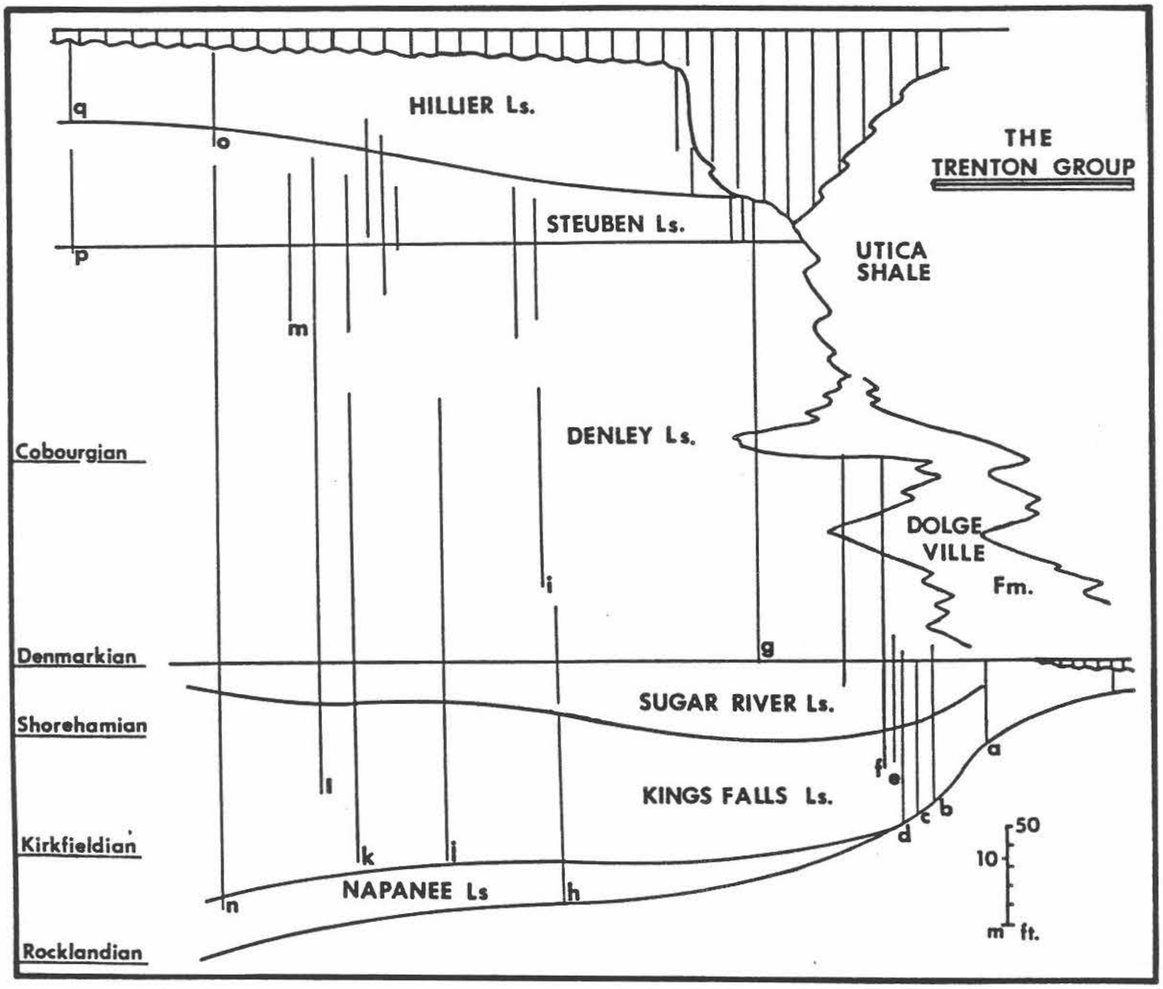


Figure 2. Stratigraphy of the Trenton Group. Letters refer to Fig. 1.

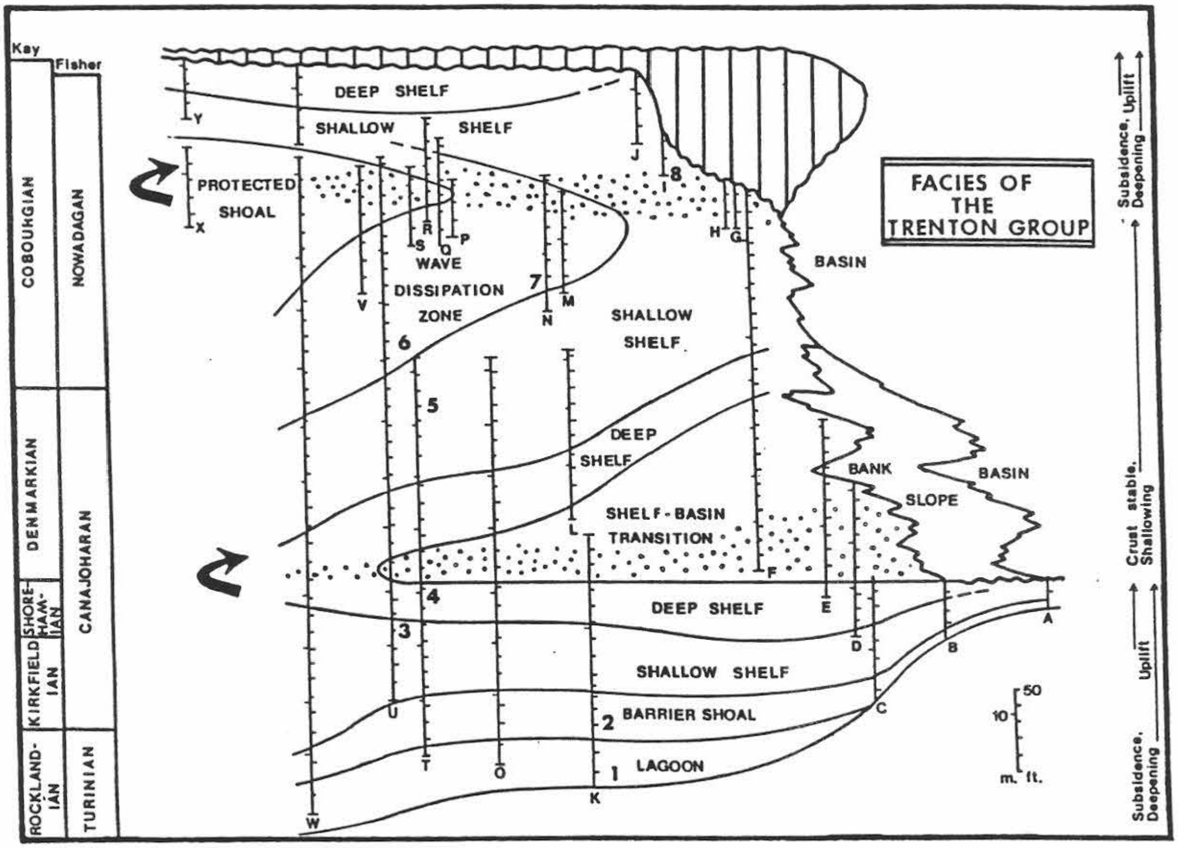


Figure 3. Facies of the Trenton Group. Arrows and stipples define the facies bottlenecks. Stippled areas display very few specimens of *Sowerbyella*. Numbers are trip stops. Stop six is approximate.

of background information. Thus any lineage can be knowledgeably traced through the entire Trenton Group and any phyletic evolutionary change can be observed.

