New York State Geological Association

77th Annual Meeting
Field Trip Guidebook

Lake Ontario Sunset, Photo by Joe LeFevre

Edited by David W. Valentino
Hosted by the State University of New York at Oswego
September 23-25, 2005

Field Trip Guidebook for the 77th Annual Meeting of the New York State Geological Association

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This guidebook was published by the New York State Geological Association.
ISSN: 1061-8724

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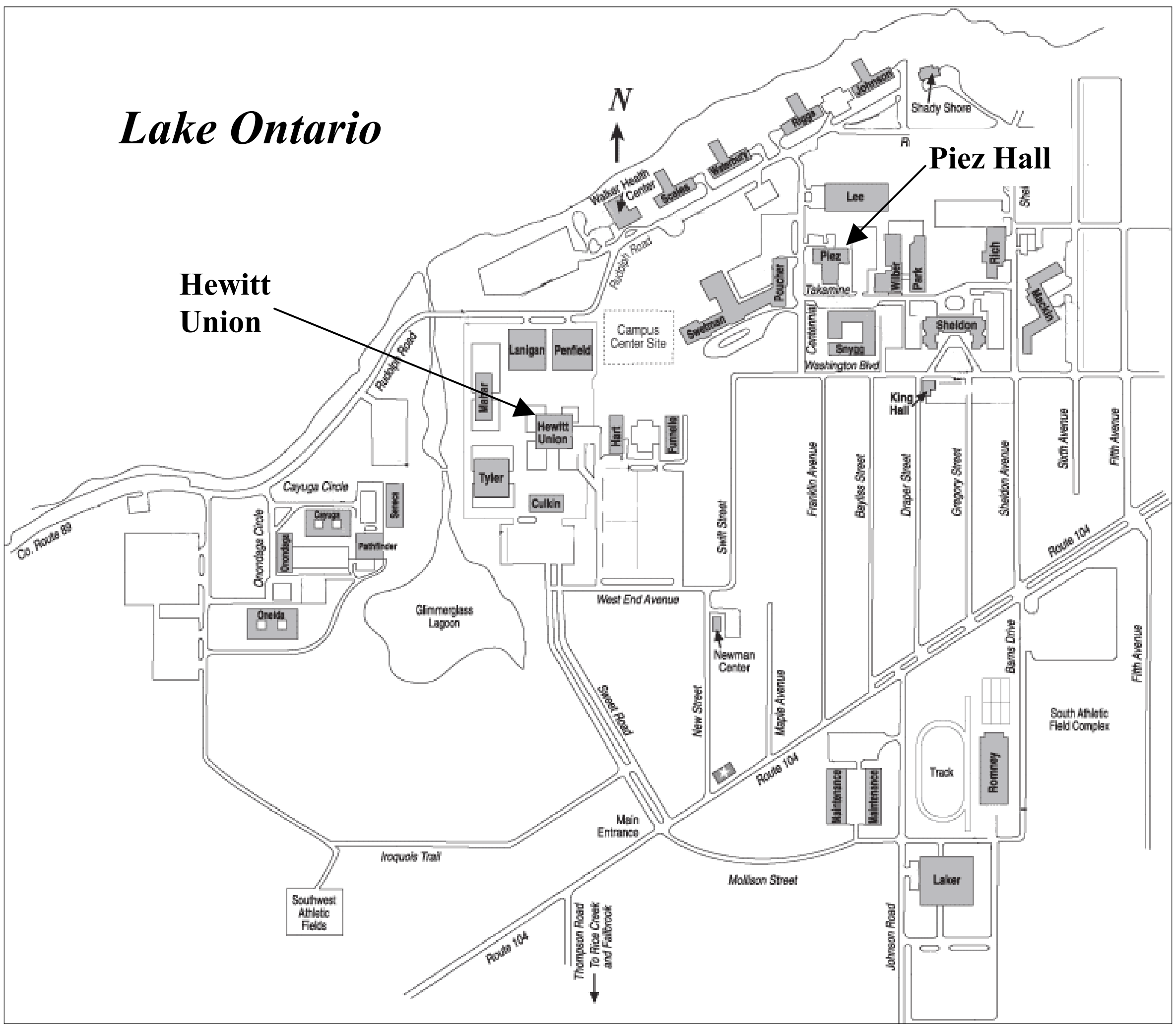
Campus Map


Lake Ontario
N
Piez Hall
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Scales
Walker Health Center
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Campus Center Site
Lanigan
Penfield
Mahar
Hewitt Union
Hart
Funnelle
Tyler
Culkin
Cayuga Circle
Cayuga
Seneca
Onondaga Circle
Onondaga
Pathfinder
Oneida
Co. Route 89
Glimmerglass Lagoon
West End Avenue
Swift Street
Franklin Avenue
Bayles Street
Draper Street
Gregory Street
Sheldon Avenue
Sixth Avenue
Fifth Avenue
Route 104
Newman Center
New Street
Maple Avenue
Sweet Road
Barns Drive
South Athletic Field Complex
Track
Romney
Maintenance
Main Entrance
Iroquois Trail
Mollison Street
Laker
Southwest Athletic Fields
Thompson Road To Rice Creek and Fallbrook
Johnson Road

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Cover photo: *In addition to intense lake-effect snow, Oswego, New York is well known for the spectacular sunsets over Lake Ontario. This photo is one such sunset taken from the campus of SUNY-Oswego by Dr. Joe LeFevre of the Department of Chemistry. The foreground includes outcrops of the Ordovician Oswego Formation, which consists of crossbedded sandstone, and is the subject of Field Trip A-3, Stop 1.*

Trip A-1

SLOPERVILLE BOG AND ESKERS

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INTRODUCTION

On this trip we will explore a large wetland centrally located in a beach, outwash, and esker-kame moraine complex that forms the Lacona-Williamstown aquifer, the largest aquifer in Oswego County. This fieldtrip will use terms commonly used to designate glacially formed topographic features, and address specific names of plants and detailed descriptions of the characteristics of relevant wetland communities.

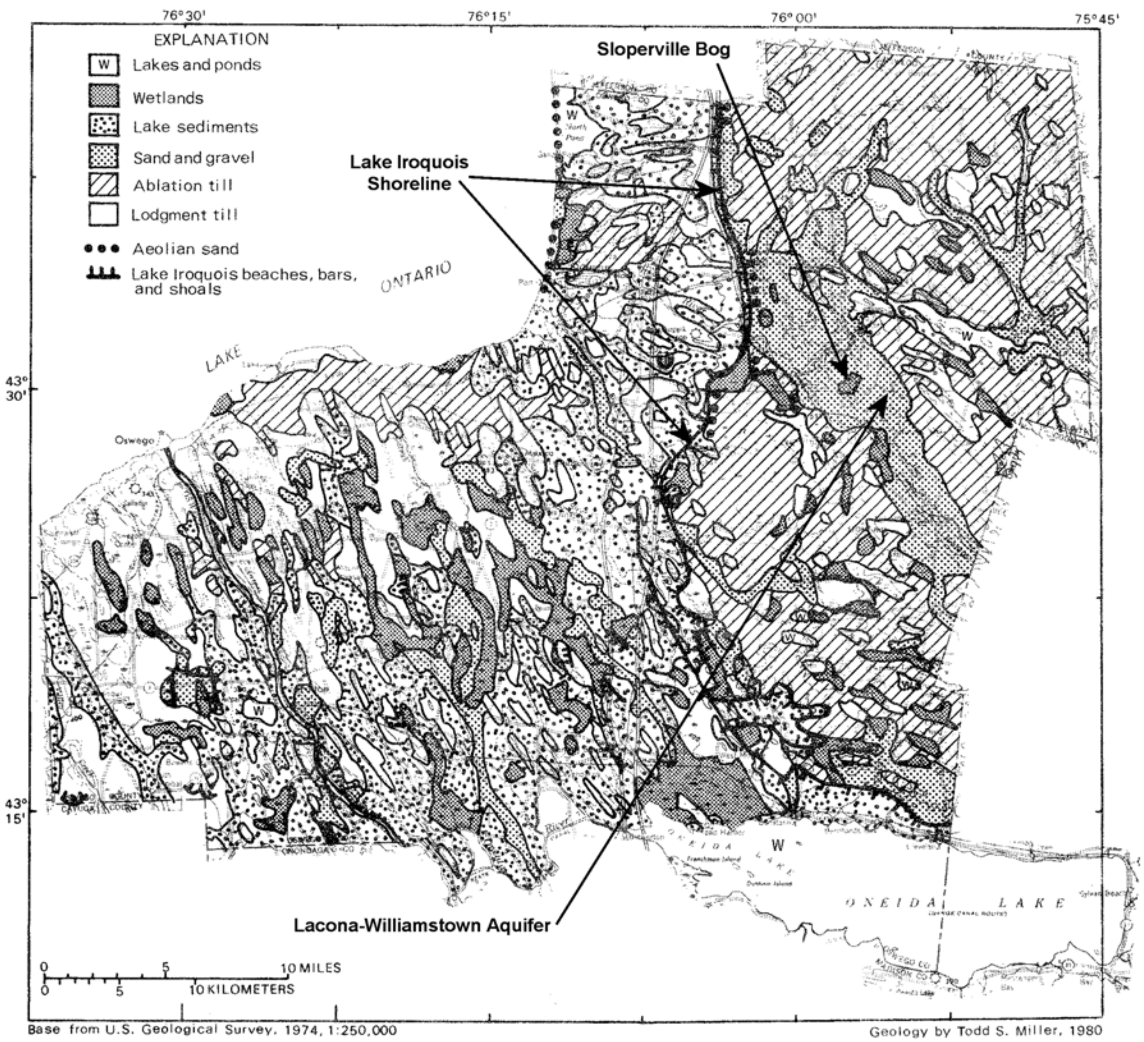

Figure 1 - Surficial geology of Oswego County (modified from Miller, 1982).

Sloperville Bog is part of a complex 73.5 ha (183.8 acre) wetland unit located in the northeast corner of the Town of Albion in Oswego County. It is situated in the adjoining west corners of the Orwell and Williamstown 7½' quadrangles. The bulk of the wetland and surrounding uplands are owned by Cotton-

Hanlon Inc., having been acquired from a private landowner in 1966[1]. The area has thus been protected from housing and other forms of development and presents a relatively unaltered landscape, flora, and fauna.

Current features of the topography and surficial geology of Oswego County are largely the result of glacial and melt water action during the most recent (Wisconsinan) glaciation and wave and wind action along the shores of proglacial Lake Iroquois and its successors. The ancient shoreline of Lake Iroquois extends south from the northern boundary of the county at approximately 76°03.75'W turning east parallel to the north shore of Oneida Lake at about 43°15'N (Figure 1). The central and western portions of the county, west and south of the Lake Iroquois shoreline, are part of the Erie-Ontario Coastal Plain, much of which features a landscape of drumlins with intervening streams and wetlands. The northeastern portion of the county is part of the Tug Hill Plateau, rising from about 150 m (500 ft) along the Lake Iroquois shore line to about 516 m (1,720 ft) at the northeast corner of the county. The topography here features a complex mix of kames, eskers, kettles, outwash plains, and wetlands. Surficial deposits include extensive areas of ablation till and sand and gravel with scattered pockets of lodgment till (Figure 1) (Miller 1982).

Oswego County is unusually rich in wetlands (sites permanently or periodically flooded or saturated by seepage and supporting emergent wetland vegetation). The 858 wetlands greater than 4.96 ha (12.4 acres) in size or of unusual local importance included in the "The Oswego County Wetlands Mapping Inventory Project" cover a total of 37,200 ha (93,000 acres) representing 15% of the county's total land area (Jones et al. 1983a).

Wetlands are commonly and broadly categorized according to the nature of the substrate and the physiognomy of the vegetation. Wetlands with a substrate of mineral soil are broadly distinguished as marshes, if dominated by herbaceous plants, or swamps, if dominated by woody plants. Where conditions are such as to limit aeration, partly decomposed plant remains may accumulate resulting in a substrate of organic peat. Most forested peatlands, like forested mineral soil wetlands, are called swamps, the different types characterized by the presence and/or dominance of different tree species. Open peatlands, dominated by herbaceous vegetation and especially those where plant roots and peat form a floating mat, are commonly called "bogs". Wetland ecologists, however, often reserve the term "bog" for peatlands where the input of water and dissolved minerals is primarily from direct precipitation with little if any contribution from ground water or surface runoff. Bogs in this strict sense have not been found in Oswego County. Floating mat peatlands fed by surface runoff, ground water, or adjacent flowing streams are referred to as "fens". Fens exhibit a gradation of features allowing identification of "poor fens", "medium fens" and "rich fens". A fen built up on the soils of a lake shore or stream floodplain where the mat is periodically inundated instead of floating with changing water levels is sometimes called a sedge meadow (Edinger et al., 2002), though some authors use this term only in association with sedge dominated marshes (Crum, 1988). Any of these open peatlands may be gradually invaded by woody plants turning a rich or medium fen into a shrub fen or a poor fen into a dwarf shrub bog (Edinger et al., 2002).

Eighty eight (88) of the 858 wetlands identified by the 1983 Oswego County wetlands inventory were cited as being or containing "bogs". Over the last ten years collaborative research with Peter A. Rosenbaum (Biology Department at Oswego) has focused on evaluation of these sites as potential habitat for species of plants and animals recognized as rare and endangered.

The majority of these peatlands ("bogs") are poor fens of the sort commonly referred to as "kettlehole bogs"[2]. Characteristically, these fens develop in the ponds formed in kettles, localized depressions resulting from thawing of isolated blocks of glacial ice that were buried in ablation till or kame deposits. These basins often have no surface drainage and are fed by surface runoff from a small watershed and/or by

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[1] Cotton-Hanlin is a timberland management company with holdings in New York and Pennsylvania. Lands in the Sloperville Bog area are leased to a private hunting club and public access is not available during the hunting season, including the spring turkey season.

[2] Authorities differ on whether a "kettlehole bog" should be called a poor fen (Edinger et al., 2002) or a bog (Johnson, 1985).

ground water. In the classic model, a floating mat of Sphagnum peat develops around the edge of the pond, gradually covers the pond, and eventually fills the basin.

Poor fens are fed by surface runoff and ground water relatively low in mineral content (weakly minerotrophic). The peat is composed mostly of the remains of *Sphagnum* (peat) mosses with a minor amount of material derived from grass-like (graminoid) plants of the sedge family (Cyperaceae) and from woody plants. pH is quite low ranging from 3.5 to 5.0 (Edinger et al., 2002). The surface of the peat is covered with a nearly continuous layer of living *Sphagnum*. The *Sphagnum* peat constitutes a chemical environment in which the low pH is maintained and the common plant nutrients, particularly calcium, nitrogen, phosphorous, and potassium, are bound in forms unavailable to most plant species. Woody plants such as eastern white pine (*Pinus strobus*), tamarack or eastern larch (*Larix laricina*), black spruce (*Picea mariana*), and red maple (*Acer rubrum*) are often present as stunted, slow growing individuals. Shrubs particularly adapted to growth in acid, low nutrient habitats, many of them in the Heath family (Ericaceae), are usually present and may be abundant. Included here are highbush blueberry (*Vaccinium corymbosum*), black huckleberry (*Gaylussacia baccata*), leatherleaf (*Chamaedaphne calyculata*), bog laurel (*Kalmia polifolia*), bog rosemary (*Andromeda glaucophylla*), and small cranberry (*Vaccinium oxycoccos*). Graminoids in the Cyperaceae commonly include certain species of *Carex* (sedge), *Rhynchospora* (beakrush), and *Eriophorum* (cottongrass). Carnivorous plants, such as pitcher plant (*Sarracenia purpurea*), sundews (*Drosera rotundifolia* and *D. intermedia*), and bladderworts (*Utricularia* spp.), are abundant. Wild orchids, such as swamp pink (*Arethusa bulbosa*), grass pink (*Calopogon tuberosus*), pink ladyslipper (*Cypripedium acaule*), white fringed orchid (*Platanthera blephariglottis*), and rose pogonia (*Pogonia ophioglossoides*), are also characteristic species of this community. Poor fens, highly localized and somewhat unusual in our area, are similar in many ways to the vast muskeg of the taiga, the boreal forest community which circles the globe in more northern latitudes. Like alpine ecological communities at high altitudes, poor fens may remain as relict representatives of more widespread communities that occupied our area as the glaciers receded in the late Pleistocene.

Thirteen (13) of the eighty eight (88) peatland containing wetlands cited in the wetlands inventory support medium fens. Medium fens are fed by moderately mineralized ground waters that, in our area, are probably enriched in calcium as the result of passing through calcareous sands or gravels. In most cases the wetland with which they are associated has a perennial outlet stream. The peat formed in medium fens is a mixture of the remains of sedges, mosses, and woody plants with sedges predominating. Medium fens have a pH of from 4.5 to 6.5 (Edinger et al. 2002). Medium fens we have explored in Oswego County are often situated along the margins of a pond. In these cases, the pH of the pond waters is usually significantly higher (7.0 to 8.8) than that of the bordering fen. Where a sedge-dominated medium fen borders an area of *Sphagnum* dominated poor fen, the transition from medium to low pH can be as sharp as the visual line that marks the edge of the *Sphagnum* mat. The dominant plants of medium fens are mat forming sedges such as *Carex lasiocarpa*, and *Cladium mariscoides* (bog-rush). *Sphagnum* mosses are usually confined to hummocks that build up around scattered trees and shrubs or clumps of royal fern (*Osmunda regalis*). A characteristic shrub of medium fens is sweet gale (*Myrica gale*). Leatherleaf and the other ericaceous shrubs also seen in poor fens may be present and abundant. The large cranberry (*Vaccinium macrocarpon*) is more likely to be more abundant than the small cranberry. Black spruce is often absent and northern white-cedar (*Thuja occidentalis*) and poison sumac (*Toxicodendron vernix*) are often present as well as eastern larch, white pine, and red maple.

Rich fens, which to our knowledge are not represented in Oswego County, usually occur in areas of carbonate bedrock. The pH of these systems ranges from 6.0 to 7.8. Graminoid peat formed from the dominant sedges may be accompanied by layers of marl. *Sphagnum* is absent or confined to hummocks. Characteristic sedges include the spikerush *Eleocharis rostellata*, the bog-rush, and the sedges *Carex flava, C. lasiocarpa, and C. hystericina*. Other herbaceous plants of rich fens are grass-of-parnassus (*Parnassia glauca*) and Kalm's lobelia (*Lobelia kalmii*). Characteristic woody plants include shrubby cinquefoil (*Potentilla fruticosa*), bayberry (*Myrica pensylvanica*), poison sumac, red maple, northern white cedar, eastern larch, sweet gale, and swamp fly honeysuckle (*Lonicera oblongifolia*).

The classifications that result from study of ecological communities represent, at best, a rough approximation of the actual situation in nature. Boundaries between different types of communities consist of gradients, sometimes abrupt and sometimes gradual, in all recognized parameters. Particular plant species that characterize a wetland community in central New York may not be characteristic of the same community as represented in the lake states or northern New England and a given wetland, like Sloperville Bog, may contain a mosaic of different wetland community types.

SLOPERVILLE BOG

Sloperville bog is part of the largest of three segments of the wetland designated as OR-29 on the Oswego County Wetlands map. Beaverdam Brook enters the wetland from the southeast, threading its way around and through a complex of eskers before expanding into an area of open water in the southern portion of the wetland and then exiting through a constriction at the southwest corner of the wetland immediately south of the south end of Sloperville Drive (Fig. 2). Water is retained in the northern portion

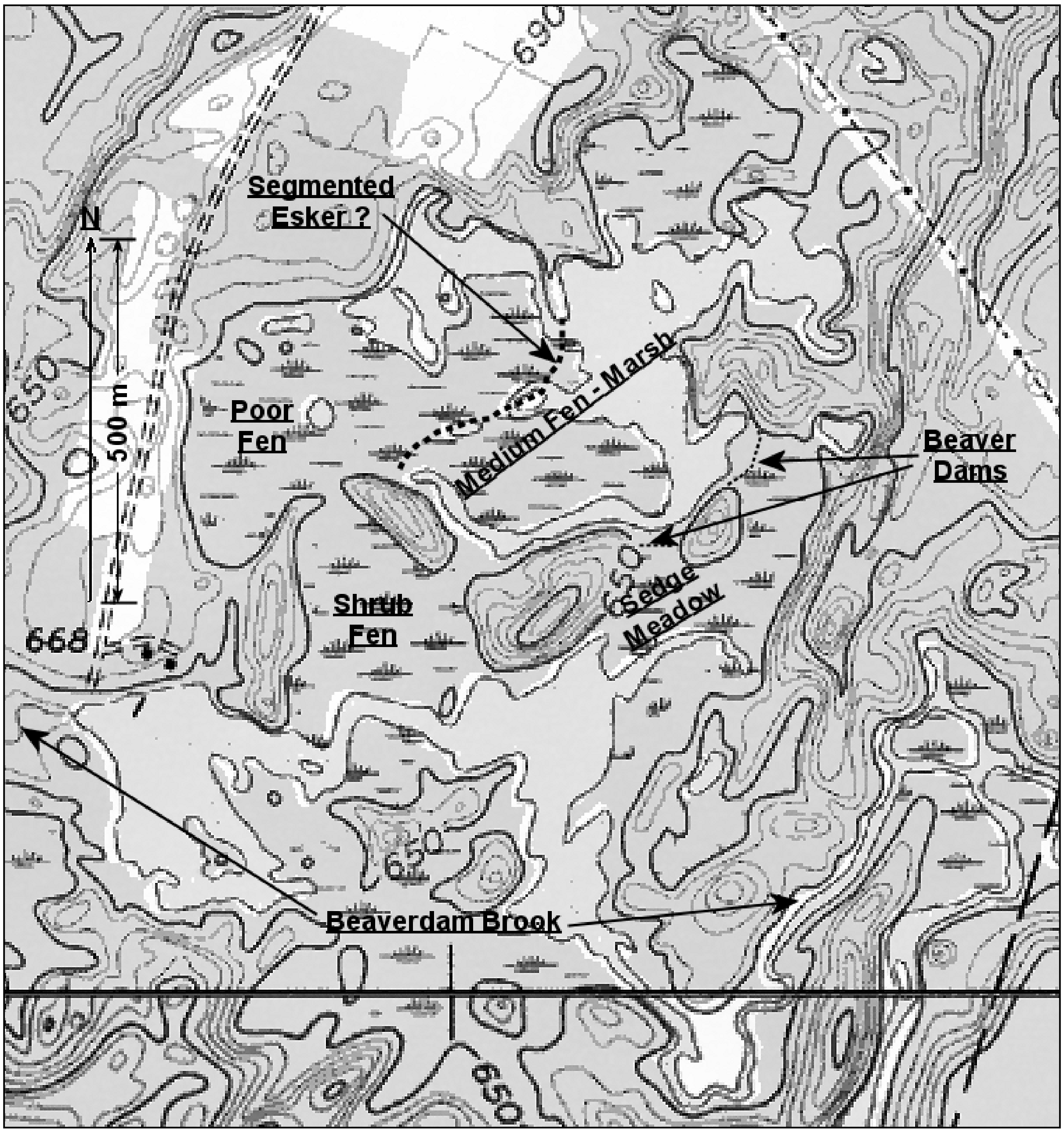

Figure 2 – Sloperville bog showing features mentioned described in the text.

of the basin by an undulating esker meandering on an east-west course across the middle of the basin. The dam formed by the esker is completed by beaver dams at two points where the esker is broached or submerged. This northern section receives runoff from a limited watershed. A steady flow of water in the stream draining from the northeastern of the two beaver dam areas suggests a significant and continuing inflow of ground water into the northern part of the basin. The northwestern lobe of the basin supports the extensive poor fen known as Sloperville Bog. Roughly half way across the basin to the east the poor fen changes over to medium fen and marsh dominated by *Carex lasiocarpa* and then gives way to open water. In the eastern part of the basin immediately below the beaver dams is a sedge meadow traversed by narrow rivulets that drain into the stream draining the northern part of the wetland down to Beaverdam Brook. Further west a shrub fen has developed in a cove bounded by the main esker that extends across the basin (Figure 2). A total of 136 species of vascular plants (clubmosses, ferns, conifers, and flowering plans) have so far been recorded from the Sloperville bog wetland. The ecological communities of the southern half of the basin have not yet been intensively explored and characterized.

The poor fen in the northwestern lobe of the basin is perhaps the most easily accessed and negotiated poor fen in the county and for that reason it is a favorite for introducing students to this type of ecological community. The surface of the mat is firm, except at the edges, and virtually flat. The vegetation on most of the fen is low and open posing no obstacle to progress over the surface.

The classic kettlehole bog or poor fen of the glaciated northeastern United States consists of a central pond surrounded by the floating mat of an open fen which gives way at its inland margins to shrubs and eventually trees, the whole comprising a series of concentric zones around the pond. The floating mat of the Sloperville poor fen, in contrast, extends to the edges of the basin separated from the mainland only by a narrow moat where incoming surface runoff supplies sufficient oxygen and mineral nutrients to facilitate decomposition of plant materials. There is a central pond, in this case immediately surrounded by a narrow floating mat consisting primarily of the sedge *Carex limosa* and dense clumps of leatherleaf. This, in turn, is surrounded by a zone of dense, head high blueberry and huckleberry bushes along with relatively tall black spruce and larch. A few other isolated, smaller groves of spruce and larch also occur in the fen. A total of fifty four (54) species of vascular plants, all more or less characteristic of poor fens in this region, have been recorded from this fen. This includes all of the orchid species so far found in the wetland and most of the carnivorous plants. Except for the thicket surrounding the central pond, the woody plants of this fen are quite stunted compared to individuals in some other poor fens of the area. Dwarf mistletoe (*Arceuthobium pusillum*), a diminutive parasitic flowering plant, can be found on the branches of some of the black spruce in the thicket on the northeast edge of the central pond area.

Roughly half way across this northern basin some isolated upland islands interrupt the fen. In this area, the poor fen of the western lobe of the basin gives way to a *Carex lasiocarpa* dominated fen, then to an area of partially submerged *C. lasiocarpa* and finally to shallow, open water. The restriction of the poor fen to the northwestern lobe of the wetland suggests that there is some limitation on drainage from this section into the rest of the basin. It is tempting to suggest that the isolated islands across the center of the basin may be the exposed portions of a segmented esker the rest of which has been buried by flooding and accumulation of peat and other sediments in the basin (Figure 2) (Jones et al., 1983b). Such a sub-surface ridge might restrict the drainage of water from the western portion of this basin. Investigation of this possibility would contribute greatly to an understanding of the nature of Sloperville Bog.

The medium fen and area of open water in the eastern part of the basin present a complex of habitats from which fifty three (53) species of plants have been recorded. Species characteristic of medium fen, marsh, shrub swamp, and aquatic habitats are included.

Directly down stream from the western of the two beaver dams currently controlling the water level in the large northern basin is a shrubby area leading to the flood plain of the stream draining the basin through the eastern beaver dam (Fig. 2). Seepage from this area flows through shallow, narrow rivulets into the stream running south below the eastern beaver dam. A small pond impounded by a second dam below the main eastern dam provides habitat from which twenty two (22) aquatic and lake shore plant species were recorded on one visit in July 2004. Thirteen (13) of these have not been recorded from other sections of the wetland north of Beaverdam Brook and an additional five (5) have been seen elsewhere only in the medium fen and open water area directly north of the beaver dams.

The flood plain of the stream draining the northern basin supports a sedge meadow dominated by *Carex lasiocarpa*. Plant species records as currently kept for this area include the shrubby margins of the flood plain as well as the open sedge meadow itself. Sixty three (63) species have been recorded from this area making it the most diverse part of the wetland surveyed to date. The sedge meadow has a level surface consisting of a firmly grounded layer of peat crisscrossed by deer trails and narrow channels draining water into the stream. In addition to *Carex. lasiocarpa*, other species found here that are characteristic of medium fens include yellow sedge (*C. flava*), marsh bellflower (*Campanula aparinoides*), marsh cinquefoil (*Potentilla palustris*), and lesser bladderwort (*Utricularia minor*). The lesser bladderwort is unusually abundant, flowering in great numbers in the small drainage channels and deer trails in this community. This diminutive plant is rarely noticed outside of its flowering season, two or three weeks in late June to early July. Although it is now known to occur in a variety of shallow water habitats, it seems to have been noticed in flower only in medium fens. Because of the scarcity of documented records of its existence in New York State, *U. minor* was listed as a S2S3 (imperiled to rare in the state) species on the New York Natural Heritage Program Rare Plant Status List for 2003 (Young and Weldy, 2003). Intensive investigations that year involved search for plants in non-flowering condition and revealed that the species is much more common than previously thought. Its status was changed to S3 (rare in the state) for the 2004 list and it has been transferred from the "active" rare plant species list to the "watch" list (Young and Weldy, 2004).

A shrubby fen occupies a cove of the wetland south across the esker from the main poor fen. It is bounded on the south by Beaverdam Brook. Shrubs and small trees are quite dense in the northern (inland) portions of this area. The community becomes more open towards the south along the brook. In contrast to the level topography of the poor fen and the sedge meadow previously described, this shrub fen area presents a hummocky surface of a sort commonly found in other medium fens in the region.

DISCUSSION

The Freshwater Wetlands Act of 1975 established procedures and regulations designed to protect and conserve wetlands in New York State. This act provides a framework within which draining, filling, and shoreline development of wetlands can be controlled. Where properly applied and enforced, these laws have done much to alleviate physical modification of wetlands and pollution through direct discharge and surface runoff. In the political and social atmosphere that has accompanied establishment of these protections, informed land management on the part of land owners, such as Cotton-Hanlin, government agencies, land trusts, and conservation organizations is encouraged by positive recognition and support of the public.

The recognition of peatlands as special places goes back to ancient times (Glob, 1969). Biologists and conservationists know these wetlands as delicately balanced and easily damaged ecosystems that provide habitat for unusual, specialized, and often rare plants and animals. Much effort has been expended on characterization of the flora and fauna of these sites and on investigation of the chemical and physiological phenomena that take place in them. Interpretation of plant microfossils – spores and pollen grains – recovered from the depths of the accumulated peat has provided clues to changes in vegetation and, by inference, climate since the last glaciation. It is clear that these are sites worthy of protection and conservation. It is also clear that protection and conservation can be achieved only with an understanding of the reasons for the formation and persistence of these unusual wetlands in the context of the broader landscape.

The existence of a particular wetland community depends on the nature of the water entering the system and the pattern and rate of flow into, within, and through the system. The nature of water entering a wetland depends on where that water came from and what it has passed over or through on its way to the wetland. The pattern and rate of flow of water within the system will be influenced by the nature and distribution of accumulated organic and inorganic sediments and by the topography and permeability of the basin. In practice, ecologists and conservationists usually limit their considerations of an ecological community to observations made at or near the surface, only occasionally sampling subsurface conditions by probing or taking cores from peat and other sediments. The surface watershed and the streams providing inflow to a wetland may be investigated in order to locate natural determinants of water quality and sources of pollution. Stream flow might be estimated and ground water inflow can sometimes be recognized. However, subsurface features such as the origin and flow of ground water and the size, shape, and permeability of the submerged basin are rarely investigated. Perhaps the trip to Sloperville Bog will suggest possibilities for hydrological and geological investigations that will help us understand and protect these and other wetlands in the region.

ACKNOWLEDGEMENTS

Special thanks to Bob O'Brien, Cotton-Hanlin forester, for providing access to the Sloperville properties and information on Cotton-Hanlin's holdings. Access to the original data sheets for the Oswego County Wetlands Inventory project and the Oswego County wetlands GIS files was provided by the Environmental Management Council of the Oswego County Department of Planning and Community Development

ROAD LOG TO SLOPERVILLE BOG

From the intersection of Interstate 81 and Route 104 proceed to Sloperville Bog as follows:

CUMULATIVE MILEAGE	MILES	ROUTE
0.0	0.0	Intersection, I 81 & Rt. 104, proceed east on 104
3.4	3.4	Turn north on Rt. 22
7.8	4.4	Turn east on Rt. 13
8.0	0.2	Turn north on Cemetary St. - continues as Rt. 22
9.9	1.9	Turn south on Sloperville Drive (Hamlin Road on USGS quad)
10.4	0.5	Park at roadside, Sloperville Bog east of road

In case of confusion in Altmar, follow the white arrow in the air photo provided below:

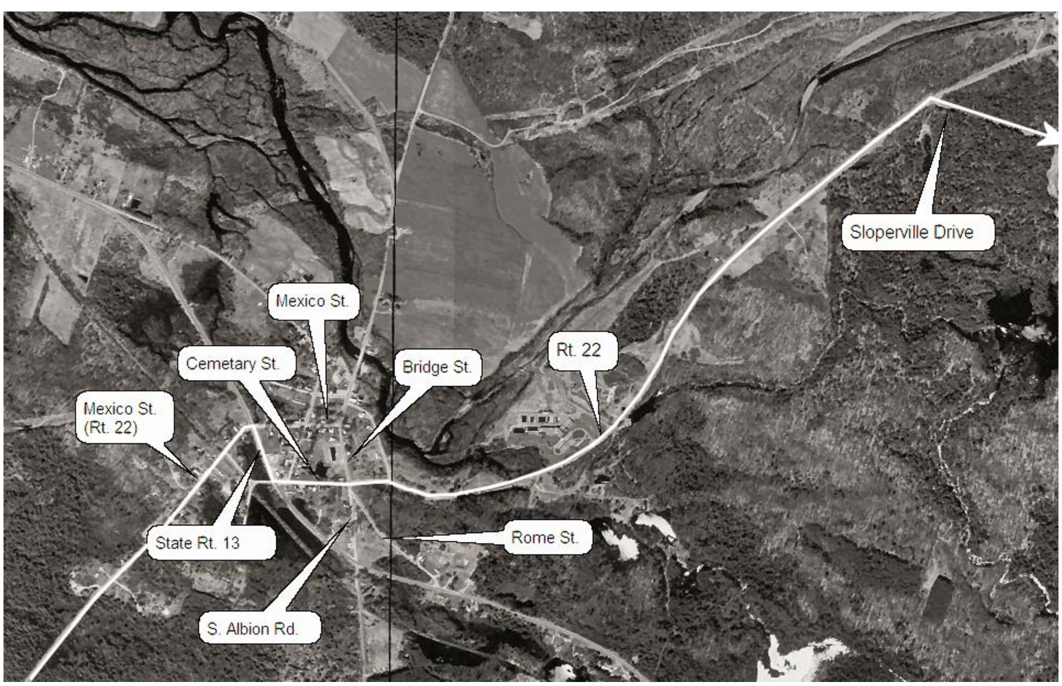

Figure 3 – Composite air photograph of the Altmar, NY area to show the location of the Sloperville bog.

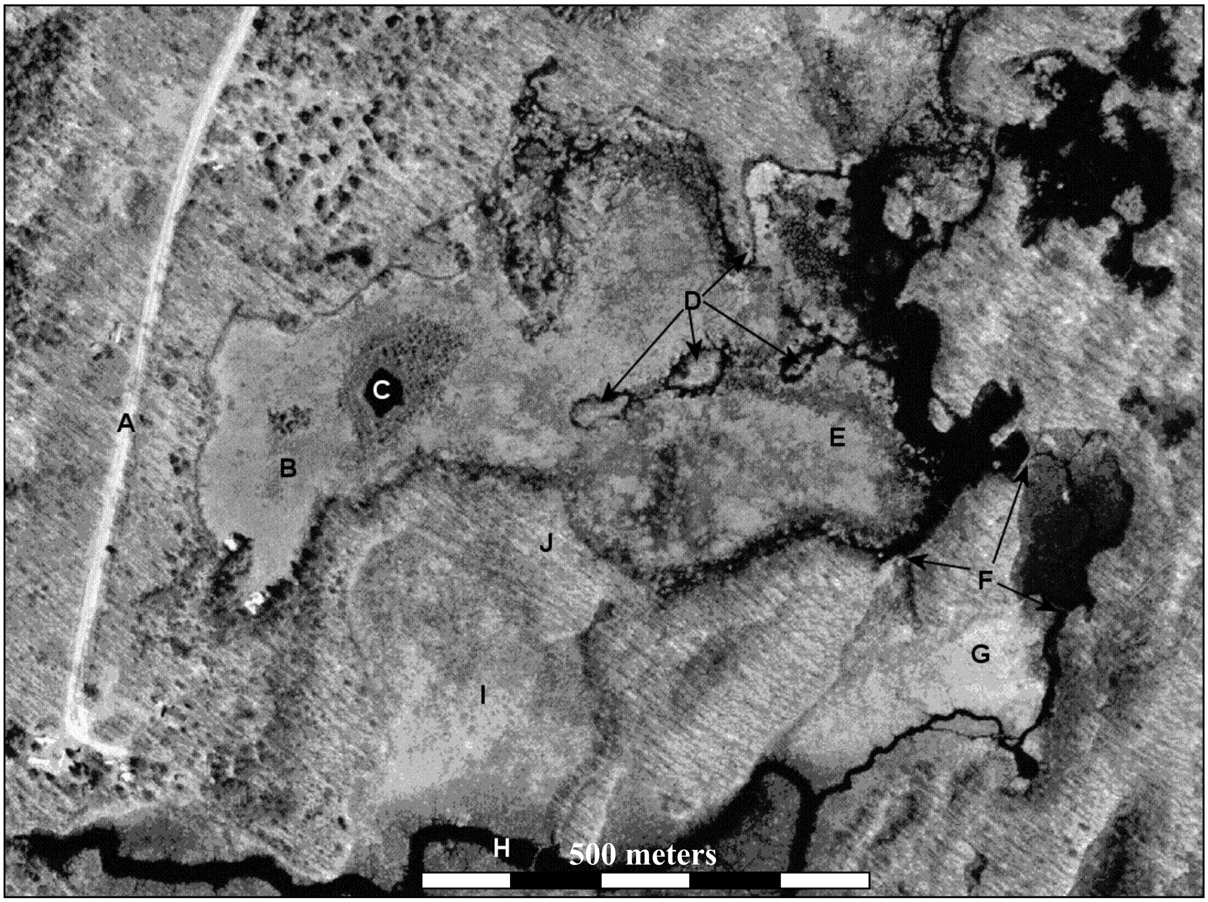

Figure 4 – High resolution air photograph of the Sloperville bog showing the field guide locations. The data source is the New York State G.I.S. Clearinghouse.

A: Sloperville Drive.

B: Poor fen. Floating mat of *Sphagnum* peat with surface layer of living *Sphagnum*. Dwarf shrubs, ferns, sedges and other bog plants make up herbaceous layer. Carnivorous plants abundant in spots.

C: Central pond surrounded by narrow floating mat of dwarf shrubs and sedges in thicket of blueberries, huckleberries, larch, black spruce (Figure 5).

D: Peninsula and islands, perhaps emergent portions of segmented esker separating western and eastern portions of northern basin and restricting drainage from western portion (Figure 6).

E: Medium fen → marsh → open water.

F: Beaver dams.

G: Sedge meadow on floodplain of stream draining northern basin. Open community dominated by the sedge *Carex lasiocarpa* on firmly grounded mat of sedge peat. Small drainage channels cross the meadow.

H: Beaverdam Brook.

I: Shrub fen. Mixture of small trees, shrubs, and herbaceous plants on mixed peat. Hummocks built up around shrubs, ferns, and sedge tussocks.

J: Main esker separating Slopperville Bog from Beaverdam Brook.

Figure 5 – Photograph of the area of open water near the center of the Sloperville bog labeled "C" in Figure 4. The view is looking southeast and the hill in the distance is labeled "J" on Figure 4.

Figure 6 – Photograph of the Sloperville bog. The view is to the north from the south side of the bog and the clusters of large trees occur at points "D" on Figure 4. The region of the photo is floating bog mat from the place where the photo was taken to the region of trees on the distant hill.

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Trip A-2

EXPLORING THE ROOT ZONE OF AN ANCIENT FAULT-DRIVEN HYDROTHERMAL SYSTEM IN THE ADIRONDACK LOWLANDS, NEW YORK

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ABSTRACT

This trip examines exposures in the Adirondack Lowlands that contain Paleozoic hydrothermal and fault-related features hosted by Mesoproterozoic 'Grenville' marbles and Cambrian Potsdam Sandstone. These features have been the source of some debate historically and are now explicable as resulting from hydrothermal activity within an ancient wrench fault system. This system is similar to the deep structures that give rise to economically important gas and oil fields in the Appalachian Basin and elsewhere. The trip will include localities where the direct effects of coupled fault-related deformation and hydrothermal alteration can be seen, and the results of recent petrographic, stable isotope and fluid inclusion work will be used to better explain outcrop-scale features. The trip concludes with a stop in Ordovician Black River Group carbonates that demonstrates on a small scale the effects of hydrothermal dolomitization related to brittle deformation that may be linked to basement faulting. The regional extent, timing and tectonic significance of the fault systems will also be considered.

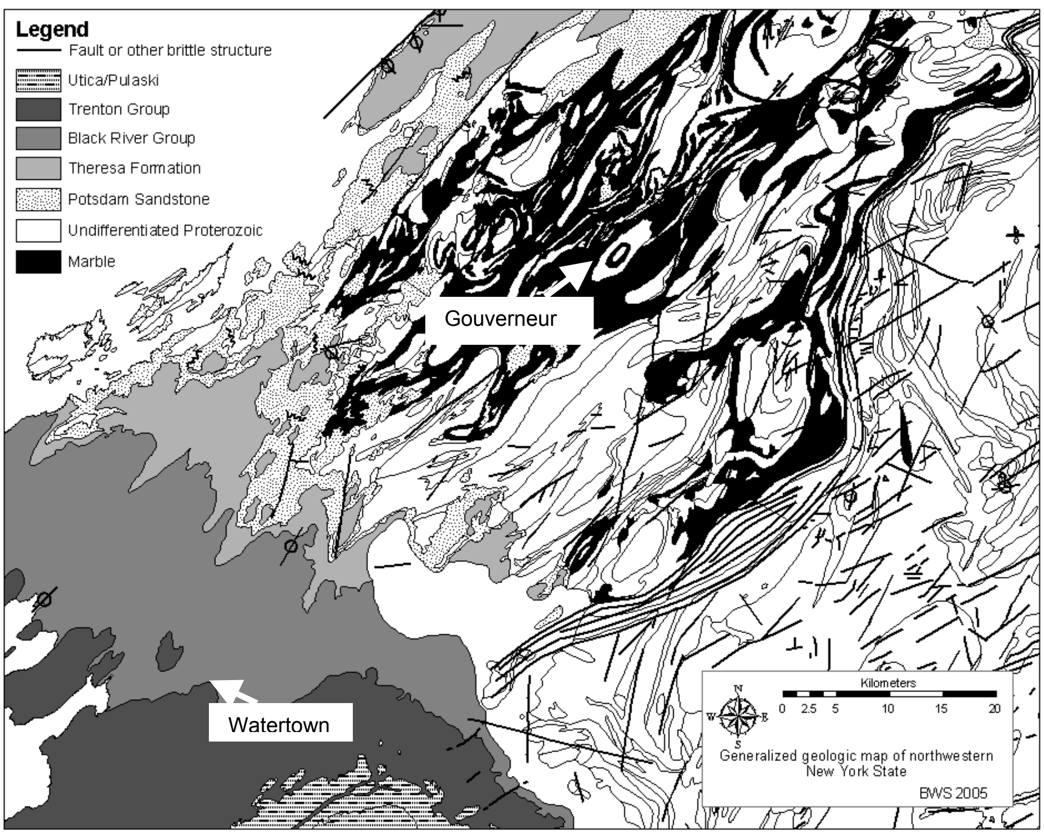


Figure 1 - Generalized geological map of northwestern New York.

INTRODUCTION

The southern Canadian Shield of Ontario and the Adirondack periphery in northern New York State offer numerous exposures of the contact between high-grade metamorphic Mesoproterozoic basement and Cambrian Potsdam Sandstone. In some outcrops in this region the pre-Potsdam erosional topographic surface of the 'Great Unconformity' is exposed at small scale (Tiller and Selleck 1992). Higher order topography, particularly in the Adirondack Lowlands and Frontenac axis region, locally reflects Pleistocene and Holocene erosional unroofing of basement that has been underneath Paleozoic cover for 500 million years. While this region has been in general tectonically inactive during that time interval, insofar as major deformation of the Paleozoic rocks is not regionally evident, it is not surprising given that long interval of time that minor structural deformation and hydrothermal alteration has occurred. Faulting and gentle folding involving the basement and Paleozoic cover have been recognized in the Adirondack Lowlands (Barber and Bursnall 1978) and modern earthquake activity is a reminder of the ongoing brittle deformation of the crust at depths of 10 to 18 km in the region (Daneshfar and Benn 2002; Mitronovas 1985).

Hydrothermal alteration caused by fluids derived from the Paleozoic sedimentary sequence interacting with basement rocks immediately underlying Paleozoic cover is well-documented in southern Ontario and northern New York (Ziegler and Longstaffe 2000). This alteration has a distinctive spatial distribution related to fluid flow from the basal sandstone aquifer and presence of reactive rocks in the underlying basement. The mineralogical signature of this hydrothermal interaction includes a range of low-temperature (<300^{o}C) phases that occur as vein and vug fills in both the basement and cover rocks, as authigenic phases in the sedimentary cover, and as distributed retrograde alteration of basement rocks. Mineral systems associated with this alteration include carbonates (dolomite, calcite, siderite, ankerite), phosphates (apatite, monazite, xenotime), sulfides (pyrite, galena, sphalerite), oxides (magnetite, hematite, anatase), phyllosilicates (kaolinite, illite, chlorite), and framework silicates (quartz, K-feldspar). This field trip will focus on the hydrothermal mineralization and alteration found in the marble belt region of the Adirondack Lowlands province. In the exposures we visit on this trip widespread dolomitization associated with Potsdam Sandstone erosional outliers and fault structures is the dominant type of basement alteration. This zone is interpreted as the structural root of an ancient hydrothermal dolomite (HTD) reservoir system. The overlying Paleozoic strata which hosted the HTD reservoir have been eroded away, but the fault-related features and hydrothermal mineralization of basement rocks and basal Potsdam Sandstone provide a window into the processes which give rise to these economically important systems which are well-known in a number of hydrocarbon provinces (Berger and Davies 1999). The timing of the coupled fault-hydrothermal activity is not well-constrained, but available data suggest activity could have occurred in medial-late Ordovician time coeval with the Taconic Orogeny or later in late Devonian time during the Acadian Orogeny.

REGIONAL GEOLOGIC SETTING

Proterozoic Basement

The Adirondack Highlands Massif and adjacent Adirondack Lowlands are part of the Grenville Province, a Mesoproterozoic collisional tectonic belt extending from Labrador to Mexico that represents the addition of crustal material to the Laurentian craton during successive orogenic episodes, with intervening sediment deposition and anorogenic magmatic events. The oldest rocks of the Adirondack Highlands are a series of ca 1300 Ma arc-magmatic systems and intervening sedimentary basins. These rocks were metamorphosed during a collisional event ca. 1180-1170 Ma. A period of magmatism marked by intrusion of anorthosite, gabbro and coeval granitoid rocks at ca. 1160-1150 Ma emplaced large volumes of plutonic rock throughout the Adirondack Highlands and Lowlands. The Ottawan compressional orogenic episode marked the final assembly of the supercontinent of Rhodinia at ca. 1070-1050 Ma (McLelland et al. 2001). This produced a Himalayan-scale mountain belt resulting in granulite facies metamorphism in the Adirondack Highlands. At that time the Adirondack Lowlands were likely at a higher crustal level, because peak metamorphic Ottawan ages are not for the most part recorded in the Lowlands, although certainly there was widespread deformation. The later phases of the Ottawan orogeny are marked by generation of leucogranite melts that were often preferentially emplaced along zones of crustal extension. One of these zones, the Carthage-Colton Shear Zone, marks the boundary between the Highlands and Lowlands, and represents a northwest-dipping, top down to the northwest extensional structure within which ca. Ma 1045 leucogranite was emplaced synchronously with extension. The zone is marked by mylonite, ultramylonite and pseudotachylite.

The Adirondack Lowlands contain (by outcrop area) much more metasedimentary rock than the Highlands but belts of metaigneous granitoid gneiss and amphibolites are widespread. The Lowlands reached peak metamorphic conditions of the upper amphibolite facies (ca. 670^{o}C, 6 kb) ca. 1170 Ma. The metamorphic expression of the ca. 1070-1050 Ma Ottawan orogeny is generally subdued in the Lowlands except in the vicinity of shear zones (Dahl et al. 2004). The Lowlands are characterized by a very strong northeast-southwest trending topographic grain that is related to the contrasts in erosional resistance of marbles, quartzites, metapelites, granitic gneisses, and amphibolites which form the majority of the rocks of the Lowlands. This erosional topography was present on the surface which was covered by the Cambrian Potsdam Sandstone and has been exhumed by Cenozoic stripping of Paleozoic cover rocks from the Mesoproterozoic basement of the Lowlands.

Paleozoic Sedimentary Strata

The Ottawan Orogeny ended in the Adirondack Highlands ca. 1040 Ma; extension and regional tilting brought the Adirondack Highlands and Lowlands into their current juxtaposition (Dahl et al. 2004). In late the Neoproterozoic (ca. 750 Ma) the breakup of Rhodinia gave rise to a rift-to-passive margin tectonic setting along the eastern margin of a new continent – Laurentia. The long period of post-Ottawan erosion removed ~ 20+ km of rock from the Highlands and by medial Cambrian time (ca. 520 Ma) northwestern New York was a low-lying coastal region with scattered aeolian dunes and braided streams which deposited quartz-rich sands of the basal Potsdam Sandstone. The region was flooded under shallow marine waters intermittently through the latest Cambrian and early Ordovician with carbonate-rich sandstones of the Theresa Formation and carbonates of the Ogdensburg Dolostone accumulating atop the Potsdam Sandstone. These younger units occasionally onlapped erosional monadnocks of Grenville basement such that no intervening Potsdam Sandstone was deposited. Early middle Ordovician Chazy Group limestones were deposited in the St. Lawrence Lowlands and Champlain Valley, and perhaps in northwestern New York State. Erosion of pre-middle Ordovician strata stripped these deposits from the western St. Lawrence Lowlands and interior of central and western New York (Robinson 1998). Widespread flooding of the continental interior in medial Ordovician time resulted in deposition of shallow marine and coastal lowland carbonate sediments of the Black River Group; these were succeeded by deeper subtidal and shelf carbonate facies of the Trenton Group. The development of the Taconic orogenic belt along the eastern margin of Laurentia in the medial Ordovician (ca. 440 Ma) produced a continental foreland basin that resulted in progressive flooding of the Trenton shelf and deposition of deeper water limestones and black organic-rich shales of the Utica and Canajoharie Formations. As the Taconic collisional event ended, late Ordovician and early Silurian deltaic and coastal plain facies infilled the relict foreland basin. These units are now found in the Tug Hill Plateau region and eastern St. Lawrence Lowlands. The Adirondack Lowlands region was likely covered by these and later Devonian, and perhaps Carboniferous, strata derived from erosion of the Acadian and Alleghanian orogens of the eastern margin of Laurentia as collision with magmatic arcs and assembly of the Pangean supercontinent closed the Paleozoic. The record of these events is based upon regional reconstructions and projections of stratal thickness patterns with little constraining data for area of this trip in northwestern New York. K-Ar and Ar-Ar dates on diagenetic illite in the Potsdam Sandstone in the Alexandria Bay area suggest elevated burial temperatures at ca. 350 Ma (Reynolds and Thomson 1993; Selleck 1995). Apatite fission track ages suggest regional unroofing was occurring in many areas of the Adirondack Highlands by ca. 120 Ma (Roden-Tice et al. 2000). All in all, as much as 4-5 km of Paleozoic strata may have covered the region by late Carboniferous time, but as little as 1-2 km is an equally likely total.

Mesozoic and Cenozoic Unroofing

Certainly dinosaurs once roamed the land which now makes up northwestern New York, but we have no record of their presence because no sedimentary deposits were laid down at the appropriate time. Based upon regional patterns, the most likely scenario is that the region was subjected to minor uplift and extension related to late Triassic-Jurassic rifting of Pangea. Subsequent thermal relaxation of the continental margin coupled with regional erosion that began in earnest ca.120 Ma lowered the region to a broad peneplain by Cretaceous time (ca. 90 Ma). In mid-Paleogene time (ca. 40 Ma) a regional uplift is suggested by changes in drainage patterns and warping of the Cretaceous peneplain. A regionally extensive dendritic drainage pattern that probably drained to the paleo-Mississsippi River was in place over the region by ca. 15 Ma. The onset of Neogene glaciation ca. 1.2 ma continued the erosional unroofing of the landscape and the modern erosional was exposed at the end of the last glacial advance retreat cycle which ended ca 12,000 ya. The modern drainage was established by perhaps 7,000 ya, as isostatic rebound of the region allowed marine waters that had invaded the Ottawa-St. Lawrence Lowlands to recede. The modern topography expressed within the region is very much a relict of a long and complex history.

Modern Topography, Structural Control, Fault/Fracture Lineaments and DEM Imagery

The Adirondack Lowlands province is often set out as exemplar of structural/lithologic control on topography and drainage patterns. Indeed, any topographic, hydrographic, geologic, or even a road map of the region clearly displays the strong northeast to southwest ‘grain’ that is largely controlled by the

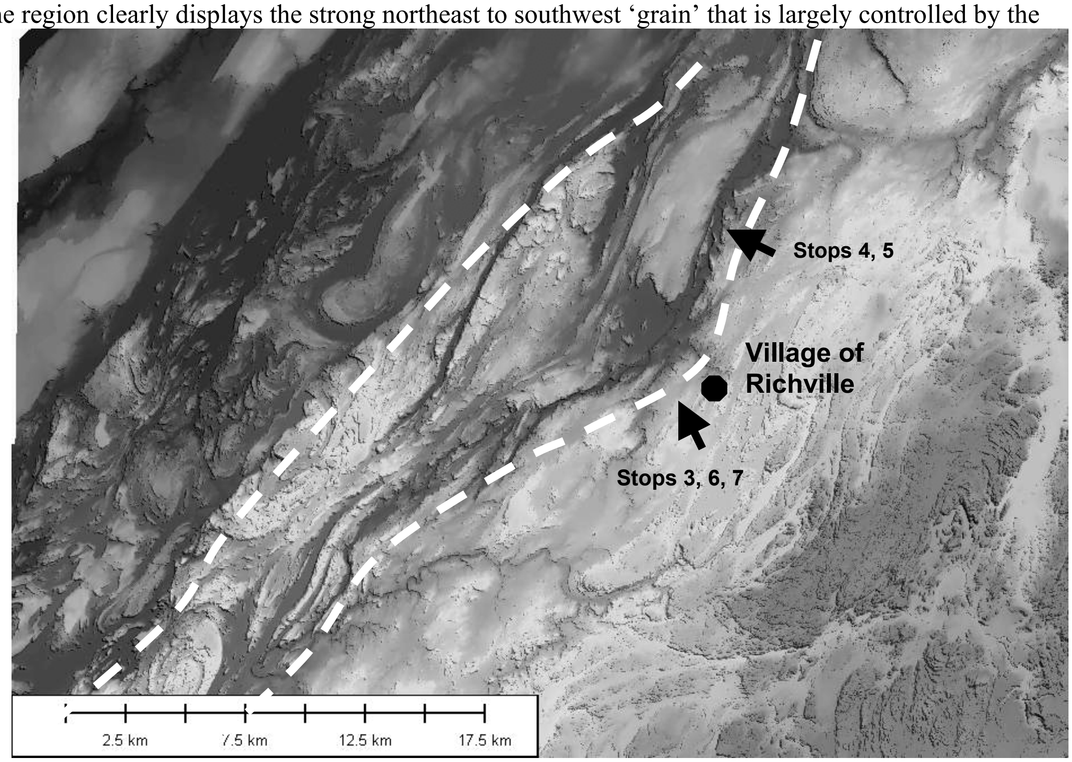


Figure 2 – Digital elevation model image of a portion of the Richville Shear Zone. White dashed lines indicate approximate boundary of zone with coupled fault/hydrothermal features.

differential erosional resistance of ductilely deformed basement rocks. In particular, the so-called marble belts often define (or at least have been mapped as such) linear valleys with intervening ridges of resistant metaigneous and metasedimentary gneiss. The well-defined linear outcrop patterns also define second-order fold structures, particularly in the vicinity of deformed plutons of ca. 1200 Ma alaskites which intrude older metasedimentary and metaigneous rocks. This topographic arrangement is a central feature of the lowlands terrain, and where the structures are masked by Paleozoic or Quaternary cover, the contrast is striking. Post-folding fault structures are evident as offsets of the linear topography or as truncations of arcuate fold ridges. Figure 2 illustrates digital elevation model (DEM) data for the region of this field trip. This imaging method has clear advantages over traditional contour maps, satellite or aerial photographs at this scale in that the image generated is based upon pure topography without the distraction of vegetation cover, roads or other anthropogenic features. The relationship between brittle structures, linear valleys and offsets in the Proterozoic structural topography are obvious and indeed a number of the mapped brittle structures are based on the existence of the lineaments without, in some cases, known geologic offset. Where these structures encounter Paleozoic or younger cover, the pattern of faulting is less clear, although fold and fault structures mapped in the Theresa, NY area by (Barber and Bursnall 1978) may be directly related to the basement features described here. One striking aspect of the terrain depicted in figure 2 is the set of lozenge-shaped structural domains outlined by shear zones, faults and lineaments. These domains are the manifestation of a 10-km broad sinistral strike-slip wrench fault system, with later minor dextral motion. The system is termed the Richville Shear Zone (RSZ) for exposures in the vicinity of Richville, NY, some of which we will visit on this trip. The along-strike extent of the RSZ is some 60 km and its continuation beneath Paleozoic cover to the northeast and southwest is not well-constrained.

FAULTING AND HYDROTHERMAL MINERALIZATION

One of the goals of this trip is to examine outcrop-scale features produced by faulting and hydrothermal activity along the Richville Shear Zone. The spatial arrangement of these features within the RSZ system and other similar fault systems in the area suggests a close linkage between faulting and hydrothermal activity. However, evidence of hydrothermal alteration can be found along the Potsdam Sandstone basal unconformity where no evidence of post-Potsdam faulting exists. These examples reflect passive fluid alteration and mineralization of the basement rocks, and these phenomena have been recognized widely in the St. Lawrence Lowlands and adjacent areas in the southern Canadian Shield and in the eastern Adirondack Highlands (Selleck 2004).

Rock and mineral assemblages

Marble. The majority of the outcrops we will examine contain calcite marble as the dominant lithology. Unaltered marble of the Adirondack Lowlands is typically coarsely crystalline (cm-scale calcite crystals are typical) white, bluish gray or gray in color with cm- to dm-scale metamorphic banding marked by segregations of calcsilicate minerals and graphite. A variety of calcsilicate lithologies (diopsidite; talc-tremolite quartzite; amphibolite; quartz-tremolite schist, etc) are found within marble and these rocks may form boudins, clots and dismembered fold hinges scattered in ductilely deformed marble.

Dolomitized marble. The outcrops we visit also contain dolomitized marble that resulted from Paleozoic alteration of Grenville calcite marble by Mg-bearing hydrothermal fluids. Dolomite may be difficult to distinguish from calcite on freshly broken surfaces, but modest weathering will usually clearly define the dolomite by its tan, buff, or yellow-gray color. The yellowish cast is due to the formation of hydrous Fe-oxides on the weathering surface produced when the dolomite dissolves slightly and releases ferrous iron from the dolomite crystal structure. Although there are dolomite marble layers that were part of the original depositional sequence deformed to produce the marble belts of the Lowlands, the Paleozoic dolomitized zones are clearly distinguished by calcite-dolomite boundaries that cut across compositional banding produced by metamorphic segregation that accompanied Proterozoic deformation. The Paleozoic dolomite is also characterized by numerous mm- to micron-scale voids that are often partly filled by crystals of dolomite, calcite quartz and sulfides. These voids are produced during the calcite to dolomite conversion process which results in a solid volume reduction. The void space could not have survived Grenville metamorphism so must be related to the Paleozoic dolomitization event. Areas of undolomitized marble adjacent to dolomite/calcite marble boundaries are often marked by a red or pink cast due to disseminated micron-scale hematite crystals in the undolomitized marble. Dolomite in the vicinity of fault zones may form lineations marked by arrays of dolomite crystals aligned parallel to slickenlines on weak cleavage or fracture surfaces.

Potsdam Sandstone. The Potsdam Sandstone in the outcrops we will visit is generally easy to identify as a coarse to medium yellow-white to red-orange sandstone with cm to dm thick pebble conglomerate and pebbly sandstone beds. Although most Potsdam Sandstone outcrops are clearly stratified, primary sedimentary structures are not easily discerned due to minor structural disruption or mineralization. Sand from the Potsdam is also found within open fractures and filling hydrothermal karst tunnels and pipes.

These fillings are often deeply colored red or maroon by abundant hematite cement, and the sand is usually tightly cemented by quartz and carbonate minerals, but rounded sand grains can usually be seen with a hand lens. Some of the conglomerates within the Potsdam contain chert clasts that are the result of silicification of marble clasts. Pebbles of jasper and clasts of laminated sandstone that had been silica-cemented, reworked and re-deposited are also present in the conglomerate and pebbly sandstone beds. A sandstone fracture-fill in granitic gneiss at Popple Hill (Rt. 58 south of Gouverneur and not visited on this trip) contains fragments of the inarticulate brachiopod *Lingulepis accuminata.* This species is a widespread form found in the Lower Ordovician Theresa Formation of the St. Lawrence Lowlands. Its presence in the fracture-fill in the Gouverneur area demonstrates that the Theresa Formation once extended over the region, and that unconsolidated sand with brachiopod fragments was sluiced downward into the fracture as it was opened during a faulting event.

Deformed Potsdam Sandstone. Within fault zones proper, the Potsdam Sandstone is extensively recrystallized and locally forms discrete fault knockers or tectonic clasts which may have an internal fabric of highly strained quartz ribbons. Clasts of Grenville quartzite may be similarly entrained within fault gouge rock and are also internally deformed and difficult to distinguish from recrystallized Potsdam Sandstone clasts. Fault gouge rock more generally consists of a mélange bearing cm-to m-scale clasts of the strongest rock types, quartzite and cemented Potsdam Sandstone, typically. The matrix of the mélange is typically a fine-grained mixture of quartz, hematite, illite and chlorite. Fabric development in the mélange is variable, but alignment of clasts with intervening weakly cleaved matrix is typical.

Hydrothermal Features. Fractures and hydrothermal leaching voids, and relict void space from partially filled tunnels, pipes and veins often contain coarsely crystalline dolomite and calcite which precipitated directly from hydrothermal fluids to form vuggy crystal void fills. Floors of voids may be filled with internal sediment derived from crystal precipitates from the contained fluid; these crystals then settled through the fluid to accumulate as sediment on the bottom of the void with a surface defined by the gravitational horizontal at the time of internal sediment accumulation. Voids that are partly filled this way are termed geopedal and record the horizontal as the flat boundary between finely crystalline internal sediment and overlying coarsely crystalline minerals precipitated *in situ* or as overlying void space. Some large individual voids at outcrop scale may record a number of internal sediment-crystalline precipitate episodes, each apparently the result of a fluid flow – fluid stagnation event. The termination of each successive episode of void-filling may be marked by thin rinds of quartz and sulfide minerals, which appear to have formed as the fluid composition or temperature changed as each mineralization/fluid flow event ended. At stop 7 near Richville we will examine multiple geopedal void-fills were tilted out of the horizontal as they formed within a fault block that was undergoing active tilting during hydrothermal mineralization. Void space not filled with mineral material is common within dolomitized marble, as well as in partially filled fractures and solution openings. This space was filled with fluid when the hydrothermal system was active. These spaces are of interest because the liquid or gas which fills them at depth can be petroleum or natural gas of economic value. Tiny voids that formed at defects on crystal surfaces as the crystals grew via precipitation from the hydrothermal fluid may trap bits of the fluid as further mineral growth occurs. These bits of trapped fluid are primary fluid inclusions and may be accompanied by inclusions that form when a crystal later fractures; the fracture fills with fluid and is healed by mineral solids that precipitate in the crack, thus forming secondary fluid inclusions. The fluid inclusions in the Richville shear zone hydrothermal system are discussed below.

Limestones and Dolostones of the Black River Group. We will make one stop at an exposure of the Pamelia Formation of the Black River Group which consists of blue-gray weathering calcite limestones and buff-brown weathering dolostones. The color distinction is very similar to that seen in dolomitized versus undolomitized marbles noted above. The Pamelia Formation outcrop also contains thin (mm to cm thick) mineralized veins that host coarse calcite and dolomite spar, pyrite and barite. These veins provide evidence for movement of hydrothermal fluids into the Ordovician carbonate sequence and reflect on a small scale the effects of structural deformation coeval with fluid migration. A key question for explorationists is the set of geological parameters that led to large-scale development of HTD in Black River and Trenton Group carbonates in the subsurface during hydrocarbon migration.

Mineral systems and geochemistry of the mineralizing fluids

Characterization of the fluid chemistry of the mineralizing system that operated in the vicinity of the Richville Shear Zone must take into account the range of minerals and the various dissolution/reprecipitation processes. The fluid must have varied at least subtly in composition and temperature through the mineralization history because some minerals are at times dissolved and at other times precipitated (e.g. calcite). The system must also have been cyclically variable in that void fills show sequences of alternating internal sediment and passive crystal growth, and chemically precipitated grains within hydrothermal pipe fill breccias show alternating oxidized iron phases (hematite) and reduced iron phases (siderite).

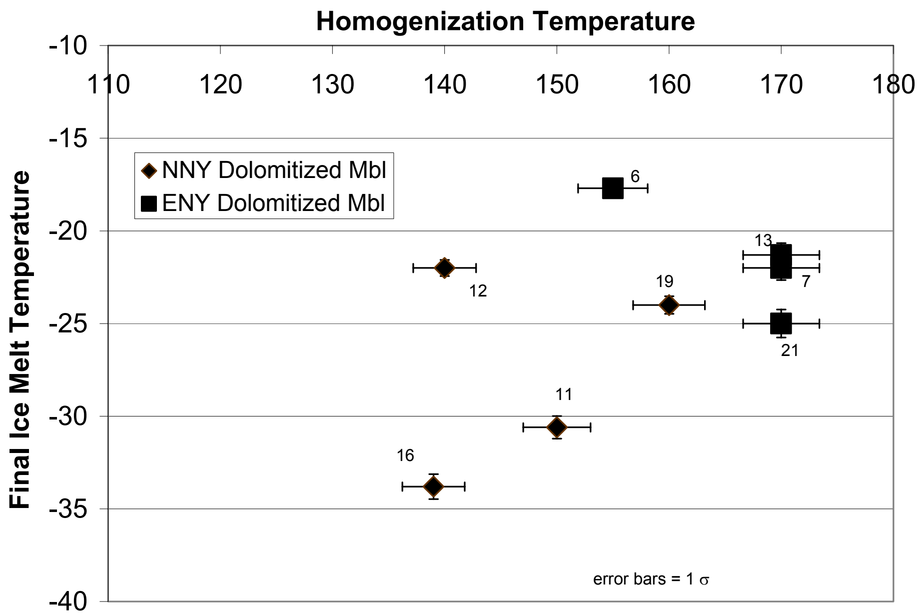


Figure 3 – Fluid inclusion data. NNY Dolomitized Mbl data include analyses of dolomite spar from stops 6 and 7. ENY Dolomitized Mbl data is from dolomite spar near Putnam Center and Fort Ann, New York. The eastern New York dolomitized marbles are similar in origin to the St. Lawrence Lowlands examples seen on this trip and the modern outcrops are near normal faults with adjacent Potsdam Sandstone. The homogenization temperatures provide an estimate of the minimum temperature of mineral crystallization. The final ice melt temperature is controlled by the salinity of the aqueous fluid in the inclusion. The numbers above each point indicate the number of individual inclusions analyzed.

Fluid inclusions. Fluid inclusion analyses were carried out on dolomite spar from void and fracture fill in marbles. As shown in figure 3, fluid inclusion homogenization temperatures range from 135^{o}C to 175^{o}C, with the eastern New York samples recording modestly higher temperatures. The final ice melt temperatures range from ~ -17 to -37 ^{o}C indicating a range of salinities from ~12 % to ~27% NaCl equivalent. These salinities are 3 to 9 times higher than the dissolved solids content of modern seawater (3.5%) and reflect derivation of the fluid from an evaporite brine, seawater or meteoric water that had received significant solute input. The salinity of the fluids in the Adirondack Lowlands marble belt dolomite is slightly higher, on average, than the Eastern Adirondack Highlands samples.

The fluid inclusion homogenization temperatures in the Richville Shear Zone hydrothermal system are beyond the typical range of the liquid petroleum window (80-120°C) and suggest that any hydrocarbons would have been present as methane or as high-molecular weight solids. Methane gas inclusions are common in quartz crystals from quartz +calcite veins that crosscut some earlier hydrothermal mineralization features. The methane-bearing inclusions are apparently primary and suggest that methane was being generated within the hydrothermal fluid system. Overall, these temperature and salinity determinations are consistent with results from other hydrothermal dolomite systems (e.g. Hulen et al. 1990; Luczaj 2001; Nesbitt et al. 1996; Qing and Mountjoy 1994; Smith et al. 2003).

Stable Isotope Data. $\delta^{13}C$ and $\delta^{18}O$ stable isotope data from unaltered Proterozoic marble, dolomitized marble, dolomite spar and calcite spar presented in figure 4. The unaltered Proterozoic marble samples are interpreted as originally sedimentary limestones whose stable isotope values are inherited from their deposition as marine carbonate sediments, and by equilibration with organic carbon as graphite during metamorphism. In general, the fields defined by the hydrothermal products such as dolomitized marble, dolomite spar and calcite spar

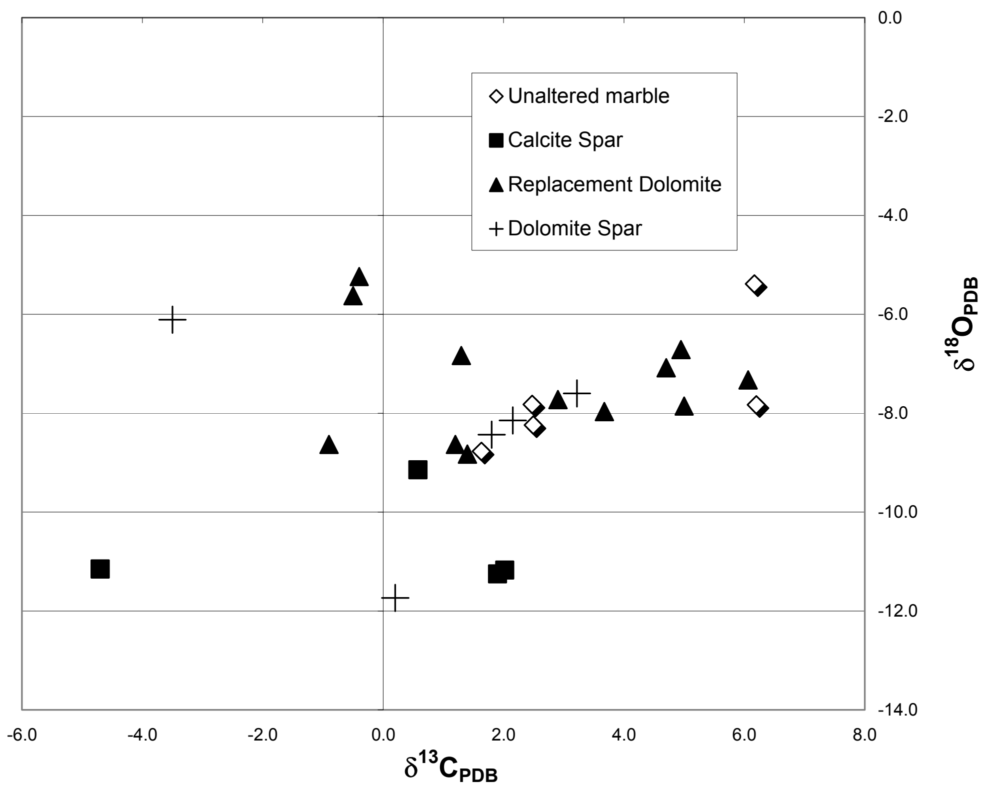


Figure 4 - Isotopic data from the Richville Shear Zone dolomitized marbles. Data from stops 1, 4 and 6 is shown.

overlap with unaltered marble. The most straightforward interpretation of this data is that the stable isotopes of the hydrothermal system are buffered by the marble. When waters interact with rock, isotopic exchange between the water and rock occurs as minerals dissolve into and precipitate from the fluid. Given time the system will reach isotopic equilibrium determined by temperature and fractionation factors of the mineral/fluid combinations. If the mass of water is much greater than the mass of minerals (a high water-rock ratio), such as in the case of carbonate minerals precipitating in the open ocean, the minerals isotope composition will reflect that of the water and the temperature at which the minerals precipitate. If the amount of rock is much greater than the amount of water (a low water-rock ratio) the stable isotope value of the water will be dominated by input from the rock, and the stable isotope signature of minerals which precipitate from that water will be controlled by rock system.

One way to assess the degree of rock-water interaction is to calculate the isotopic composition of water that was in equilibrium with the precipitating minerals at the temperature of mineral precipitation. In this case we can use the fluid inclusion homogenization temperatures in dolomite, recognizing that these represent minimum temperatures of mineral precipitation since the true trapping temperature may be somewhat higher than the homogenization temperature. As shown in figure 5 the water in equilibrium with hydrothermal dolomite spar in the Richville Shear Zone and eastern Adirondack Highlands marble

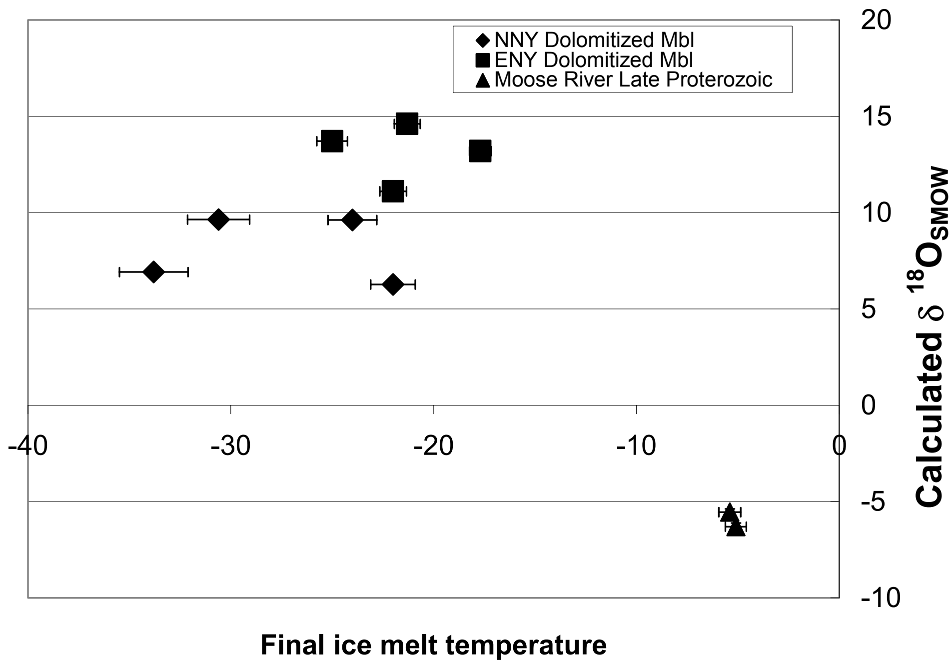


Figure 5 - Calculated oxygen isotope signature of water in equilibrium with dolomite and calcite. Final ice melt temperature is a proxy for salinity; less saline fluid from Moose River vein is much lighter isotopically and is interpreted as meteoric (freshwater). Fluids in equilibrium with dolomitized Proterozoic marble were much heavier isotopically.

had $\delta^{18}O_{VSMOW}$ ($_{VSMOW}$ is the standard used for water isotopic composition) values in the +8 to +12 range. This water is enriched in heavy oxygen relative to seawater, which has a $\delta^{18}O$ $_{VSMOW}$ of around 0. The hydrothermal waters are also enriched relative to meteoric waters, which have negative $\delta^{18}O$ values since they contain relatively more light isotopic oxygen than seawater. An example of a meteoric hydrothermal fluid is plotted on Figure 5. Moose River calcite+quartz vein data from Selleck et al. (2004) and shows $\delta^{18}O$ water values calculated from quartz and calcite stable isotope systems. The positive or enriched values of calculated hydrothermal waters seen in the RSV are interpreted as resulting from a low water-rock ratio because as a little water reacts with lots of hot rock, the oxygen isotope values of the water become more positive. We use the term ‘evolved’ to refer to water whose isotopic composition has been changed by interaction with rock. Another interpretation consistent with this data is that the hydrothermal waters were derived from evaporite brine that made its way downward into the basement rock. Evaporation of seawater in an enclosed basin can produce water with a heavy isotopic signature as the isotopically light water evaporates to vapor preferentially. Many samples of water from deep wells in the Canadian Shield have this heavy isotopic signature and are very salty, like the fluids in the RSZ system, and some workers interpret this water as evaporite brine that descended from saline basins that once overlaid the Shield. Equally plausible is that this brine results from long periods of water-rock interaction during which the water picks up high levels of solute and is progressively modified isotopically to its 'heavy' state. In the RSZ hydrothermal system it is not difficult to visualize the extent of rock- water interaction as small amounts of water were squeezed through finely crushed rock within fault gouge and filtered through tiny intergranular cracks and fissures in marble. The pattern of dolomitization of marble in the vicinity of fractures illustrates that significant rock-water interaction occurred. Along those flow paths the water would have had time to exchange isotopically to produce heavier signatures, and pick up high concentrations of dissolved constituents. The high chloride content could have been derived from evaporite brine or from minerals in the enclosing rocks, including halite, that contain chloride. The widespread shield brines discussed above similarly have high salinity with chloride the dominant anion.

A model for petroleum and gas exploration

Lithological, mineralogical, structural and geochemical aspects of Richville Shear Zone system are best explained as resulting from coeval faulting, hydrothermal karsting and hydrothermal mineralization. As discussed below, the system is an example of a fault zone that could have generated hydrothermal dolomite reservoirs resembling those that are important in the Appalachian Basin (Lavoie and Chi 2003; Martel and Durling 2003; Smith et al. 2003), the western Canada Basin (Berger and Davies 1999; Nadjiwon et al. 2000; Qing and Mountjoy 1994) and elsewhere (Waddell 1996; Westphal et al. 2004). Because it is now unroofed and at the surface, the RSZ hydrothermal system affords an unusual opportunity to examine the scale and style of deformation and hydrothermal alteration and mineralization at the interface between basement and sedimentary cover. The physical dimensions of the zone and its structural features provide a model of the subsurface character of hydrothermal dolomite reservoirs and potential keys to identification of these from seismic and subsurface geological data. The following section outlines elements of the system that are critical for development of km-scale hydrothermal alteration coupled with transcurrent, wrench-fault tectonics.

Fault/seismic pump dynamics. The occurrence of vigorous hydrothermal mineralization within a shear zone, the variation in hydrothermal fluid chemistry from oxidizing to reducing, the precipitation of layered void fills adjacent to fault zones, and evidence of high and/or fluctuating fluid pressures all point to a system evocative of the seismic pumping mechanisms described by (Sibson et al. 1975). In this scenario, motion along an irregular fault zone produces regions of extension/dilation that locally induce dramatically lowered fluid pressures causing surrounding fluids to flow into potential void space. These fluids are then expressed from the extensional zones as fault irregularities close to form compressional domains. The expressed fluids have interacted with crushed rock within the fault zones and thus contain solutes obtained from mineral alteration reactions. Fluids exiting from compressional zones move upward and laterally under high fluid pressure to alter and mineralize surrounding rock. Expressed fluids might also carry heat from the deeper part of the fault system, and conceivably could find permeable conduits that would permit outflow at the surface as hot mineral waters, as seen in many modern fault zones. The seismic pumping model is particularly attractive for the Richville Shear Zone system in that most of its critical features are explicable within this framework. The seismic pumping model may also apply to many hydrothermal dolomite reservoir systems, and indeed more widely to a number of diagenetic phenomena including color-banding of iron oxide cements in the Potsdam Sandstone (Selleck 1978).

Interaction of fluids with labile Mg-bearing minerals. The widespread dolomitization of marble in the Richville Shear Zone and indeed the dolomitization of overlying carbonate strata of the Theresa Formation and Ogdensburg Dolomite of the Ottawa-St. Lawrence Lowlands require significant importation of dissolved magnesium in the circulating pore fluids. While seawater itself can be a source of Mg for dolomitization, hydrothermal fluids not immediately connected to the seafloor may require another source of Mg. In the Richville Shear Zone system the interaction of hot salty fluids with Mg-bearing silicate minerals (e.g. biotite, hornblende, diopside, garnet, etc) in the Grenville metamorphic rocks commonly resulted in the conversion of these minerals to low-temperature phyllosilicates such as illite, kaolinite and Fe-chlorite which are Mg-poor (Tiller and Selleck 1992). The Mg liberated from this alteration process, likely taking place most effectively within the crushed rock of the fault zone as fluid was emplaced during dilational events, would then be transported into adjacent and overlying rock. Where this Mg-rich fluid interacted with calcite marble, dolomitization resulted. This fluid was also capable of directly precipitating dolomite spar and dolomite internal sediment, perhaps as fluids injected upward from below were cooled during ascent into cooler rocks. The presence labile Mg-bearing minerals interacting with circulating fluids may be a key factor in the development of hydrothermal dolomite reservoirs in deep-seated basement-cover settings.

Availability of basal sand aquifer. The Potsdam Sandstone and other basal Paleozoic equivalents served as a regional fluid aquifer throughout the Paleozoic and later history of the Appalachian Basin. In the Richville Shear Zone system, this sand served locally as part of the fluid transport network and its presence was likely critical in maintaining a fluid supply to the system. In New York State Potsdam outcrop and subcrop is absent over a region extending from the western Adirondack margin west to the Lake Ontario basin and south into the northern Finger Lakes region. This 'Potsdam absent" region is well-known in gas and oil exploration when deeper targets are explored. The link between the Potsdam basal aquifer and dolomitization in the Richville Shear Zone suggests that hydrothermal dolomite reservoirs might be not be as prevalent in regions where the basal sandstone is absent.

Appropriate burial temperatures. The dolomitization and related mineralization in the RSZ occurred at temperatures generally in the range of 140-170^{o}C. The depth of mineralization is not well-constrained because of the possibility that hot fluids from depth were rapidly expressed upward into cooler rock in a seismically pumped hydrothermal system. These temperatures lie beyond the normal range of liquid hydrocarbon generation, but were within the range of methanogenesis. The shallower, now eroded portions of the section overlying the RSZ may have contained petroleum liquids at one time. A critical feature of the RSV system was the deep circulation of fluids at temperatures sufficient to rapidly alter Mg-bearing minerals and then transport solutes and heat into adjacent rock.

Evolving hydrocarbon system. The evidence for active hydrocarbon generation in the RSZ fluid system is limited to the occurrence of methane fluid inclusions in quartz and solid organic matter inclusions in dolomite and calcite spar. The dissolution of calcite marble that preceded mineralization events was likely due to CO_2 and organic acids produced during decarboxylation of organics as kerogen compounds were converted to saturated hydrocarbons. These acidic components would then be available to promote calcite dissolution and alteration of labile silicate minerals. This process is widely recognized as a mechanism for evolution of secondary porosity in deeper carbonate and siliciclastic reservoirs, and its inferred presence in the RSZ suggests that hydrocarbons were generated in the immediately overlying strata and that fluids from these penetrated downward into the RSV during dilational seismic events. The interrelationship between hydrocarbon maturation and hydrothermal fluid movement to produce HTD reservoirs is a key factor for explorationists to consider.

Timing and regional tectonic perspective. The timing of hydrothermal activity in the Richville Shear Zone system is important because it may be tied to the development of HTD reservoirs elsewhere in the northern Appalachian Basin including a number of current exploration targets. In addition, the understanding of regional patterns and evolution of fault and fracture systems in the basement of eastern North America has implications for the tectonic evolution of the continent. Based upon a number of studies of dolomitization and fluid alteration of the Cambrian and Ordovician strata of and eastern New York and adjacent Ontario we know that the fluid systems of the RSZ type are found in other areas, and may signal widespread development of similar basement fault/hydrothermal alteration systems during the Paleozoic history of the region. The timing of these fluid alteration events is not well understood and few radiometric dates have been determined within the RSZ proper; however some studies of minerals in the region that can be radiometrically dated provide constraints. One source of potentially useful dates derive from Ar-Ar analyses of authigenic K-feldspar in altered basement and overlying Paleozoic cover from northern New York and southeastern Ontario. In general, these dates suggest a fluid alteration system linked to the medial-late Ordovician Taconic Orogeny (Ziegler and Longstaffe 2000b). If the RSV system was active during medial to late Ordovician, it would suggest that the fault system developed after the otherwise N-NE trending normal faults that were related to foreland basin subsidence elsewhere in northern New York. If the RSV system developed in response to late Taconian compressional tectonism, the burial depth of the system may have been no more than one km, based on regional isopach patterns. It is easier to explain the co-incidence of oxidizing fluid signatures with higher-temperature features if the system was recharged from near-surface oxidizing fluids that were heated at depth and then injected into the shallower parts of the regime.

Relatively younger ages of burial mineralization are indicated by K-Ar (Tiller and Selleck 1992) and Rb-Sr (Reynolds and Thomson 1993) geochronological studies of illite in the Potsdam Sandstone in the western St. Lawrence Lowlands . These dates indicate a fluid alteration event in late Devonian-early Carboniferous time (ca. 365-355 Ma). This likely represented a time of maximum burial beneath sedimentary cover and was co-incident with compressional tectonism along the eastern margin of North America. Deeper burial alteration is consistent with the temperatures determined from fluid inclusion work as temperatures of 170^{o}C would imply a burial depth of 4 km if a geothermal gradient of 40^{o}C/km existed at that time. However, as noted above, the temperatures recorded by fluid inclusions may reflect transient conditions related to upward seismic pumping of hot brines into shallower, cooler rock and thus temperatures in this system may be a wholly unreliable guide to burial depth. Alleghanian faulting during the late Carboniferous/Permian is also possible but no mineral ages of latest Paleozoic age have been determined in the region.

Related phenomena

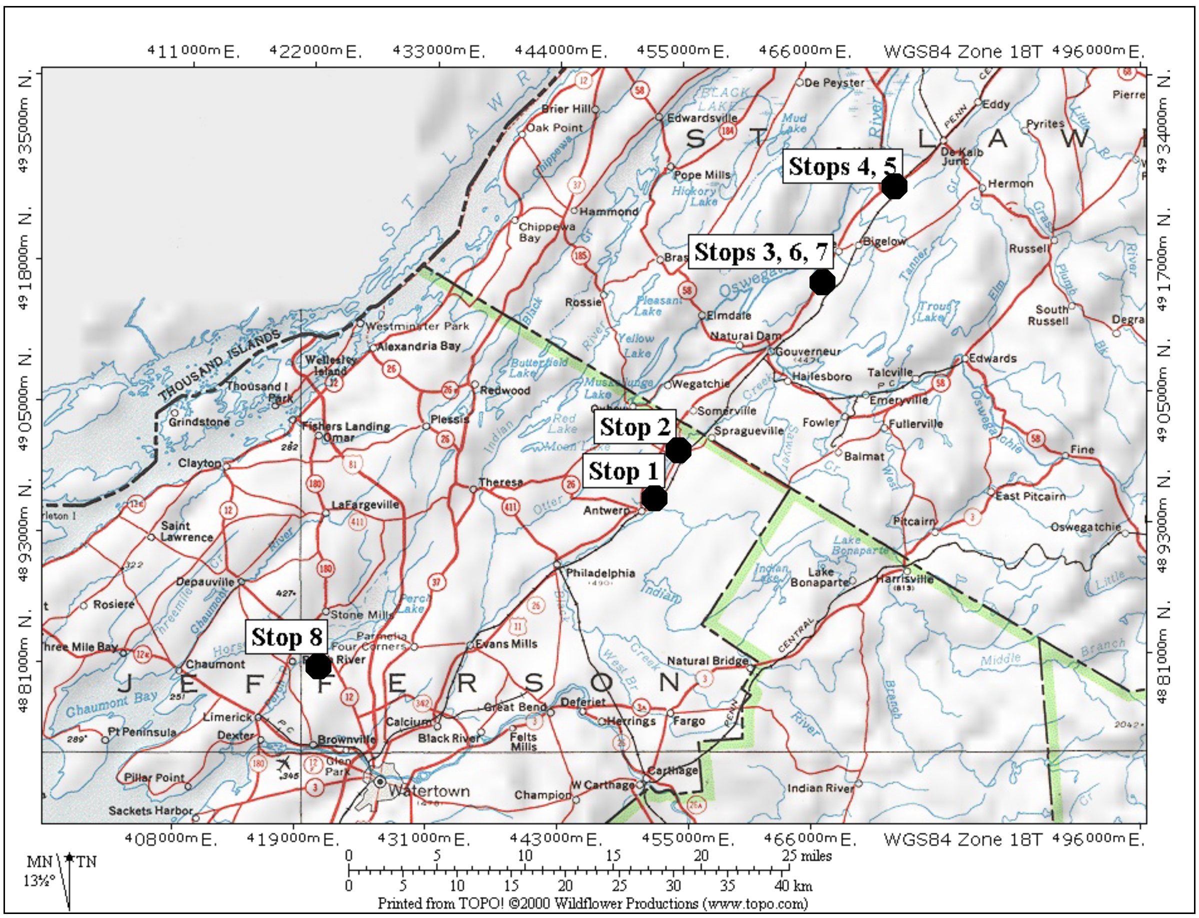
Rossie-type galena-calcite veins. The well-known Rossie-type galena-sphalerite-calcite veins, exposed 15-20 km WNW of the Richville area, were emplaced after the Potsdam Sandstone had been tightly cemented, based on the sharp wall-rock vein contacts and the absence of infiltration of vein material into the adjacent sandstone. Fluids inclusion data indicate homogenization temperatures of 120-140^{o}C (Foley et al. 1985) These veins have a NW-SE orientation and appear unrelated to the stress system of the RSZ, although further study is needed.

Hematite deposits of the Adirondack Lowlands. A discontinuous belt of hematite deposits extends from WSW to ENE across the Adirondack Lowlands from the vicinity of Antwerp to Russell, NY. A small number of these deposits were worked as open pit mines for locally consumed hematite ore for small forges and paint pigment in the 19th century, and were investigated in the 1920's and 1930's for possible development. The complex mineralogy of the Sterling deposit has been described by Robinson and Chamberlain (Chamberlain 1984; Robinson and Chamberlain 1984). In general, these deposits lie along the same trend as a suite of sulfide ore (largely pyrite) deposits that are part of the metasedimentary sequences of the St. Lawrence Lowlands. The trend of the hematite deposits is also generally parallel to the Richville Shear zone system although most of the hematite deposits lie somewhat to the southeast of the zone. The hematite deposits also share the common presence of nearby or directly overlying inliers of Potsdam Sandstone (Chamberlain 1984). The paragenesis of the hematite deposits is generally interpreted as multi-stage with pre-Potsdam surface weathering of Proterozoic iron sulfide leading to accumulation of locally thick gossans of limonite/hematite prior to Potsdam Sandstone deposition (Chamberlain 1984). Post-Potsdam reconstitution of the iron oxides involved hydrothermal fluids that dissolved and re-deposited hematite within Potsdam Sandstone as thick botryoidal masses, specular crystalline aggregates, veins and disseminated cements in sandstone and highly altered Proterozoic basement gneiss. In addition, a wide variety of oxide, sulfide, sulfate, phosphate and carbonate minerals were deposited, leading to a rich assemblage of hydrothermal origin (Chamberlain 1984; Robinson 1998; Robinson and Chamberlain 1984). Fluid inclusion temperatures from calcite indicate a minimum temperature of 140^{o}C for some of the mineralizing fluids (Robinson and Chamberlain 1984), and the paragenetic sequence of the complex mineralogy suggests that geochemical conditions varied from oxidizing to reducing during active mineralization. The mineral systems at Sterling mine include sparry dolomite and coarse calcite spar as major void-filling phases, and outcrops of marble and Proterozoic granitoid rocks immediately adjacent to the mine contain fracture-related dolomitization and sandstone injection 'dikes'. The similarity in between the Sterling-type mineralizing system and the hydrothermal mineralization in the Richville Shear zone is striking, and suggests that the fluids involved were similar. Further study of fluid inclusions and stable isotopes in the hematite deposit mineral system and dating of mineral assemblages are needed to resolve the relationships.

Sandstone, quartzite and 'enigmatic enclaves'. Many geologists who have worked in the Adirondack Lowlands have called attention to quartzite and sandstone outcrops that are separated from the main mass of Potsdam Sandstone and lie within areas mapped as marble. One interpretation suggested that some sandstone occurrences are the result of 'protection' of sandstone from penetrative deformation during Grenvillian tectonism by the presence of ductile marble (Bloomer 1965), thus making the sandstones part of the original Mesoproterozoic sedimentary sequence. This interpretation has been vigorously challenged (Brown 1967) and a more viable interpretation suggests emplacement of Potsdam sands into karst caverns along the pre-Potsdam unconformity with later diagenetic burial alteration. However, the occurrence of strained quartzite clasts within breccias, and authigenic minerals that have been interpreted as metamorphic in origin have caused some workers to call upon a more complex origin and have left some aspects of these occurrences unresolved (Bursnall and Elberty 1993). The coupled hydrothermal/fault system paradigm provides a reasonable explanation for these unusual features and is consistent with both field and laboratory data.

ACKNOWLEDGEMENTS

William Peck and James McLelland graciously agreed to act as readers for this manuscript.



ROOT ZONE OF AN ANCIENT FAULT SYSTEM – ROAD LOG

This road log begins at the traffic light on Route 11 in Antwerp, NY. To reach Antwerp from Oswego, NY, take Route 104 east to Route I81; I81 north to the Route 11 Exit north of Watertown; Route 11 East/North toward Gouverneur. This drive will take approximately 90 minutes from Oswego. Mileage begins at the traffic light intersection in Antwerp. Note that the total round trip from Oswego is approximately 200 miles.