The Trenton Group is not well suited for the documentation of cladogenesis. The carbonate platform is not likely to have any significant reproduction barriers and thus allopatric separation and cladogenesis is unlikely. It has not yet been observed. No examples of punctuated equilibrium have yet been demonstrated, although several possibilities are being investigated.

PHYLETIC EVOLUTION IN THE TRENTON GROUP

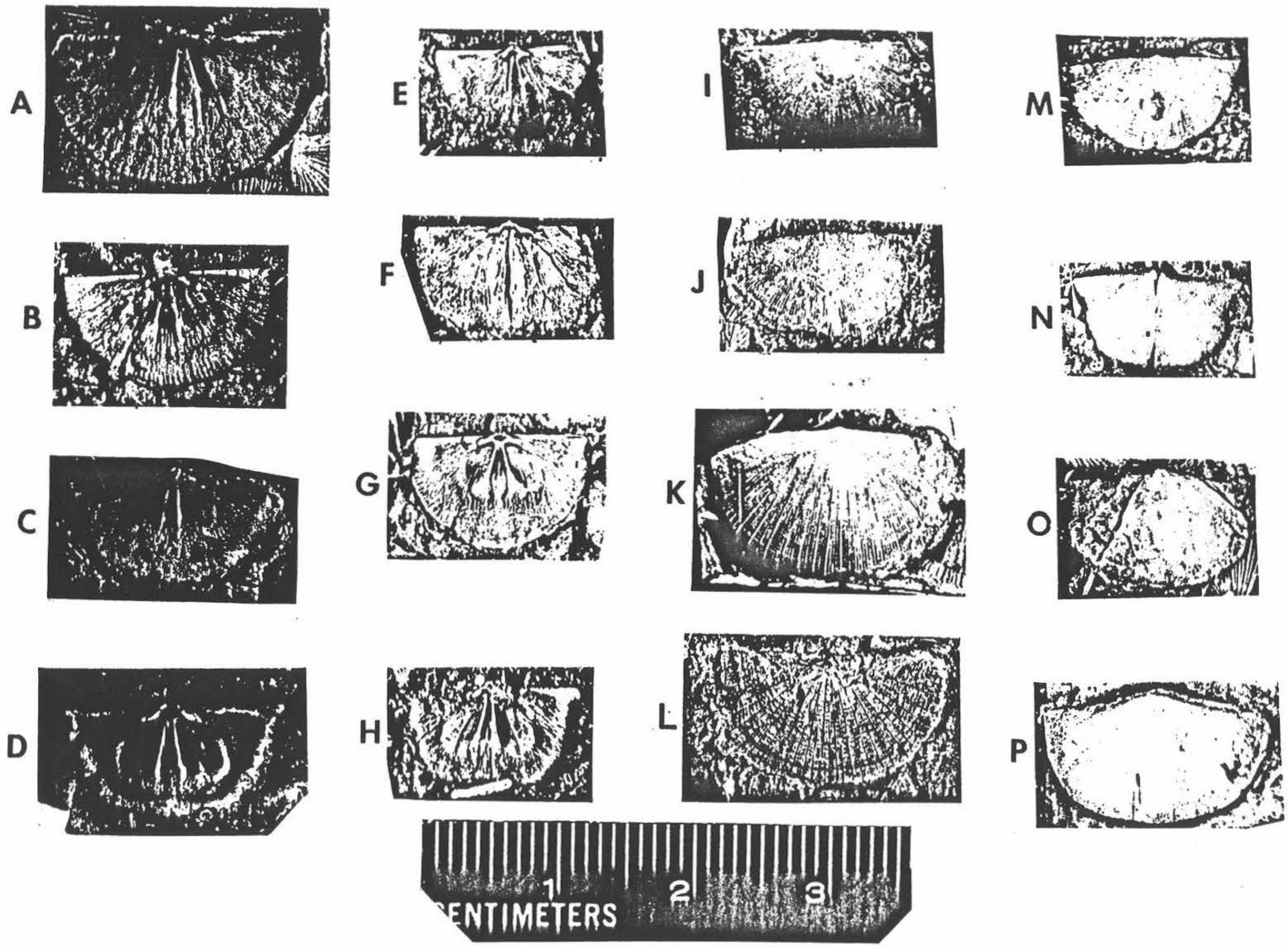
Phyletic evolution, as observed in the Trenton Group, can be described as continuously changing clinal variation in a continuously changing facies mosaic. Clinal variation is systematic geographic or ecologic variation within a species. Although such variation is commonplace among modern species, it has not often been reported in the fossil record (however, see Cisne et al. 1980a, 1980b & 1982 for other Trentonian examples). Recent studies in the Trenton Group suggest that clinal variation plays an important role in phyletic evolution: directional evolution through substantial episodes of time.

The following three models relate clinal variation to phyletic evolution: First, during times when facies are expanding and diversifying, species widen their ranges by evolving clinal variants adapted to the newly available environments. They become eurytopic by polymorphism. Second and conversely, at times when there is restriction of available facies, species experience cline sorting. Polytypic species pass through a facies bottleneck. Morphologies linked to disappearing environments are selected against and the species become stenotopic. In the third variation species can experience a form of clinal orthoselection when environments are deteriorating at one end of a species range while opportunities are expanding at the other end. Continuous, adaptive and directional selection occurs and the species evolves.

EXAMPLES OF PHYLETIC EVOLUTION

Model One

The lower Trentonian brachiopod *Sowerbyella curdsvillensis* illustrates the first model. This brachiopod evolved considerable clinal variation when a variety of environments became available during the lower Trentonian transgression. In quieter, mud-bottomed environments the species evolved a small morphology with a simple brachial valve interior (Fig. 4a). The pedicle exterior has a medial fold and alate corners (Fig. 4i).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CENTIMETERS

This form prevails in the nearshore lagoon and offshore deeper shelf facies (Fig. 3). In the adjacent wave swept, barrier bar facies *Sowerbyella* is much larger and its brachial interiors are quite ornate (Fig. 4d). They display septa, subperipheral thickenings, muscle scars and abundant and well developed vascular markings. The pedicle exteriors are often flattened or broadly rounded. These morphologies grade into each other in what is called a sloping cline. In the past this intergradation was not recognized and the plain, quiet water form was called *S. punctostriatus*.