Cum Mi.	Road Log (UTM coordinate datum NAD83)
0.0	Traffic light at intersection on US Route 11, Village of Antwerp, New York; continue northeast on Route 11
3.1	Stop 1 (18 452455E, 4898768N) – Outcrop on S side of Route 11 exposes "Antwerp" granite syenite intruding marble, which was followed by dynamothermal metamorphism prodcuing upper amphibolite facies calcsilicate mineral assemblages, reaction boundaries between granitoid and marble, and deformation including boudinage, folding and fabric development. The high ductility contrast between the granitoid and marble results in disruption of original depositional and intrusive relationships. Paleozoic dolomitization of marble is evident along near vertical fractures and as diffuse zones extending away from fractures. An open fracture filled with dolomite-cemented sandstone is found on the top surface of the E end of outcrop. The old workings of the abandoned Sterling Mine, part of belt of hematite deposits found in this region (see text) are located on private property approximately 1 km SW of Stop 1. (20 minutes) The exposures on the N side of Route 11 contain brecciated and dolomitized marble, coarse dolomite and calcite spar and minor sulfide and quartz mineralization. Some calcite spar crystals are cored by deep brown or black calcite that contains abundant organic (petroleum) inclusions. The arrangement of dolomitized marble and void-filling minerals suggests multiple episodes of collapse brecciation and mineralization. Depending on the size of the group, we may visit the locality on the N side of the road when we return via same route.
4.7	Stop 2 (18 453868E, 4900891N) Low roadcuts on the SE side of Rt. 11 expose Proterozoic marble with cm- to dm-scale compositional bands defined by variations in abundance of graphite and calcsilicate minerals such as phlogopite and diopside. The banding dips gently to the SE. Vertical dolomitized zones are associated with joints that served, apparently, as conduits for dolomitizing fluids. The dolomitized zones are accentuated by minor weathering which imparts a brownish-tan or buff color to the dolomite in contrast to the white-blue-gray of the calcite marble. The contact between dolomitized and undolomitized marble represents a fluid alteration 'front' and minor hematite staining of partially dolomitized marble is present along this front. Note also the development of minor, millimeter-scale porosity in the dolomitized marble. The porosity is caused by the net reduction in solid volume during the $2CaCO_3 + Mg^{2+} > Ca,Mg(CO_3)_2 + Ca^{2+}$ reaction. Why are dolomitized limestones often good hydrocarbon reservoirs? Could a dolomitized marble like this serve as a reservoir rock? Continue NE on Rt. 11 through the village of Gouverneur
18.8	Stop 3 (18 468148E, 4917434N) Roadcut on NYS Rt. 11 immediately NE of intersection with Richville Road exposes a fault contact between red-orange Potsdam Sandstone and Proterozoic basement. The Potsdam Sandstone here contains pebbles and small cobbles of quartzite, jasperite and cemented sandstone. Slickenlines on the E-W trending fault surface indicate reverse motion with Proterozic marble moving up to the west over Cambrian Potsdam Sandstone. Note the dolomitization of calcite marble adjacent to fractures in basement rocks adjacent to the fault. The marble and calcsilicate rocks near the fault give way to the east along the roadcut to coarsely crystalline quartz-albite pegmatite. The exposures on the NW side of Rt. 11 will be examined as Stop 6 when we return via this route. Continue NE on Rt. 11
24.4	Stop 4 (18 473726E, 4923995N) The exposure on the SE side of Rt. 11 consists of near-horizontal beds of Potsdam Sandstone. Depending on the condition of the road ditch at the NE end of the outcrop, a contact with Proterozoic marble is exposed. The basal Potsdam is locally a conglomerate,

	and near the base of the exposure cobbles of quartzite and jasperite are common. The upper portion of the exposure is medium sandstone. Note the irregular, disrupted lamination and cross-cutting color banding in the Potsdam. We will examine the exposures on the NW side of Rt. 11 as Stop 5.
26.0	Continue NE on Route 11 to parking area opposite Hermon-DeKalb Central School. This is our turn-around. Head SW on Rt. 11
27.6	Stop 5 (18 473726E, 4923995N) A series of outcrops on the NW side of Rt. 11 expose complex fault and hydrothermal solution collapse contacts between marble and Potsdam Sandstone. Proceeding from the NE end of the exposures, the first outcrop is comprised of intricately folded gray marble with a low-angle lineation marked by thin dolomite segregations. The folding of the marble is most likely the result of Proterozoic dynamothermal metamorphism but the dolomitic fabric may have resulting from strain during Paleozoic wrench faulting. Diffuse dolomitization of the marble, localized dolomitic halos around veins and dolomite + calcite void fills are also present. Continuing to the SW along the road, a fault contact between marble and a hydrothermal/fault breccia is poorly exposed near the end of the main marble outcrop. The breccia and adjacent deep red-brown sandstone here contains Potsdam sand grains and clasts of cemented sandstone, clasts of apatite-dolomite-quartz that represent rinds detached from hydrothermal pipe walls, spheroidal grains consisting of concentrically alternating hematite and siderite, and fragments of calcsilicate minerals including ‘bleached’ phlogopite; xenotime, tourmaline and monazite grains in the breccia have authigenic overgrowths. Complexly slumped or faulted sandstone adjacent to the contact is cemented by barite, pyrite and quartz. Mm to cm-scale arborescent masses of the Fe-sulfate jarosite (found by Opportunity on Mars) are present on rain-protected surfaces where pyrite is weathering. The outcrop on the SW end of the series of exposures consists of less disturbed and altered Potsdam Sandstone, resembling the lithology exposed across Rt.11 visited as stop 4. Continue SW on Rt. 11.
33.2	Stop 6 (18 468148E, 4917434N) (Opposite stop 3). The lower portion of the NE outcrop at this stop consists of a hematite-stained fault gouge that grades progressively upward into more strongly layered and folded quartzite. The fault gouge consists of cm to meter scale blocks of quartz cemented sandstone and quartzite and deformed clasts of hematite-chlorite quartzite. The cleaved matrix consists of hematite-chlorite-illite quartz. Near the top of the exposure, the quartzite contains relatively unstrained layers of quartz-cemented sandstone with rounded grains. This outcrop is a relatively thick fault gouge zone associated with the Richville Shear Zone. The relatively resistant capping quartzite has prevented erosion and thus the weak fault gouge rock is exposed at the surface. Note the flattened, boudinaged quartzite fragments and the intense hematite staining. When the fault zone was active, material of this sort was flushed by reactive hydrothermal fluids. The second major outcrop at this stop is directly opposite the fault exposure seen at Stop 3 across Rt. 11. The outcrop is somewhat unstable so use caution. Note the diffuse dolomitization of marble, large voids with dolomite, calcite and minor quartz and pyrite. Barite occurs in some void fills. Sparry dolomite and calcite alternate in some void fills, and tilted void floors apparently represent progressive rotation of voids as faulting and hydrothermal dissolution and mineralization occurred. Note the veins that crosscut former calcsilicate pods in altered marble and the residue of illite/chlorite and kaolinite which is left as a soft material on the void floors. Hematite staining and dolomitization is localized along fractures and become more pervasive as the fault surface is approached at the SW end of the exposure. The next stop is located on Rt. 11 approximately 0.2 mi SW.
33.4	Stop 7 (18 468003E, 4916963N) These exposures offer an excellent opportunity to examine marble penetrated by hydrothermal pipes and tunnels, the material filling these voids, and patterns of dolomitization and hematite staining of the marble altered by hydrothermal fluids. Note that some pipes (near-vertical orientation) represent solution-enlarged fractures, indicated by the lack of ‘fit’ of opposing walls. The pipe filling is typically Potsdam sand tightly cemented by hematite and quartz, with minor apatite. Barite, leucoxene, siderite and pyrite also occur as cements. Note that some pipes are filled with red-brown sandstone at the base but the upper portion of the pipe contains sparry

	dolomite, calcite and quartz as mineral precipitates. Tunnels (horizontal orientation) show a similar pattern of fill, and some show multiple sandstone-dolomite-quartz-calcite internal layering, representing multiple episodes of sluicing of unconsolidated sand followed by mineral precipitation. Some larger pipes and tunnels contain fragments of cemented sandstone and quartz-apatite-dolomite clasts interpreted as mineralized rinds spalled from the walls pipes and tunnels elsewhere in the hydrothermal system and carried by rapidly flowing hydrothermal fluids. The outcrop also displays a range of dolomitization patterns associated with the pipe and tunnel network and fractures. Note the 'islands' of undolomitized marble, and relict calcite 'eyes' in otherwise dolomitized rock. How can we reconcile the presence of hematite associated with reduced iron minerals, sometimes as concentrically alternating spheroidal grains? What is the origin of the pervasive hematite staining associated with the dolomitization of marble? End of road log for Stops 1-7 Continue on Rt. 11 SW through Gouverneur, Antwerp and Philadelphia to intersection of Rt. 342 and Rt. 11. Turn NW (right) onto Rt. 342, continue west to Rt. 12, crossing over I81. Turn right (NW) onto Route 12. Proceed NW to Stop 8, located 4.0 miles from the Rt. 12/Rt. 342 intersection. Stop 8 (18 422107E, 4881242N) The outcrop on the NE side of Rt. 12 is in the upper portion of the Pamelia Formation of the Black River Group. The lithologies here include dolostone, dolomitic shale, lime mudstone, and wackestone. In general, these sedimentary facies represent deposition on a broad, relatively low-energy carbonate platform of Medial Ordovician age. This platform covered much of the interior of eastern Laurentia (North America) prior to the onset of subsidence and foreland basin development in later Medial Ordovician time. Key sedimentary features to observe are cryptalgal/microbial laminations, bentonite (volcanic ash) beds, mudcracks, evaporite (gypsum/anhydrite) mineral casts, and storm event beds represented by mud rip-ups clasts, intraclast breccias, stylolites and solution voids. The SE end of the outcrop is capped by a massive, tan-weathering laminated dolostone. Immediately beneath the dolostone, beds of grey limestone are arrayed in a series of low, irregular folds with axes that trend N80W, approximately parallel to the highway. Thin (mm-cm thick) dolomite+calcite+pyrite+barite veins trending N70E are common in the limestone, and cm-scale calcite spar filled voids are present in the dolostone and some limestone beds. The mineralized fractures here are small-scale hydrothermal features that may be related to the dolomitization and related mineralization seen in the earlier stops on the trip. Note that the void fill minerals in the limestone beds are the same as those in the mineralized fractures. Examine the limestone - dolostone contact and see if you can trace mineralized fractures into the dolostone. Is the dolomitization here 'early' (preburial – related to Mg-rich brines derived from seawater that had precipitated gypsum) or is the dolomitization related to the later mineralized fractures? Why is some rock dolomitized whereas other rock remains nearly pure calcite limestone? Was there significant porosity in this rock at any time during its diagenetic history? Is the folding related to the minor mineralization or a later phenomenon? Continue NW on Rt. 12 to the Perch River Refuge parking area (visible from Stop 7), turn back onto Rt. 12 heading SE. Continue on Rt. 12 to I81 south (3.4 miles); take I81 south through Watertown to Rt. 104 exit. Take Rt. 104 west to Oswego.

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Trip A-3

FRACTURES AND FAULTS IN THE EASTERN LAKE ONTARIO BASIN, OSWEGO COUNTY, NEW YORK

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INTRODUCTION

The central Appalachian basin is host to a number of regional joint sets associated with the late Paleozoic Alleghanian orogeny (Engleder, 1979; Engelder and Geiser, 1980; Evans et al., 1989; Zhao and Jacobi, 1997; Younes and Engelder, 1999; Engelder et al., 2001). A number of researchers demonstrated the orthogonal relationship between fracture sets and the trend of broad-open folds in the Paleozoic strata, and concluded that a major set are cross fold joints (Engelder and Geiser, 1980; Engelder, 1985). Cross fold joints are fractures that develop perpendicular to fold axes, and occur as non-parallel sets (Zhao and Jacobi, 1997) in the Appalachian basin of Pennsylvania and New York due to the arcuate shape of the orogen. Generally, the cross fold joints radiate perpendicular to the acuate trend of the orogen. In central New York, the northern extent of the Appalachian basin, steeply dipping fractures that strike generally northwest are interpreted to be cross fold joints (Zhao and Jacobi, 1997). Development of these joints requires a component of orogen parallel tension. In a systematic study of regional joints, Engelder (1982) proposed a tie between specific joint sets and the contemporary stress within the lithosphere. This set of steeply dipping joints strikes generally east-northeast and were documented throughout the Appalachian basin from Ohio to central New York and it was proposed that they are related to the contemporary stress field (Engelder, 1982).

Along the southeastern shoreline of Lake Ontario (Figure 1) there are abundant outcrops of Ordovician Oswego Formation. As well, the eastern rivers within the basin are down-cut into the underlying Pulaski Formation forming excellent bedrock exposures in the Salmon River gorge (Figure 2). Both of these rock

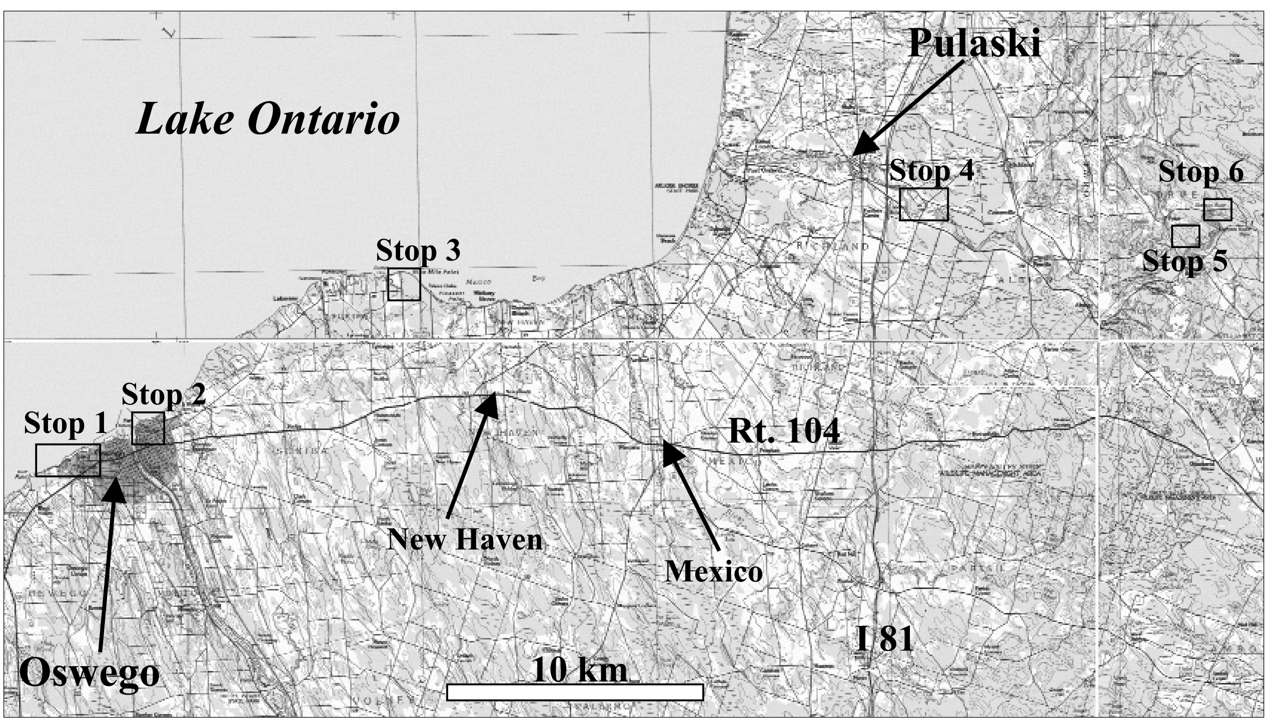


Figure 1 – Map of the eastern Lake Ontario region with stop locations shown.

formations are dominated by clastic sedimentary rocks, are generally flat lying and contain a number of systematic joint sets. A recent study of these joints documented sets with orientations similar to the cross fold joints described above. Additionally, a set of east-northeast striking joints have evidence for sinistral shear including displacement of earlier joints and parallel zones of en-echelon fractures. During this field trip six locations will be visited to show the regional and local fracture sets in the Ordovician Oswego and Pulaski Formations in the eastern Lake Ontario basin, with special emphasis on evidence for sinistral shear.

BEDROCK GEOLOGY OF THE EASTERN LAKE ONTARIO BASIN

The central region of Oswego County, New York is underlain by the Ordovician sedimentary rocks of the Pulaski and Oswego Formations. The Pulaski Formation consists of interlayered gray-red sandstone, siltstone and shale with the amount of shale decreasing upward in the section. The best outcrops occur in the Salmon River gorge in the eastern part of the county. The Pulaski Formation is overlain by the Oswego Formation, named for the excellent bedding-plane exposures in the Oswego River and along the shoreline of Lake Ontario in the city of Oswego, NY. The Oswego Formation is characterized by interlayered thin to thick beds of gray-green sandstone and siltstone containing large cross beds, abundant ripple marks, channel structures, and rare trace fossils (Patchen, 1978). The contact between the Pulaski and Oswego Formations is gradational with the Pulaski Formation containing substantial shale and the Oswego Formation containing only minor shale partings (Fisher and Laird, 1978). Taken together, these formations represent a regressive package of sediments associated with the transition from marine to terrestrial depositional environments and the onset of the Taconic orogeny (Patchen, 1978). The Pulaski and Oswego Formations are generally flat-lying with only minor eastward dip of a few degrees locally. No major faults are shown to break these formations in Oswego County, however, both formations contain extensive joints and minor faults.

Figure 2 – Photograph of the Pulaski and Oswego Formations at the Salmon River Falls, Altmar, New York. Most of the cliff face is the Pulaski Formation with the transition to the Oswego Formation in the uppermost section. The transition is marked by a decrease in shale and the absence of fossils (Patchen, 1978).

BEDROCK JOINT DATA

Joints were studied at outcrops of the Oswego Formation along the Lake Ontario shoreline, in the riverbed of the Oswego River, and in the Pulaski Formation in the Salmon River gorge. Outcrop maps were completed for all the exposures along the lake-shore and in the Oswego River. These maps include the attitude of every major joint (joints that transect the entire exposure) and a systematic sampling of typical discontinuous joints. As well, these maps document joint density for specific sets, the distribution of en-echelon fracture zones, and domains of high-density joints. The joint density was quantified at most outcrops by counting the number of joints that occur over the outcrop distance, measured perpendicular to the joint set, and the joint density was determined for each joint set separately. Where available, joint patterns were interpreted from high-resolution air photographs (Figures 3 & 4). The length and strike of individual joint traces were collected from the photographs. Finally, any kinematic information, such as slicken-sides, en echelon zones, or off-set markers were noted to assess potential displacements associated with shear fractures.

Regional Joints Patterns

In general, two joint dominant joint sets occur in the Oswego and Pulaski Formations in Oswego County. At any given outcrop, just one may be present, both present with either one as the dominant (higher density) set. Figure 5 shows the typical distribution and geometry of joint sets in the Oswego Formation at SUNY-Oswego. The Oswego Formation contains thick beds of sandstone with large cross beds common (Figure 5A), and the Pulaski Formation consists of interbedded shale, siltstone and shale (Figures 2 & 6). Generally, the two joint sets are subvertical and strike NW and ENE. The regularity of the spacing and planar morphology of two joint sets produced a “diamond” shaped pattern in the bedding planes (Figures 5B, 5C & 6C). Few plumose structures were observed on joint surface in the Oswego Formation, however, there are few favorable exposures to view joint surfaces, and many are severely weathered. On the contrary, the exposures of the Pulaski Formation exhibit excellent plumose structures (Figure 7). Some well exposed joint surfaces exhibit irregular jagged, or “stair-shaped” asperities (Figure 5D).

Systematic sampling of hundreds of joint orientations from the Lake Ontario shoreline reveals the consistency of joint attitude in the Oswego area. Rose diagrams were produced to portray the strike of the joints, and the obvious joint populations that are viewed at the outcrop dominate the plots (Figure 8), however, the attitude of the joints varies about 10^{o} to 20^{o} from the Oswego shoreline and in the Oswego River bed. A third population of joints with a generally N-S strike occurs in the rose diagrams, but are less apparent at most outcrops. The trend of joint traces interpreted from high-resolution air photographs (NYS GIS Clearinghouse) taken immediately offshore at SUNY-Oswego show nearly the same pattern that is observed at the outcrop, with the exception that the individual joints are much larger than the average shoreline exposures. These data was collected using a scale-calibrated image in a computer-mapping program (Canvas GIS), where interpreted joint traces were quantified for attitude and length. A map of the interpreted joint traces appears in Figure 4. The interpreted joint traces were sorted according to length and attitude, and the rose diagrams of Figure 9 show the systematic variation. Most very long joints are parallel to the east-northeast striking set.

EVIDENCE FOR SINISTRAL SHEAR

En-echelon Fracture Zones

En-echelon fracture zones have been found in the Oswego and Pulaski Formations. The individual fractures that make up the zones strike about 25^{o} to 30^{o} anticlockwise to the general east-northeast strike of the fracture zone (Figure 10). The boundaries of these en-echelon zones are roughly parallel to the joints that strike east-northeast. In some cases, the trace of an individual fractures terminates by curving into the leading en-echelon fracture in a zone. Figure 10B shows an example of en-echelon fractures where the individual fracture tips are curved and terminate on the adjacent fracture producing an

Figure 3 – High-resolution air photograph of the shoreline of Lake Ontario at the campus of SUNY Oswego. Air photograph data was obtained from the New York State GIS Clearinghouse internet database. The locations of bedrock exposures discussed in the text are shown. The image also shows major underwater fracture sets.

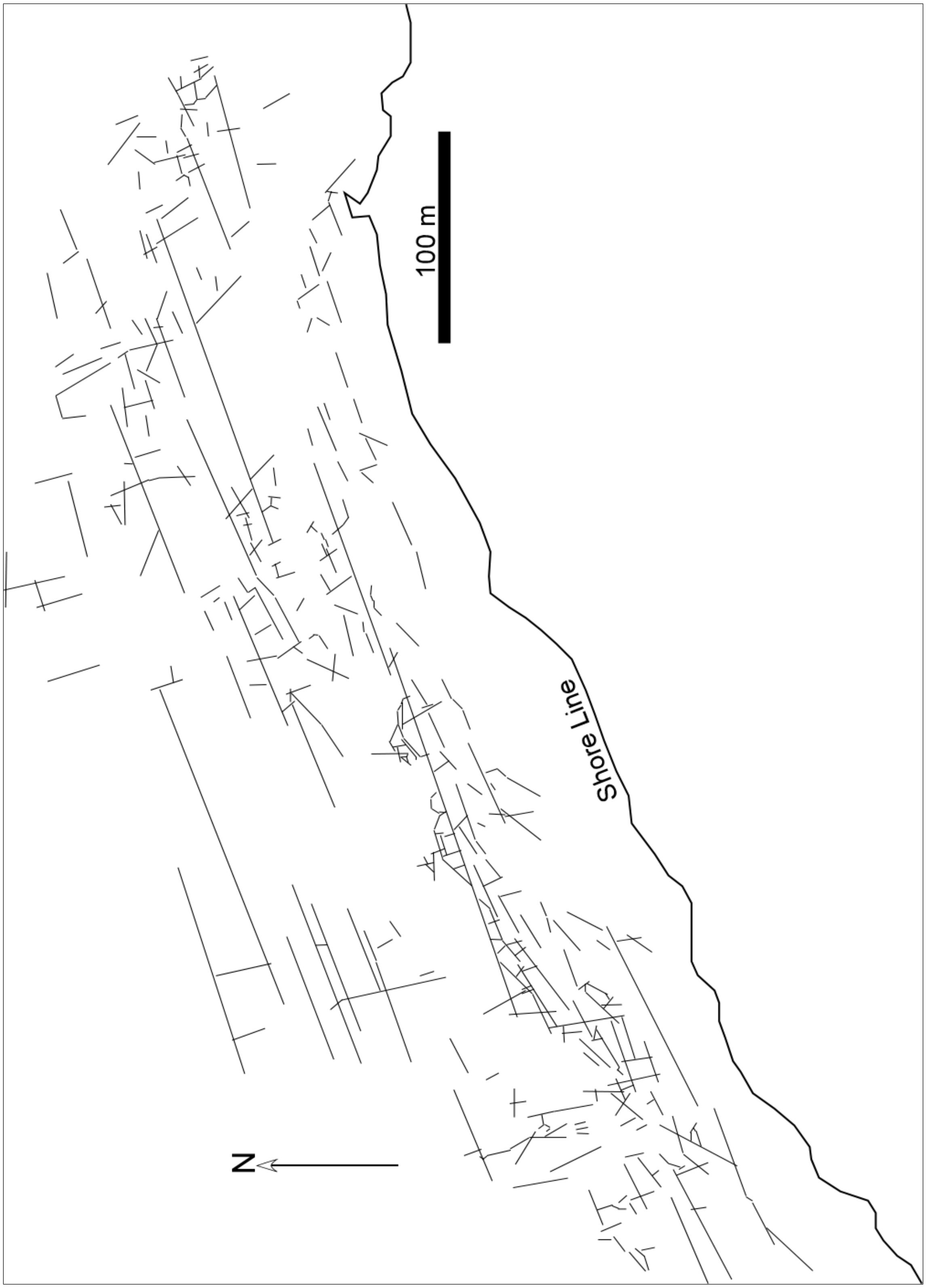


Figure 4 - Fracture trace map interpreted from the air photograph image of Figure 3. Note the continuous fractures that strike east-northeast. They range in length from 10 meters upward to more than 200 meters. These fractures are believed to be the same set that shows sinistral shear in the lake side bedrock exposures.

overall sigmoidal shape for each fracture. The orientation of these en-echelon domains, and the distribution of fractures relative to each other is consistent with a component of left-lateral shear. Often these en-echelon domains are confined to discrete beds in the Oswego Formation, and terminate where the they intersection with the bedding plane. Although the en-echelon fracture zones have been seen throughout the study region, they are best developed in the Oswego Formation along the shoreline of Lake Ontario.

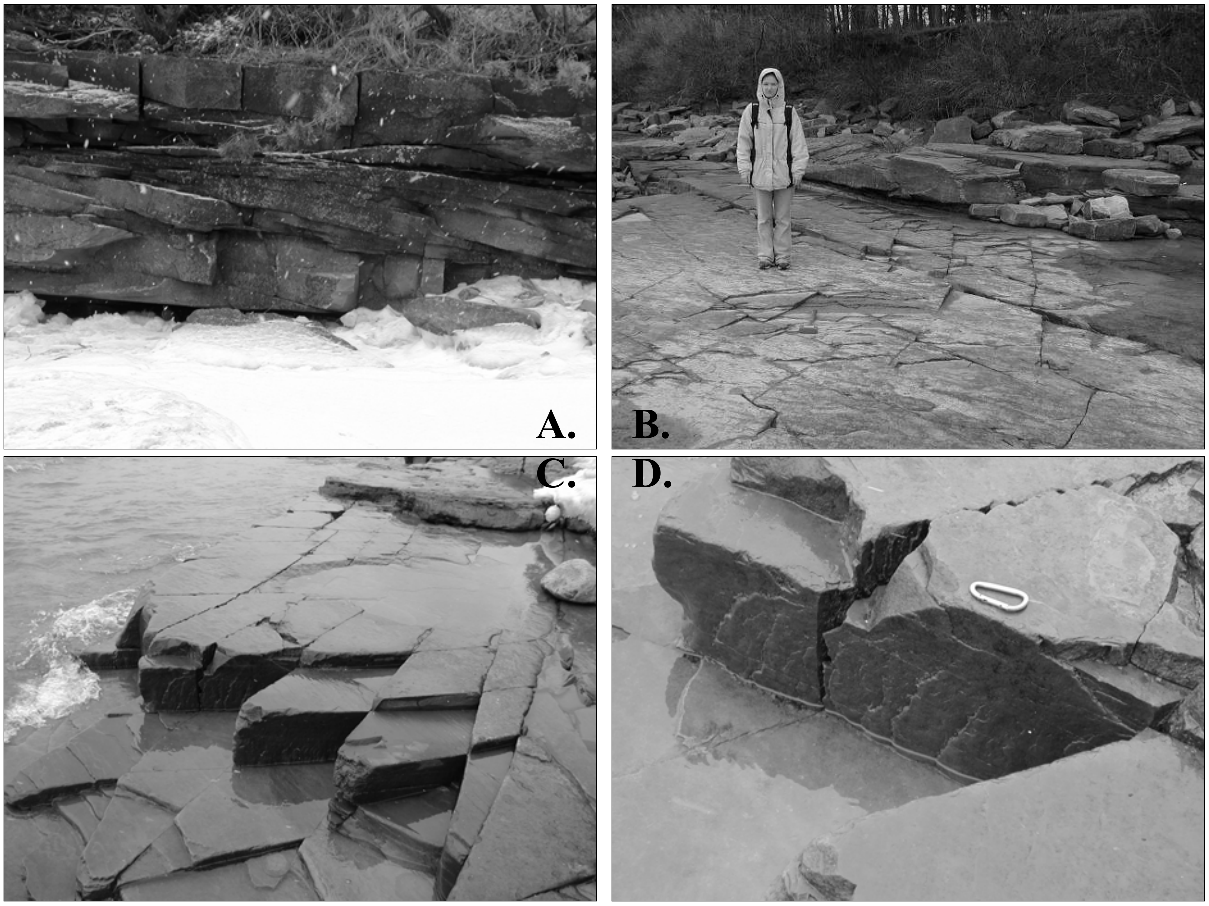


Figure 5 - Outcrop photographs of the Oswego Formation on Lake Ontario. A. Meter-scale crossbeds in sandstone. B. Bedding plane exposure with multiple joint sets. The view is looking southeast. C. Two intersecting joint sets. View is looking easterly. D. Close up view of the NW striking joint showing the irregular joint surface.

Meso-scale Sinistral Faults

Small faults have been documented in the Oswego Formation along the Lake Ontario shoreline. These faults strike about 070^{o} and are subvertical (Figure 11). They are approximately parallel to the dominant east-northeast striking joints and parallel the local en-echelon fracture zones. As well, they are parallel to the longest fractures that were interpreted from the high-resolution air photograph of Figure 4. Due to limited exposure, the length of these faults has not been accurately determined, however, the longest one observed was traced more than 70 meters parallel to strike. Sometimes these faults occur as discrete breaks in the bedrock, but more often they occur as fracture zones 10 to 20 centimeters wide. These faults appear to cross cut all other joints in the Oswego Formation and the northwest striking set have been used as offset markers (Figures 11B, C & D), that document a left lateral slip history. The most amount of displacement inferred for any one of these faults was about 1 meter, and the average displacement is about 20 centimeters. The dominant joints appear to be mutually intersecting at most outcrops with no apparent displacement, however, in the vicinity of these minor faults, both sets of fractures show evidence for left lateral displacement of the other joint set. Figure 12 shows two examples of northwest striking joints with apparent left lateral displacement of the east-northeast striking joints.

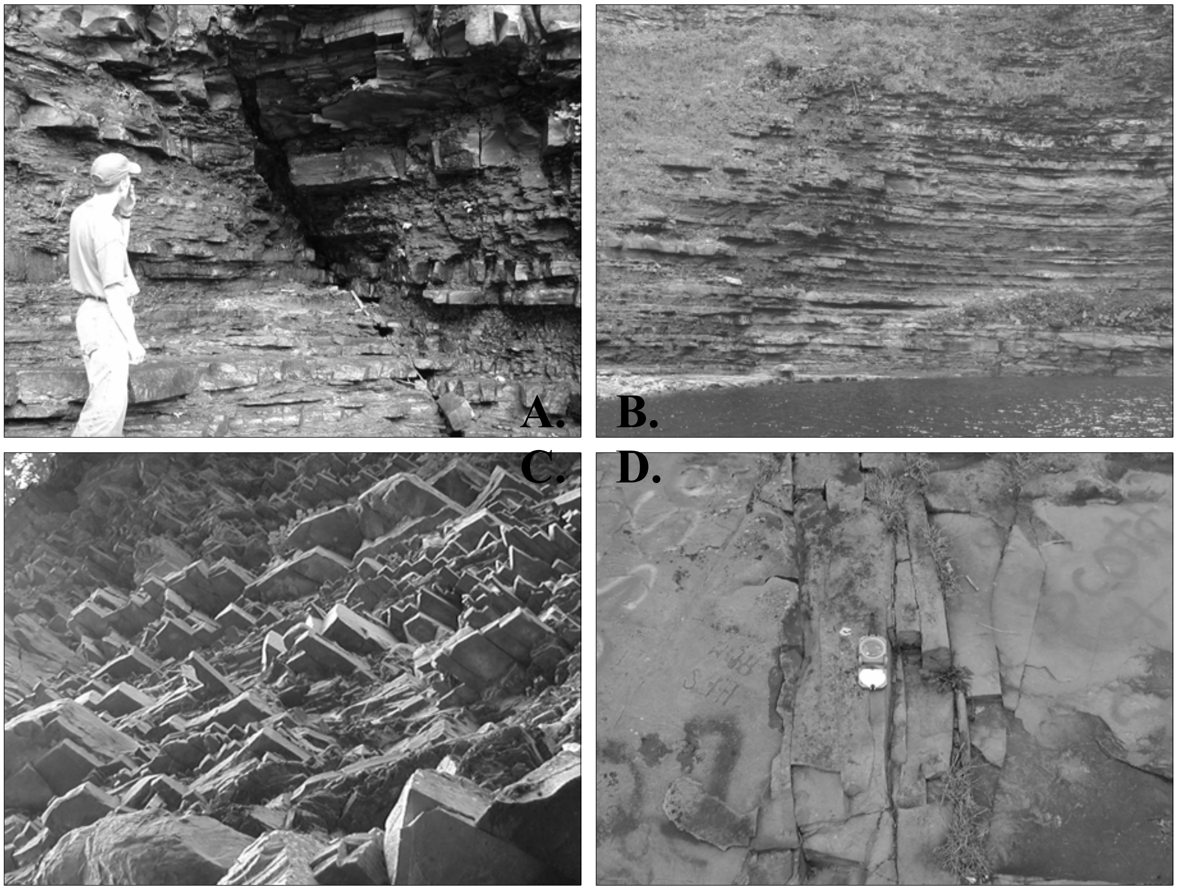


Figure 6 – Photographs of the Pulaski and Oswego Formations in the Salomon River gorge. A. and B. Outcrops of interlayered sandstone, siltstone and shale from the Salmon River falls with a minor normal fault in A. C. Irregular surface in the Pulaski Formation showing the two dominant joints. D. Pavement outcrop of the Oswego Formation from the top of the Salmon River falls with a fracture zone.

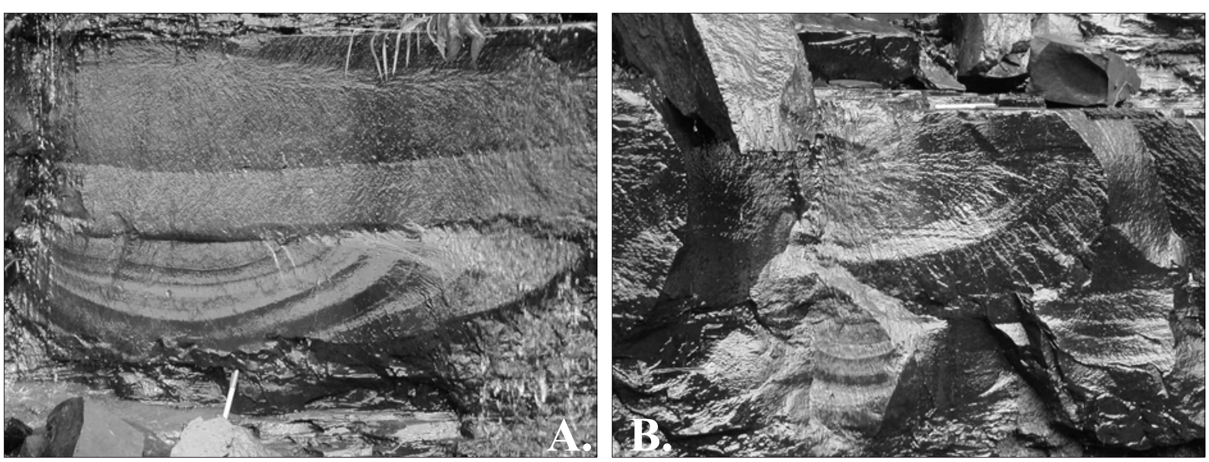


Figure 7 – Joint surfaces in sandstone layers from the Pulaski Formation with plumose structure. The plumose structures are almost always associated with the northwest striking, subvertical fractures. Pencil for scale in both photographs, and the view is looking northeast at subvertical joint surfaces.

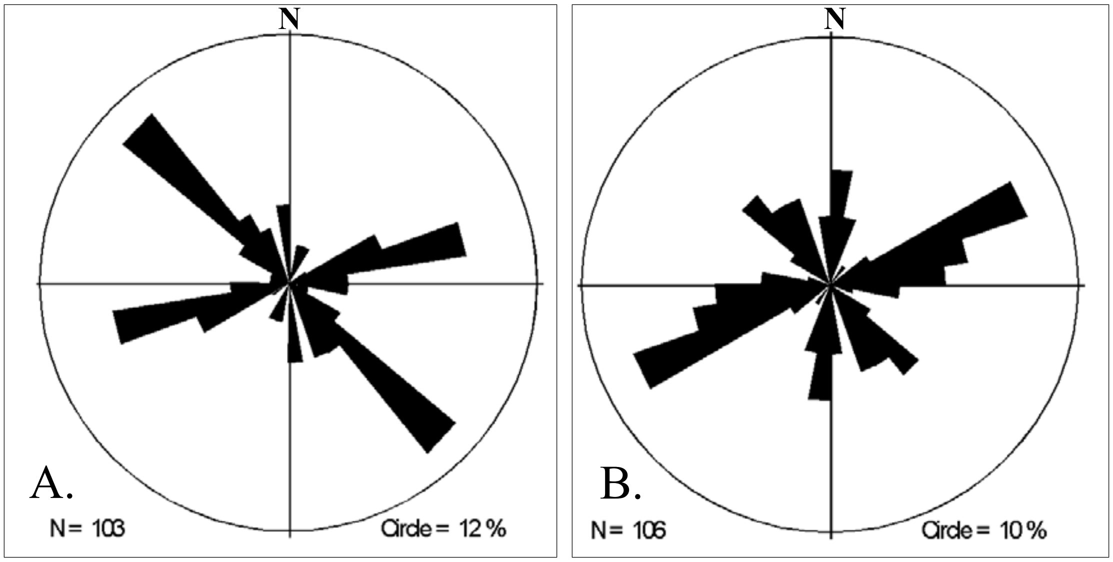


Figure 8 - Rose diagrams for joints that occur in the Oswego Formation in the area of Oswego, NY. A. A composit of shore line outcrops along Lake Ontario at SUNY-Oswego; B. Pavement outcrops that occur in the bed of the Oswego River south of the Utica Street Bridge (these exposures are only available during the lowest water times and will not be viewed during this trip).

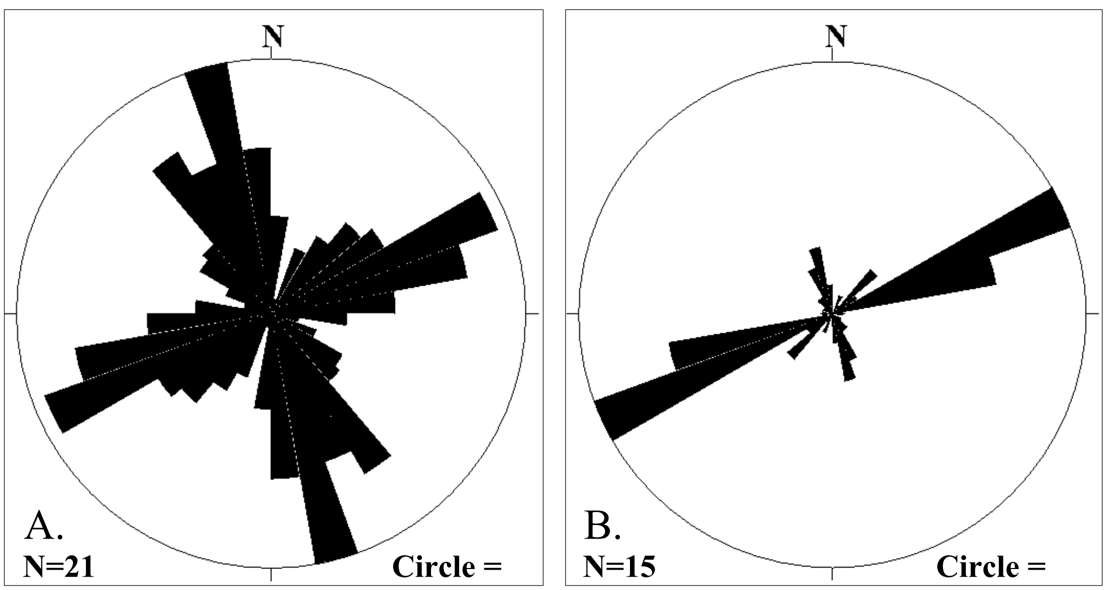


Figure 9 - Rose diagrams for joint traces interpreted from the high-resolution air photograph of Figure 3. The scale of the image was calibrated in a computer mapping program, and the length of joint traces were measured in addition to the attitude. A. Joint traces less than 10 meters in length; B. Joint traces greater than 10 meters in length.

DISCUSSION AND CONCLUSIONS

From out observations of regional and local joint patterns along the southeastern Lake Ontario shoreline and in the Salmon River gorge, there are joints related to those described by earlier researchers in central New York. Specifically, the northwest striking set of subvertical joints is probably related to the late Paleozoic Alleghanian orogeny which produced regional cross fold joints in the Paleozoic strata

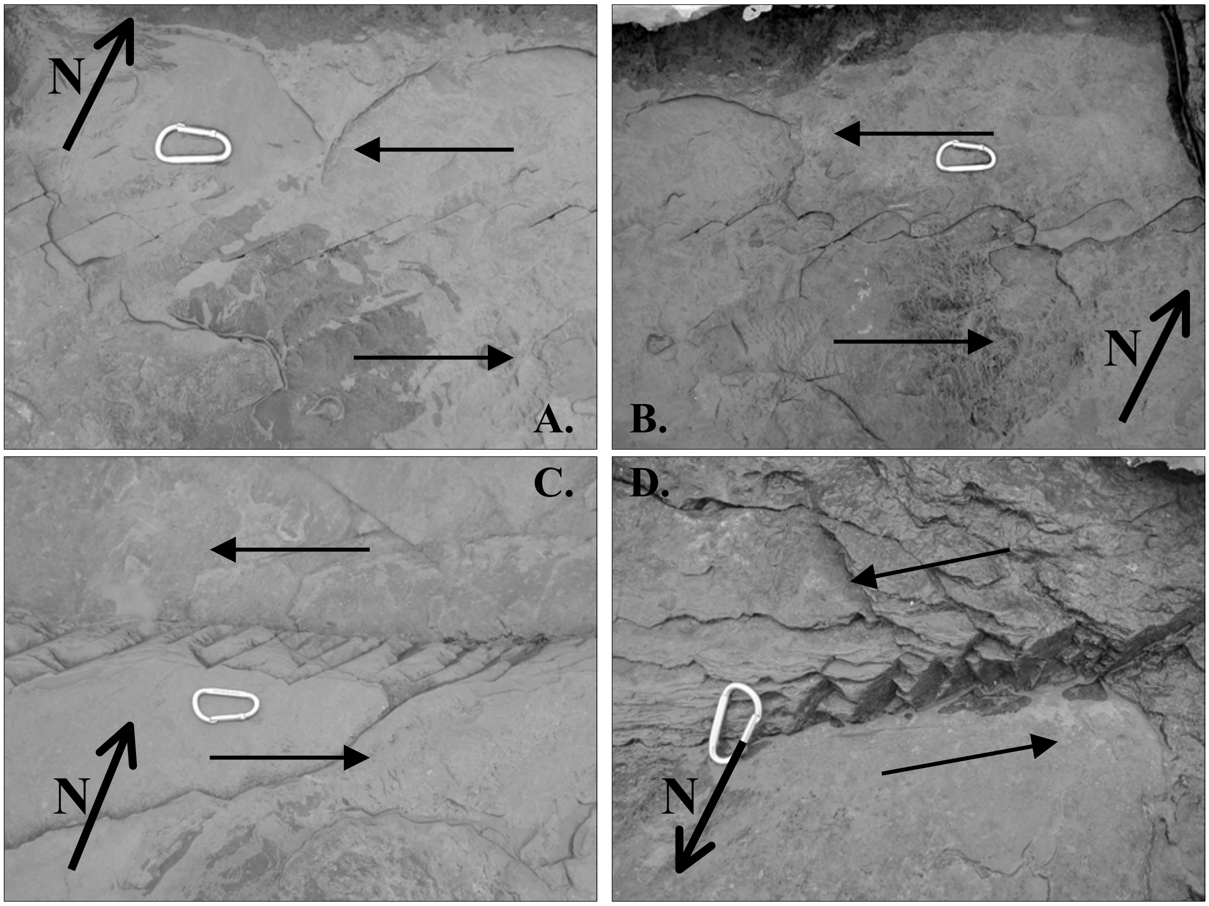


Figure 10 – Photographs from the Oswego Formation showing examples of en-echelon fracture zones. The view for each photograph is looking down at a pavement exposure and the direction of north is shown with the arrow. The metal ring shown in each photograph is approximately 10 centimeters long.

(Engelder, 1979; Engelder and Geiser, 1980; Zhao and Jacobi, 1997). The other dominant joint set, east-northeast striking, are parallel to joints described by Engelder (1982) and proposed to be associated with the modern stress field in the crust. At most outcrops of the Oswego and Pulaski Formations, it is difficult to determine the relative timing between these two joint sets because they appear to be mutually intersecting. However, the east-northeast striking left lateral en echelon fracture zones and minor left lateral faults appear to displace the apparent Alleghanian joints (NW striking set). Engelder et al. (2001) demonstrated minor displacement on joints in the Finder Lakes region, but attributed it to layer parallel deformation during the Alleghanian orogney. Only in the vicinity of these minor faults do the northwest striking fractures show evidence of shear. We interpret this local shear near the left lateral faults as sympathetic reactivation on older joints, possibly those with a locally favorable orientation relative to the stress field that produced the overall left lateral slip. Engelder (1982) proposed that the origin of the east-northeast striking joints could be the result of the contemporary stress field in the lithosphere. If these joints are directly associated with the en-echelon fracture zones and minor left lateral faults, then the local stress field should be consistent with this type of displacement. Overcoring data were collected for Ninemile Pont during the construction of the nuclear power stations (Dames and Moore, 1978). Engelder and Geiser (1984) compiled these data and show that the maximum horizontal compressive stress trends northeast-southwest. The orientation of te inferred stress would be consistent with left lateral slip on east-northeast striking failure surfaces.

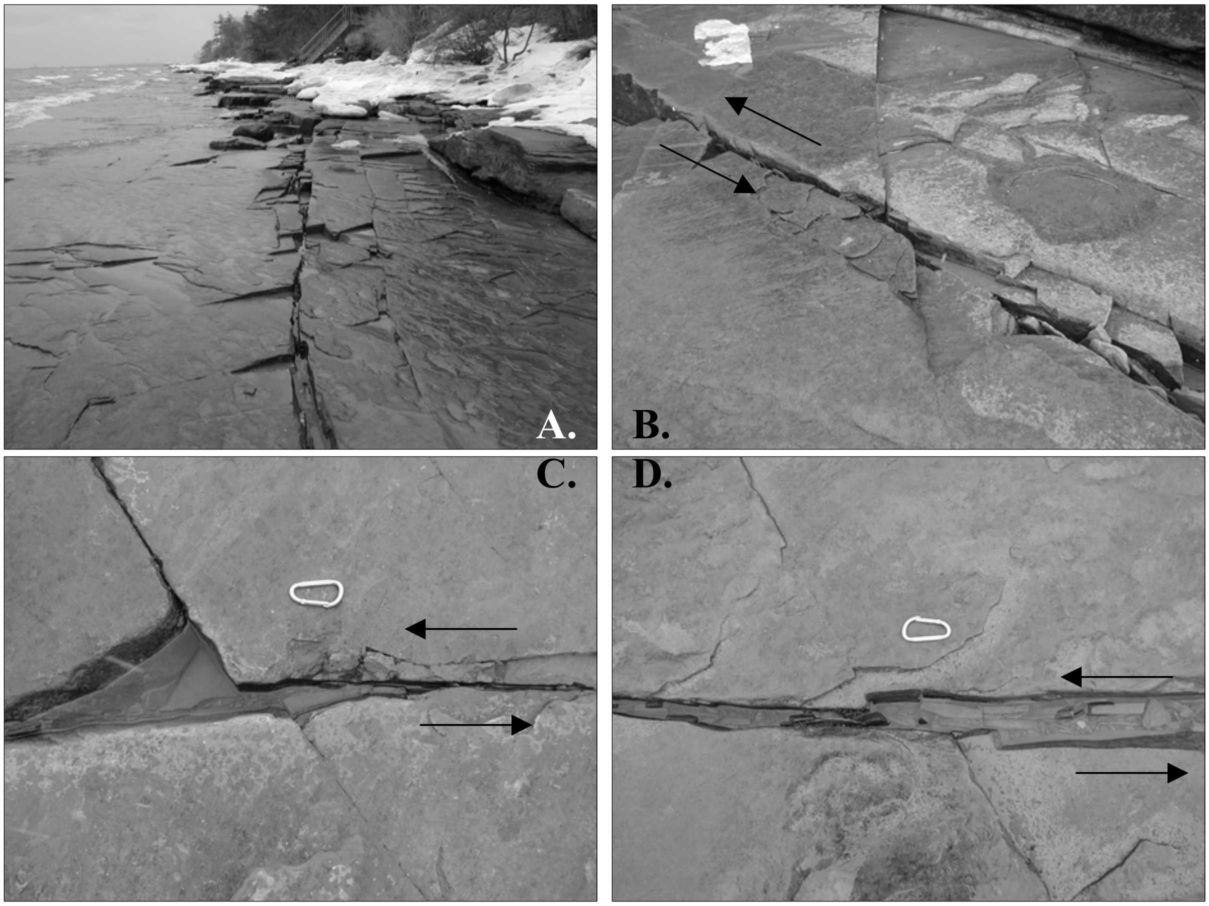


Figure 11 – Outcrops of the Oswego Formation showing a minor sinistral fault with parallel joints at the Lake Ontario shoreline on the campus of SUNY-Oswego. The offset on the fault was inferred from the apparent displacement of the northwest striking fracture set that occurs in the Oswego Formation. The view for A is east parallel to the shoreline. Photo B is a closer view of the central region of photo A. C & D show the apparent left lateral displacement of the northwest striking joints along the same fault.