During deposition of the middle Trenton Group a diversity of environments appeared in shallowing facies pattern (Fig. 3). Once again a model one pattern of clinal variation evolved. Plain brachial interiors occur in the quieter, muddier bottom environments while ornate brachial interiors are found in more agitated settings. This is a striking parallel to the lower Trentonian cline. There are differences, however. A new cline is seen in the morphology of the pedicle exteriors. In the middle Trentonian, deep and shallow shelf facies pedicles tend to have medial folds as was the case in similar lower Trentonian facies. In the barrier facies, however, pedicle exteriors tend to be more rounded and inflated. In the protected shoal facies there tends to be a broadly rounded medial fold. Thus, the overall structure of the middle Trentonian *Sowerbyella* cline is different and this cline should be recognized as a new species.

Model Two

Model Two occurs when there is cline sorting which accompanies times of facies restriction. Such an event would affect a species when facies change reduces its suitable habitats to a minimum. Perhaps a transgression would reduce suitable shallow water habitat, greatly restricting a species range. If polytypic forms pass through such a facies bottleneck, they are subject to intense natural selection. Inadaptive traits are selected against; these disappear and the species emerges from the facies bottleneck altered.

This is illustrated in the *Sowerbyella* populations of the middle Trentonian. Twice, there were times when the range of *Sowerbyella* was greatly restricted (stippled zones on Fig. 3). First, at the close of the lower Trentonian transgression, shallow water habitats had disappeared. The large forms with ornate brachial interiors were selected against and they are not seen above the facies bottleneck (Fig. 3). The middle Trentonian *Sowerbyella*, as noted above, closely parallels the lower Trentonian *S. curdsvillensis*. Both have plain, quiet

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Figure 4. a-d, *Sowerbyella curdsvillensis*, brachial interiors; e-h, *Sowerbyella*. n. sp., brachial interiors; i-l, *S. curdsvillensis*, pedicle exteriors; m-o, *S.* n. sp., pedicle exteriors; p, *S. subovalis*, pedicle exteriors.

water forms grading into ornate types found in agitated facies. However the middle Trentonian *Sowerbyella* is not as big as in the lower Trentonian form and it only very rarely displays well developed vascular markings (Figs. e-h). Large size and vascular markings are traits associated with shallow waters and selected against, within the deep water bottleneck, when the shallow facies disappeared.

A similar facies bottleneck is found at the top of the middle Trentonian Denley Limestone. During deposition of the Denley a well defined, east-to-west and deep-to-shallow *Sowerbyella* cline had evolved (see above). *Sowerbyella*, however, is virtually absent in the lower Steuben Limestone (stippled area, Fig. 3) except in the far west at Rodman. Lower Steuben depositional environments were evidently too agitated for *Sowerbyella* except in the quieter western protected shoal facies (Fig. 3). Only one clinal variant is found there. This form is recognized by the broadly rounded medial fold on the pedicle exterior (Fig. 4p) and plain brachial interior. This clinal variant made it through the western facies bottleneck (arrow on Fig. 3) and became a founding population for *Sowerbyella subovalis*, the most common species of *Sowerbyella* seen in the upper Trenton Group. The broadly rounded pedicle exterior is what most characterizes *S. subovalis*. The trait was, evidently, inherited from the bottleneck population.

Model Three

The columnals of the crinoid genus *Ectenocrinus* record an example of Model Three clinal variation. *Ectenocrinus* is first found in the deeper shelf facies of the Sugar River Limestone (Fig. 3). There it is commonly recognized by its nearly triangular columnals with triangular lumina (Fig. 5n-s). It passed, without effect, through the same facies bottleneck that generated the middle Trentonian species of Sowerbyella (arrow on Fig. 3). During the middle Trentonian, in a Model One case, it evolved a variety of clinal variants, which ranged in distribution from deep shelf to shallow water extremes (Fig. 5). The newly evolved shallow water clinal variants forms have round columnals with pentagonal lumina (Fig. 5a-c). The middle Trentonian records a long episode of shallowing facies. This included the carbonate bank margin which was shallowing and steepening (Titus, 1986, Fig. 7). Thus while new shallow water facies were opening up in the west, the old, deeper shelf facies were being restricted in the east. The result was a form of clinal orthoselection. As deep water environments shrank, the triangular forms disappeared. At the same time shallowing facies favored the round columnal forms and they eventually were the survivors. The transition is between the species *E. triangulus* and *E. simplex*. The transition is gradual and both facies and clines are central to this event.

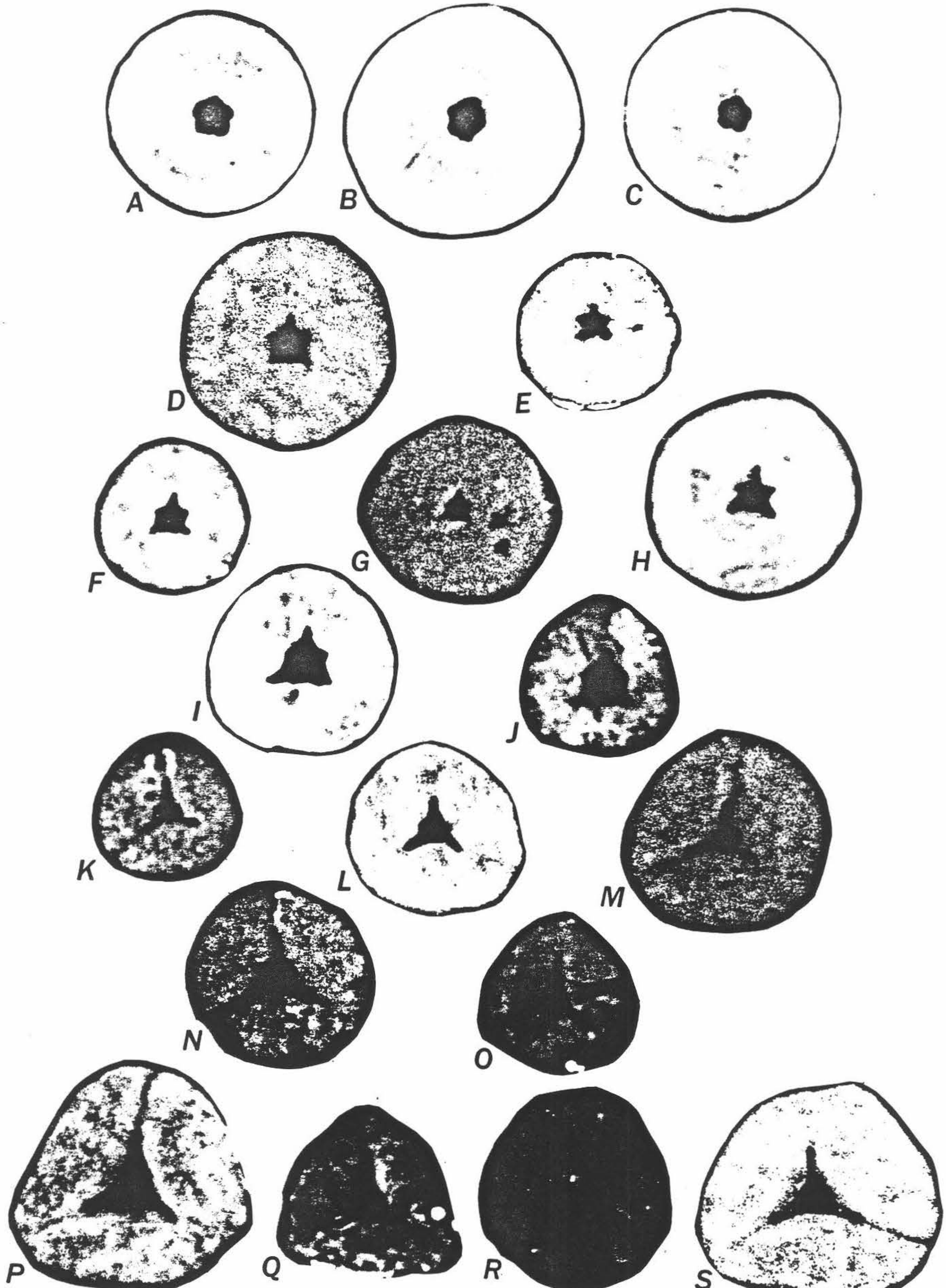

FIGURE FIVE

CRYPTIC VARIATION

An interesting problem which was raised but not solved in Titus (1992) was the disappearance and reappearance of the ornately sculptured brachial valve interiors that Sowerbyella possesses in the agitated facies in the lower and, then again, in the middle Trenton Group. These valves with their well-developed muscle scars are adapted to life in a rugged environment where individuals, from time to time, are overturned and must right themselves (there is no pedicle in *Sowerbyella*. Ornately sculptured valves are commonplace in the barrier shoal facies of the lower Kings Falls Limestone. They become rare in the deeper water facies of the upper Kings Falls and disappear in the deep shelf Sugar River Limestone. Later the trait reappears in the agitated facies of the Denley Limestone. Where did this trait go during the interval? Did it endure genetic extinction? Did it then re-evolve? If not, then where was it? The problem is a difficult one for a paleontologist who normally does not investigate the genetics of fossil species. Inferences, however, can be made.

A potential solution to the problem is to regard the ornate traits as going through cycles of being manifest and cryptic. Cryptic traits exist in the genotype but are not expressed in the phenotype. There are numerous genetic mechanisms which can preserve the genetics of a trait while, at the same time, preventing its expression. Traits can exist in a cryptic state as recessives. Stabilizing selection can keep a trait cryptic by eliminating the deleterious recessive homozygotes which express the trait. Heterosis will preserve cryptic diversity. Modifier genes can serve to mask the expression of traits while preserving them. There are ecological means of preserving cryptic variation as well.

Cryptic traits serve to maintain a maximum of species genetic diversity with a minimum of risk to the individual (any deleterious effect is masked). The benefit to the species is that the cryptic genetic diversity may sometime prove valuable as the environment shifts. any concept of species diversity thus has two facets. Expressed diversity is the sum total of traits which work successfully in whatever environment does exist. Cryptic variation represents a reserve of traits which may be potentially valuable in some future environment. Genes and species cannot foretell the future, but natural selection is most likely to preserve those species which ahve enough diversity to sustain themselves through the vicissitudes of change that always occur with time.

The on and off distribution of the ornate brachial interiors of *Sowerbyella* may well represent cycles of being manifest and cryptic traits. Indeed this may be key to accounting for the long-term success of this taxon. This issue of cryptic variation deserves much more attention. It is better illustrated in the strophomenid genus *Rafinesquina*. Future work will focus upon this.

CONCLUSIONS

In the half century since clinal variation was first described by Huxley, paleontologists contributed little to our understanding of this concept. There needs to be a greater awareness of clinal variation in the definition of fossil species and as a factor in phyletic evolution. To date, biologists have done most of the work but they have viewed clinal variation only in terms of geography and ecology. Paleontologists can explore the temporal dimension. This is being done in the Trenton Group where we can observe clines through time as well as through space. We see in the Trenton Group that, by the standards of geologic time, clinal variation is immediately adaptive to changing environments. It is through clinal variation that species can track facies change through time. Finally it is the linkage of clinal variation and facies change through time which appears to define phyletic evolution.

FIELD DISCUSSION TOPICS

Our trip gives us an opportunity to examine, in the field, evidence pertinent to some of the most difficult issues confronting paleontologists as we try to understand the fossil record of evolution. At the end of this trip we might well take time to discuss some of them. Two issues are outlined below.

This field trip illustrates some of the basic problems that paleontologists face in recognizing fossil species. We have always relied too much on the typological approach to species recognition. The two lower Trentonian species of *Sowerbyella*, which have been traditionally recognized, are *S. punctostriatus* and *S. curdsvillensis*. As we have seen, they turn out to be intergrading clinal variants of a single species. The typological approach has failed us, but when we employ a polytypic approach to species recognition other difficulties appear. The polytypic middle Trentonian form, *Sowerbyella*. n. sp., is an example. This form has a clinal variant which is identical to the deep water forms of the lower Trentonian species. It also has a clinal variant which is identical to the upper Trentonian form, *S. subovalis*. What is unique about the middle Trentonian form is the *combination* of clinal variants. Is this a proper criteria for species definition? I think so. How do you react?

A number of workers vigorously deny that phyletic evolution can produce new species. Species, instead, are regarded as sharply bounded spatiotemporal entities. No matter how much evolution may occur within a phyletic lineage, they argue that only one species should be recognized. Adherents of this point of view would thus argue that only one species of *Sowerbyella* can be recognized in the entire Trentonian sequence. I believe that I am observing phyletic evolution producing change at the species level and thus that phyletic change does result in macroevolution. I suspect that species generally are not spatiotemporally bounded entities and therefore that species are emphemeral, evolving continuously through time. You have seen some of the evidence basic to my argument. How do you react to this issue?

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ROAD LOG

The first three stops are designed to display a model one pattern of clinal variation. The brachiopod *Sowerbyella curdsvillensis*, by evolving several different clinal variants, was able to occupy most of the environments seen in the lower Trenton Group. It became a polytypic and eurytopic form.

TOTAL MILES	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0		Trip meets and organizes at the

		prominent outcrop of the Napanee Limestone just west of the bridge where Rt. 12 crosses the Sugar River, 3 miles west of Boonville. The group will walk up the west side of the stream and gather at the outcrop of the Napanee beneath the railroad bridge.

STOP 1 THE NAPANEE LIMESTONE

The strata here are primarily thick bedded, black micrites which have been interpreted as a nearshore lagoon facies. *Sowerbyella* is abundant. This form has traditionally been identified as *S. punctostriatus*. It is herein reinterpreted as the quiet water, mud bottom clinal variant of *S. curdsvillensis*. Examine the brachial interiors and see the preponderence of simple forms. Medial septa can be seen but muscle scars, vascular markings and subperipheral thickenings are uncommon. The pedicle exterior usually displays alate corners and a well developed medial fold giving this valve a peaked appearance.