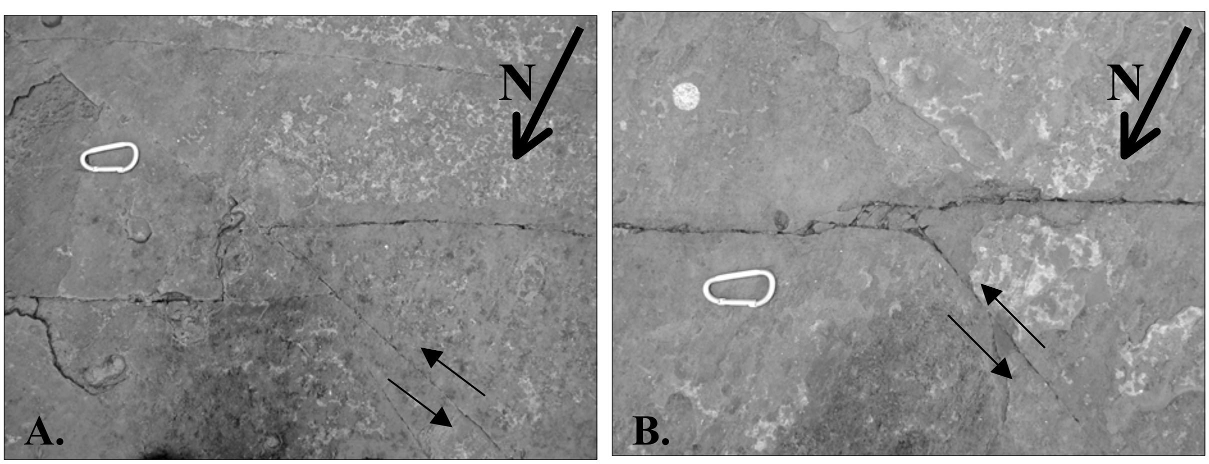


Figure 12 –A. The east-northeast striking joint is displaced about 15 centimeters toward the left along the northwest striking joint. B. The northwest striking joint merges with the east-northeast striking joint at a place where the through-going joint appears to be kinked by left lateral slip on the northwest striking joint.

FIELD TRIP DESCRIPTION AND ROAD LOG

Road Log:
Mileage:

0.0	The trip begins in the eastern parking lot of Piez Hall on the SUNY Oswego campus. Park vehicles at the lot and walk to the lake shore to view the outcrops for Stop 1.

STOP 1 – Pavement outcrops of Oswego Formation at the shoreline of Lake Ontario at SUNY-Oswego.

The shoreline of Lake Ontario at SUNY-Oswego has abundant outcrops of the Oswego Formation (see Figure 4 for the air photograph of this location). At this location, the Oswego Formation is a thick bedded green-gray sandstone (locally red) with minor siltstone and shale partings. Large cross beds and ripple marks are abundant. The fracture patterns described in the text can be seen at most outcrops, but this stop includes the pavement exposure located about 50 meters west of the westernmost metal staircase. This location contains abundant northwest and east-northeast striking joints, en echelon fracture zones, and one of the best exposed left lateral faults. See the text for details.

0.2	Return to the vehicles, turn left out of the parking lot onto Takamine Road and follow to the intersection with Washington Blvd.
0.3	Turn right onto Sheldon Ave. and proceed to the intersection with Washington Blvd.
0.6	Turn left onto Washington Blvd and follow to the intersection with Bridge Street.
2.1	Follow Bridge Street through the city of Oswego, cross the bridge over the Oswego River and continue to East 9th Street.
2.6	Turn left onto East 9th Street and follow to Crisafulli Drive. Turn left.
2.7	Turn right onto the drive that follows around Mcrobie Ball Field.
2.9	Turn right onto the short drive that leads to the parking area.
3.0	Parking area for the Fort Ontario Cemetery. Walk through the cemetery to the path that leads to the lake shore. Follow the railroad tracks west (toward the fort) until you reach the lake side outcrops.

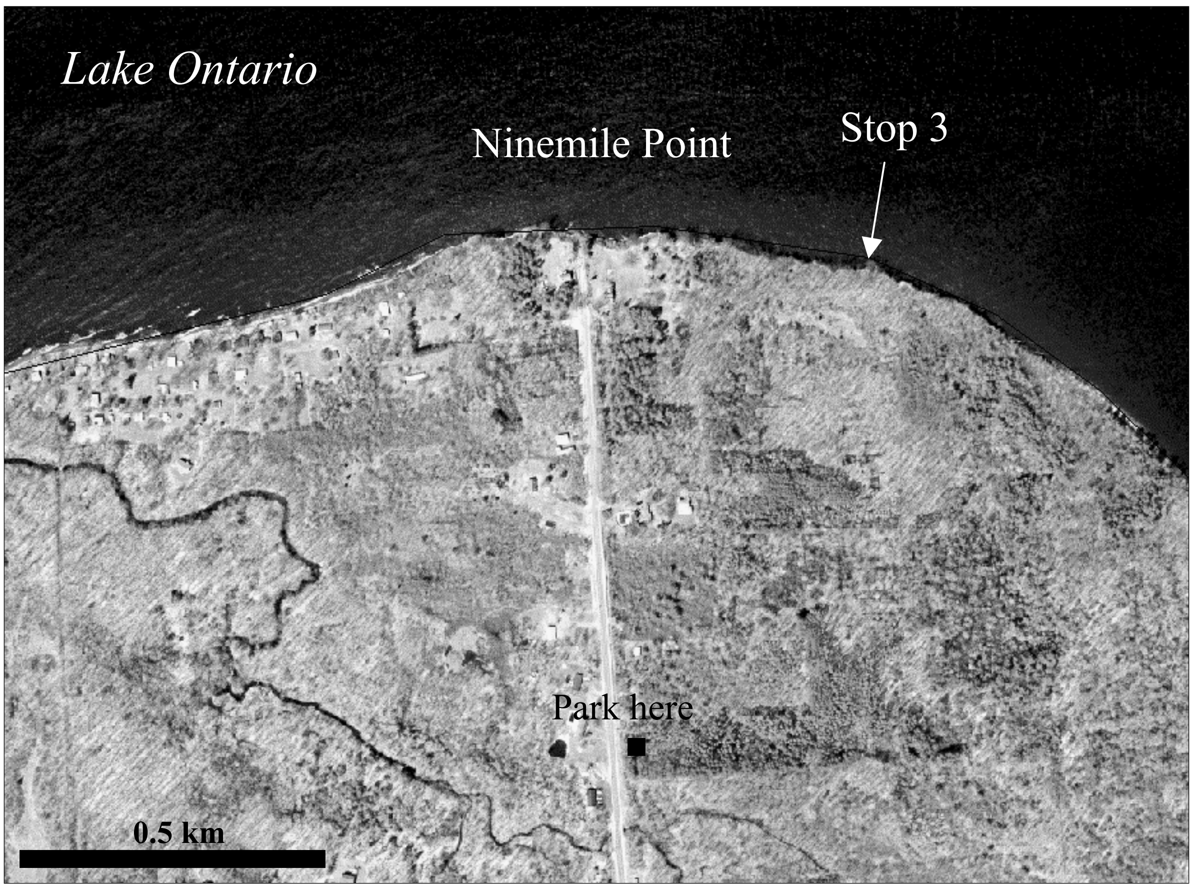
STOP 2 – Fort Ontario, Oswego Formation

A large exposure of the Oswego Formation is located at the lake level below Fort Ontario. If time permits, you are encouraged to visit the historical Fort Ontario, which dates back to the French & Indian War and the War of 1812. The air photo below shows the location of the parking area, the outcrop and Fort Ontario. At this location, the Oswego Formation is again a thick bedded sandstone with abundant joints.

4.9	Back track to the intersection of East 9th Street and Crisafulli Drive and proceed straight onto Mitchell Street to the intersection with East Seneca Street and proceed straight to the intersection with Rt. 1.
10.6	Follow Rt. 1 to the intersection with Ninemile Point Road and turn left.
12.5	Follow Ninemile Point Road to the parking area on the right.

STOP 3: Ninemile Point, Lake Ontario.

From the parking area, follow the foot trail through the forest to Lake Ontario. The walk is about 1.2 km from the parking area to the place where the lake can be easily accessed. At lake level, carefully walk west to the outcrops that form a small cliff. At this location, the bedrock is interlayered sandstone, siltstone and shale, and the relative quantity of shale is more than at Stops 1 and 2. This outcrop is the lower part of the Oswego Formation, and may even represent the transition between the Oswego Formation and the underlying Pulaski Formation.



The attitude of joints at this location varies from the attitude at the previous stops. The rose diagram of Figure 13 shows a composite of 118 joints measured at this location. Fracture density measurements are represented on the detailed stratigraphic column of Figure 13, and show more fractures in the silty and shale portions of the outcrop.

14.4	Back track to Rt. 1 and turn left.
19.1	Follow Rt. 1 to the intersection with St. Rt. 104B, and turn left.

22.1 Follow St. Rt. 104B to the intersection with Co. Rt. 3 and turn left.

26.2 Follow Co. Rt. 3 north to the intersection with St. Rt. 13 in Port Ontario, turn right.

31.3 Follow St. Rt. 13 through the town of Pulask (intersection with Rt. 11) and continue east. At the intersection with Co. Rt. 2A, turn left.

32.4 Follow Co. Rt. 2A to the first Railroad crossing. Park vehicles along the road near the crossing and follow the track west toward the Salmon River. Do not cross the bridge over the river. The outcrop is located in the hill-side along at river level to the right of the tracks. Proceed back to the vehicles.

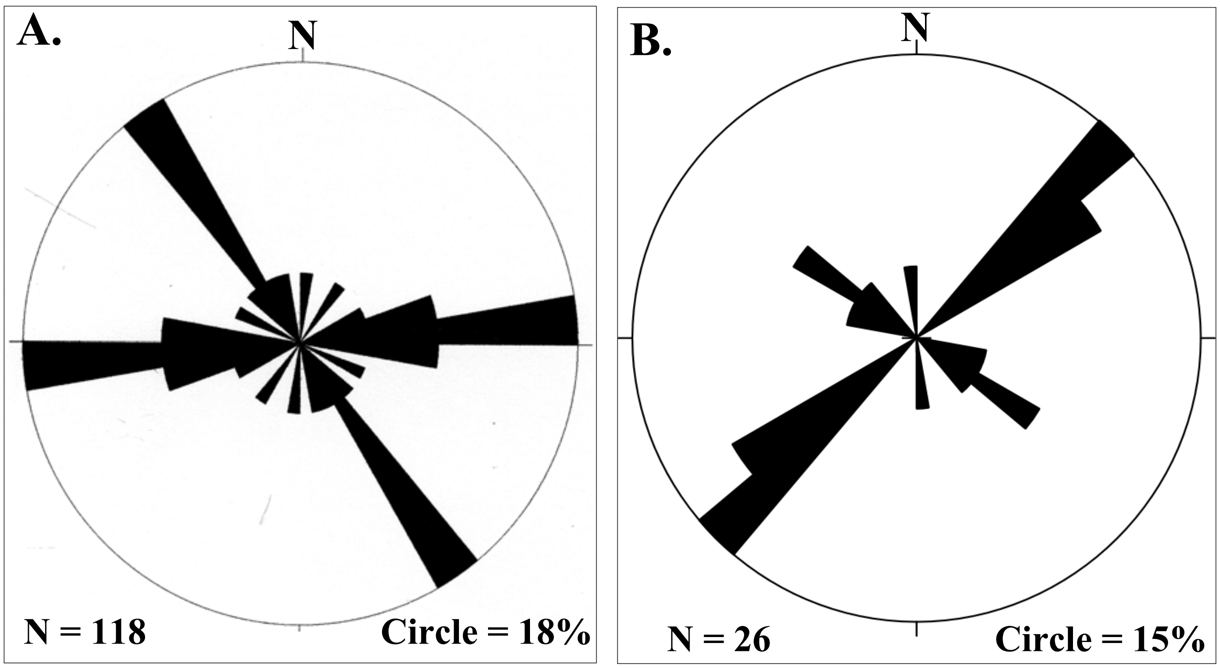
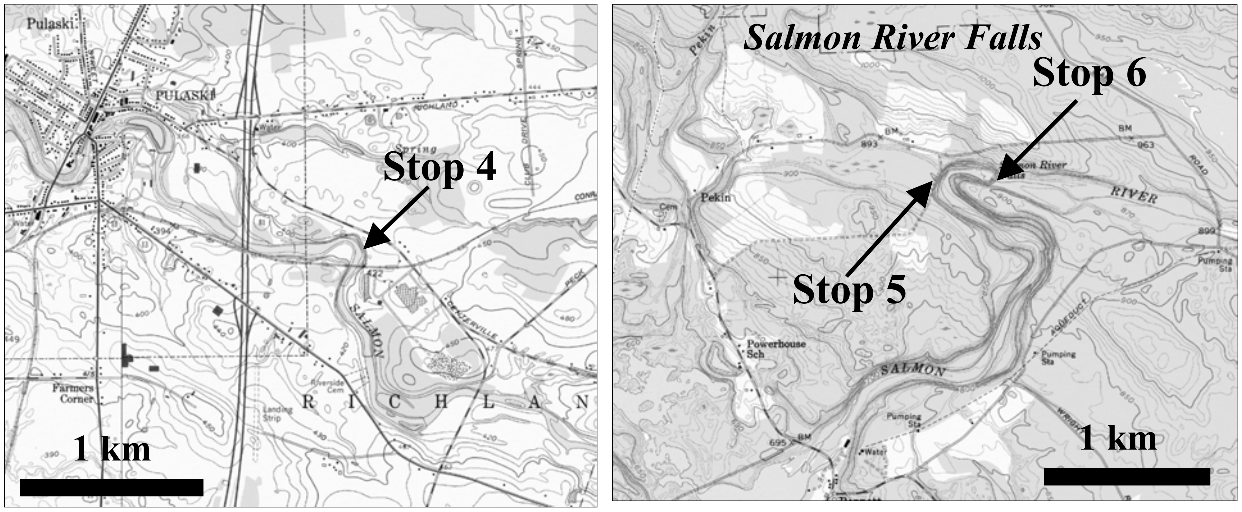


Figure 13 – Rose diagrams for the strike of joints in the lower Oswego (A.) and upper Pulaski (B.) Formations. A. Data collected at Ninemile Point on Lake Ontario, Scriba, NY. B. Data collected at Stop 4 on the Salmon River in Pulaski, NY. Although both dominant joint sets are present, the orientations vary considerably between each location.

STOP 4: Outcrop on the Salmon River under the railroad bridge east of Pulaski, New York.

Once the parking area is reached, where the railroad intersects County Route 2A, walk west along the railroad to the bridge over the Salmon River. Do not enter the railroad tracks, or cross the railroad bridge. At the eastern side of the railroad bridge, carefully traverse down the slope to the river level. There is a steep trail on the south side of the railroad tracks. The outcrop is located north of the railroad bridge on the eastern side of the river. Although this exposure is deeply weathered, there are few easily accessible outcrops of Pulaski Formation. Here the formation comprises interlayered sanstone, siltstone and shale. Sandstone beds range from 0.5 to 1 m thick and the shale contains cm-thick beds of siltstone. Occasional tool marks, ripple marks, and flute-casts are present in the sandstone beds, and the shale beds contain fossils. Joints are best observed in the sandstone beds, and are dominated by a northeast striking set (Figure 13B).



These topographic maps show the locations of Stops 4, 5 and 6 near Pulaski and Altmar, New York.

33.4	Back track on Co. Rt. 2A to St. Rt. 13 and turn left heading southeast.
39.0	Follow St. Rt. 13 to the town of Altmar and turn left onto Cemetery Street.
39.1	Follow Cemetery Street to the first intersection and continue straight. The road turns into Co. Rt. 22.
43.6	Follow Co. Rt. 22 past the NY State Fish Hatchery, past the lower Salmon River Reservoir, and turn right onto Falls Road.
44.9	Follow Falls Road up the hill to the parking area on the right.

STOPS 5 & 6: Below and above the Salomon River Falls.

Follow the steep path (with stone stairs) into the Salmon River gorge. The path splits in two directions at the river level. Follow the river bank to the right (southwest) about 200 m to the waterfall. Stop 5 is the series of outcrops that start at the waterfall and end at the sharp bend in the river downstream. At the river level, thick beds of sandstone comprise the lower part of the Pulaski Formation. Long joints (up to 30 m long), strike about 025 and are steeply dipping. These joints are spaces about 1-2 meters and intersect discontinuous joints that strike northwest (Figure 14A). The thick sandstone beds are overlain by interbedded sandstone, siltstone and shale that is progressively more dominated by fine grained sedimentary rocks upward. This sequence of rocks is best viewed in the cliff face that occurs in the sharp bend in the Salmon River. Walk back to the path and climb up the gorge. Follow the path eastward to the top of the Salmon River Falls. At this location, please obey all safety signs and stay in the designated areas. Stop 6 is the pavement outcrop at the top of the falls of Oswego Formation. Here the rocks are green-gray sandstone with abundant ripples and large cross-beds. Joints transect the entire outcrop for many meters and occur in complex zones. The dominant joint set strikes about 060, but is associated with some complex joint patterns. In places the individual joints with an average strike of 035 curve and are continuous with the 055 striking joints. Additionally, there are en echelon fracture zones with apparent left lateral geometry that merge with the dominant joint set. Joint spacing varies from a few meters to well developed fracture zones with spacing less than 10 centimeters.

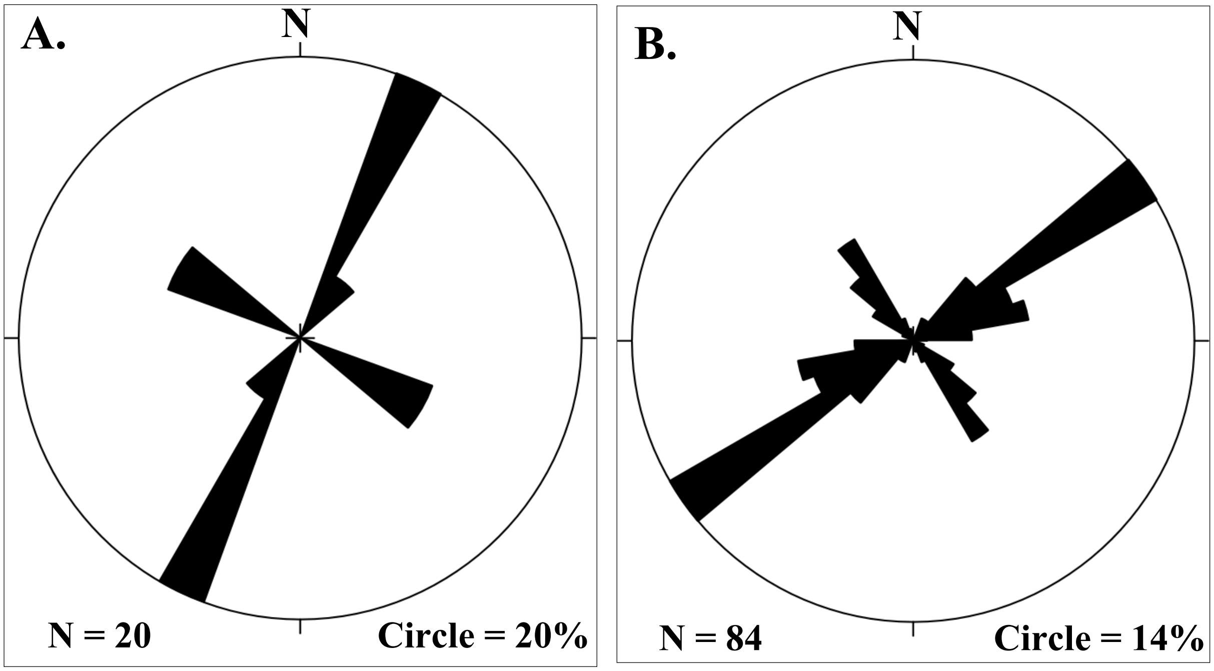


Figure 14. Rose diagrams for the strike of joints in the Pulaski (A.) and Oswego (B.) Formations at the Salmon River gorge near Altmar, New York. A. Data collected at river level down stream of the fall about 200m. B. Data collected at the pavement outcrops above the falls.

END OF TRIP.

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Trip A-4 & B-1

CLASSIC LOCALITIES OF THE BLACK RIVER AND TRENTON GROUPS (UPPER ORDOVICIAN) IN THE BLACK RIVER VALLEY: REVISITED THROUGH TRADITIONAL AND SEQUENCE STRATIGRAPHY

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INTRODUCTION

The late Ordovician (Mohawkian Series) Black River and Trenton groups of the northern New York State consist of some 250 m of highly fossiliferous, well-preserved limestones with some intervening shales. These rocks are exposed along a southeast-northwest trending outcrop belt for more than 250 km in central New York and neighboring Ontario, where they closely flank the Precambrian Grenville Province (Figure 1). These units

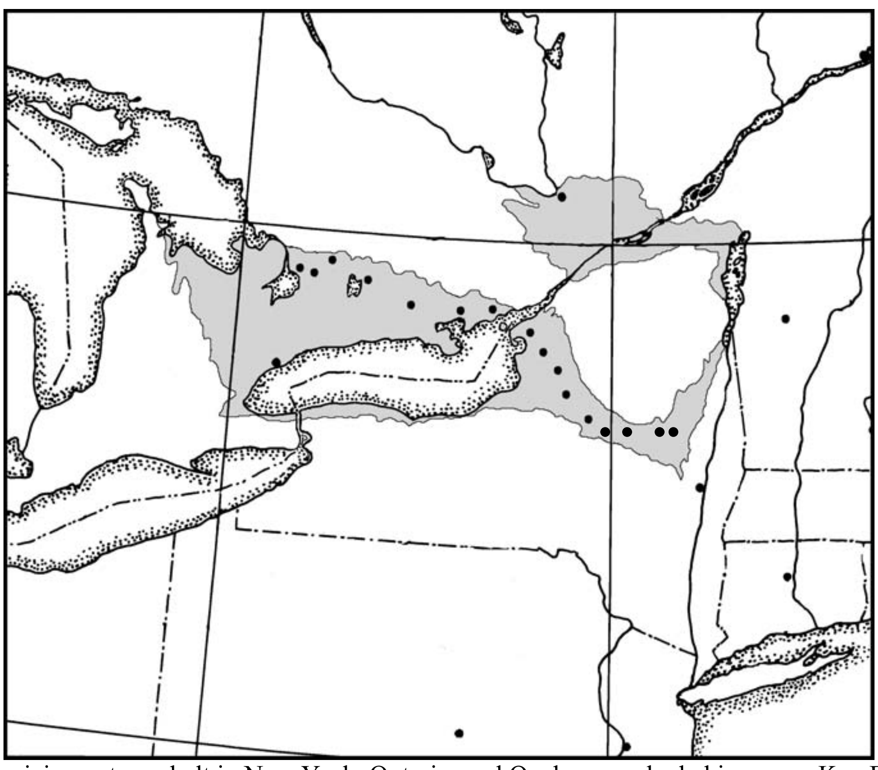

Figure 1 - Ordovician outcrop belt in New York, Ontario, and Quebec are shaded in grey. Key Black River & Trenton Group outcrop localities are indicated within the outcrop belt.

deposited from ~457-450 million years ago, have been studied for nearly two centuries and have become very important units for the study of paleontology, carbonate sedimentology, stratigraphy, paleoclimatology, paleoceanography, paleoecology, structural geology, and more recently, petroleum and natural gas geology. Mostly known from the renowned Mohawk Valley sections including those at Trenton Falls on West Canada Creek, the Black River and Trenton play (and have played) an important role in our understanding of depositional processes, environmental change and even mountain building and plate tectonics. The purpose of this field trip is to revisit the classic stratigraphy of the Black River and Trenton groups, and to introduce some of the modern research ideas that are being applied to answer long standing questions regarding the history of these rocks.

GEOLOGIC BACKGROUND

Absolute and Relative Time Scale:

The Black River and Trenton units, originally defined in central New York State (Vanuxem, 1842), are among the earliest formally designated stratigraphic units in North America (Figure 2). As a consequence, the names of these well-known intervals were applied in a variety of localities not only within the New

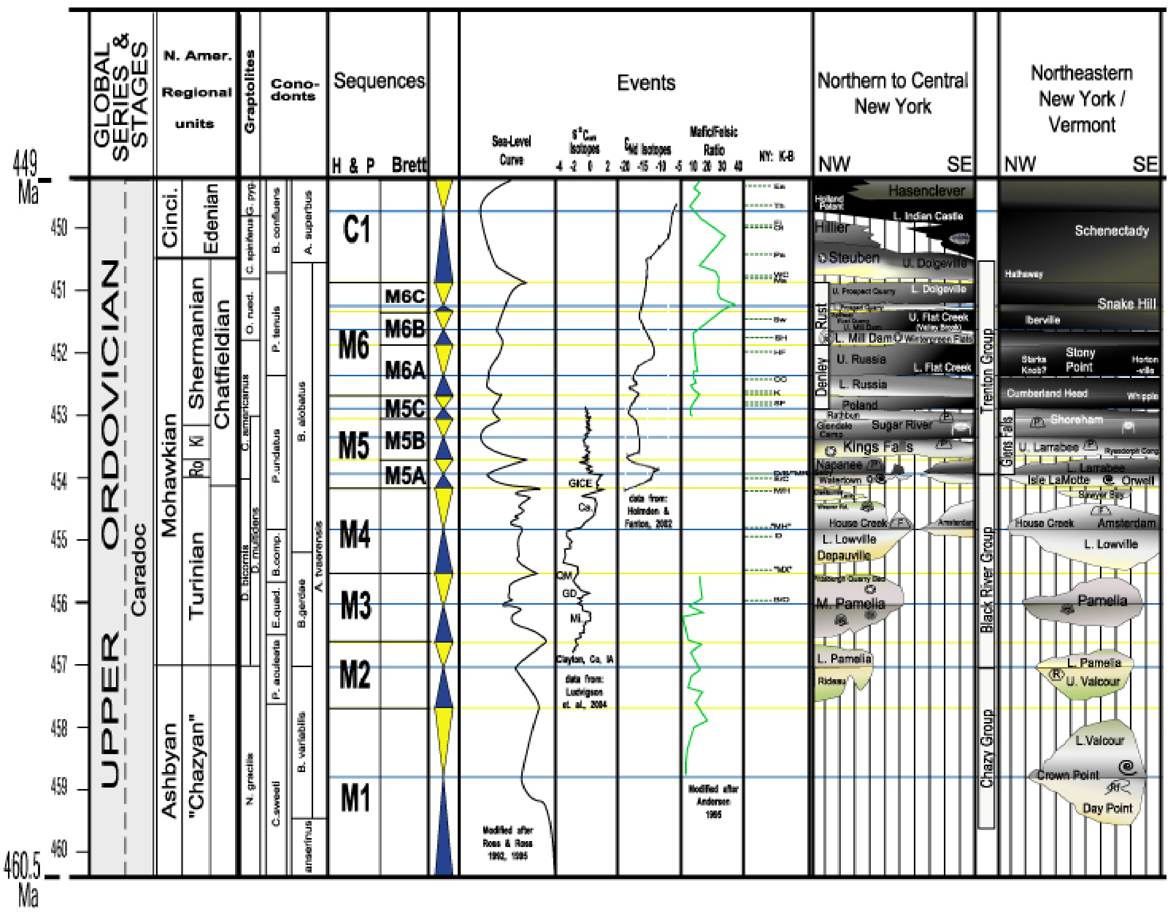

Figure 2 - Upper Ordovician Relative and Absolute Chronostratigraphic Chart. Illustrated are the boundaries of the Mohawkian Series and internal stage boundaries. In addition, positions of key biostratigraphic zones for graptolites and conondonts, as well as 3rd-order depositional sequences of Holland & Patzkowsky (1996) and Brett et al., (2004) and important chemo- and event stratigraphic horizons are established. Abbreviations: Ro. = Rocklandian; Ki. = Kirkfieldian.; Cinci = Cincinnatian

York region, but throughout much of eastern North America. Originally these units were described on the basis of gross lithology, however, subsequent investigations looked at these not just as rock-units, but also as a series of distinctive biostratigraphic units which were considered as time-rock units that were sometimes used interchangeably with the original lithologic rock units. The practice of using rock-terminology and time-rock terminology as interchangeable was eventually challenged as it was surmised that rock units themselves did not have a relative time component and indeed in some cases were diachronous through time according to Walther's Law. Not surprisingly, stratigraphers have discouraged the use of lithostratigraphic terms for time-rock terms. As such the older lithology-based rock terminology (i.e. Black River, Trenton) are strictly to be used as descriptors of lithology, while new nomenclatural terms are established to reflect relative chronology. Among these are the stage level classifications of Kay (1937, 1960) which include the Rocklandian, Kirkfieldian, and Shermanian stages for the Trenton Group, and in order to accommodate a return of the older Black Riveran (time-rock) to Black River Group (rock-unit) as was originally intended, Fisher's (1977) term Turinian, has been instituted. More recently, what was called the Trentonian Stage is roughly equivalent to what is now termed the Chatfieldian Stage (Leslie and Bergström, 1995), and collectively the Turinian and Chatfieldian Stages make up the Mohawkian Series as part of the North American time-rock classification. However, the alternative stage classification, using Kay's (1937) more precise subdivisions of the upper Mohawkian Series: the Rocklandian, Kirkfieldian, and Shermanian stages, are still useful in the type region, and are also still in common use in neighboring regions despite some recent opposition.

The interval of time embraced by the Black River and Trenton Groups lies now entirely within the Late Ordovician, and not the Middle Ordovician as previously defined. It should be noted that the IUGS Subcommission on Ordovician Stratigraphy has recommended that the interval previously termed Middle and Upper Ordovician (i.e. the Llandeilo, Caradoc, & Ashgill stages; which include the Chazy, Black River and Trenton interval) all be assigned to Upper Ordovician (Webby, 2004). In the British classification the entire Upper Ordovician interval is assigned to the Caradoc Series (now in emended use as the Caradocian Global Stage) (*see* Figure 2).

Absolute age dating for this interval has become a more realistic endeavor in recent years as multiple K-bentonites have yielded datable phenocrysts both inside and outside of the New York region (Samson *et al*. 1988, Haynes, 1994; Kolata *et al*., 1996, Min et al., 2001; McLaughlin *et al*., 2005; McLaughlin *et al*., in press). Two widely distributed K-bentonite horizons, the Deicke and Millbrig, have been dated by a number of geochronometric systems (*see* Figure 2). The resulting ages using U/Pb dating methodology are established at about 454.5 +/- 0.25 Ma for the Deicke K-bentonite; and to 453.1 +/- 0.65 Ma (for U/Pb dates of zircon) for the Millbrig (Min et al., 2001). Moreover, recent dating of the High Falls K-bentonite (from Trenton Falls) has yielded a highly concordant date from U/Pb zircon date of 451.8 +/- (McLaughlin *et al*., 2005, McLaughlin *et al*. in press). Based on these assessments and by extrapolations from dates on the Ordovician-Silurian boundary, the entire Black River to Trenton Group interval ranges from approximately 457 to 450 million years providing for 6-7 million years of deposition.

Tectonic Setting:

During the late Middle Ordovician, the majority of ancestral North America (Laurentia) was covered by shallow tropical seas commonly referred to as the "Great American Carbonate Bank." At this time, the eastern portion of Laurentia was rotated approximately 60 degrees clockwise relative to its present configuration, and the New York region was located approximately ~30º south of the paleoequator (Scotese *et al*., 1994). Due to the onset of Sloss's (1963) Tippecanoe Megasequence in the earliest Upper Ordovician, sea-levels were, over the long-term, in a transgressive phase such that during the deposition of the Trenton very little land was exposed on the continent. In fact, the only exposed areas were portions of the modern Canadian shield and areas along the Transcontinental Arch that were in general of very low relief (Figure 3). Although we have no epicontinental seas comparable to the size of the Laurentian Great American Carbonate Bank, modern regions such as Florida Bay and the Bahamian Platform are commonly used as analogs for understanding the conditions of the Laurentian craton at that time.

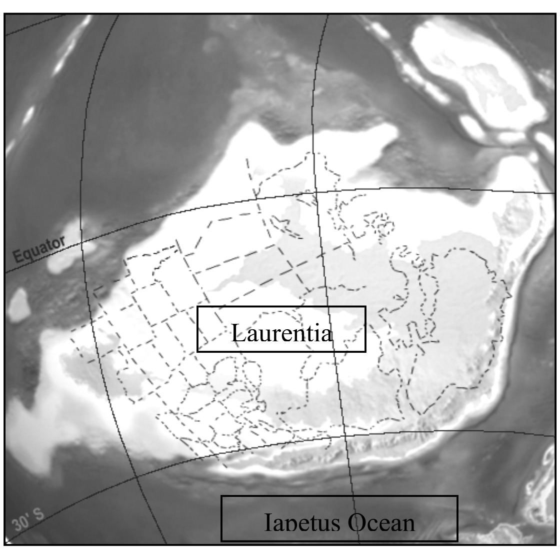

Figure 3 - Paleogeographic reconstruction of Laurentia. Dashed lines indicate the relative position of modern political boundaries for reference (modified from Blakey, 2003).

Moreover, at the time the Black River and Trenton limestones were deposited, the eastern margin of Laurentia had begun to undergo a major paleoceanographic, paleoclimatic, and paleoenvironmental transition associated with the onset of the Taconic Orogeny. The subsidence of portions of the Laurentian craton in Chazy time during the early or Blountian tectophase of the Taconic Orogeny ended a period of over 30 million years (U. Cambrian to M. Ordovician) where virtually no tectonic activity occurred on the craton. However, in the 6-7 million year period that ensued, the Laurentian margin of the present day eastern United States was impacted by the development of two pulses of collision (or tectophases). The first tectophase was located south of New York in the vicinity of Tennessee and impacted sedimentation in the Sevier Basin and neighboring regions throughout the Turinian (Figure 4). The second tectophase, referred to as the Vermontian Tectophase collided in the vicinity of the New York Promontory producing the Martinsburg Basin during the deposition of the Trenton Group in the Upper Mohawkian. This later event ultimately influenced changes across the entire eastern Laurentian Craton as far west as Minnesota.

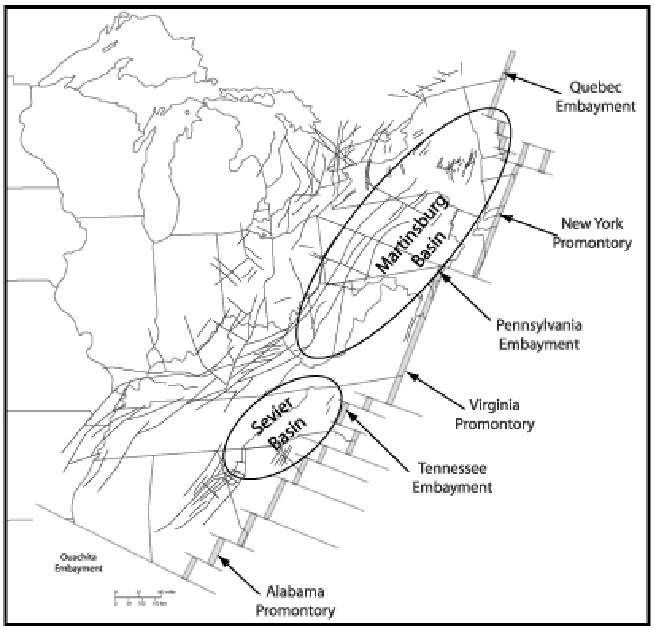

Figure 4 - Extent of the post-Iapetan rift margin of eastern Laurentia upon which the Great American Carbonate Bank Developed. Highlighted are the position of key transform and rift segments that are responsible for the development of an irregular series of promontories and embayments (after Thomas, 1991). In addition, the position of two basins induced by the Taconic Orogeny: the southern or Sevier Basin (Blountian Tectophase) and the northern or Martinsburg Basin (Vermontian Tectophase) is shown. Also indicated are the position of key fault structures that are either Pre-Cambrian rooted or have been demonstrated to have had Taconic-related movement.

Although the exact relationships of these tectophases have been studied (Ettensohn, 1991, 1994) many circumstances of their temporal-spatial relationships are still being worked out. In spite of this, however, it is now fairly certain that the development of orogenic activity on the eastern Laurentian margin resulted from the subduction of the cratonic margin of Laurentia under the closing Iapetan Oceanic plate resulting in the development of a foreland basin (Shanmugam & Lash, 1982; Rowley & Kidd, 1981) or deep-water moat exterior to the cratonic interior termed the Champlain Trough (Kay, 1937) into which siliciclastic sediments derived from the collisional belt were deposited (Figure 5).

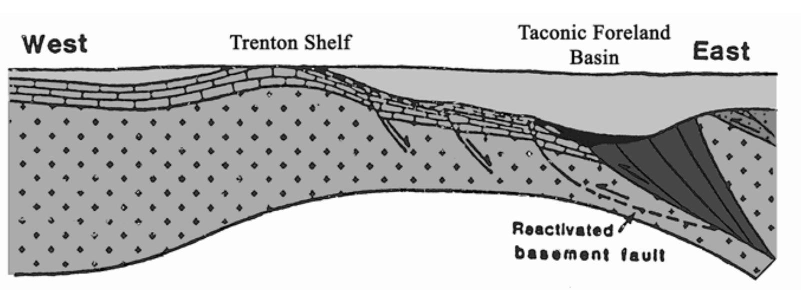

Figure 5 - Schematic Cross-Section of Trenton Shelf and Taconic Foreland Basin during early Kirkfieldian Time. Notice the development of an accretionary prism introducing a deformational load to the eastern margin. Outboard of the accretionary prism, is the location of the volcanic arc which during the Trenton delivered significant quantities of sediment into the foreland basin and up onto the craton (Modified after Shanmugam & Lash, 1982)

These siliciclastic sediments were dominated initially by clays (hence the predominance of shaly limestones in the Trenton) but in later phases of each orogeny clays were replaced upward and laterally by silts (siltstones) sands (sandstones) and even gravels (conglomerates) as tectonic activity subsided and the foreland basin was filled in. Although both tectophases produced similar basin fill packages, provenance studies of these siliciclastic sediments have highlighted a major difference in the source of Blountian and Vermontian sediments (Mack, 1985; Andersen, 1995; Bock et al., 1998). These data from both sandstones and mudrocks, suggest Blountian sediments were derived from accretionary prism sources that were scraped off the cratonic margin during the initial subduction and collision of the Blountian tectophase (Figure 4). On the other hand, sediments derived from the Vermontian tectophase (including those of the Trenton Group) indicate a significant shift to volcanic source rocks indicating a fairly major difference in tectophase processes (*see* Cornell, 2005).

Within the field trip region, these tectonic settings are expressed through a variety of observations: First within both adjacent Ontario and in the central and southern Black River Valley, Precambrian basement rocks are surrounded and onlapped by the carbonate rocks of both Black River and Trenton affinities. This suggests that there were some small islands of low lying granitic basement exposed near the shore of eastern Laurentia (Brookfield, 1988). Although dominantly carbonate, initially, the relatively flat low-lying shield complex allowed the deposition of discontinuous siliciclastic-rich sediments in topographically low areas and at times, especially during lowering sea-levels these quartz-rich siliciclastics were remobilized and carried out across the carbonate platform for some distance. In the ensuing transgressions, these siliciclastics were extremely scarce. As such, the general lack of siliciclastic sediments in the almost pure carbonate succession of the Black River indicates very little transport and introduction of terrigenous craton-derived sediments into the local depositional regime. Therefore, this suggests a fairly arid climate and/or a very low elevation with a broadly peneplained configuration for the Grenville Shield and neighboring islands.

Second, within the foreland basin succession and on the adjacent shelf, early in the depostion of the Black River and continuing throughout the deposition of the units are the occurrences of K-bentonite or altered volcanic ash. The predominance of these K-bentonites supports the theory of a volcanic island arc located offshore. Responsible for the largest volcanic eruptions in earth's history (Kolata *et al*., 1996; Huff *et al*., 1992), this volcanic terrain referred to variably as the Ammonoosuc Arc (Rowley & Kidd, 1981) or Shelburne Falls Arc (Karabinos *et al*., 1998) was beginning to migrate towards the northeastern margin of Laurentia as it was subducted.

Climatic Setting

The Black River and Trenton Group sediments of New York State accumulated in subtropical to sub-temperate latitudes, ranging from about 25° to 30°S of the paleoequator (Scotese, 1990). Although very much dominated by tropical carbonates throughout much of the Ordovician, Mohawkian strata record the critical change from the widespread shallow carbonate platform of the Great American Carbonate Bank into the actively subsiding Taconic foreland basins. In the New York region, the impact of Blountian-aged siliciclastic sedimentation is lacking as is the case within much of the region of the subsequent Taconic Foreland Basin. As such clean, peritidal to shallow subtidal Black River carbonates were widespread in the New York region and contained algae and coral-dominated faunas. Subsequently with the onset of the Vermontian phase of the orogeny, the Black River rocks give way vertically and laterally to mixed carbonate/siliciclastic units with

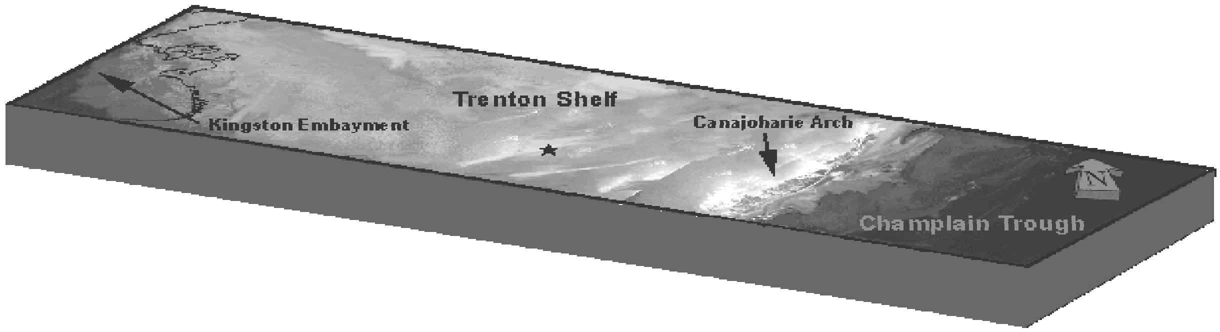

Figure 6 - Paleogeographic reconstruction for the New York (Trenton Shelf Region) during the earliest Trenton (L. Rocklandian and Kirkfieldian) just prior to the major advance of the Martinsburg foreland basin (Champlain Trough) into the eastern New York region. Notice the Canajoharie Arch, a topographically high region during deposition of most of the Black River and lower Trenton. This exposed arch was initially submerged intermittently during periods of high sea-level (i.e. Upper Rocklandian) and exposed and eroding during periods where sea-level was low. Moreover, the Canajoharie Arch acted as a local barrier to open ocean circulation and may have influenced deposition during the latest Turinian through early Shermanian. Thereafter, the Canajoharie Arch became submerged (inverted?) and no longer influenced carbonate sedimentation in that region.

diverse brachiopod-bryozoan-echinoderm biotas and, finally, to black shales. As a result, Laurentia witnessed oceanographic, environmental, and climatic changes which are presumably responsible for concomitant faunal change. In this regard, observations of various outcrop regions suggest an important difference between the Black River and Trenton interval was a difference in climatic regime possibly related to Taconic tectonism. Brookfield (1988) has argued for a cool water (temperate) carbonate depositional regime extending through the entire Black River and Trenton interval in southwestern Ontario (Lake Simcoe), based largely on sedimentologic evidence (e.g., lack of reefs, calcareous algae, and ooids; common occurrence of peloids, and increasing Sr values from Black River into Trenton rocks). Whereas Holland and Patzkowsky (1998) suggested that the Trenton interval was transitional from tropical Black River to temperate post-Trenton intervals. In contrast to Brookfield's observations in southwestern Ontario, and supportive at least in part of Holland and Patzkowsky, many workers in New York and southeastern Ontario have recognized the dominance of many calcareous algae, ooids, and even in some horizons evaporite crystal molds and primary dolomites. These observations suggest that deposition was at least tropical to sub-tropical (at times fairly arid?) and occurred under warm water conditions for at least part of the interval.

STRATIGRAPHY OF THE BLACK RIVER AND TRENTON GROUPS

As mentioned previously the Black River and Trenton Groups have been studied for well over a century and have been, almost since the beginning, a topic of contention and debate for generations of geologists resulting in many revisions and reassessments in stratigraphic nomenclature (Figure 7). Part of this contention and debate stems from: 1) their predominance in early geologic studies of North America such that these units were often used as "guinea pigs" for the education of students, 2) the implementation of new stratigraphic methods (using advanced science) which required the re-evaluation of some of the older long-standing ideas, and to some extent 3) the complex and disconnected outcrop regions which offer minimal continuous exposure and apparent geographic variability. Despite their long-standing history of research, the Black River and Trenton groups are once again the subject of major new discoveries within the immediate region of New York, and within the larger cratonic setting of eastern Laurentia. Moreover, newly recognized chemostratigraphic events (*see* Figure 2) have been studied over fairly broad regions of eastern North America and even in Scandinavia and Estonia (Ludvigsen *et al*., 1996; 2004; Saltzman *et al*., 2001; Bergström *et al*., 2004) and suggest that the transition from the Black River into the Trenton was a fairly major event that was certainly trans-Iapetan in nature and may have even been global.

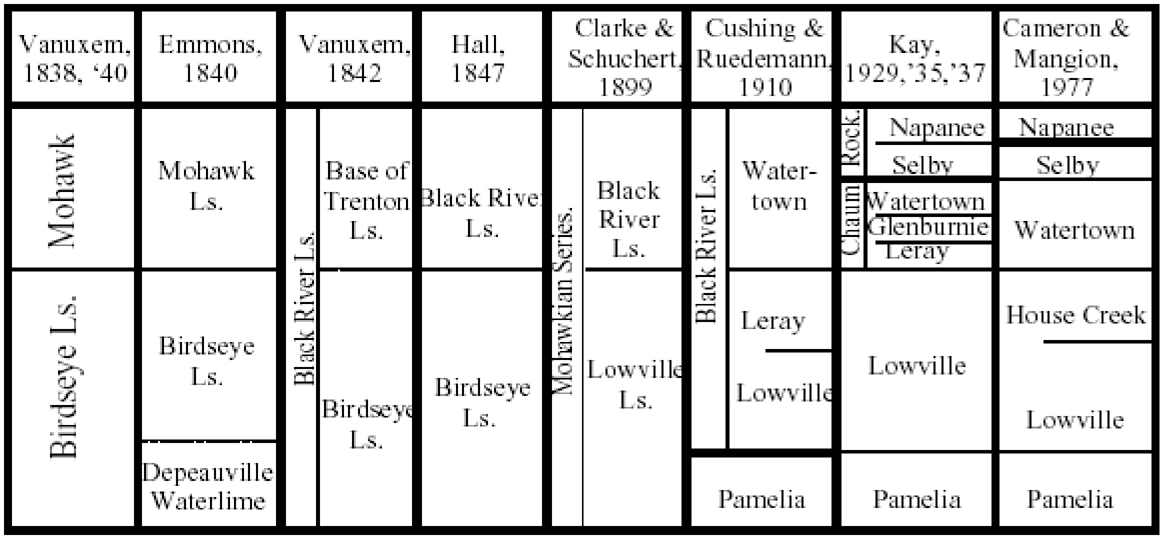

Figure 7 - Historical nomenclature for the type Black River to lower Trenton Group interval for New York State. Note the position of the Black River-Trenton Group Boundary (drawn in bold black line) as drawn by Kay at the base of the Selby, and Cameron & Mangion at the base of the Napanee (from Cornell, 2001).