0.0	0.0	Return downstream and cross the Rt. 12 bridge to the outcrop of the Kings Falls Limestone east of the road.

STOP 2 THE KINGS FALLS LIMESTONE

The lower strata here are primarily thick bedded, coarse grained biosparites which have been interpreted as a barrier shoal facies, lying offshore of the Napanee lagoon facies. Toward the top of the outcrop the facies grades into a shallow shelf facies. While, at this location, these strata overlie those of the lagoon facies, presumably there were contemporaneous lagoonal facies elsewhere. *Sowerbyella* is very abundant at the base of this outcrop. This form is larger than the one in the lagoon facies. Most of the brachial interiors are quite ornate. They have medial septa, muscle scars, vascular markings and subperipheral thickenings. The pedicle exteriors have blunt corners. They don't often show medial folds. Instead they tend to be flattened or broadly rounded. This form has traditionally been identified as S. *curdsvillensis*. It is herein reinterpreted as the agitated facies clinal variant of a more broadly defined *S. curdsvillensis*. By collecting enough brachial interiors at these two outcrops, an assemblage of forms can soon be put together that show the gradation between the lagoon and the shoal clinal variants.

21.1	21.1	Drive north on Rt. 12 to Lowville. Turn left in Lowville to remain on Rt. 12.
22.0	0.9	Drive north on Rt. 12. Cross Mill

Creek bridge and park immediately. Climb down to creek and walk to exposure east of and about 50m below the bridge.

STOP 3 THE SUGAR RIVER LIMESTONE

There is a prominent re-entrant below the bridge which can be traced downstream. This is a thick bentonite which lies 10 m (32 ft.) below the top of the Sugar River Limestone. *Sowerbyella* is rare in the deeper water facies above this bentonite, but common in the shallower facies below. Examine the strata several meters below the bentonite. The common form of *Sowerbyella* here is the mud bottom, quiet water form. Although some of the brachial interiors are ornate, most are plain. Medial septa are seen but other features are uncommon. The form is relatively small. Pedicle exteriors often display a medial fold and alate corners. This location records a return of both quiet water conditions and the quiet water clinal variant of *Sowerbyella*. Upstream into the highest strata of the Sugar River Limestone, and for a considerable distance into the overlying Denley Limestone, *Sowerbyella* is quite scarce. This zone is part of the first *Sowerbyella* bottleneck.

The next location is designed to display an example of model 3 clinal variation. The middle Trentonian crinoid genus *Ectenocrinus* occupied a shallow to deep shelf range of environments. Because of bank margin steepening, the deep end of its range was deteriorating while shallowing facies offered opportunities for *Ectenocrinus* in shallow water environments. The result is a kind of clinal orthoselection.

23.3	1.3	Turn around and return, on Rt. 12 through Lowville. Take the right fork onto Rt. 26.
26.7	3.4	Follow Rt. 26 through Martinsburg. Turn left onto Glendale Ave.
27.5	0.8	Turn left and enter Whitikers Falls Town Park. Park and follow one of the trails down to Roaring Brook and then climb down to the top of Whitikers Falls.

STOP 4 THE SUGAR RIVER AND DENLEY LIMESTONES

At the base of the falls a re-entrant can be seen. This is the same bentonite which was observed at Mill Creek in Lowville. Here it is also 10 m (32 ft.) below the top of the Sugar River Limestone. The bentonite is a marker bed which is found at the same level at all outcrops from Mill Creek, Lowville to Sugar

River. It has not been found at Deer River or at any of the Mohawk Valley locations.

Examine the several meters of strata above the falls. *Ectenocrinus* is represented by moderately abundant columnals. Look carefully as these columnals are small. It is very helpful to bring a water container and pour water on columnal rich beds. This brings out the contrast between the columnals and their micritic groundmass. *Ectenocrinus* is easily recognized by its trimeric morphology. Each columnal is composed of three elements fused together along faintly visible sutures. The most common form is a triangle with very rounded corners. At first glance the lumina appear to be triangular, and many are. But you will soon notice that many also have very poorly developed 4th and 5th points, along with three long and attenuated points (Fig. 5i-5m). All specimens at this level belong to the species *E. triangulus*.

Climb to the level above where the trail intersects the stream. Here and for quite a distance upstream *Ectenocrinus* columnals are different. Now they are generally rounder and the lumina are composed of more equal sized points (Fig. 5s-5v). Forms at this level are intermediate between *E. triangulus* and its descendent *E. simplex*. Continue upstream and observe the increasingly abundant columnals of *Ectenocrinus*. The transition from *E. triangulus* to *E. simplex* is not just one of morphology. *E. triangulus* was a relatively stenotopic form which never became especially abundant. The descendent, *E. simplex*, was altogether different. It was eurytopic and one of the dominant forms in the upper Trenton Group, a "weed" crinoid. This evolutionary event records an ecological transition from the K-strategist ancestor to the r-strategist descendent.

The last locations are designed to illustrate model 2 examples of clinal variation. *Sowerbyella* passed through two facies bottlenecks, one at the base of the middle Trentonian and one at the base of the upper Trentonian (Fig. 3, arrows). The best place to study the *Sowerbyella* of the middle Trentonian is at Mill Creek, Lowville, but as that outcrop extends for such a long distance along the stream, it is not practical to include this location in a one-day trip. We will examine middle Trentonian forms at other locations, stops 5 and 6.

		Return to the cars and drive back to the park entrance.
27.7	0.2	Turn left and travel to bridge where Glendale Road crosses Roaring Brook. Climb to the exposure immediately above the bridge.

STOP 5 THE DENLEY LIMESTONE

At this level the transition to *Ectenocrinus simplex* is complete. Virtually all *Ectenocrinus* columnals, upstream from the bridge, are round with pentagonal lumina (Fig. 5 z-5aa). *Sowerbyella* is not common here but the specimens which can be seen are of interest. They are nearly identical to the quiet water lower Trentonian forms. They have plain brachial interiors; the pedicle exteriors have medial ridges and alate corners.

28.9	1.2	Turn around and travel back towards Martinsburg. Turn right on Rt. 26.
30.8	1.9	Turn left onto B. Arthur Rd.
32.7	1.9	Turn left onto W. Martinsburg Rd.
32.8	0.1	Stop at large road outcrop on the right.

STOP 6 THE DENLEY LIMESTONE

The exposure here is of the middle Denley Limestone. The facies are the shallow shelf possibly grading into the barrier (Fig. 3). Several types of *Sowerbyella* can be recognized from the brachial interior structure. Many have plain interiors. While medial septa are present, none of the other internal structures are seen. The other type has an ornate brachial interior. Medial septa, muscle scars and subperipheral thickenings can be seen, some or all, on the same interiors. The plain forms found at this outcrop are typical of the Denley's deeper shelf facies while the ornate forms, although never very common, are typical of the more shallow and agitated facies. Significantly, vascular markings are quite unusual and these ornate forms are never as large as their lower Trentonian equivalents. Pedicle exteriors are more inflated and rounded. This is *Sowerbyella* n. sp.

33.9	1.1	Continue south on West Martinsburg Road. Turn left onto West Road.
51.4	17.7	Head east on West Road. At Whetstone Gulf State Park bear right onto Rt. 26. Continue east on Rt. 26 past its junction with Rt. 12D. Continue on 12D until the bridge over Sugar River at Talcottsville. Park, descend dirt path to outcrop.

STOP 7 THE UPPER DENLEY LIMESTONE

The best location to see *Sowerbyella* next is at the outcrop of the lower Steuben Limestone on Rt. 177 just west of Rodman, but this is too far away for our trip today. The Rodman exposure shows *Sowerbyella* within the second bottleneck (Fig. 3, arrow). Specimens there are small and have a well developed, broadly rounded medial fold. At Talcottsville we can see *Sowerbyella* in the barrier facies just below this second bottleneck. The form is abundant up to the base of the falls and then quite a bit less common in the bottlenecking beds above. The specimens here are, for the most part, characterized by well inflated pedicle exteriors which sometimes display the broadly rounded medial folds typical of the second bottleneck. Brachial interiors are generally plain. Notice the magnificent display of large symmetrical ripple marks (parapipples) here.

55.4	4.0	Continue south on Rt. 12D. Enter Boonville; enter Rt. 46 at its junction with Rt. 12D (no turn).
69.7	14.3	Follow Rt. 46 south to Frenchville. Turn left onto Rt. 274.
70.8	1.1	Park just west of the bridge over Big Brook. Descend to outcrop east of the brook.

STOP 8 THE HILLIER LIMESTONE

The Hillier Limestone is exposed all along Wells Creek and Big Brook. The environment of deposition was quiet deep shelf and the deposits are mostly biomicrites. This location is above the second facies bottleneck and a third species of *Sowerbyella* is exposed. The form, *S. subovalis*, is characterized by a greatly inflated pedicle exterior with a broadly rounded medial fold. Brachial interiors are mostly plain. Medial septa are seen, and sometimes poorly developed subperipheral folds can be found, but muscle scars and vascular markings are absent. These traits are apparently inherited from the clinal variants of S. n. sp. which made it through the second bottleneck. The ones on Big Brook are larger than their middle Trentonian ancestors.

END OF FIELD TRIP

Surficial Geology and Soils of Southern Madison County, New York

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Introduction and General Bedrock Geology:

Southern Madison County, New York lies south of the Onondaga-Helderberg Escarpment on the northern limits of the Appalachian Plateau. Local relief is approximately 600 feet, with valley floor elevations of 1100 feet in the vicinity of Hamilton, New York. The region is underlain by middle Devonian shales, siltstones and sandstones of the Hamilton Group. These fossiliferous rocks are rather poorly resistant to erosion, and surface outcrops are rare except in stream valleys and along steeply-sloping valley walls. One prominent sandstone unit in the Hamilton Group, the Chenango Sandstone, was quarried in the area as dimension stone. The older buildings on the Colgate University campus are built of Chenango Sandstone quarried on the hill south of the main campus.

Progressively older Paleozoic rock units are exposed to the north, including lower and middle Devonian limestones of the Onondaga and Helderberg Groups, the lower Devonian Oriskany Sandstone, shales, sandstones and dolostones of Silurian age, upper Ordovician shales, and lower and middle Ordovician limestones. Upper Cambrian sandstones and dolostone overly Proterozoic basement rocks in the Mohawk Valley. Lithologies representing all of these rock units can be found within the glacial drift which mantles the bedrock of southern Madison County. The dominant exotic (lithologies other than Hamilton Group) rock types in the drift are limestone, dolostone, chert, sandstone and various Proterozoic rocks of Adirondack origin.

Glaciation and Deglaciation:

Pleistocene continental glaciation profoundly affected the topography and surficial geology of the area. Pre-glacial topography, dominated by generally north-south dendritic valley systems, was altered by enlargement and reshaping of valley cross-sections, and sculpting of upland bedrock surfaces. Although no direct evidence of pre-Wisconsin glaciation is observable in the area, it is assumed that multiple advance-retreat cycles occurred. The presence of reworked clasts of cemented glacial gravels in Wisconsin-age drift may imply the presence of pre-Wisconsin glacial deposits. However, such clasts may have been generated within relatively short time periods during the last deglaciation.

Four phases within the deglaciation history can be identified:

1. Upland ice phase - No upland areas are know to have escaped glacial coverage during the Wisconsin maximum advance. Upland lodgement tills are often the only glacial deposits on upland ridges. In this region, the onset of deglaciation was characterized by thinning of the ice sheet to expose upland regions while active ice tongues occupied progressively lower elevation of the valleys.

2. Valley ice tongue phase - Active ice flow in the valleys is documented by the presence of glacial trim lines on valley walls, and deposition of ice-contact drift on valley margins. Kame terrace landforms characterize the valley walls in the region, and multiple terraces may indicate progressive lowering of the active ice surface within the valleys.

3. Fluvio-glacial outwash and stagnant ice phase - As the active ice margin retreated to a position near the present Onondaga Helderberg Escarpment, a complex period of minor advance and retreat of the ice sheet culminated with development of morainal deposits within the northern terminus of the major north-south valleys of the Appalachian Plateau. These deposits represent the so-called Valley Heads Moraine. In Madison County, these ice-margins fed major glacial streams which deposited extensive outwash blankets to the south. It is important to note that during this phase the major drainage from the ice sheet was to the south, and thus significant amounts of water and debris were transported through, and deposited in, the north-south valleys. In this area, the southern limit of the ice margin was approximately 15 kilometers north of the Village of Hamilton, at the approximate latitudes of Stockbridge Falls and Oriskany Falls.

Outwash was deposited around and above stranded stagnant ice that remained to the south of the active ice margins. Subsequent melting of these ice masses gave rise to kettle depressions and kettle lakes rimmed by steep, angle of repose slopes. Such depressions often interrupt relatively flat outwash plain surfaces.

4. Modern Drainage Phase - As the ice margin withdrew from the present Mohawk River Valley and Oneida Lake Plain, drainage through the Mohawk and Hudson Rivers was established. (These rivers would have carried much greater discharges than at present, because the St. Lawrence River Valley was still ice-covered.) An abrupt decrease in the discharge of both sediment and water in the valleys of southern Madison County was the consequence of this newly-established drainage pattern. With the onset of essential present-day discharge, major transport and deposition of fluvial-glacial sediment ended, and minor incision and terracing of outwash plain surfaces ensued.

The transition from Phase 2 to Phase 3 in the study area generally correlates with the "pre-Valley Heads" glaciation in the western Mohawk Valley (Ridge, Franzi and Muller, 1991). Valley Heads moraine deposition began approximately 15.5 ka.