History of the Black River - Trenton Group Boundary

Given the predominance of previous stratigraphic studies, the stratigraphic details of the Black River-Trenton Group boundary is still problematic in the type section areas and this has led to major correlation controversies outside of the New York State type region. A major dilemma in this issue has been the paucity of outcrop exposures in the specific boundary interval within the Black River Valley thus limiting their interpretation. As such, most stratigraphic studies of the boundary have been completed in the Mohawk Valley region where the Black River Group itself is highly modified and abbreviated by surfaces of non-deposition, extreme facies change, and erosional truncation-karstification. These details are only now being worked out, but historically the position of the Black River/Trenton boundary has been debated (Kay, 1937; Fisher 1962; Titus & Cameron, 1976; Cameron & Mangion, 1977) for nearly a century (*see* Figure 7).

Most recently, the boundary has been synonymous with top Turinian or base Rocklandian in the type area (Fisher, 1962). This placement corresponds to a position at the base of the Selby Formation (of earliest Rocklandian Age) and at the top of the Watertown Formation of uppermost Turinian Age (Figures 2 & 7). This particular boundary interpretation is drawn on a biostratigraphic basis and in the Black River Valley region, this biostratigraphic interpretation is born out through the development of a subtle facies dislocation across a mineralized hardground surface. When traced into the abbreviated and substantially thinned sections of the Mohawk Valley, the equivalent of the sub-Selby facies dislocation is represented by a distinctive lithologic boundary separating off-shore deeper-water facies from underlying shallow water facies at a marked discontinuity (Kay, 1935; 1937). This very shallow-water equivalent of the Watertown and the overlying deeper-water Selby is coincident with the same biostratigraphic transition as in the Black River Valley, but as the lithostratigraphic distinctiveness of this contact is minimized it has not been accepted universally. On this basis, Cameron and Mangion (1977) have established the boundary position not at the biostratigraphic base of the Rocklandian (as in the Mohawk Valley), but at the more pronounced lithologic break at the top of the Selby and base of the Napanee. In both the Mohawk and Black River Valleys this particular surface is quite pronounced by the first appearance of shales and interbedded calcisiltites, and micritic wackestones. However again, this particular delineation is also complicated due to the thin condensed and mineralized nature of the underlying Selby (25-30 cm), as such, it has commonly been overlooked.

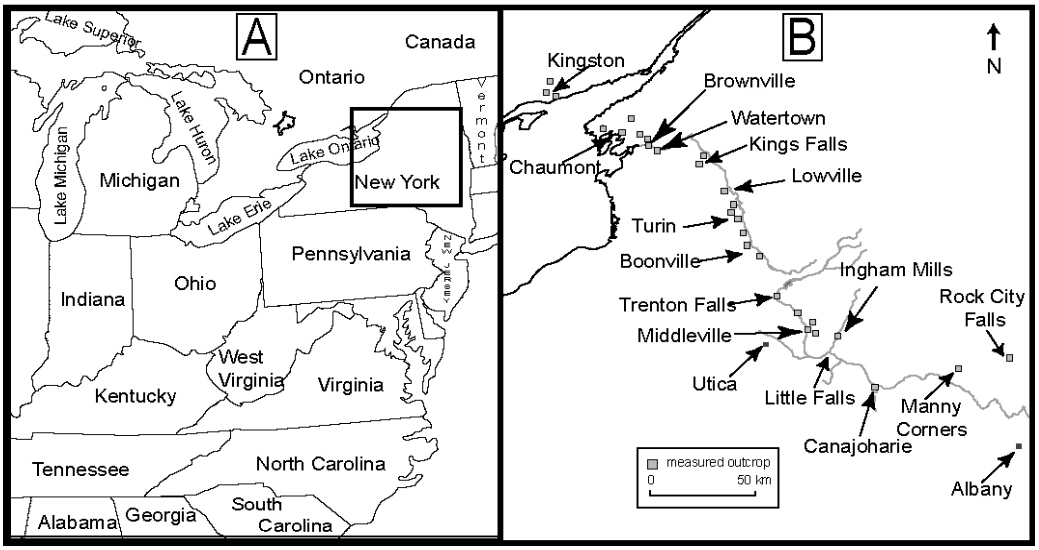

Figure 8 - Important Black River to Trenton Group outcrop localities in New York State. Field trips stops are located in the vicinity of: Chaumont, Brownville, Kings Falls, Lowville, Boonville, and Turin.

Adding to the controversy, another contact (recognized by 19th century geologists) occurs at the top of the "Birdseye Limestone" and base of the overlying Mohawk or Trenton Limestone. For decades this lithologic contact was the important lithologic delineation recognized by geologists. This particular contact was de-emphasized in later studies because the massive-bedded Watertown Limestone, although not equal to the Birdseye, was more similar to it than the overlying thin-bedded and argillaceous limestone of the basal Trenton. In the Mohawk Valley where the Watertown was not previously recognized, the basal Trenton contact approximated this stratigraphic position which suggested that the entire Watertown had been truncated in this region at the base of the Rocklandian – a position held by Kay. However, through detailed correlations Cornell, (2001) was able to identify a lateral (isochronous) transition in the Watertown from off-shore to more onshore intertidal facies in a southeasterly direction. Thus at the base of the shallowest Watertown equivalent facies in the Mohawk Valley, a distinct channelized and karstic surface is documented at Inghams Mills where several meters of underlying Birdseye have been truncated by pronounced erosion (*see* below). The same contact in the Black River Valley is also documented by truncation of underlying units, although the amount of truncation becomes less and the contact is more subtle where the basal Watertown grainstone is not readily apparent. Thus although the boundary at the top of the Birdseye has since previously been abandoned, there is the possibility (and precedence) for establishing the lithostratigraphic and biostratigraphic boundary at the base of the Watertown and not at its summit (*see* below).

Outside of the New York type region, the complexity of this argument has promulgated an unenthusiastic view of New York nomenclature and there have been attempts to apply, rework, or otherwise abandon the type region classifications for lithostratigraphic and time-rock or stage nomenclature (which means the abandonment of Kay's time-rock terms: Rocklandian, Kirkfieldian, and Shermanian) in lieu of a newer stage called the Chatfieldian (Leslie & Bergström, 1995; *see* Figure 2). The establishment of this particular stage also has furthered an additional complexity to the argument: in that the base of the Chatfieldian Stage is drawn at the position of the Millbrig K-Bentonite. This particular volcanic ash is not yet firmly established in New York State (*see* Mitchell *et al*., 2004; and Brett *et al*., 2004 for this particular controversy). Yet regardless of its placement, if the type New York stage classifications can be highlighted and identified on a more regional scale in relationship to the Millbrig using other chronostratigraphic methods such as sequence stratigraphy these stage classifications have precedence over the Chatfieldian.

General Stratigraphy

The Black River Group, as originally defined in the lower Black River Valley near Watertown, northwestern New York State (*see* Figure 8), is about 60 m thick and has been subdivided on the basis of

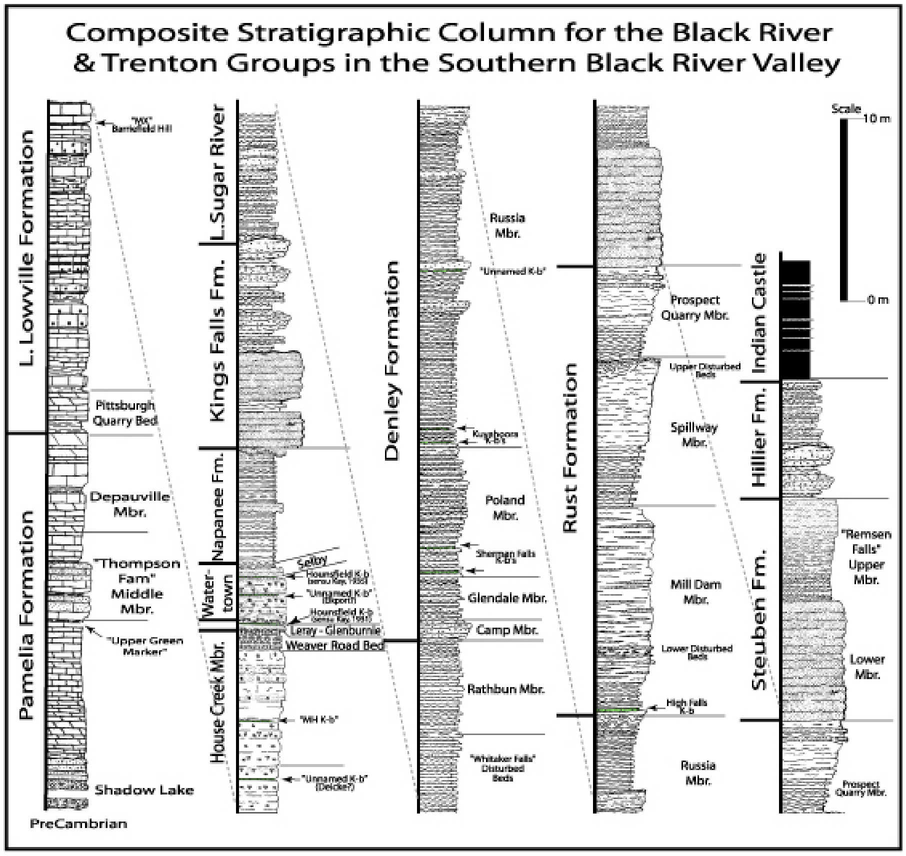

Figure 9 - Standard composite stratigraphic section for the Black River and Trenton Groups in Northwestern New York State. Important stratigraphic units and distinctive markers including K-bentonites, and other unique correlation horizons are identified in the sections (compiled from: Ross, 1968; Brett & Baird, 2002; Brett, et al., 2004, Cornell et al., 2004).

lithostratigraphy into three formations: the Pamelia (40 m), Lowville (17 m) and the Watertown limestone or upper Chaumont (3 m) formations (Fisher, 1977) (Figure 9). Additional units were subsequently named based primarily on biostratigraphic criteria and came to take on a chronostratigraphic connotation. Hence the term Chaumont, originally used for medium to thick bedded carbonates with distinct faunas typified by a wide variety of brachiopods, echinoderms, and cephalopods, was replaced in part by the lithostratigraphic term Watertown, and was modified to be used as the Chaumontian Stage (Kay, 1937). This term is not used at present, but it is included in the uppermost portion of the Turinian below the transition to the Rocklandian. As is illustrated here, many of the early workers including Kay (1937), and Young (1943) considered many of the Black River Groups (and overlying Trenton) to be both biostratigraphically and lithostratigraphically controlled thus often equated time-rock terms and rock-terms and used them interchangeably. Although this practice was discouraged these authors considered many of these units to be fairly isochronous and layer-cake in style (*see* Figure 10: model A) over fairly broad distances and therefore using either a lithostratigraphic or biostratigraphic classification was often sufficient for correlation.

Unfortunately without the specific “total package” understanding of these early workers, combined with loosely-constrained lithostratigraphic boundaries between these major Black River sub-units, many subsequent authors focused on gross lithology (and not fossil content) to establish spatial and temporal relationships of each of these stratigraphic units. Moreover without a major effort to trace out these units from their type sections using substantive internal marker horizons, only minor progress was made in correlating larger facies packages (Johnsen, 1971; Walker, 1973; Textoris, 1977). Thus without fairly well-correlated sections that were constrained using some geochronologic system, the facies distribution of Black River strata came to be viewed as a mosaic of large-scale diachronous facies, with little to no continuity of individual facies over long distances (Figure 10: model B, C). Thus, these strata were once viewed as time transgressive facies, with component facies grading laterally to each other in a large-scale,

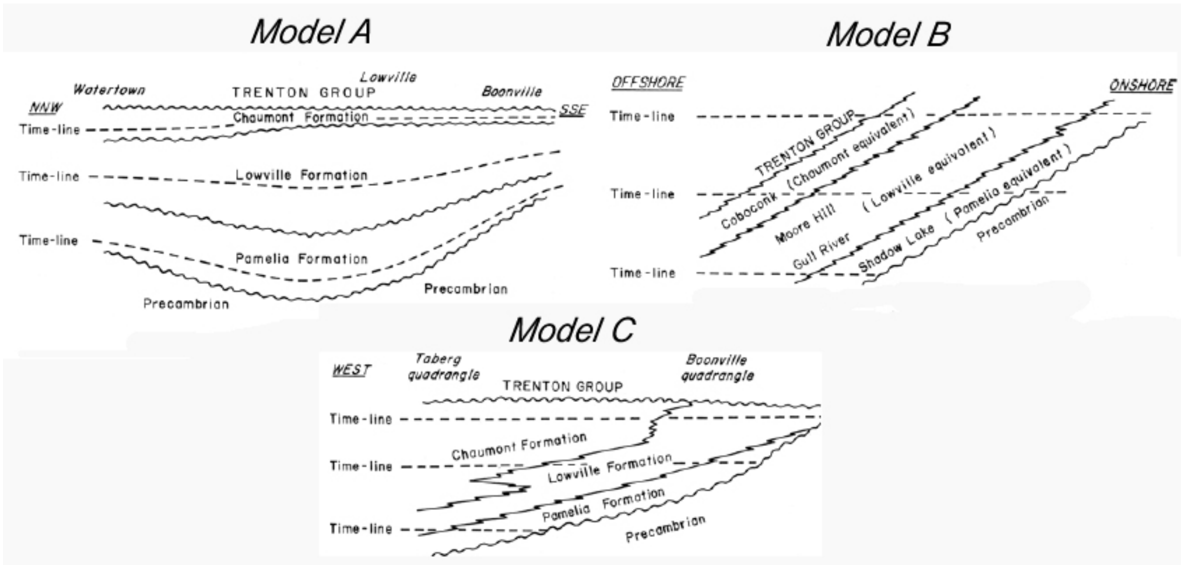

Figure 10 - Depositional models used to explain the relationship of individual Black River and Lower Trenton units. Model A: Time parallel facies model (after Kay, 1937; Young, 1943), Models B & C: Diachronous progradational facies models (B: after Winder, 1960 and C: after Fisher, 1962 and Walker, 1973)

time-transgressive Waltherian manner (*see* Figure 10). These models have been tested in recent studies (Cornell, 2001a, and b) on the basis of sequence stratigraphy, K-bentonite and other marker horizon correlations. Although facies patterns do show evidence for lateral time-equivalent variation, it is possible to differentiate individual units lithologically and biostratigraphically based on the re-evaluation of "type-section units." These studies suggest that isochronous horizons run parallel to formational contacts without crossing them (Figure 10, model A) and in some cases formational and intraformational contacts are even isochronous surfaces. Clearly, however there is also evidence for diachronous facies patterns (i.e. progradation/retrogradation) within formations and ultimately help in the recognition of shallowing and deepening between successive packages.

<u>Pamelia Formation.</u> In northwestern New York, the Pamelia Formation is the lowest unit exposed in the Upper Ordovician. The Pamelia itself is early Mohawkian in age and rests either on Precambrian Grenville basement rocks or on Cambrian-Lower Ordovician sandstones and carbonates. In either case, the contact at the base of the Pamelia represents a significant unconformity ranging in age from nearly 600 million years to less than 10 million years. Described from exposures in the Town of Pamelia in Jefferson County, the Pamelia Formation itself is a fairly significant unit and can be divided into several internal units. These are generally referred to as the lower, middle, and upper members. In total the Pamelia in the type region reaches ~45 m in more northwestern New York and Ontario and thin substantially to the southeast into the southern Black River Valley where the thickness varies but is generally less than 20 m and thins to extinction in some regions south of Boonville, only to reappear in a few locations in the western Mohawk Valley.

In terms of composition, the Pamelia is a heterolithic unit containing a wide range of siliciclastic influenced carbonates. Where clearly exposed the lower member is generally in contact with either the Theresa Formation or the Precambrian, nonetheless in either case, the base is consistently dominated by quartz-rich sands with fragments of underlying beds imbedded in its matrix. In the upper part of the lower Pamelia, massive planar & crinkly laminated dove-gray micritic limestones appear and are often interbedded with dolomitic caps and green shaly micrites that show an overall shallowing upward pattern to dolostone-dominated beds and mud-cracked argillaceous micrites near the top. The middle member of the

Pamelia formation, informally referred to as the "Thompson Farm Member" for the exposures in the Thompson Farm Quarry at Lafargeville, Jefferson County, New York, is distinctive from the lower or upper Pamelia in that although it contains some quartz-rich beds at its base, it does transition upward into fairly coarse-grained micrite-dominated carbonates exhibiting wackestone to packstone textures in some beds. This unit is clearly a deeper water facies and is recognized through the appearance of bryozoans, large-cephalopods, brachiopods, a variety of bivalves, and some crinoid ossicles and occasionally an entire crinoid itself. Moreover, in places this unit may contain *Tetradium* coral colony thickets and an occasional stromatoporoid and in the southern Black River Valley also contains intervals of ooid grainstones. As near the summit of the lower Pamelia, the upper portion of the "Thompson Farm" contains a distinctive argillaceous green dolomitic limestone containing small well-rounded, poly-crystalline quartz grains and sometimes a fairly red or magenta shale sitting on top of a subjacent dolomitic limestone. This bed has been referred to as the "Upper Green Marker" in southeastern Ontario or the Pittsburgh Quarry bed informally by Conkin & Conkin (1991). The uppermost contact of this bed often contains some evidence for truncation at the base of the overlying upper or "Depauville Member." The uppermost member of the Pamelia, referred to again as the "Depauville Member" after the Depauville Waterlime of previous geologists. Historically, the Depauville was described as gradational out of the middle member and developed from the first massive earthy weathering dolomitic limestone up through the uppermost dolomitic limestone. Although this expanded Depauville does contain all of the main dolomitic beds, there are significant patterns (shallowing-upward followed by deepening-upward) within the expanded Depauville. Moreover it also contains several key contacts and correlatable marker horizons internally. Thus herein the Depauville is restricted to the most substantial massive-bedded dolostone with extensive vug development sitting on the Pittsburgh Quarry Bed. The vugs are typically filled with celestite (strontium sulfate) nodules and / or other evaporitic pseudomorphs. In some case, although no mineral remains, it is not uncommon in some bedding cross-sections to see evidence for moldic preservation of cubic or bladed evaporite textures similar to halite or gypsum impressions.

<u>Lowville Formation.</u> Previously, the Lowville was subdivided into a lower member, about 13 m of sparsely fossiliferous, fenestral dove gray micrites and minor shales and an upper House Creek member (which in its upper part is intermittently called the Leray Limestone) comprising about 4-5 m of medium gray burrow mottled, locally cherty wackestone to packstone. The Lower Lowville, is the classic "Birdseye" Limestone of 19th century geologists and exhibits numerous fenestral spar filled voids, laminated micrites, often containing the vertical burrow *Phytopsis tubulosum* which is the characteristic trace fossil in this particular unit. Other faunal components are minimal but generally the unit contains small gastropods, ostracods, fragmented bryozoan colonies, and a few *Bathyurus* sp. trilobites all of which suggest a fairly restricted and slightly saline environment of deposition. The latter Lowville or House Creek Member is clearly a more off-shore yet still of a shallow depositional environment. The unit commonly contains abundant *Tetradium*, large tabulate corals, stromatoporoids, and associations of gastropods, bivalves, and a few brachiopods. Overall, the biofacies and the lithologic character of the House Creek indicate an overall shallowing upward from the base. A particularly important marker horizon in the uppermost House Creek Member contains a thin interval of mud-cracked gray-yellow weathering shales, platy micrites and domal stromatolites, designated informally as the "Weaver Road." In most of the Black River Valley successions this unit is present although it may lose the characteristic domal stromatolites in lieu of less prominent LLH style or laterally linked types in the southern Black River Valley. It does however become truncated below the overlying unit southward.

<u>Chaumont Formation.</u> A 0.5-l m bed of crinoid gastropod grainstone and welded coral brachiopod richpackstone separates the Weaver Road bed shales from a higher 0.5 to 1 m shaly nodular limestone interval with a distinctive fauna. This couplet of beds occurs in the position of Kay's (1929; 1931) Leray Limestone which formed the basal portion of the Chaumont Formation. Thus as a member, the Leray would sit beneath the Glenburnie member and overly unconformably the underlying House Creek. Although not well developed in most localities south east of Brownville, New York, the shaly-nodular facies of the Glenburnie does show up in the Martinsburg region and exhibits some of the faunal similarities to the underlying Leray although it contains many more bryozoans. This shaly interval is preserved in southeastern Ontario, and in a few portions of New York, but is generally thin to absent south of Watertown but reappears in the vicinity of Martinsburg. This particular member was also described (sensu Kay, 1931) to contain the Hounsfield K-bentonite, which subsequently in Kay (1935) was relocated

to a higher stratigraphic position, although indeed a K-bentonite does occur in this position. The upper contact of the Glenburnie Member with the overlying Watertown is demarcated by a fairly sharp, yet in weathered outcrops, subtle transition into much more massive beds of the Watertown which may contain at its base a series of intraclastic conglomeratic beds indicating that there may be erosional truncation of underlying units. In the type area, a 2 to 3.5 m horizon of massive, ledge-forming condensed packstone to micritic grainstone interval occurs above the Glenburnie/Leray interval. With a distinctive coral, algae, brachiopod, and cephalopod fauna, this unit was termed the Watertown Limestone by Cushing et al., (1910) and reserves this designation today. The Watertown has generally been assigned to the Black River Group as it appears lithologically more similar (more carbonate, less shale) to the underlying strata than the overlying Trenton. Conkin and Conkin (1991), however, have argued that this unit should be assigned to the Trenton Group based on faunal and lithologic evidence and this corroborates other long-standing evidence discussed previously. When traced into the southern Black River and Mohawk River Valleys, the Watertown becomes substantially thinner, more fine-grained, and even becomes intertidal as at Ingham Mills and a few other western Mohawk Valley localities. Lithologically, these Watertown–equivalent limestones appear very similar in appearance to the lower Lowville and have been referred to as such in many localities, yet by tracing out individual packages and marker horizons it is possible to differentiate Watertown equivalent shallow water facies. This lateral facies change is identifiable and follows the diachronous concepts of Fisher (1962) and Walker (1973) but only within high-order cycles, and not on the formation-scale as previously proposed.

Selby Formation. In the Watertown, New York to Kingston, Ontario area the massive and often coarse grained Watertown limestone is overlain by about 0.5 to 3 m of nodular bedded, bituminous dark gray packstone, termed the Selby. Although a complete outcrop section of the Selby is lacking in Ontario, this interval becomes substantially reduced and thins to a single condensed packstone to wackestone observed in localities south of Martinsburg, New York. This particular unit exhibits typical faunas found in the Watertown, but becomes dominated and relatively enriched in both planispiral and orthoconic cephalopods, numerous brachiopod species, and occasional crinoids. A prominent K-bentonite horizon is located at the contact with the underlying Watertown, and in 1935, Kay reassessed the position of his Hounsfield K-bentonite to this horizon in an attempt to establish more regional correlations with sections in the Mohawk Valley. The uppermost contact of the Selby is again a fairly sharp lithologic break with the sudden change to thin shaly interbedded wackestones and calcisiltites of the overlying Napanee Formation. This surface shows evidence for substantial deepening across the contact and will be discussed as a flooding surface later.

Napanee Formation. As in most sections, especially to the northwest and southeast of Middleville, this interval is overlain by deep-water thin bedded, platy calcisiltites, and interbedded wackestones and shales assigned to the Napanee Formation. Collectively, the Selby and Napanee by definition have been assigned to the Rocklandian Stage, based on lithologic and biostratigraphic evidence, and form the base of the Trenton Group. Despite its rather distinctive lithologic appearance, the Napanee was first defined on the basis of its faunal composition. Originally, Kay (1937) designated this unit (the upper member of the Rockland) on the presence of the brachiopod *Triplesia cuspidata* (Hall). These *Triplesia*-bearing strata were superjacent to the Selby Limestone and subjacent to the Hull (or Kings Falls Limestones). In addition, the Napanee tends to be heavily dominated by the small orthid brachiopod *Paucicrura (Dalmanella) rogata* (Sardeson).

However, lithologically, the Napanee Formation can be recognized independently of faunal evidence based on its stratigraphic position above the underlying Selby and beneath the overlying crinoidal grainstones of the Kings Falls Formation. Unfortunately, due to the softer weathering shaly interbeds the Napanee-Selby contact is poorly exposed in most regions where the Napanee is weathered back along terrace plains and covered by colluvial debris as in Roaring Brook. Some key sections can be observed including those at Sugar River near Boonville.

Kings Falls Formation. Above the Napanee Formation, limestones once again become coarse-grained and dominated by brachiopod-crinoid grainstones and skeletal packstones. The contact between the Napanee and the overlying Kings Falls Formation (Kirkfieldian Stage) is very sharp and often dramatic in weathered outcrop sections. The Kings Falls weathers in stark contrast to the underlying Napanee and in some places channels and erosional features indicate that the Napanee is truncated at the base of the Kings Falls. This is evident at several localities including those at Ingham Mills, and Middleville in the Mohawk Valley and at Boonville on the Sugar River where there is evidence for the truncation of underlying beds. As a unit, the Kings Falls Formation is dominated by medium to thick bedded crinoidal grainstones especially in the base. These basal beds transition upward into mega-rippled grainstones and packstones that become substantially interbedded by additional thicknesses of shale and slightly deeper water taxa. As such, the Kings Falls is composed of several shallowing-upward cycles of thick ledges of calcarenite or crinoidal grainstones and brachiopod coquinas. Interbedded with these very coarse-grained lithologies are a series of thin dark gray to black shales. Most often the grainstone beds show excellently preserved hummocky cross-stratification, and large-scale pararipples on the top surfaces of many beds. As mentioned the distinctive component of this particular interval is the dominance of echinoderm skeletal remains. In the Ontario region, this unit is equivalent to the Kirkfield Formation (uppermost Bobcaygeon) and is demonstrated to contain many classes of primitive and not-so primitive echinoderms including asteroids, ophioroids, cystoids, edrioasteroids, carpoids, crinoids, etc. While most of these have not been recorded from entire specimens in New York, there are numerous plates found that are thought to be derived from many of these taxa.

<u>Sugar River Formation.</u> In the Black River Valley region and in the nearby West Canada Creek Valley, the upper Kings Falls transitions rapidly into thinly interbedded shales, and fine-grainstones, with thin nodular wackestones and calcisiltite stringers common. This lithology is more similar to the Napanee Formation subjacent to the underlying Kings Falls, than it is to the distinctively coarse-grained Kings Falls Formation. Although the Sugar River is dominantly interbedded shales and fine grainstones, it becomes distinctively more nodular and finer-grained. Near the top of the succession, the Sugar River becomes more medium-bedded, calcilutite facies with fewer fossiliferous horizons. This upper unit has been recognized as a separate unit and named the Rathbun Member of the Sugar River Formation. To the southeast into West Canada Creek and Mohawk Valley regions, the Rathbun Member is traceable into a coarser-grained, crinoidal packstone lithology with minor grainstone stringers, as is exposed south of Middleville, New York. This upper member shows evidence of thinning, and shallowing into the Middleville region and then transitions into the shale dominated facies of the lower Flat Creek Shales east of Little Falls. The Sugar River carries a distinctive fauna that enables its easy field recognition wherever it is exposed. Generally speaking this interval is established by the occurrence of two major faunal elements in the New York sections. First, the blind lace-collared trilobite *Cryptolithus tesselatus* (Green) is present in large numbers at several horizons but most predominantly in the medial Sugar River. Second, the largest occurrence of the gumdrop-shaped bryozoan, *Prasopora* is within the lower portion of the Sugar River. Both of these faunal elements are abundant and often compose very dense accumulations on some bedding planes.

<u>Denley Formation.</u> Although it has been recognized as a very distinctive stratigraphic interval**,** it has undergone a substantial transition in stratigraphic nomenclature. Initially, Kay applied the term Denmark Formation to the fossiliferous interval of dominantly fine-grained carbonates and interbedded shales, overlying the Sugar River Formation in the central Black River Valley in and around the town of Denmark. In 1968, Kay relegated the term Denmark for more biostratigraphic purposes and applied the term Denley Formation to denote rock terms. In common usage, the Denley Formation is recognized lithologically and on the occurrence of two distinctive marker horizons at the lower and upper contacts of the formation. These two marker horizons include the basal "*Trocholites* beds", or Camp Member, and the High Falls K-bentonite from Trenton Falls. Thus delineated, the Denley Formation now encompasses only a portion of the original Denmark Formation, and now contains two members: the lower or Poland Member and the upper or Russia Member. Collectively grouped the Denley is generally considered to be composed dominantly of fossiliferous nodular fine-grained limestones which show evidence for shallowing-upward cycle development and numerous internal and distinctive marker horizons. Generally speaking the Poland Member is defined between the basal City Brook or Camp Member and the lowermost Kuyahoora K-bentonite. As bracketed**,** this interval is a relatively condensed, nodular calcilutite interval containing abundant *Trocholites* and transitions to more tabular-bedded barren calcilutites which are interbedded with buff weathering shales and thin coquinal beds of the Glendale or upper Poland. This unit is moderately fossiliferous with species of trilobites especially common. The planar middle to upper Poland interval grades upward into coarser-grained, rippled calcarenites that often contain intraclastic conglomerates showing significant evidence for early cementation and development of storm clasts. This interval is

relatively resistant to weathering, and makes up the ledge forming the cap of many waterfalls in the region including Sherman Fall at Trenton Falls.

The upper member of the Denley Formation was originally named by Kay in (1943) for the upper calcilutite-dominated facies above the Poland and below the coarse-grained carbonates of the overlying Cobourg (now called Rust Formation). This particular member of the Denley is very distinct in that it shows several well-developed cycles with shallowing upward motifs. Moreover, as the lower series of cycles show slight facies differences between the lower, middle, and uppermost beds, three sub-members have been designated. Although these will not be considered here, the uppermost submember of the Russia is represents a very distinctive change in lithofacies with the basal 2 m composed of very tabular and even-bedded calcilutites that grade upward into more standard nodular wackestones and packstones showing well-developed bioturbation. The cap of this interval is rather distinctive in that the uppermost beds show evidence of soft-sediment deformation. This unit is sometimes mistakenly referred to as the "Lower Disturbed Zone" which occurs in the base of the overlying Rust Formation. This bed is immediately overlain by a condensed, quartz-rich, crinoidal grainstone that truncates portions of the underlying interval.

Rust Formation. As originally described, Kay (1943) introduced the term Rust for the lower member of the Cobourg Formation. Recognizing two divisions within the Cobourg in West Canada Creek Valley, he named the lower unit Rust, after exposures of the well-known quarries on the William Rust Farm just east of Trenton Falls. Up to that point, the interval near the top of the Trenton was not differentiated, but because of its rather distinctive facies, it was separated from the overlying Steuben or upper Cobourg. Recent review by Brett & Baird, (2002) has redefined the nature of the Rust such that, in contrast to the finer-grained Russia below, this unit is generally considered to be composed of nodular to wavy-bedded coarse-grained packstones to grainstones which can internally be divided into three lithologic divisions. These divisions although very similar in overall composition, again can be distinguished as shallowing-upward packages with similar motifs to the underlying Russia. However because of its distinctive internal stratigraphic intervals, as well as its coarser lithology, the Rust is very easily recognized in the West Canada Creek Valley. Further to the east, however, correlations show that this interval becomes substantially modified to the east and transitions into a condensed turbidite-shale succession and then finally into basinal black shale facies. Internally, there are many interesting horizons useful in correlation and include K-bentonites, and deformed or "disturbed" horizons.

In addition to lithologic characteristics, the Rust Formation is one of the best-known stratigraphic intervals in the Trenton Group because of the well-known collection of trilobites, echinoderms, and other fossil specimens collected from this interval, by William Rust and Charles Doolittle Walcott. With respect to biostratigraphic zonation, the base of the Rust Formation is now clearly established as containing the boundary between the *Corynoides americanus* and *Orthograptus ruedemanni* graptolite zones.

Steuben Formation. The Steuben Formation represents the uppermost unit exposed in the Trenton of the Mohawk Valley and was used for the upper portion of the Cobourg of Kay (1943). This unit is exceptionally coarse-grained and has been recognized since the very first geological surveys of New York. The Steuben Formation is represented by massively bedded, very coarse-grained skeletal grainstones that commonly show evidence for cross-bedding and sediment reworking. In the Trenton Falls region the Steuben is represented by two main coarse-grained intervals interrupted by a thin interval of shaly, wavy-bedded wackestones and packstones. Thus delineated, the Steuben is separated into two main units separated by the middle shaly interval. Due to its association with the underlying Prospect Quarry Member of the Rust Formation, the Lower Member of the Steuben appears as the upper half of a shallowing or coarsening-upward succession. The Upper Member, referred to here as the Remsen Falls Member displays a similar upward-coarsening pattern from the base of the middle shaly interval.

Modern Stratigraphy: Correlation Tools

Historically, correlations have been hampered by the low resolution of conodont biostratigraphy, a lack of readily identifiable index taxa, and the practice of broad-scale lithologic grouping. Although useful for mapping stratigraphic units, the practice of lumping units by lithology, without regard to internal marker units, has been demonstrated to be an extremely simplifying practice that has sometimes overlooked major surfaces and unconformities. Modern stratigraphers spend much time looking for and identifying these particular intervals so as to establish a more precise-level of correlation that provides for higher spatial-temporal resolution of rock units.

Within the Black River and Trenton groups, the presence of a number of K-bentonites and other event beds distributed throughout the succession provide the basis for establishing more regional correlations within previously recognized "lithostratigraphic or rock-units". K-bentonites are generally deposited instantaneously, with respect to geologic time, over a very large region and are profoundly useful in establishing chronostratigraphic correlations (thereby when they are correlated) which help to delineate time-rock units independent of biostratigraphy and lithology. Other types of unique marker beds used to establish regional high-resolution correlations include a variety of event beds such as sharp facies disjunctions, erosional contacts, mineralized surfaces and hardgrounds, unique taphonomic beds, stromatolite beds, epiboles, unique lithologic units, and distinctive patterns of bedding. Given the development of the practice of stratigraphic packaging using predictive patterns of deepening and shallowing of facies (i.e. sequence stratigraphy), it is now possible in association with the variety of correlation tools mentioned, to provide the discriminating stratigrapher a variety of tools from which detailed high-resolution correlations can be based (*see* Figure 2).

SEQUENCE STRATIGRAPHY

Recent studies of graptolite and conodont biostratigraphy, and K-bentonite and isotope geochemistry have resulted in a refined chronostratigraphic framework for the Black River and Trenton groups in central New York State (Goldman et al, 1994; Mitchell et al, 1994; Baird et al, 1992; Brett & Baird, 2002; Baird & Brett, 2002), and southern Ontario (Armstrong, 1997; Melchin et al, 1994; Brookfield & Brett, 1988; Cornell, 2001). Progress has also been made toward developing a high-resolution sequence stratigraphy within the framework of chemo- and biostratigraphy (*see* Figure 2). Nonetheless, while the Black River/Trenton-equivalent interval has recently become the focus of study from the standpoint of sequence stratigraphy and cyclicity in other parts of North America; similar studies have only recently been undertaken in the type area. Holland and Patzkowsky (1996, 1998) have identified and correlated six Mohawkian depositional sequences (M1-M6) throughout the southern Appalachians and eastern Midcontinent area (*see* Figure 2). Until recently, these sequences and their components have not been recognized or studied in the classic New York-Ontario sections (Cornell, 2001; Brett *et al*. 2004; Figure 11).

Within this context, the following sequence stratigraphic interpretations have been delineated for the Black River to Trenton interval (Figure 11). As is illustrated, the Black River Group forms three entire depositional sequences and portions of a fourth. These correspond to the M2, through M4 sequences of Holland & Patzkowsky on the basis of lithostratigraphic, biostratigraphic, and chemostratigraphic correlations. Moreover, although K-bentonite correlations are controversial, these also can be demonstrated to show at least local correlations. Emphasis here will be placed on the uppermost complete 3^{rd}-order depositional sequence in the Black River. This sequence, referred to as the M4 sequence begins with the lower Lowville deepening up to a maximum flooding surface at the base of the House Creek Member. Subsequently, the offshore bioturbated wackestones of the House Creek shallow upward through the stromatolitic and mudcracked Weaver Road Beds before being capped by the thin grainstone-wackestone couplet – nodular carbonate of the Leray/Glenburnie. Although very thin in nature, and similar to some individual parasequences, this interval is interpreted as a mini condensed sequence that may have developed and been truncated prior to the development of the larger transgressive phase of the overlying M5 (*see* Figure 2).

The overlying Watertown, Selby, and Napanee form yet another sequence commencing with the deepening and condensing-upward Watertown and Selby formations. These formations are much more fossiliferous than the underlying Lowville, yet are transitional between the House Creek and the Napanee above. The traditional Black River/Trenton lithostratigraphic boundary, at the top of the Selby, is interpreted as a flooding surface separating the TST below from the HST above. In fact, this flooding surface is very distinctive and represents an important correlation surface across the outcrop belt and in other outcrop regions as well. In some cases in the Mohawk Valley, this sequence boundary becomes composite where the sequence boundary at the base of the Kings Falls comes down to merge with this one through truncation of the Napanee Formation. Based on the position of the M4 sequence, this boundary interval sequence forms a portion of the M5 sequence (of Holland & Patzkowsky, 1996), as it is recognized that there are several sequences represented within the entire M5 of these authors. This sequence is herein referred to as the M5A sequence.

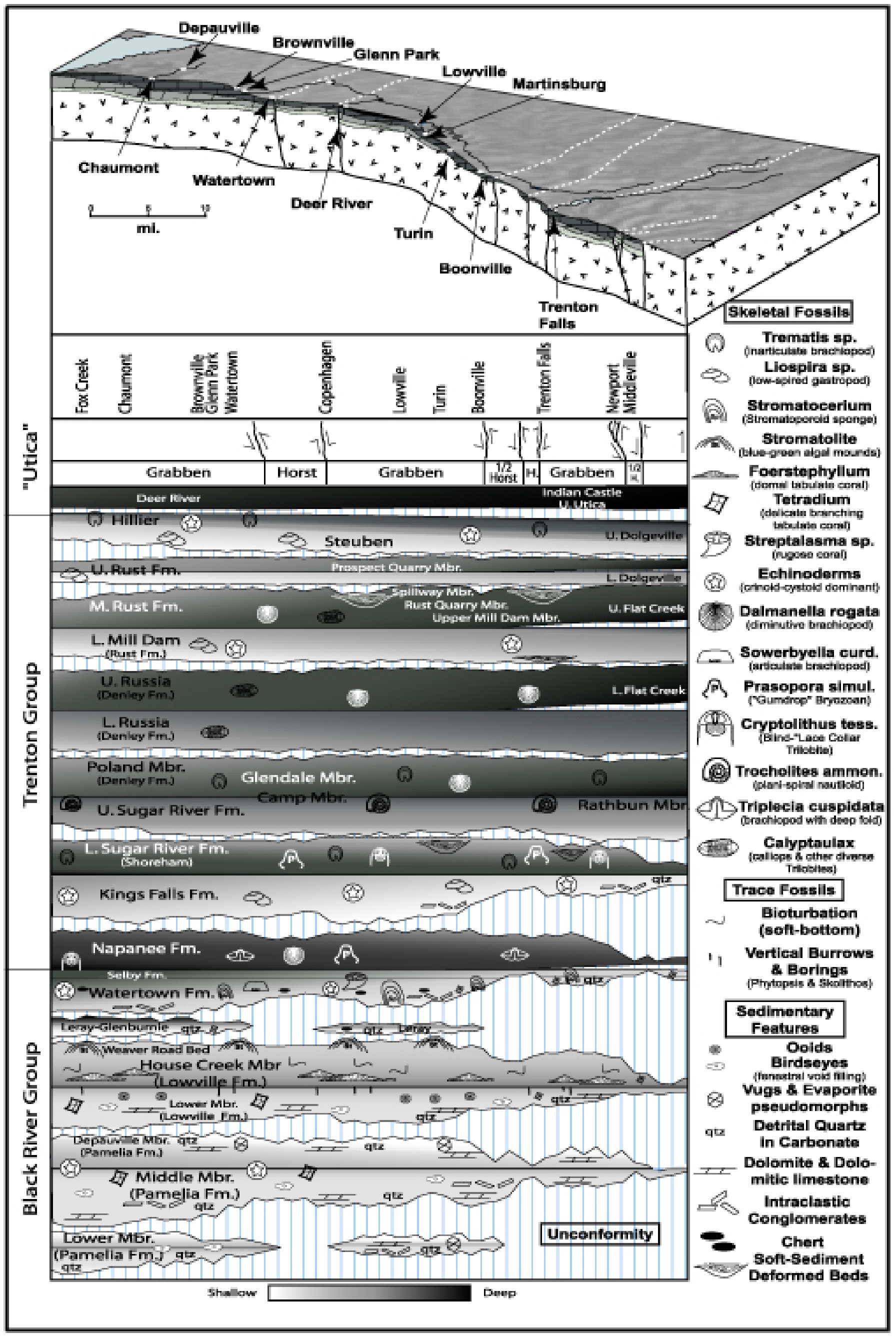

Figure 11- Sequence Stratigraphic Interpretations for the West Canada Creek to Black River Valley Region of New York State. Illustrated are the position of key stratigraphic units, lithologic marker horizons, and characteristic body and trace fossils. Vertical lineation represents unconformity, which in most cases represents the development of sequence boundaries, although in some cases maximum flooding surfaces demonstrate minor truncation and fairly sharp discontinuities. Sequence delineation follows Brett et al., (2004), and Cornell, (2005).

The base of the Kings Falls terminates the M5A sequence with fairly substantial erosion in some localities. Coupled with the lower Sugar River Formation, the Kings Falls forms an entire sequence (M5B) before it transitions upward into the Rathbun Member of the Sugar River Formation. Thus constructed this M5B sequence begins in the top of the Kings Falls which has historically been referred to the Kirkfieldian Stage and terminates in the Shermanian at the summit of the lower Sugar River Formation. The extremely coarse grainstones of the Kings Falls fine and deepen upward to more shaly interbedded nodular limestones rich in *Prasopora* bryozoans and the trilobite *Cryptolithus*. Thus it is interpreted that the deeper water nature of the Sugar River indicates that the contact is a maximum flooding type contact. Subsequently, the Sugar River itself begins to shallow upward to its summit where it transitions into the Rathbun Member at the base of the next overlying sequence.

The last sequence of the M5 succession of Holland & Patzkowsky (1996) is entirely within the basal Shermanian Stage. Composed of the Rathbun and basal Denley, Poland Member, this sequence is much less distinctive than either of the underlying sequences. However changes in sedimentary and biofacies character enable the recognition of important patterns that help delineate this sequence. Important observations in this sequence include the widespread occurrence of channel-filled slump-folded strata just below the upper Sugar River which shallows upward to a subtle discontinuity surface with slightly shallower water facies juxtaposed on the top. This is interpreted as a sequence boundary although it was probably not a type I or exposed sequence boundary. The maximum flooding interval is recorded by the condensed Camp or "*Trocholites*" beds. These beds are sharply overlain by much deeper and finer grained lutites of the Upper Poland.

The uppermost sequence of the Trenton Group, using the Holland & Patzkowsky model, is the M6 sequence. On the broad-scale, this sequence is divisible into 3 subordinate sequences (M6A, B, and C) with each entirely within the middle to upper Shermanian Stage. The M6A, sequence is entirely developed within the Upper Denley or Russia Member and represents fairly deep water mixed siliciclastic carbonate assemblages. Due to the widespread and correlatable nature of many of the Russia cycles, it is possible to pick out more subtle progradational and retrogradational patterns in lithology that enable development of the sequence interpretation. The M6B sequence represents a fairly substantial change in the overall M6 sequence. The development of the fairly coarse-grained Rust Formation, indicates a fairly substantial shallowing over the Russia and given the nature of the contact, there is evidence for truncation at the base of the Rust. Thus a sequence boundary is established at this position. The distinct occurrence of deformed or "disturbed" horizons within this sequence is indicative of widespread tectonic activity (earthquakes) which impacted the sediments deposited on the sea-floor. Within the Rust, cycles in the Mill Dam member show an upward deepening pattern to a substantial flooding surface in the upper Mill Dam and lower Rust Quarry interval. This particular drowning event is responsible for some of the most spectacular preservation in the trilobites and echinoderms collected by Walcott and Rust. The final sequence of the M6 is the M6C. This particular interval is substantially thin on the shelf, but in the nearby basin thickens. This sequence is entirely constructed in the Prospect Quarry Member of the Rust and documents a fairly substantial amount of deformed and brecciated channel-fill fabrics. Like then condensed mini-sequence at the base of the M5A sequence, the M6C sequence also demonstrates some remarkable condensation, again on the shelf while expanding out in the foreland.

The shaly nodular wackestones and packstones of the Prospect Quarry are sharply overlain by a much more substantial body of coarse-grained crinoid-brachiopod grainstone of the Steuben Formation. Due to its biostratigraphic assessment and the comparable sequence stratigraphic position, the Steuben is classified as equivalent to the basal portion of the C1 sequence, although it is not yet Cincinnatian in age. Nonetheless, even more pronounced than the shallowing in the Mill Dam member of the Rust, the Steuben Formation records a very sharp change to extremely shallow-shoal type conditions from the West Canada Creek Valley northwestward through the Black River Valley. As this unit is traced further northwest into Canada it does become less massive and becomes interbedded with minor shales. Regardless, upward from the base, the Steuben shows several large-scale and many small-scale cycles illustrating a deepening upward pattern to a maximum flooding interval. The overlying Cincinnatian aged black shales of the Indian Castle sit atop a fairly major unconformity interpreted by most authors as a drowning-type unconformity.

REGIONAL SYNTHESIS

As mentioned above, the transition from Black River to Trenton is coincident with the onset of the Vermontian Phase of the Taconic Orogeny. During this time, the cratonic margin was again influenced by plate tectonic processes to establish yet another renewed phase of tectonism. Within the successions of the Black River Valley through the Mohawk Valley, these strata exhibit an abrupt transition from the fairly passive continental margin to the tectonically active, rapidly subsiding foreland basin. Although this transition has historically been considered to be straight forward and coincident with major deepening in the "Trenton transgression," our recent work in the upper Black River to lower Trenton groups demonstrate a more complex pattern of activation. Using the spatially and temporally constrained framework of sequences first delineated by Brett and colleagues (2004), it has been possible to establish patterns of uplift and subsidence within the Trenton Shelf region. When patterns of deposition are investigated in relationship to structural features of the region, it appears that activation of the Vermontian phase was manifest on the shelf through the reactivation of ancestral faults resulting in a series of uplifted blocks and associated down-dropped regions (*see* Figure 12).

Using onlap/offlap patterns, the amount of incision/truncation across sequence boundaries, and within sequence facies change, it is possible to establish a chronology of events leading up to the main phase of subsidence in the New York region associated with the cratonward migration of the northern extension of the Martinsburg foredeep basin (Champlain Trough). The Black River platform in New York has been studied in detail by Young (1943) and it had been determined to have extended from southern Ontario into south-central New York State. Through this region, the platform was characterized by a region of major subsidence in the Kingston-New York region. Although never formally named the "Rideau Trough" was a region of subsidence showing onlap patterns in the lower Black River both: to the north and west (onto the Frontenac Arch and Canadian Shield), and to the south and east onto the Adirondack Arch of Kay (1937). This pattern of onlap is clearly documented in Figure 11 where the M2, M3 and M4 sequences clearly show the progressive shift of sedimentary facies to the southeast so that by the M4 highstand (U. Lowville, House Creek), most of the Adirondack Arch had become submerged and deposition occurred in a fairly shallow lagoonal system from Ontario through eastern New York. Paleocurrent analysis of aligned orthoconic nautiloids yield a northeast-southwest trend indicating that a minor barrier may have been present in the eastern Mohawk Valley (Canajoharie Arch) and may have acted as a protective barrier but it was of minimal extent during the M4 highstand.

Near the end of the M4 highstand and during the M5A sequence boundary, there is evidence that the majority of the region became exposed during sea-level lowering, however, in the southern Black River Valley to Mohawk Valley region, this lowstand was accentuated by the uplift of horst structures allowing for the truncation and karstification of the M4-M5 sequence boundary. In some regions in the southern Black River Valley and in the Mohawk Valley, especially at Inghams Mills, New York it is possible to identify several meters of truncation at the base of the M4-M5 sequence boundary.