Glacial Sedimentary Facies:

The surficial deposits of the study area can be broadly subdivided into sedimentary facies whose characteristics were controlled by the environments of deposition. These characteristics are briefly summarized below:

1. Lodgement Till: Surface exposures of lodgement tills are generally encountered un upland surfaces and the upper portions of valley walls. These materials were deposited beneath active ice as compact, poorly-sorted silt and clay-rich sediments. Angular, striated boulders are common, and such tills are dominated by locally-derived shale, siltstone and sandstone of the

Hamilton Group. Lodgement tills generally form thin veneers in areas of shallow bedrock on uplands, and are assumed to be present in the subsurface in lowlands. These tills have low hydraulic conductivity, and are therefore relatively poor aquifers.

2. Ablation Till: Ablation tills cover extensive areas of the uplands, and were deposited relatively passively during ice melt. These tills are often intercalated in complex fashion with fluvial-glacial deposits. Ablation tills are generally less compact and somewhat more well-sorted than lodgement till, although extreme ranges of particle size are characteristic. Locally derived lithologies dominate.

3. Ice-Contact Stratified Drift: Water-lain sands, gravels and silts deposited in contact with ice are common along lower valley walls and valley floors. These materials were deposited in subglacial, englacial, and proglacial streams within and adjacent to active ice margins, and in proximity to stagnant ice. Well-developed stratification and well-preserved primary sedimentary structures are typical, and post-depositional deformational features, such as soft-sediment folds and faults, are present. These sediments are generally moderately well-sorted, and form high-quality aquifers.

4. Fluvial-glacial outwash: Well-sorted sands and gravels deposited by braided streams are the dominated surficial material in the valley floors. Pebbles and cobbles in these deposits are well-rounded, and clast suites contain relatively high proportions of exotic lithologies. Outwash gravels are highly desirable aquifer materials, and are the source of good quality aggregate.

5. Proglacial lake and pond deposits: Proglacial lake delta deposits are often associated with ice-contact stratified and outwash facies. These deposits generally consist of well-sorted sand and gravel, with silts and clays comprising lake-bottom facies. Lake sediments are not abundant in the study area.

The thickness of the glacial sedimentary cover is highly variable in the study area. On some upland ridges and on steeper valley walls, drift may be absent or but a few meters in thickness. In the valley floor areas, thickness of the total drift cover is commonly in excess of 40 meters.

Soil Development and Surficial Geology:

Postglacial soil development in southern Madison has been controlled by the typical factors of climate, slope, drainage vegetation and parent materials. In addition, clearing and tilling of land for agricultural purposes, which began in the early 19th century, has increased erosion rates, changed vegetative cover and altered near-surface portions of soil profiles. The great majority of soils are relatively well-buffered in the subsurface because of the abundance of carbonate minerals in the parent materials. However, as will be explored on this fieldtrip, acidic surface horizons are often present in areas of coniferous forest canopy, and, locally, in organic soils of swamp and marsh origin.

Major soil orders: The dominant soils in the area are Alfisols and Inceptisols, with subordinate Entisols and Histisols. Alfisols are characterized by well-developed organic-rich surface horizons

(A-horizon) and relatively clay-rich B-horizons with significant enrichment of iron and aluminum. Inceptisols have less well-developed profile definition, and significant iron enrichment in the B-horizon is absent. Entisols, which have relatively little profile development, are found on steeper slopes, areas of erosion and on modern stream floodplains where sedimentation occurs regularly. Histisols, which represent the accumulation of plant debris in areas with permanent high water table, are found in marshy and swampy areas, and along the margins of some lakes and ponds.

Parent Materials and Soil Characteristics: The typical soils of the region can be very broadly divided in terms of parent material into those soils that developed on silt and clay rich parent materials, such as lodgement and ablation tills, and those that developed on more well-sorted sands and gravels of outwash and ice-contact stratified drift origin. The tills are typically rich in local shales and siltstones, which provide significant amounts of clay-size materials during chemical and physical weathering of larger clasts. The primary clay mineralogy of the local shales and siltstones consists of chlorite and well-crystallized illite/muscovite. These minerals represent the starting materials for the development of the clay minerals now seen in modern soil profiles.

The relatively high clay content of till parent materials leads to soils with significant subsurface accumulation of translocated clay, and hence, such soils are often poorly drained, and perched water tables are common, particularly during wet periods. The most typical representative of soils developed on till in the area is the Mardin soil series (USDA soil mapping unit). Mardin series soils are found extensively on the uplands area where lodgement till forms the underlying parent material.

Soils that developed on more well-sorted gravels and sands exhibit better drainage and significant development of clay-rich subsurface horizons which impede throughflow of water is uncommon. However, these soils have abundant clay in the B-horizon because sand and gravel-size clasts of shale and siltstone were initially present in the parent material. These clasts release clay during chemical and physical weathering, and some clay accumulates in the B-horizon. These soils do drain more quickly in the spring, and are, in general, better agricultural soils. Typical soil series developed on gravelly and sandy parent materials are the Howard, Palmyra and Chenango map units.

Soil Clay Mineralogy:

Chemical weathering processes that accompany soil formation result in the transformation of phyllosilicate minerals. In the soils of the study area, the most common clay mineral weathering sequences involve the alteration of detrital chlorite and illite/muscovite into vermiculite. Both chlorite and illite/muscovite are abundant in the C-horizon (relatively unaltered parent material) of local soils, but are progressively less abundant toward the soil surface. Chlorite transformation proceeds by oxidation and hydration of ferrous iron in the chlorite and release of magnesium into soil waters. The transformation of illite/muscovite results from removal of potassium from the mica structure, with subsequent expansion of the structure to accommodate hydrated interlayer cations.

The relative abundance of the clays vermiculite (in surface horizons), illite/muscovite and chlorite leads to the relatively high cation exchange capacity of most local soils. Base saturation of the soils is moderate to high, and pH is generally neutral to very slightly acidic, because of the buffering capacity provided by carbonate minerals in the parent material. These soils are therefore not significantly affected by acid precipitation.

REFERENCES

On this fieldtrip we will examine the relationships between glacial sedimentary facies, landforms and soil development in southern Madison County. Those participating on the trip will receive maps and areal photographs for guidance interpretation. If you are using this guide subsequent to the meeting, the following items are suggested. Some of these materials also served as bibliographic sources for this article.

1. Hamilton, N.Y. 7 1/2 min. Quadrangle

2. Hanna, W.E, (1981) "Soil Survey of Madison County" USDA/SCS/Cornell Agricultural Experiment Station; - This very useful compilation is available from the Co-operative Extension Office in Morrisville, N.Y.

3. Cadwell, D. and Muller, E. (1986) Surficial Geologic Map of New York, Finger Lakes Sheet: New York State Geological Survey, Map and Chart Series #40 - Available from the New York State Geological Survey, Albany, New York

4. Ridge, J.C., Franzi, D.A. and Muller, E.H. (1991) Late Wisconsin, Pre-Valley Heads Glaciation in the Western Mohawk Valley, Central New York and Regional Implications: Geological Society of America Bulletin, vol. 103, p. 1032-1045

ROAD LOG

Trip mileage log begins at the main traffic light in downtown Hamilton.