ROAD LOG FOR CLASSIC LOCALITIES OF THE BLACK RIVER AND TRENTON GROUPS (UPPER ORDOVICIAN) IN THE BLACK RIVER VALLEY: REVISITED THROUGH TRADITIONAL AND SEQUENCE STRATIGRAPHY (Part I)

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	SUNY Oswego campus entrance and intersection with US104. Head East on US104.
20.8	20.8	Maple View. Continue East on US104, past the overpass for I81.
22.3	1.5	Turn left (N) onto I81.
51.7	29.4	Exit 42 (Adams Center). Turn right (E) on Rte 177.
52.4	0.7	Intersection with US11. Continue East on Rte 177.
77.3	24.9	Junction with NY12. Turn right (S) on NY12.

80.1	2.8	Lowville. Turn right (S) on NY12&26.
80.4	0.3	Turn left (E) on River St.
81.0	0.6	Turn right (N) on State St. and take and immediate left into parking area for Joan's School of Gymnastics. Walk down to exposures in stream bed.

STOP 1: LOWVILLE, NY, MILL CREEK GORGE
[UTM 18T 0461517 4847863]

This locality represents the type locality for the Lowville Formation and although the original quarry section is flooded (it is just across the road from the gorge) the gorge offers excellent exposures and weathered outcrops for the review of sedimentary structures and fossiliferous strata. The section starts in the Pamelia Formation and intermittently in the banks at the lower end of the gorge. It is possible to observe the uppermost Pamelia Formation. Exposures of the "Depauville" Member and overlying Pittsburgh Quarry Bed of Conkin and Conkin (1991) can be identified.

Up in the narrows of the gorge, exposures of the lower Lowville Member or "Birdseye" of early workers can be seen in a series of well-developed meter-scale cycles, each with a flooding surface at their base and a shallowing upward motif. The Upper Lowville, or House Creek Member (of Walker, 1973), is fairly well exposed in the southern wall of the gorge and although it can be difficult to access, this member of the Lowville is distinctively less argillaceous and more massively bedded. It is composed almost exclusively of bioturbated wackestones with minor stringers of intraclastic packstones, and a few ribbon-bedded horizons showing some ooid development. Overall the House Creek represents protected, but marine conditions typical of shallow water lagoon areas. Although it is much more subtle and is occasionally recognized as one more cycle of the Lowville, the uppermost beds of the House Creek, or the "Weaver Road Beds" show the distinctive development of domal stromatolites in fairly argillaceous platy micrites. Overlying the Lowville Formation is a series of more massive-bedded limestones previously referred to the Watertown Limestone. Although much of this particular interval is extremely cherty, the basal few feet are less so and are fairly coarse-grained. In stark contrast to the underlying lithologies, the basal few feet of the upper massive unit are characterized by large colonies of *Tetradium*, crinoids, brachiopods, a variety of high-spired gastropods and numerous other taxa. Given the faunal and lithologic composition of this bed and through correlations with sections to the northwest in the Watertown area, it is believed that this thin interval of beds represents what is left of the Leray-Glenburnie interval. Without the occurrence of the Hounsfield K-bentonite (*sensu* Kay, 1931), in this succession it is surmised that the Glenburnie and the K-bentonite have been eroded by the overlying sequence boundary and were removed prior to the deposition of the Watertown. The Watertown is the last well-displayed unit is this portion of Mill Creek. Just below the Rt. 12-26 bridge over Mill Creek, the south wall of the gorge displays the massively bedded and chert-rich equivalent of the Watertown Limestone. Approximately 4 m of Watertown are exposed and are typically well bioturbated wackestone to packstone facies. Like the House Creek, the Watertown contains many coral taxa, sponges and stromatoporoids, and more diverse brachiopod and crinoid assemblages. Based both on lithologic and faunal patterns, the Watertown itself demonstrates a significant and even more substantial upward deepening pattern. Although many of the internal cycles have become amalgamated, they can often be identified through the recognition of firmground and hardground surfaces. These surfaces become especially prominent near the top of the succession where they show evidence for phosphate, pyrite, glauconitic staining, and may even contain K-bentonites sitting on them. Moreover, these surfaces often are accentuated by hard-substrate borings and in some instances have been found to contain crinoid & bryozoan holdfasts indicating that these surfaces were well cemented on the sea-floor prior to their final burial. Collectively then, the Watertown itself is clearly an upward deepening unit and although it does not represent the deepest portion of the sequence, it is the deepest facies within the entire Black River to this point.

Near the top of the gorge section below the bridge, a small poorly-exposed succession of rubbly weathering wackestone-packstone lithology stands in stark contrast to the underlying massive bioturbated beds of the Watertown. This particular interval becomes very condensed in places and shows evidence for sediment starvation through the accumulation of numerous current aligned cephalopods (with a variety of taphonomic signatures), and a range of diminutive orthid, and strophomenid brachiopods (*Dalmanella*,

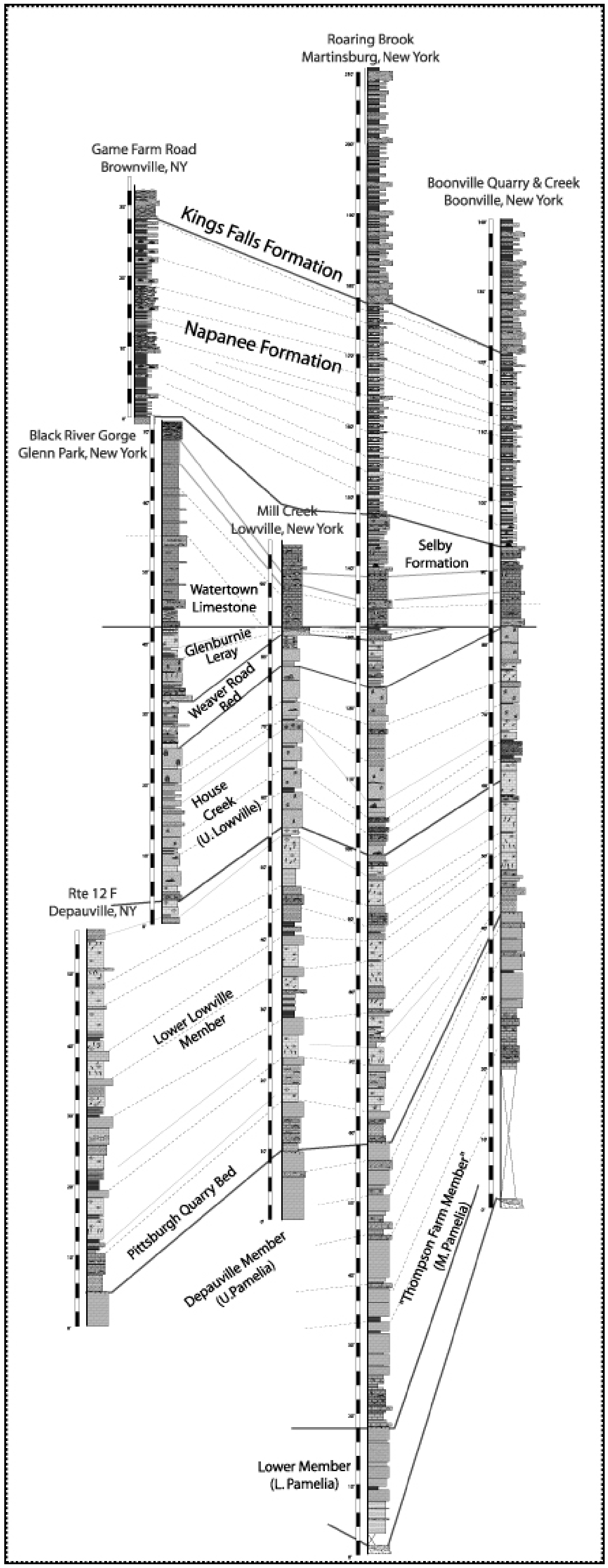

Figure 12 - Correlation of stratigraphic sections for Upper Ordovician field trips (Parts I and II).

Sowerbyella respectively). This unit is referred to as the Selby, and is included in the Rocklandian Stage. Lithologically and faunally, the Selby is transitional out of the Watertown into the overlying Napanee Formation. Unfortunately, poor exposures above the bridge do not permit the observation of the upper Selby to Napanee interval.

81.1	0.1	Turn right back onto State St. At STOP sign turn left onto River Rd.
81.8	0.7	Turn right on Markowski Rd.
82.7	0.9	Turn left (S) on NY12. Note Precambrian outliers on West side of highway.
85.3	2.6	Turn right into Linda's Roaring Brook Restaurant. Park. Proceed to Roaring Brook and walk downstream to Precambrian contact.

STOP 2a: EAST MARTINSBURG, NY, ROARING BROOK, LOWER SECTION
[UTM 18T 0465811 4843142]

Just east of Route 12 on the lower section of Roaring Brook is the unconformable contact of the Pamelia Formation with the underlying Precambrian Grenville Gneiss. Although exposure here is covered for some distance above the contact, a few feet of the Pamelia are exposed and document the transition from a siliciclastic sandy (dominated by quartz and feldspar fragments) carbonate upward into more medium-bedded argillaceous limestones. Although very few fossils are found in this unit, occasionally a few gastropods, ostracods, and small fragments of bryozoans are represented. These are commonly of poor preservational style and many are coated with greenish glauconitic material. Although the depositional history of this lowermost Pamelia is not well constrained, it is very similar to strata exposed in southeastern Ontario where this particular interval is located on the top of the Shadow Lake and is time equivalent with the basal Gull River Formation. Here in New York, the basal Pamelia sandy limestones are thought to represent the deepest phase of the Shadow Lake or Rideau sequence and overall remain fairly shallow but mudcracks a few feet off the basal sands suggest that this may be a shallowing succession off the basal transgressive surface. That is the red and green shaly carbonates of the Shadow Lake (which are not developed here in the central Black River Valley) were probably deposited during the transgressive phase, while the overlying lower Pamelia was deposited during the maximum inundation of that particular sea-level rise event.

85.3	0.0	Walk upstream to additional exposures of the Pamelia and of the Lowville and Watertown.

STOP 2b: EAST MARTINSBURG, NY, ROARING BROOK, MIDDLE SECTION

Up section, the remainder of the Black River Group is exposed through the top of the Watertown Limestone. The middle-upper division of the Pamelia is well represented and overall represents one transgressive-regressive sequence. Although not formally named, the middle or "Thompson Farm" member is a distinctive facies of the Pamelia and is well developed through the transition from laminated argillaceous micrites and dolomicrites to more medium-bedded (and distinctively less argillaceous) wackestones containing a fairly restricted fauna but occasionally contains a few echinoderm fossils (crinoids), a few large strophomenid brachiopods, very large endoceratid cephalopods, and several trilobite taxa. In contrast to the more restricted ostracod-bivavle-gastropod assemblages of both the lower and upper Pamelia, the middle member is characteristically more normal marine in faunal composition. Although it is still considered a fairly shallow peritidal deposit, the presence of these faunal elements indicates that the middle Pamelia was deposited in deeper water. A fairly rapid and substantial transition upward into the upper Pamelia or "Depauville Waterlime" of early workers is documented by the transition back into dolomitic limestones and sandy dolomites characteristic of this interval. In addition, distinctive sedimentary features suggest that this particular interval shows evidence of hypersaline conditions. In some beds evaporite crystal pseudomorphs are distinguished as are the development of large vugs containing celestite and/or gypsum. Although it is clear that many of these have been recrystallized during later diagenesis, the moldic preservation of some of these crystal faces indicates that these environs may have been highly saline at the time of deposition.

Above the Pamelia, both the lower and upper Lowville members are well exposed and with some investigation are readily compared to sections visited at Mill Creek in Lowville. Unfortunately, due to extreme solution along joints in the stream bed, many of the sedimentary and faunal details of the Lowville are obscured but many are still readily observable. In this region, the Lowville is approximately 25 m thick. Near the top of the succession, an argillaceous, mudcracked interval at the base of the Watertown is recognized. Although there are no large domal stromatolites (as in the northern Black River Valley), there is evidence for low, broad stromatolitic mounds (LLH style) in this interval. These stromatolites border on thrombolitic textures where they are burrowed and are evidently grazed.

Continuing upstream are exposures of the Watertown Limestone. Again as below, some local small-scale structures are developed in these exposures. Although most appear to be related to differential compaction over shallow Precambrian basement features, it is also evident in this section that there are minor faults that offset strata locally. It is not clear at this time if these are related to more recent uplift associated with Neogene to recent Adirondack tilting or if these are representative of local faulting at or near the time of deposition in the Upper Ordovician. Nonetheless, the Watertown and overlying Selby Formation represent a continuous deepening showing the upward transition from typical bioturbated cherty wackestones and packstones of the upper Watertown into more nodular, rubbly weathering Selby. The Watertown itself is substantially thinner than the Watertown in the type region and it is believed that only the upper portion of the Watertown is here by overlap. At the top of the section, as is the case in most of the region, the top Selby forms a broad terrace where the overlying shaly bedded Napanee Formation is eroded back almost a half mile. Although we will not trek overland to the next exposure, the basal beds of the Napanee are intermittently exposed just upstream.

	0.0	Return to vehicles
89.3	4.0	From parking lot, turn right (W) onto Tiffany Rd.
90.5	1.2	Turn right (NW) on Glendale Rd.
95.7	5.2	Turn right into Whittaker Park.
96.1	0.4	Drive into park to designated parking area. Excellent view of the Adirondacks to the East. LUNCH. Proceed down trail to exposures in stream. We will examine the lower exposures first, then walk upstream and exit at the bridge on Glendale Rd. Vans will ferry participants back to vehicles.

STOP 3: WHITAKER FALLS PARK, MARTINSBURG, NY, ROARING BROOK UPPER SECTION [UTM 0463947 4842616]

Upstream from the previous stop, we start just above the lower waterfall in Whitaker Falls. Although we could travel downstream, for the sake of time, we will focus on the interval from the lower falls upward through the upper Trenton. This portion of Roaring Brook exposes the entirety of the Shermanian from the Sugar River at the base through the Denley, Rust, and finally the Steuben formations. We will not see the Kings Falls or Napanee up close at this locality although they are exposed downstream (we will see them at the STOP 4 in Boonville, NY). Exposed in Whitaker Falls Park are three closely spaced waterfalls in this portion of the gorge. Just below the lip of the first waterfall, and exposed in the south wall of the gorge are the uppermost beds of the Kings Falls (forming the floor below the falls) and the Lower Sugar River formation. The shaly nature of the Lower Sugar River allows for it to be slightly less resistant and allows for the notch under the waterfall. Also exposed in the wall of the gorge at this point are at least two recessive weathering notches and a deformed channel. The channel and the layer above the waterfall show evidence for brecciated fabrics suggesting syn-sedimentary movement of materials on the sea-floor after some period of lithification. Although it is not yet clear as to the regional relationship of this feature, both of these features are found in many local sections throughout the region and are prominent horizons useful for correlation.

The lowermost waterfall is capped by the Rathbun Member of the Sugar River Formation and near the top of the middle waterfall is the transition to the upper Poland. Just above the lip of the middle falls and below the bench leading to the upper waterfall is the transition to the Upper Denley (Russia Member). Within the face of the third falls are the well developed cycles of the Russia and the basal Mill Dam Member of the Rust Formation. The fairly coarse cycle caps of the Mill Dam help to hold up the face of the falls and are exposed for some distance upstream where excellent large-scale ripples are well developed on them. Like the coarse-grained Kings Falls some distance downstream, the Mill Dam in this region was substantially shallower than the underlying Denley Formation, although it too shows an upward deepening pattern.

Continuing upstream some distance exposures in the wall of the gorge afford views of the Rust Formation and its members. In the walls of the gorge before the Glendale Road Bridge, exposures of the Mill Dam and Spillway members show characteristic coarser-grained facies with well preserved faunas, extremely large brachiopods and some crinoids. Another deformed interval is located here and is tentatively correlated with the uppermost Spillway member slump structure at Trenton Falls. Just above the bridge, exposures of the Prospect Quarry member give way to the basal coarse-grained Steuben Formation. Although an additional 25-30 meters of section are exposed upstream near the Martinsburg town garage (as is the contact with the overlying black shales), owing to time constraints we will defer their investigation at this time.

96.5	0.4	Return to Park entrance and turn left on Glendale Rd.
99.4	2.9	Turn right (S) on NY12. Again, note Precambrian outliers and roadcuts on East and West sides of highway.
105.6	6.2	Hell's Kitchen Rd. to right. Do not turn. Notice canal locks for the Black River section of the Erie Canal.
110.2	4.6	Four canal locks on left (E).
111.1	0.9	Pull into Barrett Quarry on left. HARD HATS! Time permitting, we will also examine the exposures along NY12, both N and S of the quarry and in the Sugar River.

STOP 4: BOONVILLE NY, BARRETT MATERIALS AND PAVING QUARRY AND SUGAR RIVER SECTIONS
[UTM 18T 0473787 4819422]

Just north of the village of Boonville, NY on the north shore of Sugar River sections are exposed in the Barrett Materials and Paving Quarry and in the adjacent stream section to the south of the quarry and along route 12. Entrance is by permission only and hard hats and steel-toed shoes must be worn at all times. Although exposures in this section extend down to the base of the Pamelia, we will focus on the uppermost Black River Group exposed in the quarry. In the older northwest corner of the pit and along the west wall of the pit, exposures start in the lower Lowville and extend upward through an abbreviated House Creek section where the Watertown Limestone, or "7' tier" of quarry operators, immediately overlies the middle House Creek with possible unconformity. Although the 7' tier is dominantly a single massive bed composed of fairly coarse micritic wackestone, in the face of the unit to the southwestern corner of the pit, the upper massive bed displays a prominent recessive notch containing a prominent clay horizon. Although it still remains to be chemically fingerprinted it sits in the same position as a K-bentonite approximately 1.5 meters below the top of the Watertown at Brownville gorge. In the older portions of the quarry, the K-bentonite has been squeezed out and it is difficult to make out, as is the case in most of the weathered stream sections.

Although no higher exposures are afforded in the quarry–the cap of the quarry wall is the Watertown–additional exposures in the stream bed allow for the continuation of the stratigraphic section. Just to the south of the quarry in the stream bed and southern bank of the Sugar River, the upper contact of the

Watertown with the Selby is observed. The massive Watertown is easily recognized in the bed of the stream and along the southern bank a several decimeters of thin-bedded rubbly weathering limestone is exposed below the typical shaly-interbedded Napanee Formation. At this particular location, only the basal 2-3m of the Napanee is exposed in the outer bend of the stream. Upstream, above the NY 12 bridge, additional sections are exposed from the Napanee up through the King Falls Formation. W will not walk upstream to see them in weathered section, but we will look at the same interval in the road cuts on both the north and southwest side of the highway. Here in the road cuts are the uppermost ~5 meters of Napanee followed by ~3 meters of Kings Falls. Although not in the best position for weathering, the outcrop shows the sharp contact of the Kings Falls on the underlying Napanee. In fact, the contact shows evidence for channeling into the underlying Napanee at the M5A-M5B sequence boundary. Additional exposures are visible upstream, including the type succession of the Sugar River Formation, and additional units up through the upper Trenton.

		Return to vehicles, turn left (S) on NY12 and proceed to Boonville.
114.1	3.0	Turn right (W) on E. Schulyer St.
114.3	0.2	Proceed to traffic light (follow signs to Rte 46/294).
114.4	0.1	Turn right (W) on Rte. 294 and follow to West Leyden.
121.2	6.8	Turn left (S) on NY26.
131.6	10.4	Stokes Center. Turn left (W) onto Lee Center Rd.
139.2	7.6	Intersection with NY69, just SW of Taberg. Turn right (NW).
147.3	8.1	Intersection with NY13. Turn left at light.
147.4	0.1	Turn right at light. Continue on NY69.
174.6	27.2	Mexico. Turn left (W) at intersection with US104.
191.1	16.5	Campus entrance. End of field trip.

ROAD LOG FOR CLASSIC LOCALITIES OF THE BLACK RIVER AND TRENTON GROUPS (UPPER ORDOVICIAN) IN THE BLACK RIVER VALLEY: REVISITED THROUGH TRADITIONAL AND SEQUENCE STRATIGRAPHY (Part II)

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	SUNY Oswego campus entrance and intersection with US104. Head East on US104.
20.8	20.8	Maple View, NY. Continue East on U104, past the overpass for I81.
22.3	1.5	Turn left (N) onto I81.
62.2	39.9	Turn off on Exit 41. Take US12 (N) to Depauville, NY.
70.4	8.2	Cross Rte 180, Gunn's Corners.
75.0	4.6	Depauville, NY
75.4	0.4	Continue to roadcuts just N or village. Park (with care!).

STOP 1: DEPAUVILLE, NY, ROADCUT ON NY12
[UTM 18T 0414505 4888355]

The village of Depauville contains an excellent succession of the Black River Group and the contact of the basal Pamelia with the underlying Theresa Formation (Upper Cambrian). The Pamelia is mostly exposed in the waterfall succession of the Chaumont River (and just to the west along Buttermilk Creek) where nearly 9m of Pamelia are exposed below the county route 170 (Depauville Road) bridge. We will not look at this succession and instead we will focus on the relatively new outcrop on the northern limits of the village along NY 12. Just above the Depauville VillageMarket, and below the town garage, a fairly long stratigraphic succession of nearly 20m is visible along both sides of the roadcut. We will use this section to discuss the long-standing debate concerning the description of the Pamelia-Lowville contact. In previous classifications the contact was drawn either at the highest dolomitic limestone or at the lowest occurrence of relatively pure "birdseye." In this locality a prominent succession of cyclically-bedded limestones shows an upward deepening pattern and is based at a very distinctive quartz-rich sandy bed. This bed, referred to as the Pittsburgh Quarry Bed by Conkin and Conkin (1991), is a prominent marker from southern Black River region through southwestern Ontario and even in the Ottawa region. Given the prominence of this marker and the very distinctive pattern of cyclically-bedded carbonates above, an argument can be made for placing the upper contact at this stratigraphic position, thus including it in the base of the Lowville Formation. Although some dolomitic limestones do occur above this position, they are few and dominated by the relatively pure micrites of the Lowville. The remainder of the succession through the cap of the outcrop is referred to the lower member of the Lowville Formation and shows the overall transgressive pattern of the M4 sequence. Near the top of the succession a noticeable reentrant occurs and contains a K-bentonite which correlates with a prominent K-bentonite in Ontario referred to as the MH K-bentonite. The recognition of this reentrant is aided by its position just below the sharp flooding surface to the overlying House Creek Member. The terrace at the top of the outcrop is developed in the basal House Creek with the remainder of the House Creek exposed in the bluffs to the north, west, and south of the village.

75.7	0.3	Continue N on NY 12 to Old Town Spring Rd.
80.6	4.9	Turn around at Old Town Spring Rd. at crest of hill. Return through village of Depauville on NY 12 and continue (SE) to Gunn's Corners (Rte 180). Turn left.
83.5	2.9	Proceed to roadcut on W side of highway. Park.

STOP 2: LIMERICK, NY, ROADCUT ON RTE 180, JUST EAST OF GUFFINS CREEK
[UTM 18T 0418165 4878863]

This stop is intended for quick look at facies of the upper Lowville House Creek that have not been observed up to this point. In the small outcrop on the northwest side of the road, a succession of ~4-5 meters of strata are exposed. The base of the succession is dominated by bioturbated wackestones containing various algae, corals, and a variety of brachiopods, the House Creek is clearly more normal marine than any of the underlying units in the Black River. Nearer the top of the section, relatively coarse-grained facies are developed and in places show evidence of cross-bedding and ooids. Intermixed with *Stromatocerium* stromatoporoids, the ooids reflect an important shift in the deposition of the House Creek suggesting a shallowing which is terminated in the Weaver Road beds (observed at STOP 3).

84.4	0.9	Turn around. Proceed N on Rte 180 to Depauville Rd (Rte 54). Turn right (S).
87.9	3.5	Continue S on Rte 54 to parking area on right side of highway at top of hill crest near Brownville Fish Game Club. Roadcut is just before entrance to the Club.

STOP 3: BROWNVILLE, NY, ROADCUT ON RTE 54 NEAR BROWNVILLE FISH-GAME CLUB
[UTM 18T 0420019 4876663]

It is unfortunate that the type section of Kay's Hounsfield K-bentonites (*sensu* 1931 and 1935; *see* STOP 4) is overgrown and poorly exposed. To alleviate this difficulty this newly widened outcrop section located on the northeast side of Rte 54 just NW of the intersection with Mullin Road, is used as a new reference section for the uppermost Lowville through basal Trenton succession. This outcrop displays the Weaver Road beds at the base of the outcrop, the overlying Leray and Glenburnie interval and the overlying Watertown Formation. In this particular succession, K-bentonites are located at several horizons, including near the base of the Glenburnie (Hounsfield *sensu* Kay, 1931) and in the overlying Watertown near the base of the Selby (*sensu* Kay, 1935). The stratigraphic succession shows evidence for mudcracked argillaceous ribbon micrites sharply overlain by the somewhat sandy and coarse-grained basal Leray. The Leray in this succession is relatively thin (~1.5m) and is overlain by the rubbly weathering Glenburnie. This particular facies is very subtle in relationship to the Watertown and is most often included within it. However, as this unit is traced into southern Ontario, the Glenburnie interval becomes distinctively more shale rich and equates to a 1.5m thick shale in the Kingston region located below the much coarser grained and cross-bedded Watertown in that region. In New York, the Watertown is less of a grainstone, but is dominated by extremely bioturbated wackestone-packstone facies. The remainder of the succession displays the cherty facies of the Watertown that generally is developed along bedding planes and within burrows. The Watertown becomes distinctively more nodular towards its top and at the position of a prominent reentrant the facies changes to a much more condensed and darker grey, rubbly weathering unit referred to as the Selby. Only a minor portion of the Selby is exposed, as it and the overlying Napanee Formation were relatively nonresistant to glacial processes that were active in this region.

88.0	0.1	Return to Rte 54, turn right (W) on Mullin Rd.
89.8	1.8	Proceed to STOP sign at ‘T’ intersection with Cemetery Rd. (Mullin Rd. becomes Game Farm Rd.). Continue S on Game Farm Rd.
90.4	0.6	Park in entrance to Browville Transfer Station (only on Sundays) to your right near the top of the hill.

STOP 4: BROWNVILLE, NY, FARR’S QUARRY AND ROADCUT ON WEST SIDE OF GAME FARM ROAD
[UTM 18T 0419612 4873577]

As mentioned at STOP 3, the original type-section for Kay’s Hounsfield K-bentonites is poorly exposed, and despite a covered interval in the quarry and adjacent lower ditch, the stratigraphic succession continues up the hill through the Napanee and into the lowermost Kings Falls Formation which forms the cap of the hill below the junk yard. This stratigraphic section is visited to continue and to illustrate the relatively close stratigraphic proximity of the new Fish-Game Club locality just northwest of here (STOP 3). Although the contact of the Selby and Napanee is poorly exposed during late winter and early spring it is possible to get near the base of the Napanee to measure and describe ~8-9m of section. In this region, the Napanee is fairly fossiliferous and contains several identifiable taxa including many key brachiopods, some crinoids, and several trilobites. Overall, the Napanee is defined as interbedded shales, calcisiltites, and thin-bedded wackestones. Nearer the top of the ditch the Napanee begins to shallow upward showing more thin-bedded coquinal packstones and grainstones. The base of the overlying Kings Falls is represented by the very sharp contact to the medium-bedded and rippled grainstones of the Kings Falls.

91.3	0.9	Continue S on Game Farm Rd. to Rte 54 (Brownville).
91.7	0.4	Turn right (S) on Brown Blvd. Proceed to intersection with Main St. Turn left (E) on Main St.
93.4	1.7	Turn right in Glen Park HydroPower Fishing Access site. [LUNCH]. CAUTION: river levels can rise 5’ in 5 minutes; seek higher ground if you hear SIREN.

STOP 5: GLEN PARK, NY, GLEN PARK HYDROPOWER FISHING ACCESS SITE, BLACK RIVER GORGE
[UTM 18T 0423760 4872110]

In the recent past, the Black River Gorge has become synonymous with rafting/kayaking and hydropower. In the narrow part of the gorge, from just below Brownville village through the parking area just above the hydroplant diversion channel, the Black River Gorge descends through about 25m of section ranging from the Lower Lowville at the western end of the gorge through the Selby at the eastern end. On the opposite or south side of the river, exposures of the overlying Napanee are exposed intermittently in the bluff. Although we saw the same succession at STOP 3, the exposures here in the gorge afford an excellent opportunity to look at well-weathered blocks and enigmatic structural features that indicate there may be some offset between the two sides of the river. Below the fish ladder (accessed via the gravel road that runs along the edge of the water diversion channel) it is possible to observe the beautifully exposed Weaver Road Beds. If water level is low enough, we will walk on them and study their complex structures. The field trip will "officially' end here; however, interested people may wish to continue on to outcrops at King's Falls. Road log is continued below.

93.4	0.0	Leave parking area. Turn right on Main St.
95.0	1.6	Turn left on NY 12F (Bridge St.).
95.1	0.1	Turn left (E) on NY 12F. Proceed to Watertown, NY
97.1	2.0	Cross over I81. Continue E on NY 12F.
100.4	3.3	Jcin with NY12/US11, turn right (S). Get in left lane.
100.5	0.1	Turn left (S) on NY 12.
102.7	2.2	Turn right. Stay on NY 12S.
115.3	12.6	Copenhagen, NY. Intersection with Mechanic St.
115.4	0.1	Turn left (NE) on Cataract St. (becomes Deer River Rd.).
118.6	3.2	Turn right on Old State Rd. (gravel surface).
118.8	0.2	Proceed to parking area. Enter Deer River and walk upstream.

STOP 6: COPENHAGEN, NY, KING'S FALLS ON DEER RIVER
[UTM 18T 0448106 486287]

Like many localities in this region, the Deer River is characterized by several well-developed waterfall successions. Among them are the beautiful sections at Kings Falls just north of the Village of Deer River on NY 26. Accessed via Old State Road, this succession begins in the upper portion of the Napanee and continues upstream to the summit of Kings Falls which exposes the upper Sugar River and basal Denley (Denmark of Kay for the region just south of this locality). Several fascinating structures are found at this locality. In contrast to the block-faulted regions of the Mohawk Valley, the Tug Hill Plateau is often considered as a tectonically quiescent carbonate platform during deposition of the Trenton. This stop, in addition to the section along Roaring Brook with the S-fold below the lower falls, and the brecciated horizons, as well as the structures at the Glen Park section, shows that this region may not have been so stable. Much interest has recently been generated in Trenton/Black River structures due to recent natural gas exploration in the Finger Lakes Region of New York State. The producing horizons are in dolomitized grabens where hydrothermal fluids have migrated along the bounding faults. While, the fault kinematics and mineralization is different here at King's Falls, it does appear that they have acted as fluid conduits. Beneath the Old State Road bridge are excellent exposures of basal King's Falls with thick-bedded calcarenites and grainstones, some showing large symmetric ripples. Downstream the lower contact with the Napanee Formation is exposed. Where the river turns to the west the beds on the north side are dipping towards the northeast, while the beds on the south side have experienced minor faulting. A small thrust fault ramps up towards the east with drag folding of the footwall. Just downstream of this thrust is a small (offset ~0.2m) NNE-trending normal fault with rollover folding on the hanging wall. While these structures are minor, there is significant mineralization in the outcrop including thick calcite veins, an iron-rich or possibly sulphide residue on joint surfaces, and significant dissolution of brachiopods in certain beds. This evidence indicates that these faults have acted as fluid conduits. Farther upstream at the base of King's Falls is another fault. This fault strikes roughly east-west, dipping 40ºS with the beds of the hanging wall drag-folded into the fault trace. The hanging wall is composed entirely of the King's Falls Formation. There is also significant mineralization near this fault in the form of thick veins and rare vugs. The King's Falls-Sugar River contact is exposed about halfway up the falls.

TRIP END.

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Trip B-2

VARIATIONS IN L- AND S-TECTONITE ON THE NORTHERN BOUNARY OF THE PISECO LAKE SHEAR ZONE, ADIRONDACK MOUNTAINS, NEW YORK

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INTRODUCTION

The subject of this field trip is the variation in deformation fabrics along the northern margin of the Piseco Lake shear zone (Gates et al., 2004) with special emphasis on the development of L-tectonite domains at various scales. To some extent, but on a much more regional scale, this was also the emphasis of field guide for the 76th NYSGA field conference (Valentino et al., 2004). The current field guide covers a geology field trip that is a continuation of the earlier field trip with overlap of a few field stops.

The Piseco Lake shear zone (Figure 1) is a major Grenvillian structure that is 10 to 20 kilometers wide and strikes generally east -west in the southern Adirondacks. Kintematic analysis in the zone demonstrated dominately low-angle sinistral shear (Gates et al., 2004). For the current study, an area of 42 square kilometers was mapped in detail and the study area spans the northern limit of the Piseco Lake zone in the area of the West Canada Creek basin (Figures 1 & 2). The objective of this study was to document the detailed rock fabric variation within the shear zone, within the transition zone and within the wall rocks to the shear zone. This study was designed to better understand the strain and metamorphic history associated with this major Adirondack structure, and document the geographic distribution of L- and L-S tectonites that were previously reported (McLelland, 1984; Chiarenzelli et al., 2000; Gates et al., 2004).

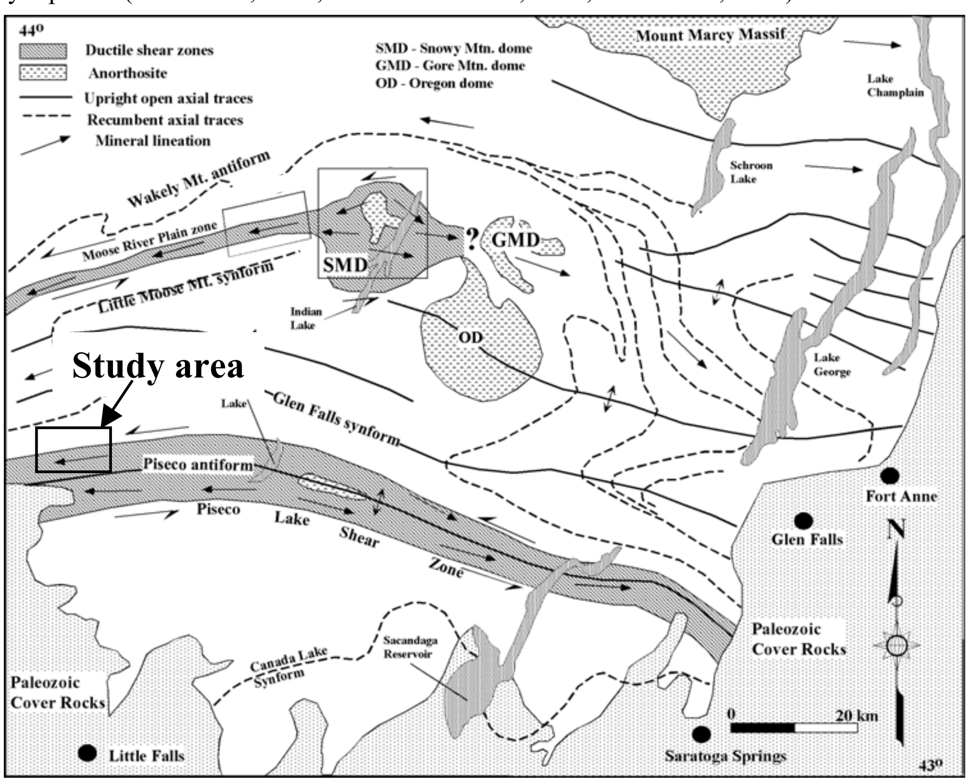


Figure 1 - Schematic structure map of the southern Adirondacks showing the Piseco Lake shear zone (modified from Chiarenzelli, et al., 2000 and Gates et al., 2004). Study area shown with box.

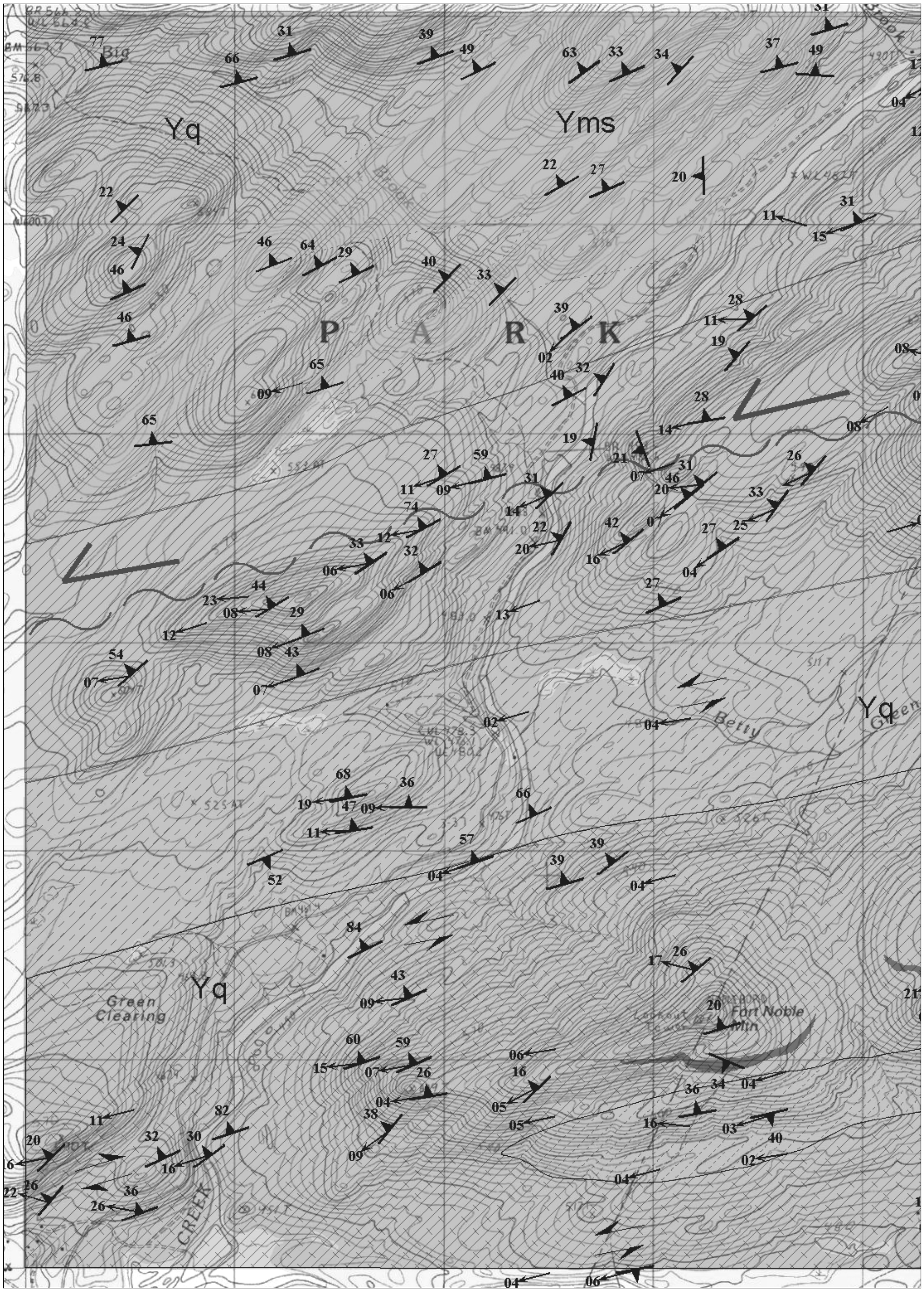


Figure 2 - Bedrock geologic map of the West Canada Creek basin in the southwestern Adirondacks. The map area spans the northern limit of the Piseco Lake shear zone. The base map is a provisional USGS metric topographic map with a 1 km grid. The next two pages show the eastern extension of the map area and the map explanation respectively.

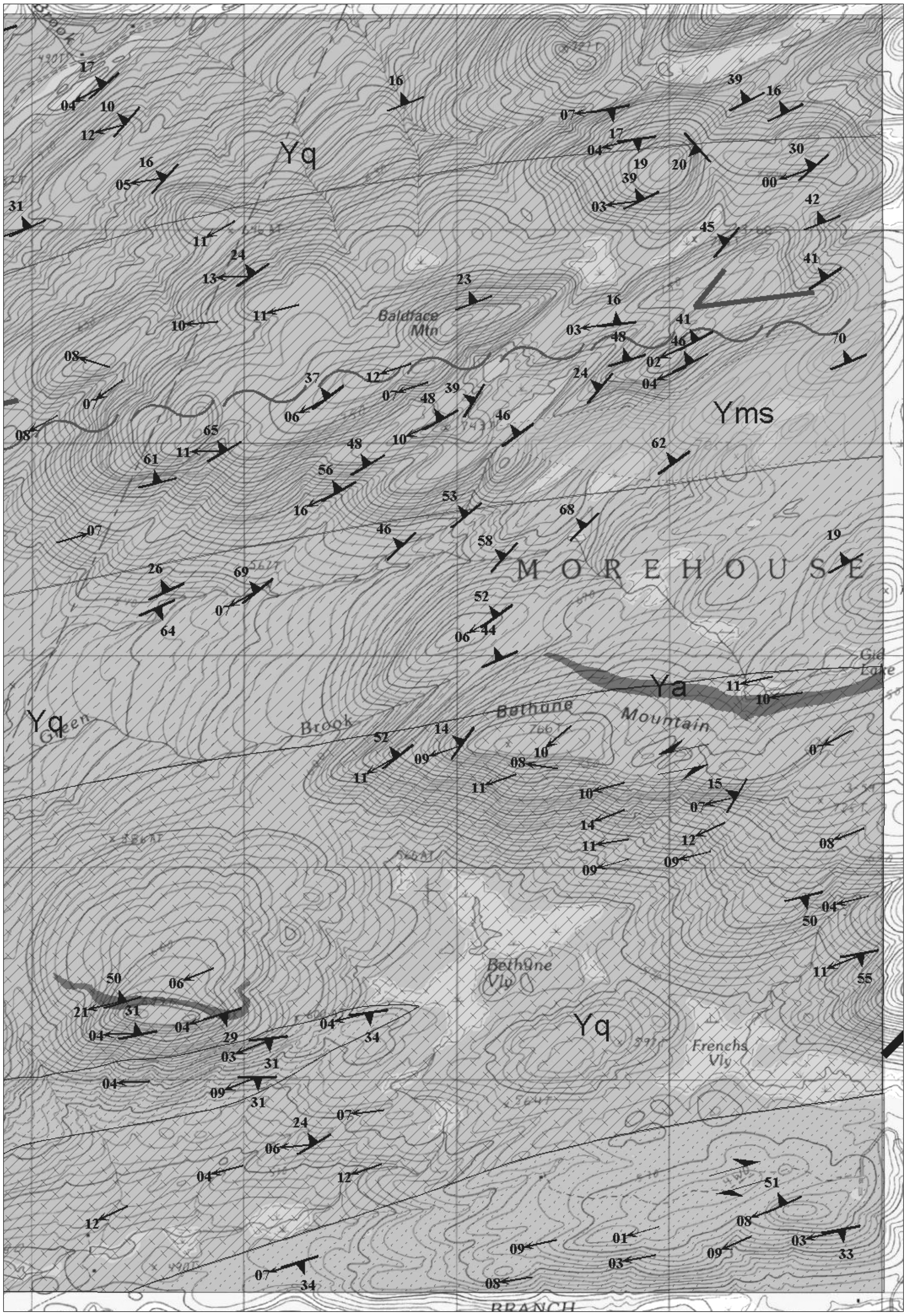


Figure 2 continued - Eastern extent of the geologic map of the West Canada Creek basin. See the next page for the map explanation.

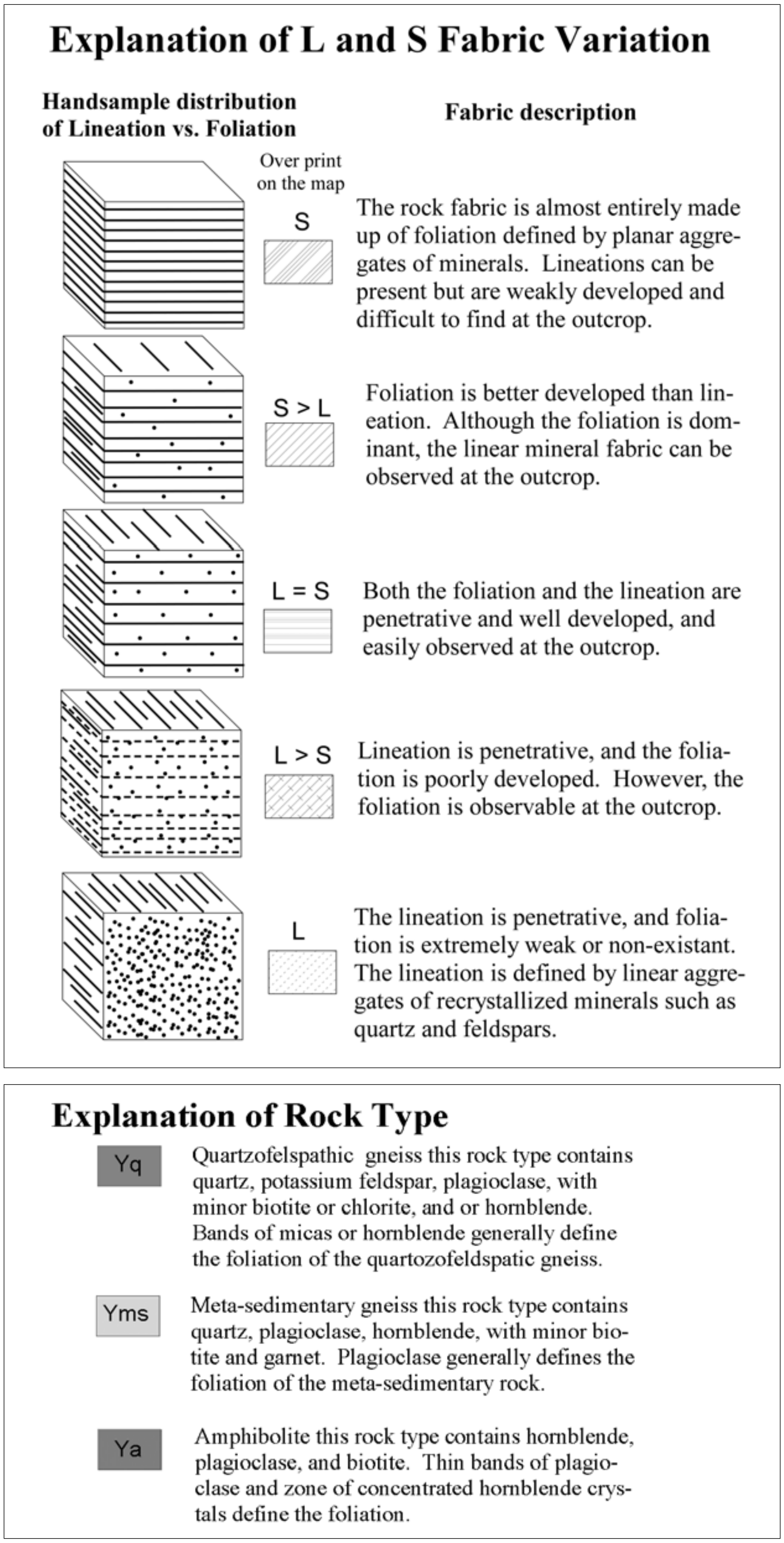


Figure 2 continued - Map explanation for the West Canada Creek basin with special emphasis on rock fabrics and lithology. Three gray shades were used to represent general rock types and the shades are overlain with patterns to represent the five categories of rock fabrics. The structure symbols on the map represent the foliation (S) and lineation (L) as described in the text.

FABRIC VARIATION IN THE PISECO LAKE ZONE

Five domains of varying fabric intensities were documented (L>>S, L>S, L-S, S>L, and S) within the Piseco Lake zone and the shear zone transition region with the wall rocks (see Figure 2 on the previous pages). The northern boundary of the Piseco zone is defined by a gradational increase in L-S fabric intensity from north to south. Both the foliation and lineation are defined by dynamically recrystallized aggregates of quartz, K-feldspar, plagioclase and minor mafic phases. This fabric transition corresponds with an increase in grain size reduction of all these minerals. Within the Piseco Lake zone the fabric variation occurs systematically from L-S dominated, to L>S and finally L>>S tectonite (Figure 2). A cigar-shaped map-scale domain of L>>S fabric, 3.5 km by 0.5 km in size, trends parallel to the linear fabric observed at the outcrop. A change in the dominate dip direction between the L>>S and L>S domains supports the presence of a foliation fold over the cigar-shaped domain.

The wall rocks to the shear zone in the study area are mostly granitic gneisses and minor dioritic gneiss containing metamorphic index minerals of hornblende and hypersthene. The granitic gneiss contains a dominant gneissocity that strikes generally east-west with very weak mineral lineations. Quartz and feldspars form course crystalline aggregates that define the gneissocity. The presence of hornblende and hypersthene, and the gneissic fabric suggest these granitic rocks were metamorphosed under granulite facies conditions as reported by earlier researchers (McLelland, 1984).

Within the zones of intense L-S deformation fabrics, the rock is generally granitic gneiss, however, it contains abundant feldspar and quartz grains up to a few cm in diameter. In places, K-feldspar grains appear to be relict igneous metacrysts. As mentioned previously, the L-S fabrics are defined by planar and linear aggregates of dynamically recrystallized quartz and feldspar grains. Additionally, the fabrics are defined by chlorite and minor biotite. This observation was previously noted by Gates et al. (2004) where they demonstrated that the occurrence of chlorite in the Piseco Lake zone to be fabric forming and parallel to the mesoscopic foliation and lineation. These rock textures and index minerals suggest two conclusions: 1. the Piseco Lake zone developed in course grained granite that is not found in the wall rocks, and 2. Cannon (1937) and McLelland (1984) described similar rock fabrics for other parts of the Piseco Lake zone, however, they did not mention the presence of low-grade fabric forming metamorphic index minerals.

Systematic look at the structural data

The structural data collected during the mapping was divided based on the fabric categories that define the five fabric domains, as shown on the map of Figure 2. These data were used to generate lower hemisphere contour diagrams for the poles to foliation and lineation. Poles to lineations are plotted at or near the perimeter of the diagram and the poles to foliation form the diffuse girdle on the interior of the diagrams. The stereogram representing the data from the L>>S domain (Figure 3E) demonstrates that the foliation is dominantly dipping to the south. But the stereogram showing the data from the L>S domain demonstrates that the foliation is dominantly dipping northward. The L-S, S>L, and S domains show foliation dominantly dipping to the north, however the general strike is consistent throughout all the diagrams.

OHIO GORGE REGION

The West Canada Creek flows through the east-west trending Ohio gorge a few kilometers south of the geologic map area of Figure 2. Nearly 90% bedrock exposure afforded the opportunity to study the fabric variation in great detail in the heart of the Piseco Lake shear zone. Access to the gorge is restricted due to private property and high water most of the year. During the Summer 2004, a detailed outcrop map was produced for the southern side of the gorge. High-resolution digital photographs were taken and assembled into a mosaic. The photo mosaic was used as the base map, and rock fabric and textural variations were overlain at the sub-meter scale. In general, the bedrock exposed in the gorge is the megacrystic granitic gneiss typical of the Piseco Lake shear zone, and there is little variation in the overall mineral content along the extent of the gorge, but, the outcrop analysis shows variation in deformation fabric at the scale of 10's of meters. Three fabric categories were observed in the gorge, L>>S, L>S and L-S tectonite as described previously. These categories demonstrated gradational and abrupt contacts between one another, and the shape of some fabric domains in the gorge show similar geometric relationships to the map-scale domains.

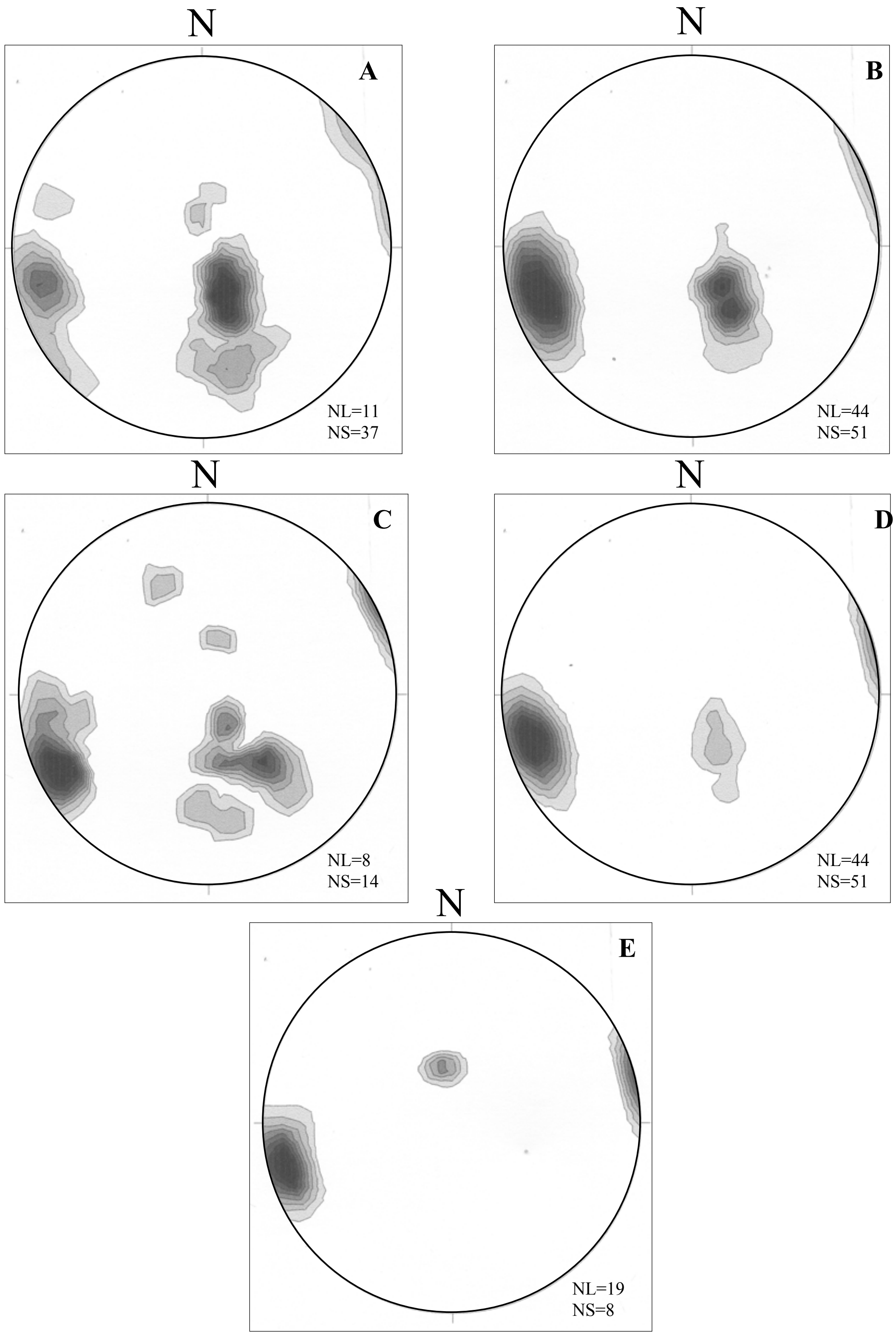


Figure 3 - Lower hemisphere contour stereograms for the L-S domains represented on the geology map of Figure 2. For each diagram above (A-E), the poles to lineations plot near the perimeter of the diagrams and the poles to foliation form the interior domains. Note the increase in the intensity of the cluster of linear data with the decrease in the occurrence of the foliation data.

Throughout the Ohio gorge there are extensive kinematic indicators consistent with sinistral low-angle shear. The L-S domains contain the best preserved porphyroclasts, with the L>>S containing few. The kinematic indicators include S-C fabrics, shear bands, asymetically broken K-feldspar grains, sigma- and delta-porphyroclasts (Lister and Snoke, 1984; Simpson and Schmid, 1983, Passchier and Simpson, 1986).

There are a number of ductile normal faults that crosscut the dominant deformation fabrics. The trace of the dominant outcrop fabric shows "drag" as a primary indicator of normal displacement. Most of these small normal shear zones contain granitic pegmatite, and there are also parallel pegmatite dikes that show no deformation. The ductile normal zone located on the east end of the Ohio gorge exhibits oblique sinistral-normal displacement, while the remaining ductile normal faults exhibit dip slip offset. Figure 4 shows a sterographic plot of the orientation of these ductile normal shear zones and other pegmatite dikes in the Ohio gorge. Both inside the ductile normal zones and within undeformed pegmatite dikes, they are composed of course grained quartz and K-feldspar with minor chlorite. These pegmatites vary in thickness from 0.5 m to 6 cm within one of the normal shear zones. Valentino et al. (2004) noted ductile normal shear zones in the upper reaches of the Ohio gorge and related it to larger-scale displacement at Speculator Mt. farther east.

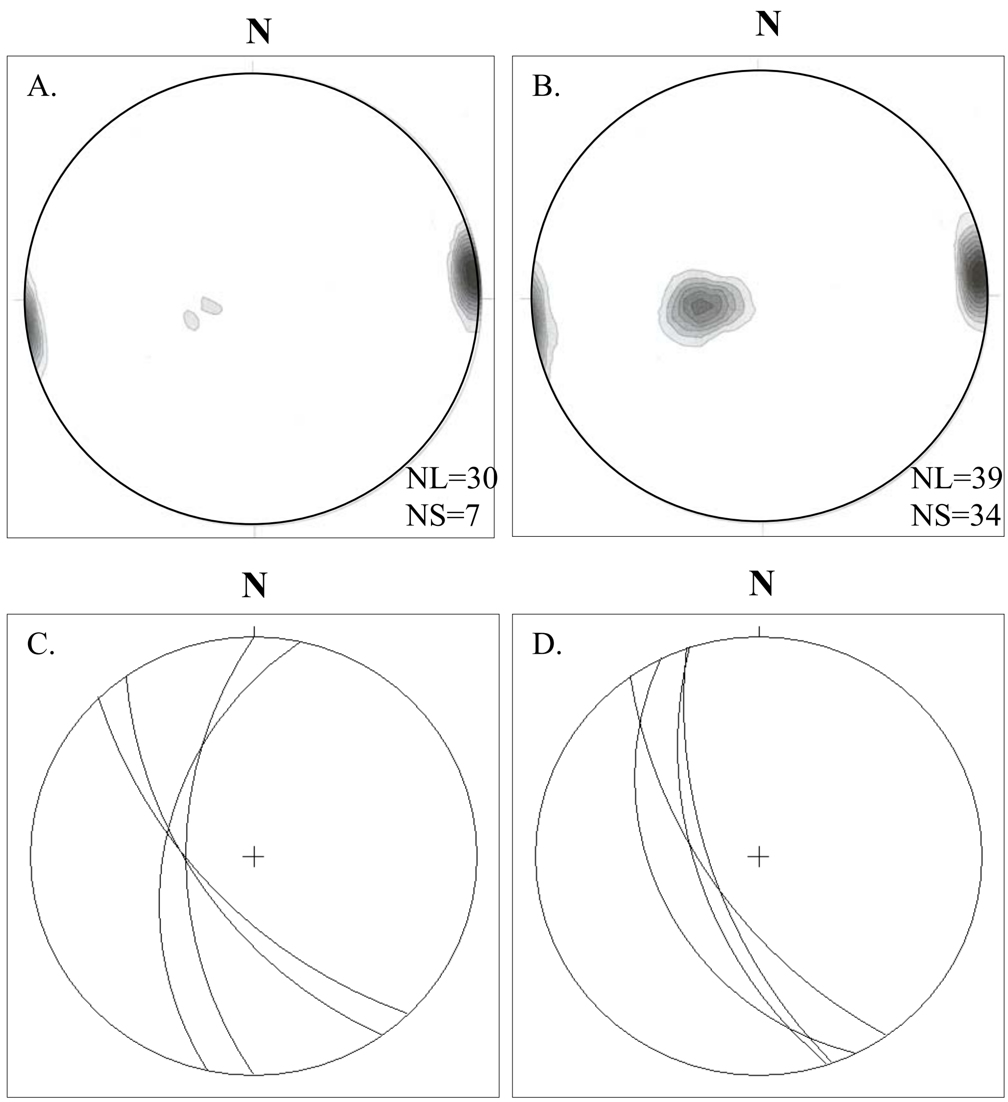


Figure 4 - Lower hemisphere sterograms for structures observed in the Ohio gorge. A. & B. Contour diagram for the poles to lineation and foliation from L>S domain (A) and L-S domain (B) parts of the gorge; C. & D. Great circle plots for ductile normal shear zones (C) and undeformed pegmatite dikes in exposed in the gorge (D).

DISCUSSION AND CONCLUSIONS

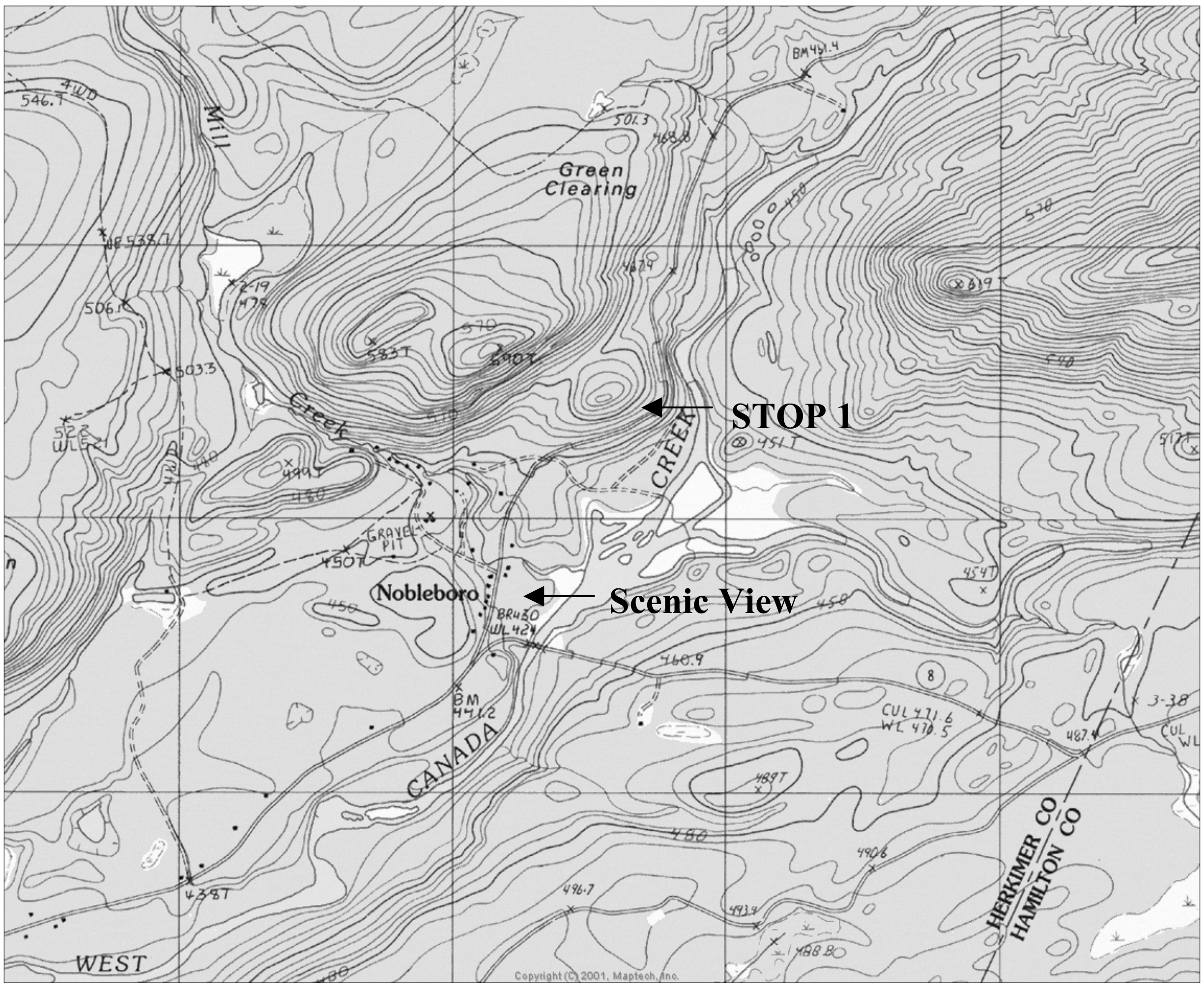
Deformation fabrics vary systematically across the northern boundary of the Piseco Lake shear zone in the West Canada Creek basin. Within the Piseco zone, all rocks contain well-formed mineral elongation lineations, however, the variation in fabric intensity appears to be controlled by the development of foliation. The attitude of mineral elongation lineations vary little, but foliation varies systematically in intensity and orientation. Rocks dominated by L-tectonite occur in a cigar-shaped domain, that occurs in broad open foliation antiform, but the foliation is only weekly developed in these areas (L>>S domain). The deformation within the zone occurs in rocks of granite protolith that are different from the high-grade granitic gneiss that occurs in the northern wall-rock to the shear zone. Geochronologic studies in the Adirondacks constrain the timing of deformation and peak metamorphism to 1090-1030 Ma. This is after the intrusion of the AMCG plutonic rocks from 1160-1100Ma (McLelland and Isachsen, 1986; McLelland et al., 1988; McLelland et al., 1996; McLelland et al., 2001). An AMCG suite age and Ottawan metamorphic age was reported by McLelland et al. (1988) for deformed granite from the core of the Piseco antiform. The location of the rock used for this analysis occurs directly along strike about 20 km east of the current study area. Although the AMCG granite within the Piseco zone is highly deformed, there is scant evidence within the study area, that it was subjected to the regional high-grade metamorphism that is preserved in the wall-rocks. It is worth considering that the gneissocity and high-grade metamorphism that occurs in the shear zone wall-rocks pre-dates the intrusion of the granite. However, it is also possible that the low-grade dynamic metamorphism that occurred in the Piseco zone entirely overprinted the high-grade metamorphism of both rock (wall-rocks and granite) units.

ACKNOWLEDGEMENTS

The fieldwork during 2003 was funded through a grant from the Association of American State Geologist. The fieldwork during 2004 was funded by an R.E. McNair scholarship to D. Piaschyk. Lab work was funded by a SUNY Oswego Scholarly and Creative Activity grant. We also thank Jim McLelland for introducing us to this project and visiting with us in the field during the Summer 2003.

ROAD LOG AND STOP DESCRIPTIONS

Road Log:
Mileage:
0.0 The trip begins at the assembly point in the parking lot of the scenic overlook on the West Canada Creek. The scenic view is located in Nobleboro off of Route 8, between Poland and Piseco.
0.6 When exiting the driveway to the overlook, turn right on Haskell road, travel about 0.6 miles and park along the side of the dirt road. Caution the banks of the road may not be stable. Cross the dirt road and walk west to what appears to be an old quarry (STOP 1).



STOP 1: S>L Granitic gneiss with strong gneissic texture

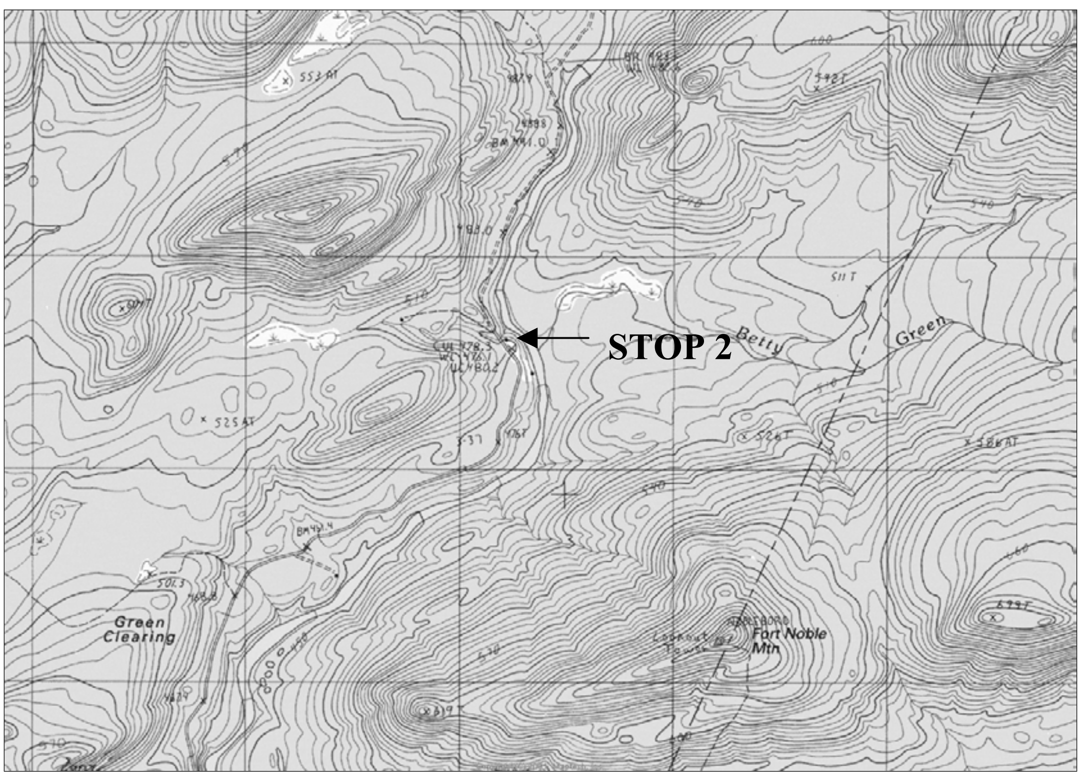
Exposures of course granitic gneiss with well developed gneissocity can be seen in the high-wall of this old quarry. This stop demonstrates S>L fabric, lineation area weakly developed and foliation dominates. (Figure 5). The grain size is uniform and few kinematic indicators are present. The foliation is penetrative and is defined by planes of K-feldspar and biotite and dips to the northwest. The lineations are defined by streaks of biotite and plunge to the southwest.

Figure 5 – Photograph, view is looking north, of granitic gneiss with well-developed foliation and gneissocity. K-feldspar-rich bands define the gneissic texture. Weakly developed lineations can be observed with closer inspection, and they are parallel to the mechanical pencil in the center of the photograph.

Mileage:

0.6 Continue northeast on Haskell road.

2.7 Arrive at a clearing, before bridge and gate turn right and park in grass. Walk east down a small hill to the West Canada Creek (STOP 2).



STOP 2: L-S and S>L Dioritic Gneiss

The southern most part of the outcrop demonstrates L-S fabric (Figure 6). Lineations are about equal to foliation in intensity. Rods of hornblende define the linear fabric, which plunges to the southwest. Plagioclase and quartz define the foliation. In thin sections of this outcrop a ceased reaction is preserved. The hornblende crystals were breaking down to form biotite in a retrograde reaction. The northern most part of the outcrop demonstrates S>L fabric with gneissic textures, similar to the previous stop. Some small faults with 15cm of displacement and boudins are also preserved at the northern end of the outcrop (Figure 6). Garnets are also visible at the northern end of the outcrop but were not observed at the southern end in hand sample or thin section.

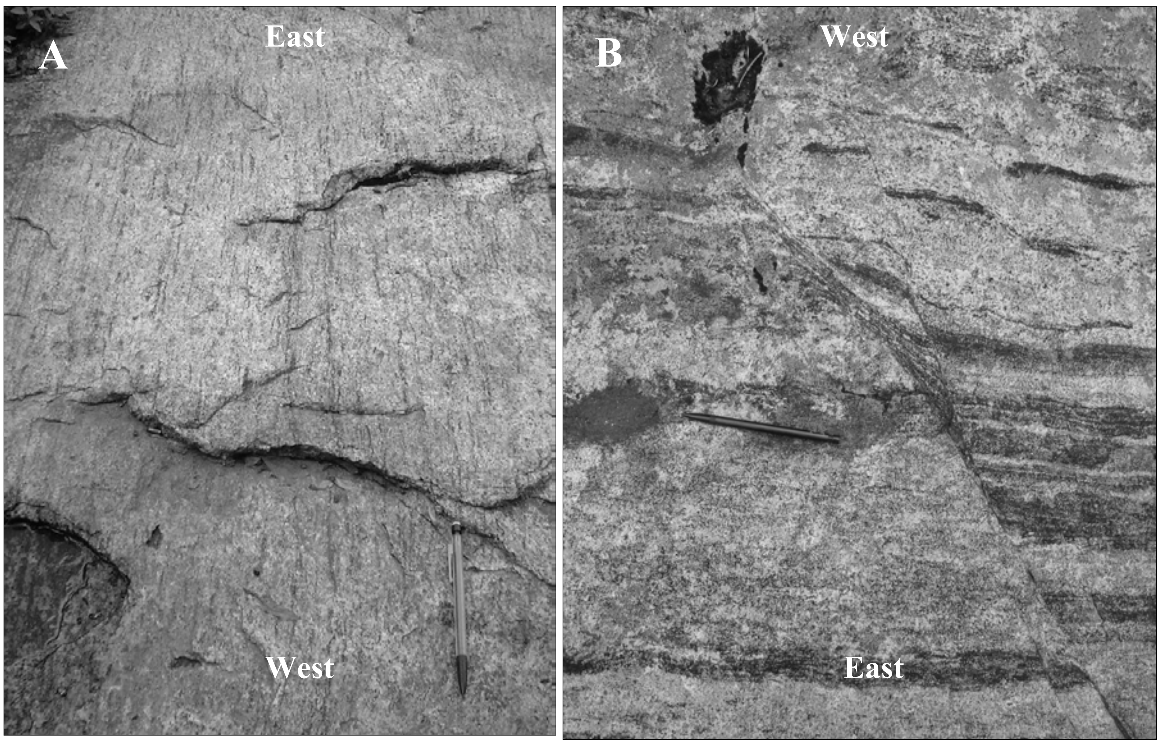


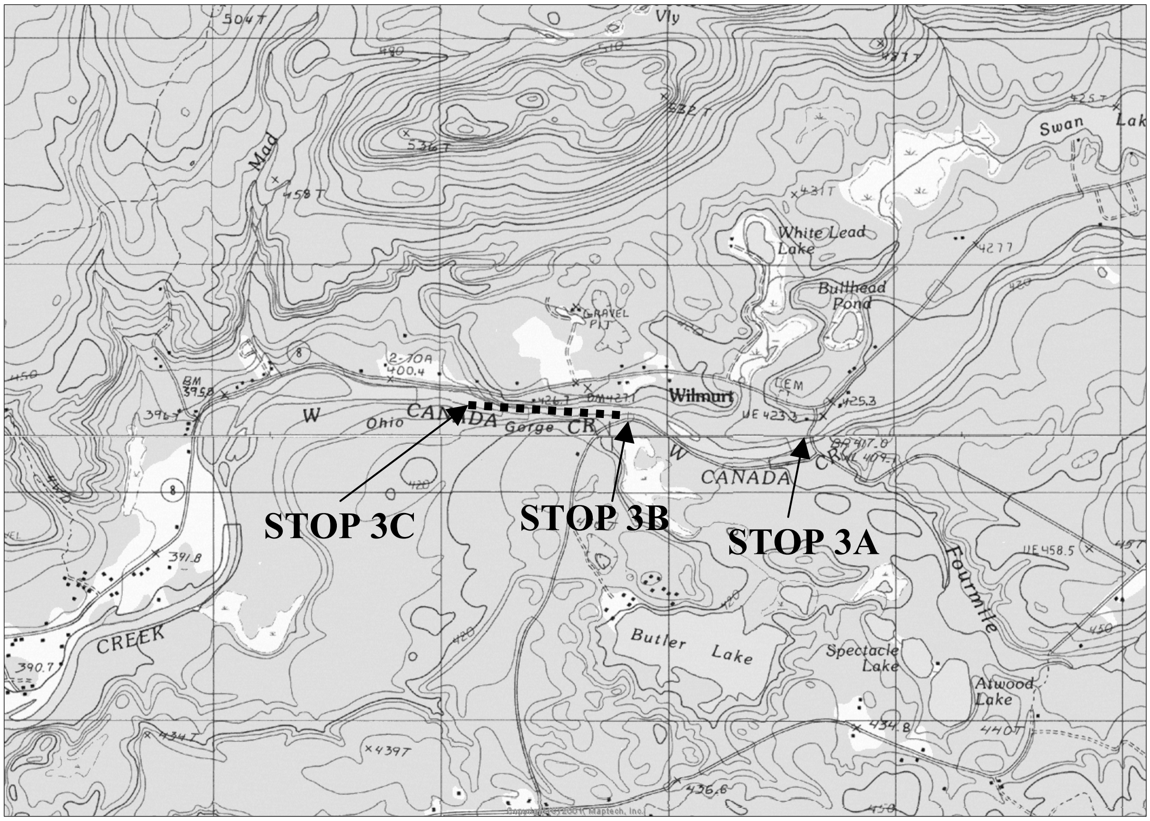
Figure 6 – (A) This is a photo of the L-S fabric at the south end of the outcrop at STOP 15. The black lines are rods of hornblende that define the linear fabric. (B) This is a photo of S>L fabric at the north end of the outcrop. A fault is shown from the top left corner to the bottom right corner. The offset of hornblende bands demonstrates a displacement of about 15cm.

Mileage:

2.7 Turn left out of the parking area heading south on Haskell road.

5.4 Pass the over look and take a right at the stop sign onto Route 8 west.

8.1 Turn left onto Gray Wilmurt Road just after right-hand curve in Rt. 8. Cross a bridge over the West Canada Creek and park at the intersection with Jones Road. Walk back toward the bridge over the West Canada Creek and down the hill to the outcrop just east (upstream) of the bridge (see location map). The outcrop forms a small waterfall on the creek. This is location 13 of Valentino et al. (2004).



STOP 3: The Piseco Lake Shear Zone at the Ohio Gorge of the West Canada Creek

The Piseco Lake shear zone traces westward through the West Canada Creek basin. Some of the best continuous exposures occur in the Ohio Gorge near Wulmurt. This stop contains highly deformed granitic gneiss in the gorge. During periods of high water, the exposures in the gorge may be covered by water. Permission is needed from the landowners at STOP 3A and 3C.

STOP 3A East of the Ohio Gorge. The West Canada Creek forms a small waterfall at the upstream part of this outcrop. Pavement exposures reveal the L-S and L>S deformation fabric in granitic gneiss (Figure 7). Foliation is gently dipping and the lineations are subhorizontal. In the region immediately down-stream of the falls, the foliation is defined by planar aggregates of recrystallized K-feldspar and quartz that alternate with dark layers containing abundant chlorite and minor biotite. The dominant fabrics are cross cut by at least three small high-strain zones. Two are steeply dipping and strike about east-west, and the third strikes south and dips moderately westward. One of the steeply dipping high-strain zones occurs in the vertical face at the southern side of the outcrop. Another occurs at the western limit of the outcrop close to the water. The north-south striking zone occurs in the low ledge near the falls. This small shear zone contains deformed pegmatite, and cross cuts the Piseco Lake shear zone foliation and lineation. Shear sense is top down to the west or normal. The other high-strain zones both contain evidence for oblique sinistral shear.

Mileage:

8.1 From the parking area head west on Wilmurt Road

8.7 Cross over a small bridge and on the first curve before driving up a hill park on the right side of the road. Walk northwest about 20 m, and STOP3B is the first outcrop on the southeast side of the Ohio gorge.

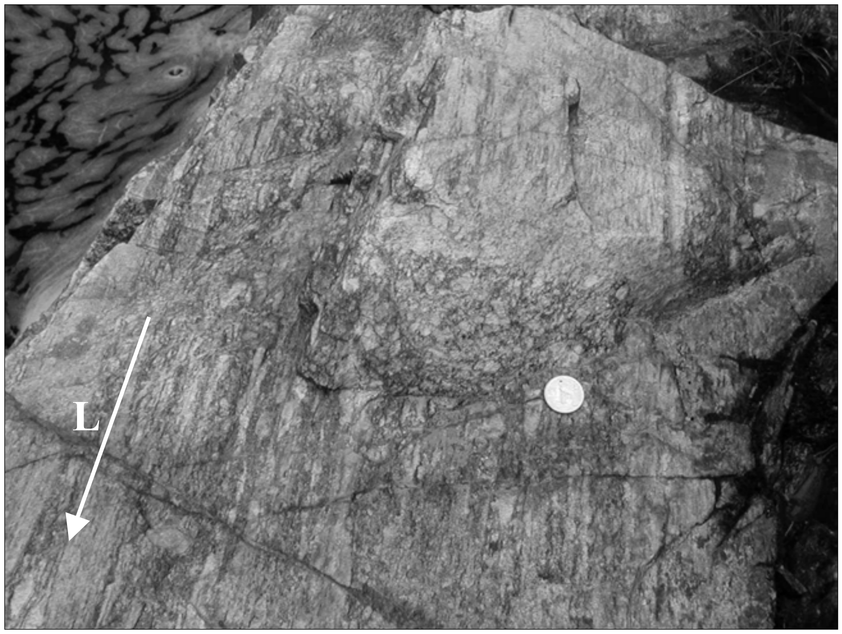


Figure 7 - Outcrop of granitic gneiss with L>>S fabric. The view is looking west. Note the textural differences in this view. The area above the coin is a subvertical surface with the ends of the mineral lineations exposed. The rest of the outcrop is broken parallel to the lineations.

<u>STOP 3B East end of the Ohio Gorge.</u> The east end of the Ohio gorge demonstrates L>S and L-S fabrics. The gorge contains granitic gneiss that varies only in the proportion of quartz, K-feldspar, plagioclase, biotite, and chlorite. The L-S domains have the larges grain size with numerous δ and σ shear sense indicators. The L-S domains also demonstrate lineation about 1 to 5 cm long, .5-2 cm wide, and 1-3 lineations per cm. The L>S domains have a smaller grain size then the L-S domains. The L>S domains demonstrate more lineations and less shear sense indicators, lineations are 1-6 cm long, .5-2 cm wide, and 3-5 lineations per cm. A contact between L>S and L-S can be observed (Figure 8). West of the fabric contact is an oblique left normal fault about 5 cm wide. This fault crosscuts the metamorphic fabrics strikes north south and dips to the east. Farther west is a .5 m wide pegmatite that strikes northwest southeast and dips southwest that also crosscuts metamorphic fabric. The pegmatite is composed of quartz, K-feldspar, and minor chlorite with grains about 1 to 3 cm in size.

STOP 3C requires a .75-kilometer traverse west along the ridge of the Ohio gorge. Please **WATCH YOUR STEP** because the walls of the gorge are vertical and a fall would likely result in serious injury or worst.

<u>STOP 3C West end of the Ohio Gorge.</u> The west end of the Ohio gorge also demonstrates L>S and L-S domains. Another ductile normal fault can also be observed. The ductile normal fault is much larger then the one observed at the east end of the gorge (Figure 9).

Mileage:

8.7	Back track to Route 8, and turn right.
12.0	Pass the scenic view where the trip began and continue east on Route 8.
16.9	Turn left onto Fayle Road.
18.5	Cross a one lane wood bridge, drive to an opening in the tree and the end of Fayle Road. Park and hike to the west about 350 meters to STOP 4A.

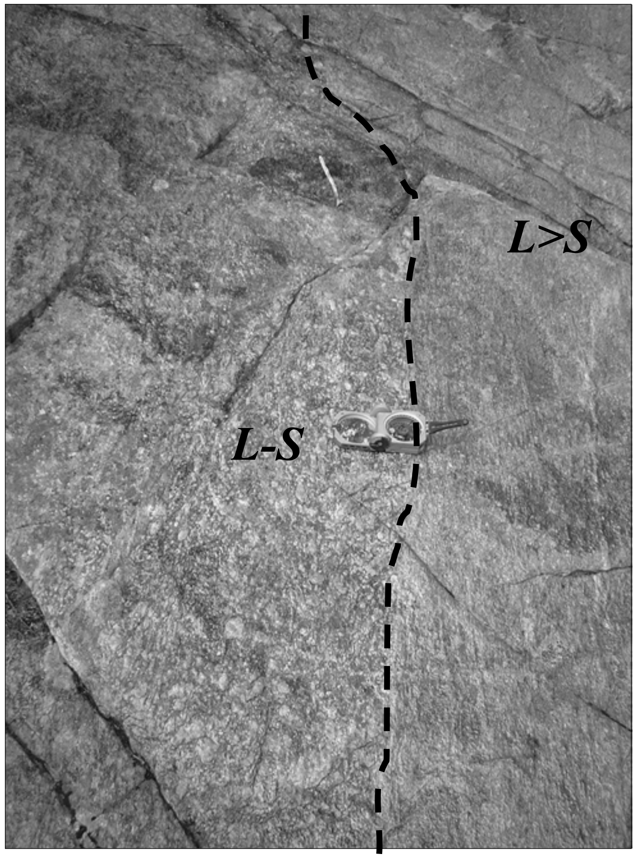


Figure 8 - This is a photo taken at the eastern end of the Ohio gorge, and the view is westward. This photo demonstrates a contact between L-S and L>S rock fabrics. In the L-S domain the grain size is larger and the lineation is less developed, while in the L>S domain the grain size is smaller and the lineations are stronger and easier to see. Note that the boundary between these fabric domains is slightly oblique to the general trend of the rock fabric. Brunton compass for scale.

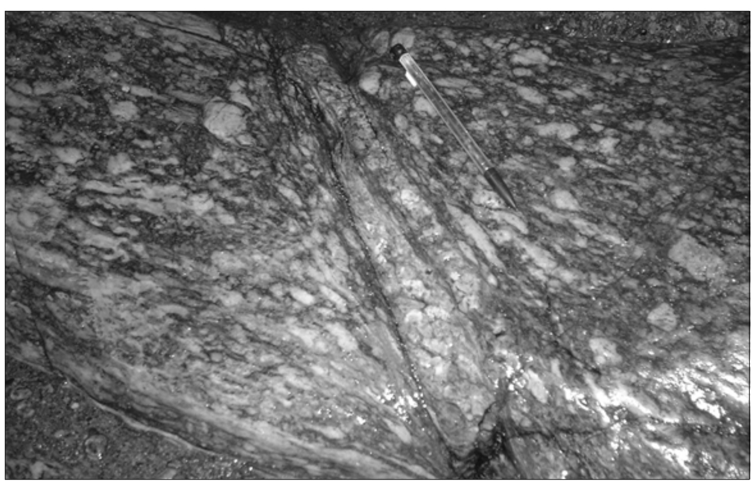
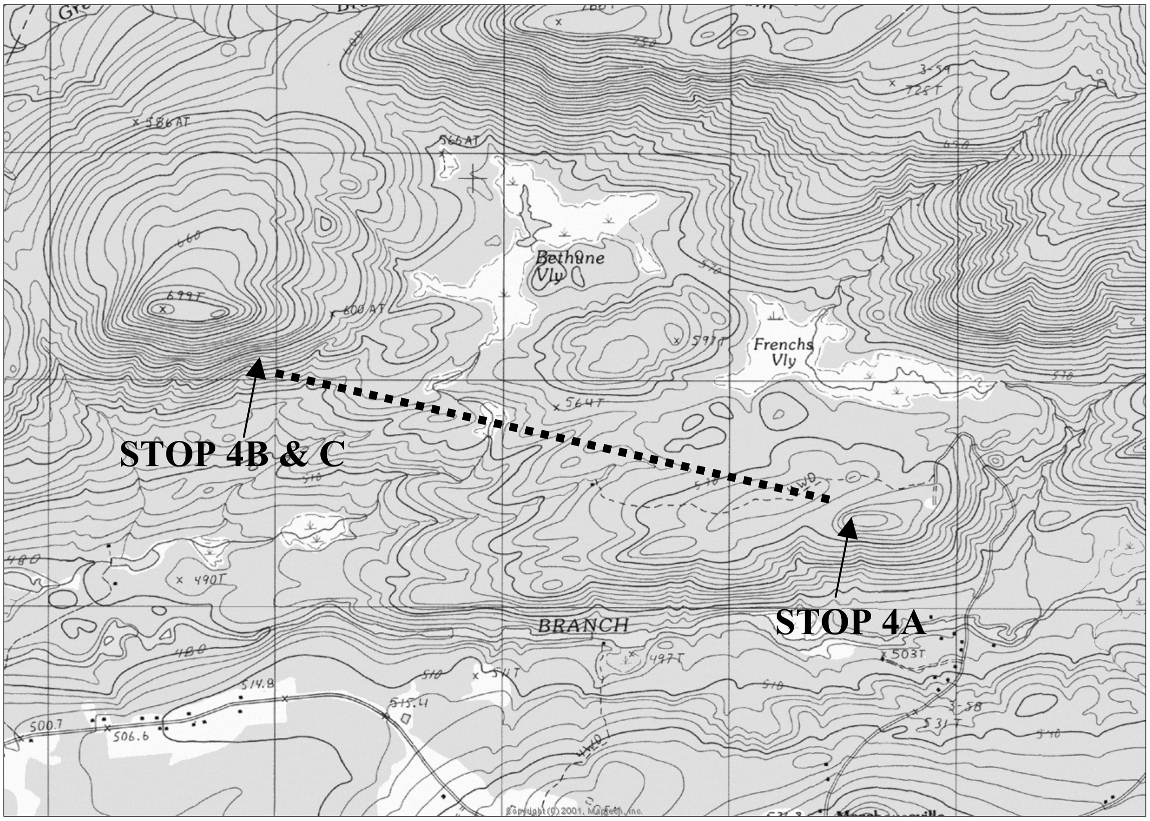


Figure 9 - This photo was taken facing south at STOP 3C. This portion of the ductile normal shear zone will only be visible if the water level is low. The small shear zone crosscuts the metamorphic fabric of the Piseco Lake zone. The footwall displays fabric drag and the hanging wall remains relatively undisturbed. The center of the zone contains deformed pegmatite with dynamically recrystallized quartz and K-feldspar.



STOP 4: L>>S Granitic Gneiss

Excellent outcrops on the northern side of a small hill just east of the parking area. Follow the dirt road to a path through the woods, and then head up hill to the south to the outcrops. This outcrop of granitic gneiss contains domains of L>S and L>>S. The L>S domains contain large and numerous σ-type shear sense indicators, some δ-type are present but are much less frequent. The porphyroclasts are large about 1-3 cm and the recrystallized porphyroclastic material is often wrapped with a quartz ribbons (Figure 10). The interpreted shear sense is low-angle and left lateral. The granitic gneiss is composed of quartz, K-feldspar plagioclase, and minor chlorite and biotite. The foliation strikes east west and dips to the south.

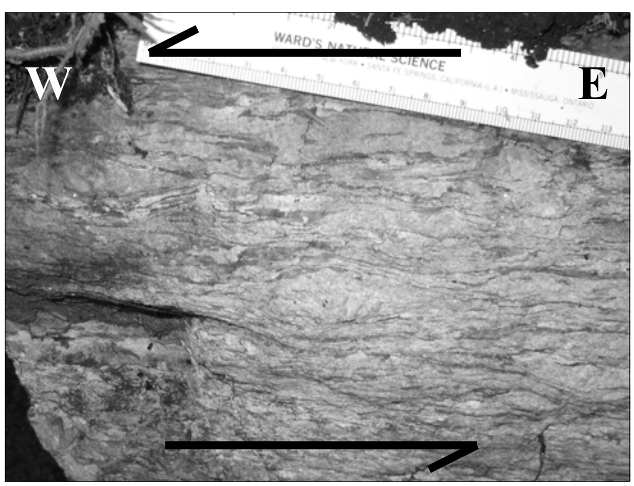


Figure 10 - The view is south at STOP 4, and looking at a surface parallel to lineation and perpendicular to foliation. Large porphyroclasts of K-feldspar and have σ-type tails which display left lateral shear sense.

STOPS 4B and 4C require about 2.5 kilometers of traverse at a bearing of about 280º. The traverse will cross a few small streams and under brush can be thick in places. There is no trail to follow, so **PLEASE STAY WITH THE GROUP**.

STOP 4B Mafic Gneiss. This outcrop is a rare mafic gneiss composed of biotite, hypersthene, plagioclase, quartz and ilmenite. The fabric is L>S, lineations are defined by rods of plagioclase and streaks of biotite. The grains size is very small about 0.5mm. This rock unit borders the L>>S domain which can be seen at STOP 4C.

STOP 4C L>>S Cigar-Shaped Domain. The final stop is a spectacular L-tectonite (Figure 11). The outcrop is granitic gneiss composed of quartz, K-feldspar, plagioclase, and fabric forming chlorite and biotite. Foliation is hard to see in hand sample but can be seen if stained for plagioclase and K-feldspar.

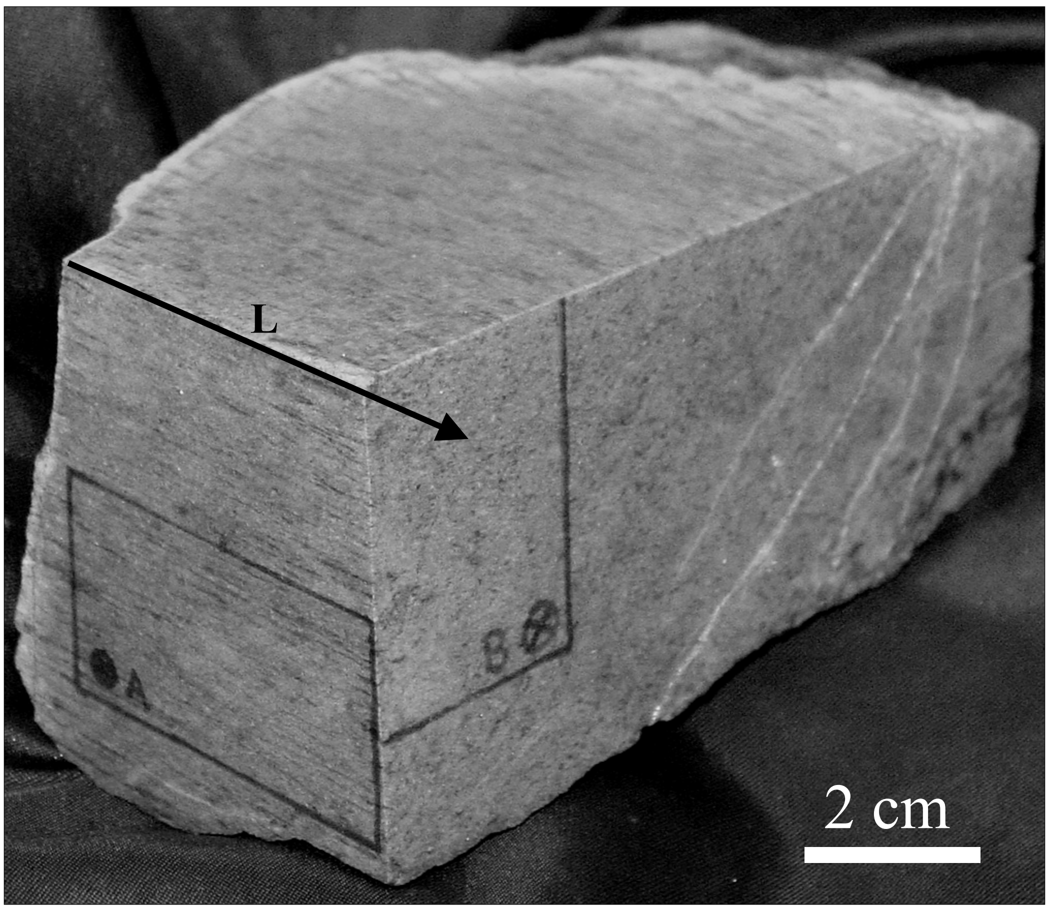


Figure 11 – Rock sample from STOP 4C with three mutually perpendicular cuts to reveal the internal fabric. Quartz rods and recrystallized aggregates of K-feldspar define the penetrative lineations, and foliation does not exist in this specimen. The dark minerals are primarily chlorite and biotite, and they too define a microscopic lineation as demonstrated by Price et al. (2003).

END OF TRIP.

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Trip B-3

MONITORING COASTAL CHANGE: EASTERN LAKE ONTARIO

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INTRODUCTION

This field trip will examine the sandy beach-dune barrier system that extends for nearly 17 miles north of the Salmon River mouth along the eastern Lake Ontario coast (Fig. 1). Barrier bars separate ponds and wetlands from open waters of Lake Ontario and are cut by inlets that connect ponds to Lake Ontario. The sandy beaches and dunes on the barrier system, together with adjacent wetlands, ponds and lake waters, provides habitat for a number of rare plant and animal species and also for fish, waterfowl and migratory birds. The beaches also afford recreational opportunities for residents and tourists, and inlets provide access to Lake Ontario for recreational craft from marinas and boat launches around ponds and tributary streams. The sand dunes here are the largest in the state, and they protect residences landward of the dunes from wind and wave erosion and flooding due to storm surge. The shoreline is a mix of public and privately-owned land, with public lands designated as state parks and wildlife management areas. The private land that amounts to roughly half the barrier system has been developed for cottages, homes, RV parks and campgrounds that are built on or just behind the dunes. Governmental and non-governmental conservation agencies have been working for decades to protect this fragile system.