Cum. Miles

0.0 Traffic light at intersection of Broad, Lebanon, Payne and Madison Streets. Proceed north on Madison Street.

0.7 Bear right at intersection with Johnnycake Hill Road

1.0 Entrance to Madison Street Quarry on right.
Stop #1. (Conditions permitting) - Madison Street Quarry - This quarry is on private property.

The exposure is at the margin of a kame terrace that continues north along the valley the holds Lake Moraine. Depending on the conditions at the working face, large-scale cross-stratification can be seen, suggesting that this portion of the terrace formed as a lake delta, indicating that ponded water was present to the south. Exposures also contain highly chaotic, poorly sorted debris flow deposits.

Note the composition of the clast suite, rounding of larger clasts, and heterogeneous textures and fabrics.

Return to Madison Street, turn right from quarry entrance, proceed north.

1.4 Intersection with Airport road. Turn left (west) onto Airport Road.

1.7 **Stop #2 Valley Overlook on Airport Road**

We stop briefly here for a view to the south of the major geomorphic features of the Hamilton area. Note the obvious terraces that flank the lower valley walls, possible glacial trim lines on middle and upper valley walls, and the subdued topography of the valley floor. Note also the feeder canal in the near foreground.

Continue west on Airport Road

1.8 Intersection with Johnnycake Hill Road. Continue west on Airport Road.

2.0 Cross feeder canal from Lake Moraine

2.3 Intersection with Route 12B. Turn right (north) onto 12B.

2.8 Note prominent terrace on left.

3.7 Intersection with Woodman Pond Road. Turn left (west) onto Woodman Pond Road.

4.3 **Stop #3 Woodman Pond and Chenango Canal**

Woodman Pond is a kettle lake that was once used as the water supply of the Village of Hamilton. The village now gains water from wells within the village that are located in highly permeable outwash gravels. This supply, available as a backup, was prone to high algal/organic loads because of overwintering flocks of 1000+ Canada Geese. Note the steep, angle of repose slopes that rim portions of the pond and adjacent marshlands.

Immediately to the west is the Chenango Canal. The canal system in this area was originally developed in the late 1830's for transport of goods from Binghamton to Utica. In the Hamilton area, a system of feeder canals captured south-flowing drainage of the Chenango River, Payne Creek and Kingsley Brook and diverted that water to the north. The Chenango Canal was essentially abandoned as a transportation system in the middle part of the 19th century as railroads were constructed. Portions of the canal system are still used, however, to supply water to the New York State Barge Canal system. The old Chenango Canal at this site carries water from Payne Creek (Lake Moraine) and the Chenango River to Oriskany Creek, which supplies the Barge Canal.

Continue west on Woodman Road.

4.6 Intersection with Eaton Road. Turn left (south) onto Eaton Road.

5.0-5.3 Howard Series soils on left.

5.4 Sharp right curve crossing feeder canal from Chenango River.

6.2 Hamilton Village limits.

6.9 Intersection with Montgomery Street. Turn right (south) onto Montgomery.

7.2 Intersection with Lebanon Street. Turn right (southwest) onto Lebanon, which becomes Randallsville Road.

8.1 Crossing feeder canal from Chenango River.

9.0 **Stop #4 Randallsville Road Terraces**

Two obvious terrace levels are present in this area. Both are underlain by relatively well-sorted gravelly sand, and the material is clearly of fluvial-glacial origin. It has been proposed that the terraces represent either glacial lake deltas or kame terraces, although these may also be erosional terraces formed during late Phase 3 deglaciation.

The soils on the terraces are excellent representatives of the Palmyra series, and are very good agricultural soils. The feeder canal at this site carries water from the Chenango River to the north, skirting the Village of Hamilton, emptying into the Chenango Canal immediately north of the village.

Turn right (west) onto Armstrong Road

9.7 Crossing feeder canal and Chenango River.

10.0 Intersection with River Road. Turn right (north) onto River Road.

10.2 Intersection with Chamberlain Road. Turn left (west) onto Chamberlain. Stone house at intersection is made of Chenango Sandstone.

11.9 Intersection with Bartlett Road. Turn left (south) onto Bartlett.

12.9 Intersection with Geer Road. Turn right (west) onto Geer.

13.3 Entrance to Bewkes Center. **Stop #5 Bewkes Center**

The Bewkes Center property was a gift to Colgate from a former chair of the Board of Trustees, E. Garrett Bewkes. We will proceed to the wooded area in the vicinity of Seymour Pond to examine soil profiles in Stockbridge and Volusia series soils, and compare soil profile development in areas of deciduous and coniferous forest canopies.

Seymour Pond may be of kettle origin. However, the surrounding surficial material is ablation till. A more likely origin is the interruption of surface drainage by the configuration of ablation till deposits at the northeastern margin of the lake. The recent sediments in the lake are almost entirely of biogenic origin (diatom and algal/organic). The lake waters are well-buffered, with a pH that varies from 6.8-6.9.

Proceed east on Geer Road.

13.6 Intersection with Bartlett Road. Turn right (south) onto Bartlett.
14.3 Intersection with Reservoir Road. Proceed south on Reservoir.
15.6 Intersection with Lebanon Road. Turn left (east) onto Lebanon.
17.9 Intersection with Rodman Road. Continue east on Lebanon Road, bearing right.
20.4 Intersection with River Road. Continue east on Lebanon.
20.6 Crossing Chenango River.
21.0 Intersection with Route 12B. Turn left (north) onto Route 12B.
21.1 Intersection with Craine Lake Road. Turn left (west) onto Craine Lake.
21.7 Bear right at Y-intersection to proceed around lake.
22.1 **Stop #6 Craine Lake**

Craine Lake is an obvious kettle lake surrounded by steep, angle of repose slopes in outwash. The soils developed on the outwash surface are Palmyra series. Craine Lake waters are well-buffered, and precipitation of fine-grained calcium carbonate occurs on aquatic plants during the warm summer months.

Continue around Craine Lake and exit toward Route 12B.

23.5 Intersection with Route 12B. Turn left (north) onto Route 12B.
25.2 Cossitt Concrete Sand and Gravel Plant on left. The quarrying is developed in outwash.
26.0 Hamlet of Middleport.
26.6 Intersection with Horton Road. Turn right (east) onto Horton. This road is steep and rough.
27.5 **Stop #7 Horton Road (time permitting)**

The fields and pastures of the uplands along Horton Road are underlain by a thin veneer of lodgement till with a high percentage of local Hamilton Group lithologies characterizing the clast suite.. Typical Mardin Series soils are developed on the till. These soils are moderately good agricultural soils, but are slow to drain in spring. As we continue east on Horton Road, and onto Preston Hill Road, you will note a number of marshy areas which attest to the rather poor drainage typical of Mardin series soils.

Continue east on Horton Road to intersection with Preston Hill Road. Turn left (west) onto Preston Hill and continue west to Route 12B to return to Village of Hamilton.

End of Trip.