A program of beach profiling has been ongoing since 2002 at 5 locations in the barrier system (Fig. 2). The goal is to identify patterns of seasonal and longer-term erosion and deposition on these beaches. This field trip will visit several locations where beach monitoring has been conducted. Since 2003, Sandy Pond Inlet has been studied by detailed mapping. This inlet is a very important access point to the lake from the numerous marinas and private residences around North Pond. The inlet becomes shallow in late summer, and larger craft have difficulty navigating the shallow channel. We will visit Sandy Pond Inlet at the end of this field trip.

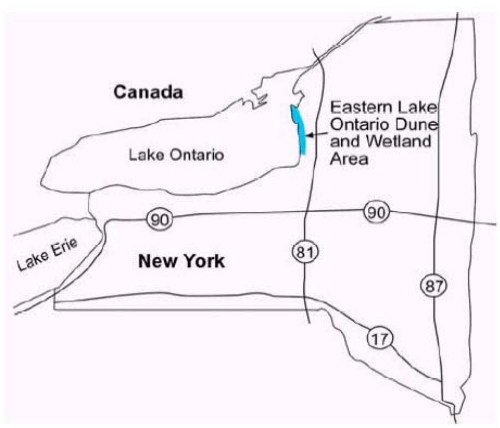


Fig. 1 - Location of the sandy barrier system.

Background

The eastern Lake Ontario barrier system formed during a period of significantly lower lake level that allowed wave action to transport sand from a shallow submerged sand body shoreward (Sutton, et al., 1974). Winds and wave action had established the beach-dune system by at least 1290 years ago (Woodrow, et al., 2002). Present-day, higher lake levels, have submerged the sand source to a depth where wave action can not reach and transport additional sand to the barrier system. Other sources (rivers, erosion of coastal lands) contribute little sand to the system today (Steadman, 1997). Therefore, when waves and wind erode sediment from the beaches and dunes, there is no significant replacement. Especially toward the southern end of the system, increasing areas of beach are becoming covered with cobbles.

Previous studies of coastal change in the barrier system utilized analyses of historical maps and aerial photographs (Weir, 1977), and focused on the barrier between Lake Ontario and North Pond. DelPrete

(1997) supervised SUNY Oswego students in mapping changes in Sandy Pond Inlet during the 1980s and early 1990s. More recently, McClennen, et al. (2000) used Geographic Information Systems (GIS) and air photo analysis to document changes along the entire barrier system. Steadman (1997) reviewed previous studies of the barrier system and identified research needed to improve conservation efforts for beach-dune ecosystems. His report identified geomorphic studies, including beach profiling, as one of 4 major research areas needed to improve understanding of coastal processes and conservation plans.

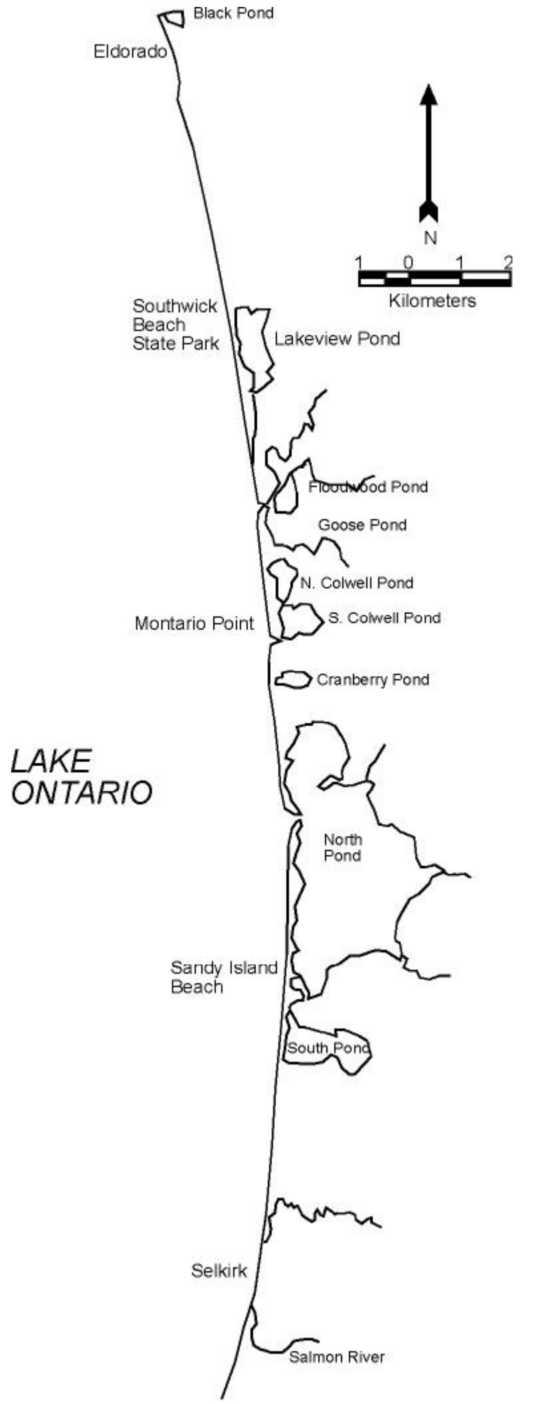


Figure 2 - Beach profiling was conducted at Selkirk,Sandy island Beach, Montario Point, Southwick Beach State Park, and Eldorado Beach.

Long-term trends in beach erosion and deposition can be understood in terms of a beach budget. If sources of sand supplied to a beach and processes that erode sand from a beach can be identified and quantified, a simple mass balance can be performed. If rate of sediment supply and rate of sediment erosion are equal, an equilibrium beach morphology will be maintained. If rates of sediment supply and erosion are not equal, the beach will either shrink or grow. The amounts and rates of sediment supply and sediment erosion on this coast have not been quantified (Steadman, 1997; Woodrow, et al., 2002), but as noted above, sediment supply is thought to be limited. Sediment is removed from beaches by wind, wave action transports the majority of sediment offshore or parallel to shore. Repeated beach profiling can detect onshore and offshore movements of sediment over time (Komar, 1998). By comparing beach profiles made over the course of several years, changes in the amount of sediment gained or lost on individual beaches can be assessed.

Lake levels vary on long-term, seasonal, and shorter time scales. During extended periods of higher than average lake levels, coastal erosion intensifies. This is a great concern for coastal property owners as well as conservation groups and land managers. Lake Ontario levels vary seasonally by 0.6m to more than a meter between high summer and low winter lake levels (Fig. 3). During the study period, wind storms caused lake level to rise by .25-.30m over time scales of hours. Beach profiles are affected by lake level variations at all these scales.

Bruun (1962) developed a simple model to predict changes in an equilibrium beach profile due to sea (or lake) level rise. The equilibrium profile is of the form

$$d(\mathrm{x}) = A\mathrm{x}^{\mathrm{b}}$$

where d(x) is water depth at a distance x from the shoreline, the exponent b is typically near 2/3 and the coefficient A is related to grain size. The "Bruun Rule" states that an increase in sea level results in erosion of a volume of sediment from landward portions of the profile. An equal volume of sediment is deposited in the nearshore zone, raising the sea floor such that the shape of the equilibrium profile is maintained, but shifted upward and landward (Schwartz, 1968). For the case of falling sea level, Bruun (1983) noted that lowered water levels would result in erosion in the nearshore zone and deposition onshore, resulting in the formation of beach ridges on steep shores and offshore bars on gently sloping shores, and thus the profile may change. Wood, et al. (1994) studied beach profile changes on Lake Michigan. They found systematic changes in beach profiles due to lake level variations, due to time lags between the rate of lake level change and profile changes, it was not clear that an equilibrium profile was maintained.

BEACH PROFILING LOCATIONS AND METHODS

Beach profiles were measured monthly during snow-free months (April or May through November) at 5 locations along the sandy eastern Lake Ontario shoreline (Fig. 2; Table 1) using the Emery method (Morisawa and King, 1973; Fig. 4). The Selkirk profiles were measured on private land just north of the mouth of the Salmon River, at the southern end of sandy barrier system. This portion of the shoreline has been extensively developed, with closely-spaced cottages built atop vegetated aeolian dunes (Fig. 5a). The beach at Selkirk is coarser-grained than the other beaches, with gravel-sized sediments dominating on the swash face. At other localities, mean grain size of beach sediments is fine sand. The Sandy Island Beach locality is now a state park, and the profile was measured on the public beach (Fig. 5b). This area lies at the south end of a zone of very tall dunes vegetated by dune grasses, bushes and mature trees. Some residences were built atop some of these dunes. At Montario Point, beach profiles were measured just north of a small inlet that connects South Colwell Pond to Lake Ontario. This beach is undeveloped, and the dunes at this location are not as tall as at Sandy Island (Fig. 5c). The profiles at Southwick Beach State Park were measured just south of the park boundary in the Lakeview Wildlife Management Area (Fig. 5d). The Eldorado site (Fig. 5e) lies just north of an area of closely-spaced cottages, but within the undeveloped Black Pond Wildlife Management Area. The profile locations at Southwick Beach and Eldorado begin on low dunes vegetated by shrubs and mature trees. At all 5 profiling locations, a datum was installed and its location and elevation relative to mean sea level was surveyed. Profiles were made from the datum along transects oriented perpendicular to shore. Profiles extended into the water until water depths became to great for wading. During cold months, the profiles ended at the water edge. In 2004, a program of surveying profiles to greater water depths was begun. Water depths were measured from a boat and the boat location was measured using an electronic distance meter. The plan is to measure profiles twice per year, at higher and lower lake levels.

BEACH PROFILING RESULTS

Monthly beach profiles were plotted for each of the 5 locations. There was good agreement among profiles taken at each location (Fig. 6), showing that the Emery technique generally yielded satisfactory results. The Selkirk (Fig. 6a) profiling locality showed the least variation in profile shape during the study period. This is to be expected, as Selkirk Beach is the coarsest-grained. The profiles at Montario Point showed the greatest variability (Fig. 6b), as this locality was affected more strongly by storm surge and perhaps due to its proximity to an inlet. All locations show seasonal changes in profile shape. Storms caused changes in profile shape at some locations. Longer-term changes in beach profiles are subtle. All of these changes are described below.

Seasonal variations in beach profiles

At all locations, the beach face shifts position as lake level rises and falls. During the study period, highest lake levels, about 75.1-75.3m during the study period, occurred in late spring/early summer. Lake levels dropped through the fall season, reaching levels of roughly 74.3-74.6m by early to mid-December. Profiles measuring during late spring, rising lake levels, showed that nearshore portions of the profiles had been eroded, and the subaerial portion of the beach was narrower compared to its position the previous fall. As lake level fell in late summer and into the fall season, the beach widened, and deposition occurred in the nearshore zone as the beach face prograded lakeward. These patterns are similar to those observed by Weishar and Wood (1983) on the Lake Michigan coast in Indiana.

Table 1. Dates and lake levels during beach profiling

Date	Lake Level (msl)	Selkirk	Sandy Island	Montario	Southwick	Eldorado
Jun 26, 00*	75.305			X	X	X
Nov. 25, 00+	74.414				X	X
Sep. 22, 02	74.596	X	X			
Sep. 29, 02	74.552			X		
Oct. 25, 02	74.413	X	X	X	X	X
Nov. 30, 02	74.364		X	X	X	
Apr. 13, 03	74.716		X			X
May 3, 03	74.845		X			
May 30, 03	75.066	X	X		X	
Jul. 9, 03	75.067				X	X
Jul. 10, 03	75.037	X	X	X		
Aug. 11, 03	75.022	X	X	X	X	X
Sep. 14, 03	74.697		X	X	X	X
Oct. 12, 03	74.624	X	X	X		
Oct. 18, 03	74.633				X	X
Nov. 14, 03	74.694		X	X		X
May 1, 04	74.955					X
May 2, 04	74.975		X		X	
Jun. 6, 04	75.151	X	X			X
Jun. 8, 04	75.151			X		
Jul. 8, 04	75.107	X	X		X	X
Jul. 11, 04	75.101			X		
Aug. 17, 04	75.009		X	X	X	X
Sep. 6, 04	75.097	X				
Sep. 25, 04	74.930		X	X	X	X
Oct. 26, 04	74.596					X
Nov. 6, 04	74.561		X	X	X	X
May 5, 05	75.151	X	X	X	X	X

* measured by J. Gibbs, SUNY ESF + measured by F. Florence, Jefferson Community College

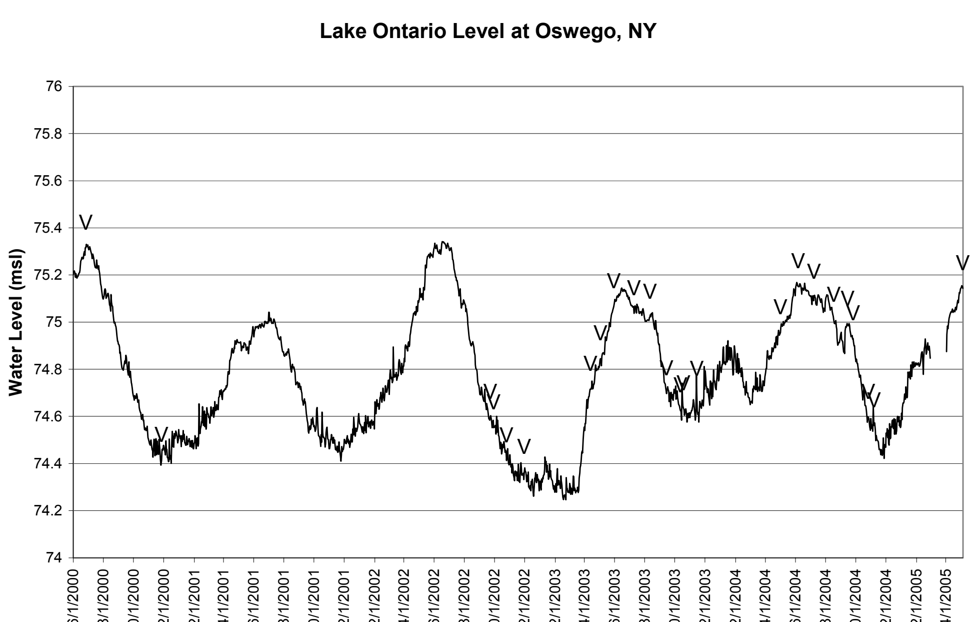


Figure 3 - Lake Ontario levels, June 1, 2001 to May 31, 2005. Arrows indicate dates beach profiles were measured.

Figure 4 - Emery technique used to measure beach profiles.

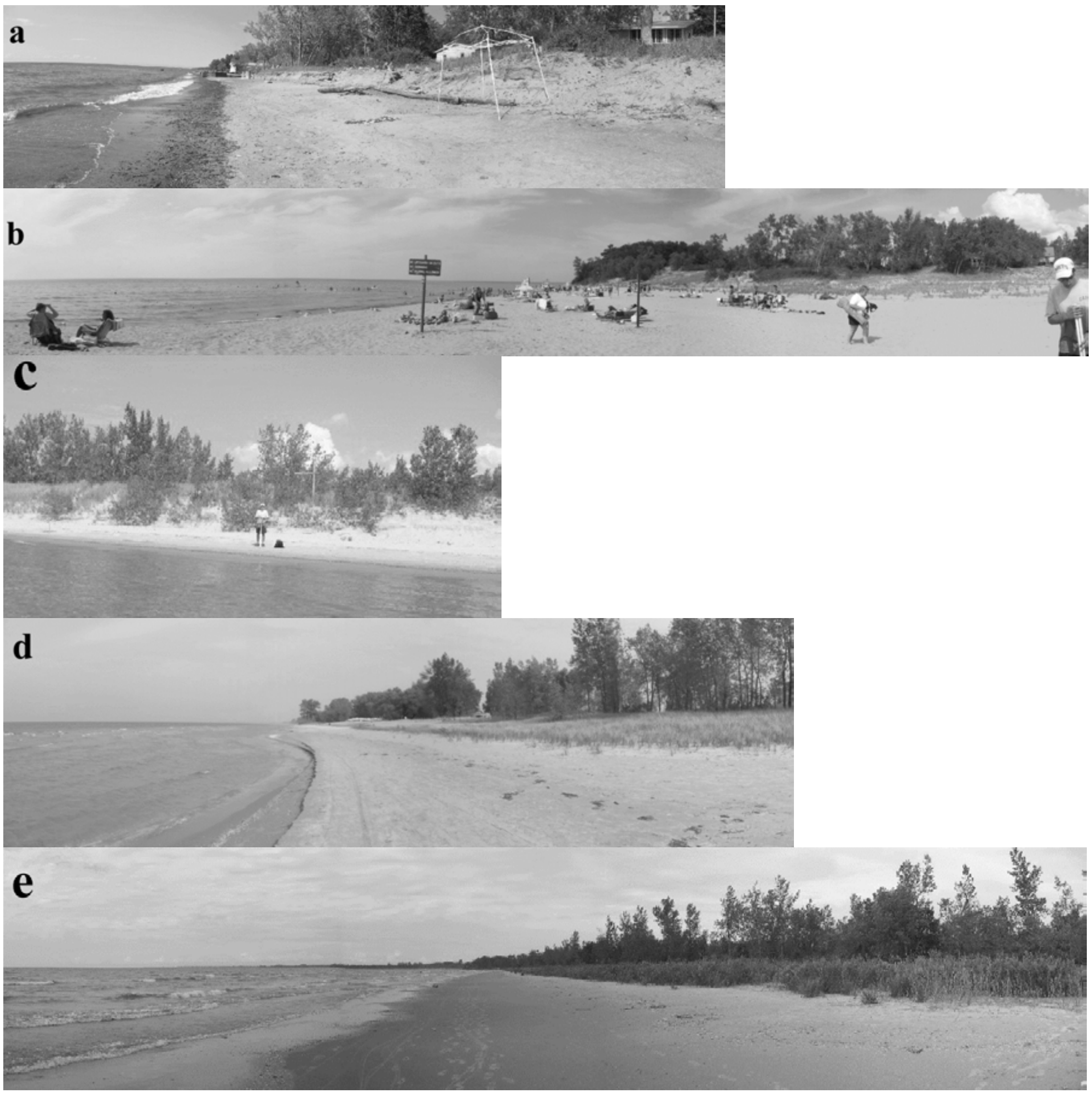


Figure 5 - Beach profiling locations: a) Selkirk; b) Sandy Island Beach State Park; c) Montario Point; d) Southwick Beach State Park; e) Eldorado Beach.

At some locations, nearshore bars developed during the summer and decreased in size with time as the beach face prograded lakeward. For example, at Montario Point, beach width was approximately 10m on July 11, 2004, but by November the beach had widened to about 36m (Fig. 7a). The 2004 Montario profiles also show that one or more nearshore bars were present in June and July and decreased in size during the summer and into fall. The 2004 profiles from Eldorado (Fig. 7b) also show the progradation of the beach face from May through November, and the August and September profiles show a nearshore bar. The landward margin of that bar was eroded between the August and September profiles.

The Eldorado profiles also show changes in the back berm portion of the beach, which at this location is vegetated by dune willow. The vegetated area increased in elevation during 2004, presumably due to the accumulation of wind-blown The subaerial portion of the beach lakeward of the vegetated area also increased in elevation during the May-September period, but was eroded before the November 6 profile was measured. This could have been the result of wave action during fall wind storms.

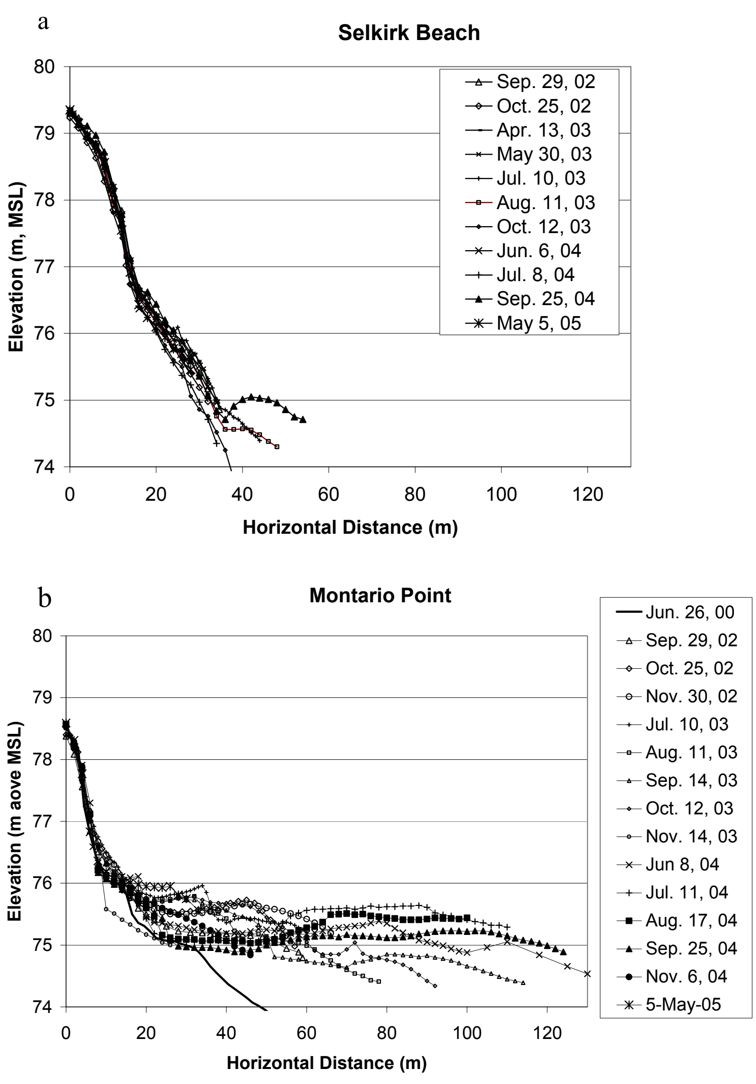
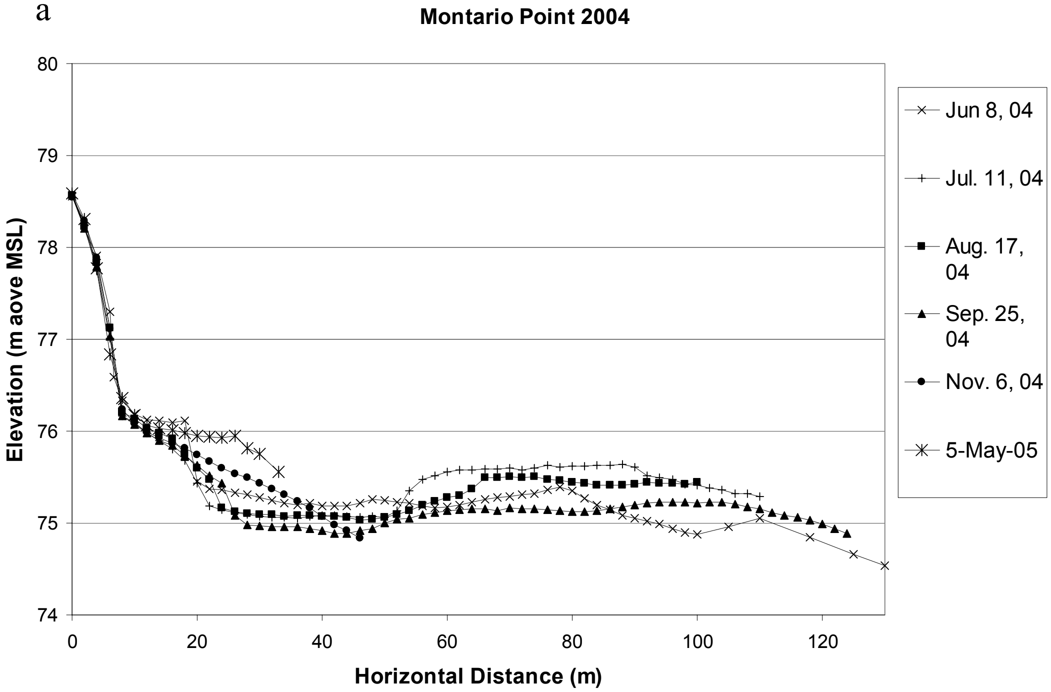


Figure 6 - Examples of graphs comparing all beach profiles measured at a particular location. a) The gravelly beach at Selkirk show very little variation in profile shape over time. b) The beach at Montario Point showed the greatest variability in profile shape during the study period.

a



b

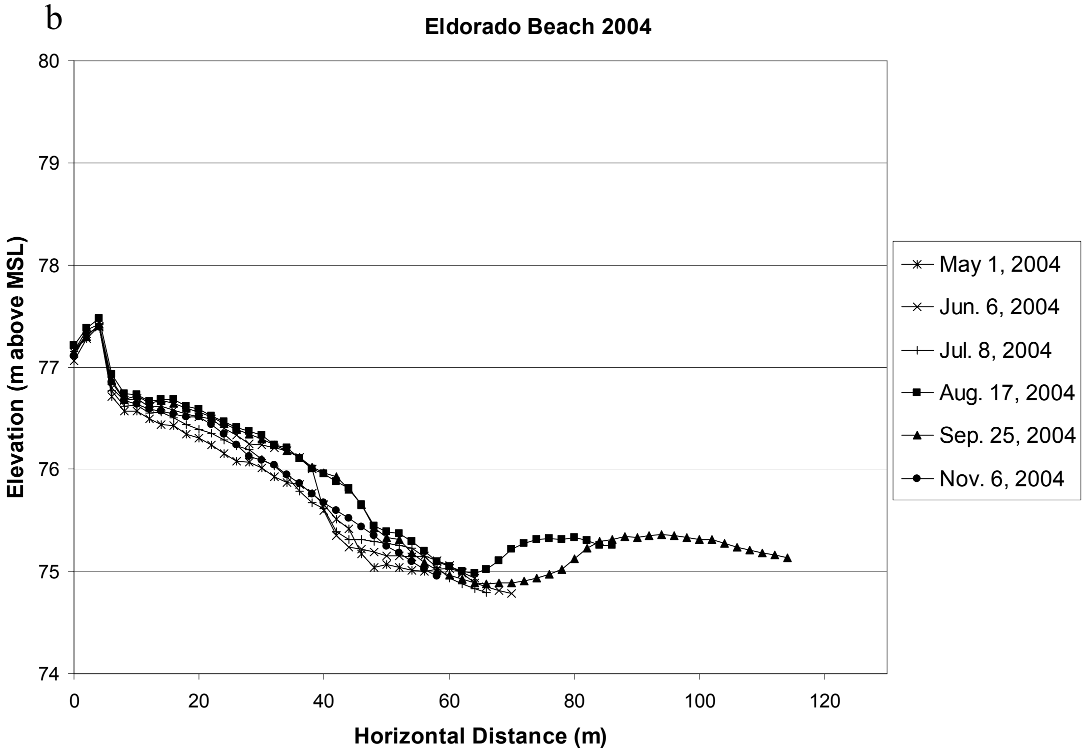


Figure 7 - a) 2004 Montario profiles showing nearshore bar from meters 60-100. b) Eldorado profiles from the same year show build-up in the vegetated area (0-10m) and a nearshore bar in late summer.

Short-term changes in beach profiles due to storms

Several wind storms occurred during the study period. Wind data retrieved from the NOAA Data Buoy on Lake Ontario north of Rochester (National Data Buoy Center, n.d.) showed that wind speeds (8 minute averages) over 40mph occurred for much of the afternoon and evening of October 15, 2003. Winds were dominantly westerly, and gusts up to 61mph were recorded. During the storm, lake levels measured at the Oswego and Cape Vincent, NY gages increased by about 0.25m. On November 13, 2003, winds increased to over 30mph during the early morning hours, and from 10am to 10pm, average wind speeds were over 40mph with gusts of 66mph (70-81mph gusts were recorded at SUNY Oswego). Winds were initially west-southwesterly, and became westerly by evening. Winds in e_cess of 30mph persisted until about noon the ne_t day, and eventually became northwesterly. During the height of the storm, lake level increased by 0.3m at Oswego and 0.45m at Cape Vincent. Beach profiles were measured at Eldorado and Southwick Beach a few days after the October storm, and profiles were measured at all locations except Selkirk on November 14, 2003.

The October, 2003 wind storm had little effect on Southwick Beach, but on Eldorado Beach, a large accumulation of zebra mussel shells was deposited as a result of the storm (Fig. 7). In places, the mussel shells formed accumulations more than 0.2m thick. Zebra mussels flourish on the bedrock at the lake bottom just a few hundred meters offshore as well as in shallow water along the rocky coast to the north. Therefore, it is fairly common to find smaller accumulations of mussel shells at Eldorado in the springtime, probably due to deposition from melting ice. Zebra mussel shells are found on all beaches, but in much smaller numbers.

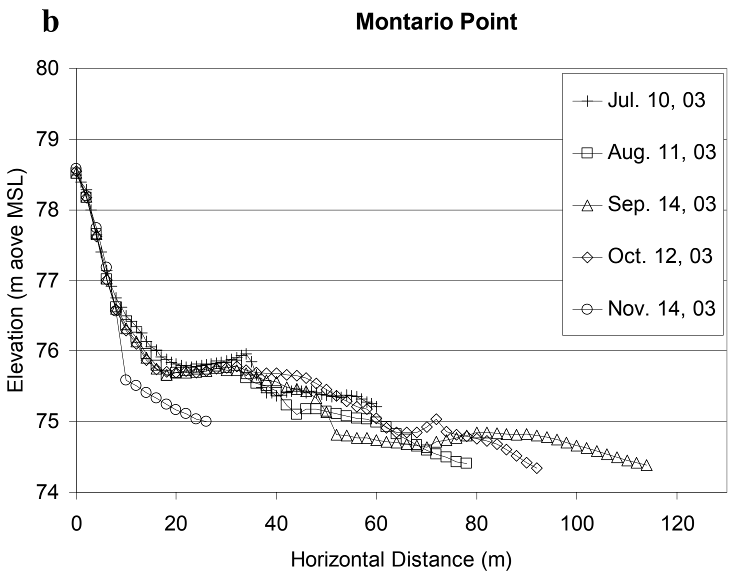
The larger, November, 2003 windstorm also affected some beaches more dramatically than others. The greatest impact of the November storm occurred at Montario Point where waves eroded the beach and the base of the dune (Fig. 8a, b). Wave ripples were observed on the beach at Eldorado (Fig. 8c) and Southwick (Fig. 8e) beaches, indicating that water had covered much of beach to some depth, but beach elevations were only slightly altered (Fig. 8c,h). Storm surge also inundated Sandy Island Beach (Fig. 8f), and cobbles were observed over much of the beach following the storm (Fig. 8g). However, the November profile at Sandy Island Beach was similar to the October profile, except that the beach face increased in elevation in November relative to October (Fig. 8i).

On November 5, 2004 there was a windstorm that produced westerly winds with speeds greater than 30 mph persisted from about 1am until midnight, with gusts up to 53 mph. Hourly lake levels rose by about 0.25m and 0.30m at the Oswego and Cape Vincent gages, respectively. Beach profiles were measured at Eldorado, Southwick, Montario Point, and Sandy Island beaches the next day, but none of the profiles showed any significant changes.

Figure 8 - Zebra mussel shells deposited during the windstorm of Oct. 15, 2003. Photo was taken on Oct. 18, 2003.

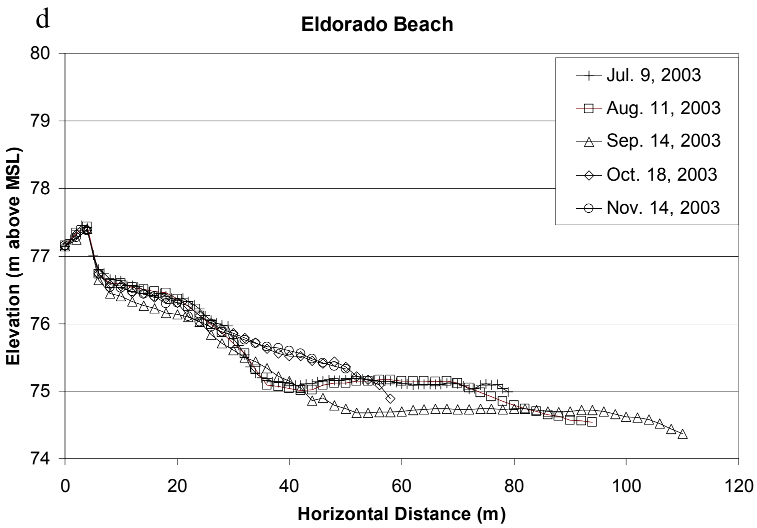

a

b

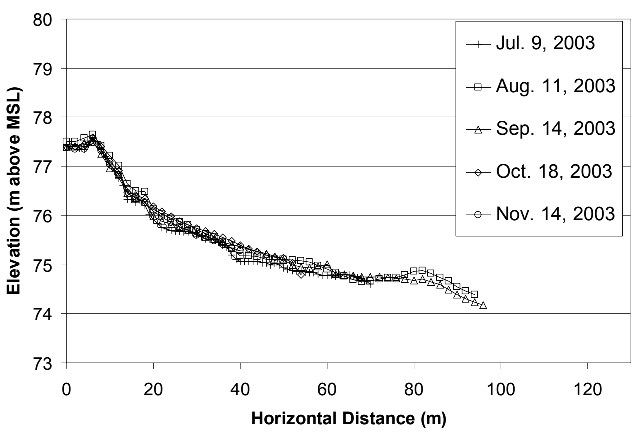

Montario Point
80
79
78
77
76
75
74
Elevation (m aove MSL)
0
20
40
60
80
100
120
Horizontal Distance (m)
Jul. 10, 03
Aug. 11, 03
Sep. 14, 03
Oct. 12, 03
Nov. 14, 03


c

d


Eldorado Beach
80
79
78
77
76
75
74
Elevation (m above MSL)
0
20
40
60
80
100
120
Horizontal Distance (m)
Jul. 9, 2003
Aug. 11, 2003
Sep. 14, 2003
Oct. 18, 2003
Nov. 14, 2003

h



i

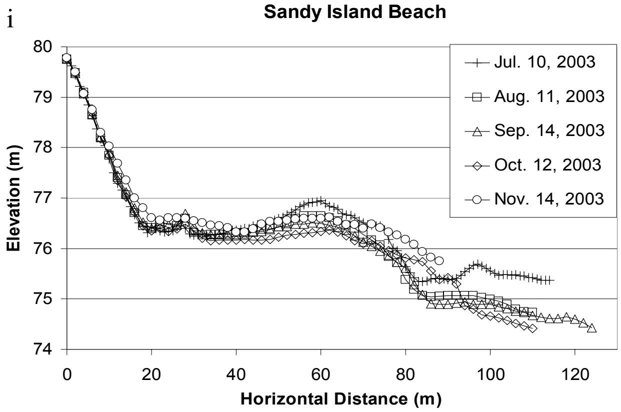


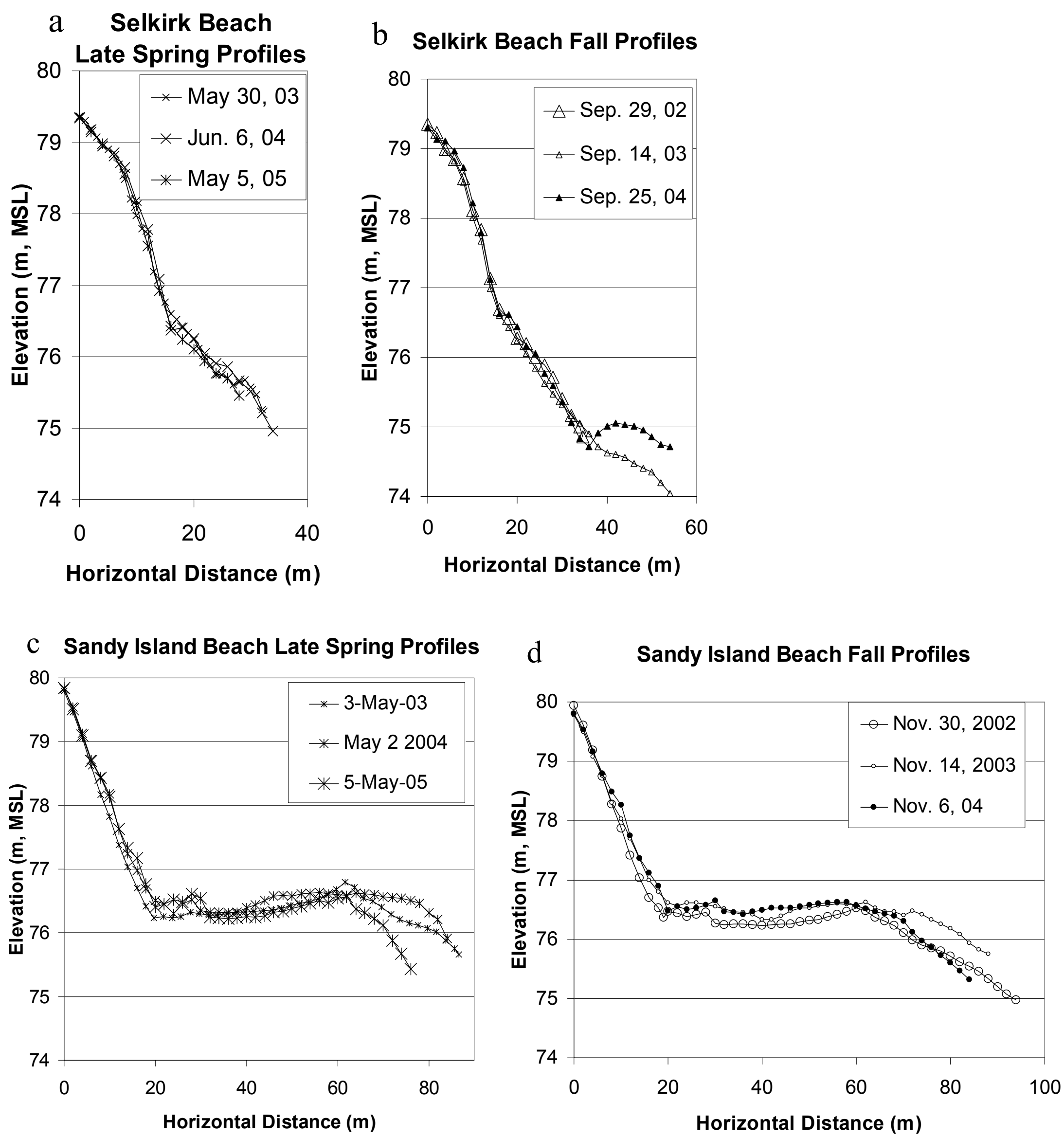
Figure 9 - Effects of windstorms on eastern Lake Ontario beaches. a) Montario Point beach, on Nov. 14, 2003. Note scarp at base of dune. b) Montario Point profile measured on November 14, 2003 shows erosion on the beach and lower dune slope. c) Wave ripples on Eldorado beach, November 14, 2003. d) Eldorado beach profile for November 14, 2003 is similar to the one measured in October. e) Southwick Beach State Park on Nov. 14, 2003. f) Sandy Island Beach at the height of the wind storm on Nov. 13, 2003 (photo courtesy S. Bonanno), and g) the following day. h, i) Comparisons of beach profiles measured on Nov. 14, 2003 with previous profiles measured at Southwick and Sandy Island Beaches, respectively.

Long-term changes in beach profiles

In order look for longer-term changes in beach profiles, comparisons were made between profiles measured in successive years at the same time of year for each beach profiling location. Fig. 10 shows profiles that were measured during late spring/early summer (higher lake levels) and fall (lower lake levels) at each location.

The Selkirk profiles (Fig. 10a,b) show the least change of all the profiling localities. The coarser grain sizes on this beach has precluded any significant modification of the beach during the study period. The increase in elevation on the Sept. 2004 file (Fig. 10b) resulted from formation of a sand bar as lake level dropped and that is not likely to have survived into the next season.

Comparisons of late spring/early summer and fall profiles measured at Sandy Island Beach (Fig. 10c,d) show that elevations on the swash face, beach berm and back-berm portions of the beach fluctuated over the 3-year period. Only on the vegetated dune slope was an increase in elevation from 2002/2003 sustained into 2004/2005. This net accretion indicates that the efforts of the conservation groups that planted the beach grasses there were effective in stabilizing the slope.



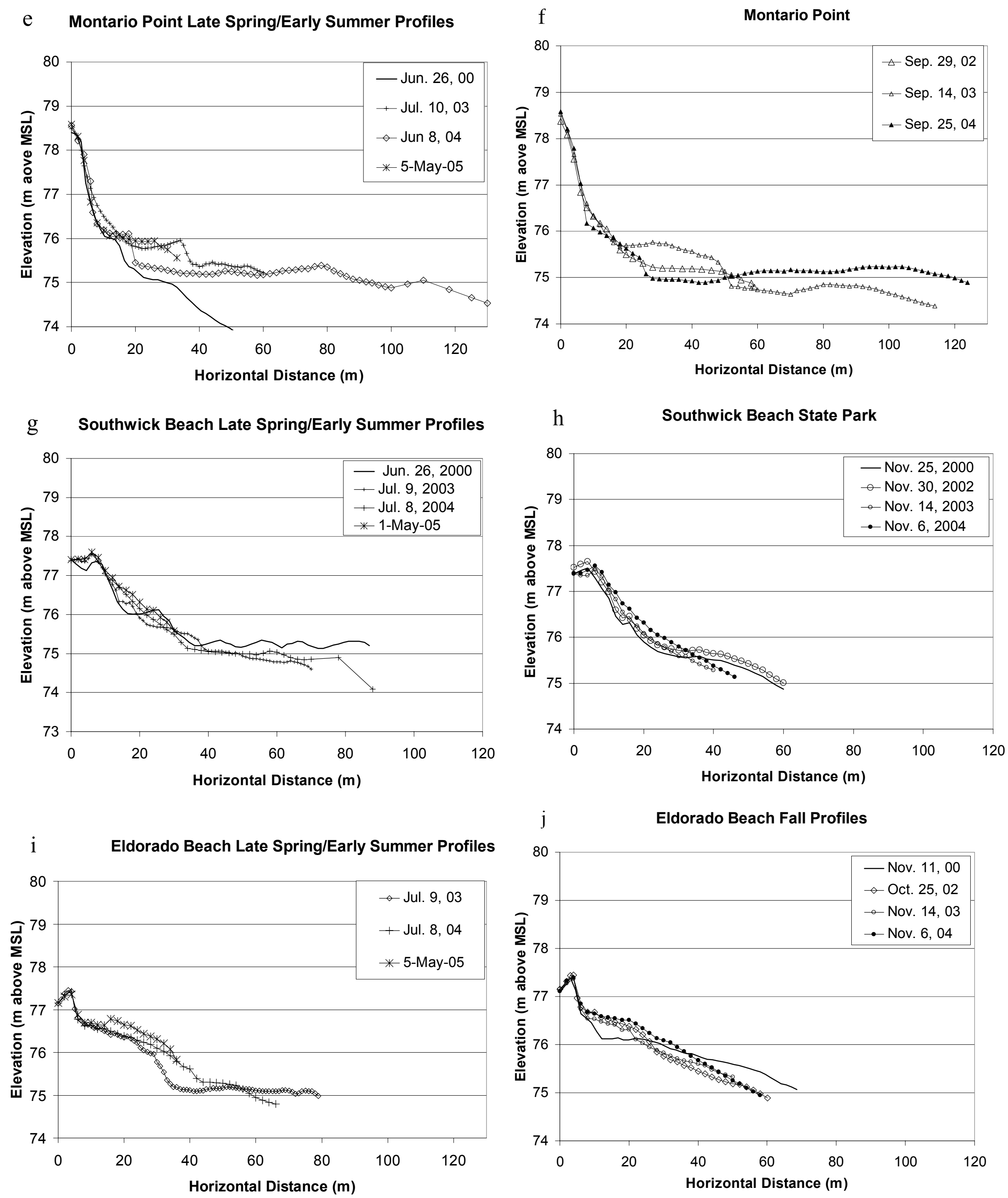


Figure 10 - Comparisons of beach profiles made in successive late spring/early summer surveys at each beach profiling location.

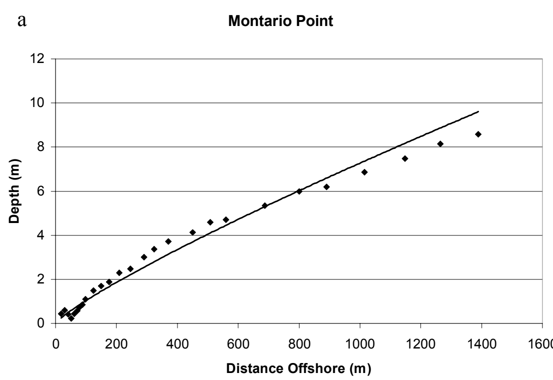
At Montario Point a beach profile measured in June, 2000 during an earlier study is available for comparison. Figure 10c shows that the landward portions of the beach and the beach face had been prograding lakeward from 2000-2003, but erosion during the November, 2003 windstorm scoured the beach and base of the dunes below the June, 2003 levels (Fig. 10d). The July, 2004 and May, 2005 profiles show that the beach has largely recovered from the storm, as the May, 2005 profile elevations are near or above those measured in July, 2003. Comparisons of the July and November, 2004 profiles with earlier ones indicates that the beach had not recovered fully by those dates.

Southwick Beach profiles show only minor changes during the 2002-2005 study period (Fig. 10g,h). One consistent trend is deposition in the area vegetated by beach grass just lakeward of the dune. The vegetation appears to be effective at trapping windblown sand, and the simple string fencing is helping to protect this area from damage by foot traffic. In the nearshore zone at Southwick Beach, elevations on late spring/early summer profiles from 2003 and 2004 show a decrease of about 0.2m relative to those measured by other workers in 2000. The profile measured in Fall, 2000 indicates that the beach was wider and higher at that time, compared to the 2002-2004 period. Seasonal changes in beach profiles seem to have been greater during 2000 than during the 2002-2005 study period.

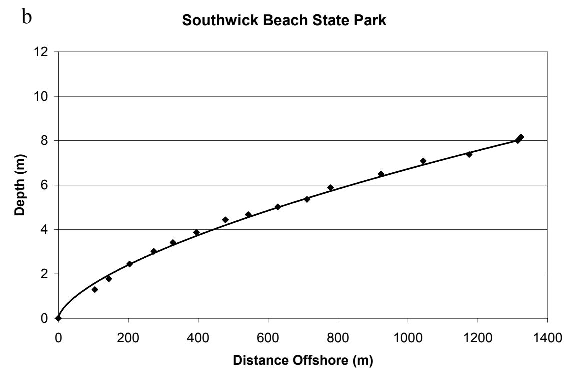
Eldorado profiles (Fig. 10i,j) show that sediment has accumulated in the area vegetated by dune willow just lakeward of the dunes during the 2002-2005 study period. A profile measured in November, 2000 by other workers shows that accumulation of more than half a meter occurred in this vegetated zone between 2000 and 2002. The May, 2005 profile shows an increase in elevation of 0.1 to 0.25m due to the accumulation of sand and zebra mussel shells in the vegetated area and on the beach berm. When the November, 2000 profile is compared with more recent ones it appears that, like at Southwick, the 2000 profile was lower in the vegetated part of the beach but higher near the water edge.

Figure 11 shows beach profiles measured from a boat at 3 of the 5 profiling localities. The profiles from Montario Point and Southwick Beach State Park generally have the shape predicted for equilibrium beach profiles, $d = Ax^b$ (Bruun, 1962). The Eldorado profile does not have this form because the sand sheet thins rapidly offshore exposing bedrock on the lake bed. Power functions fit to the Montario Point and Southwick profiles yielded values for the exponent b of 0.8458 and 0.6403, respectively. These values are in the range found by Wood, et al. (1994) for Lake Michigan profiles, and the Southwick exponent is quite close to the predicted value for equilibrium profiles of 0.67. The values obtained for A were 0.0211 and 0.0806 for Montario and Southwick, respectively. The R^2 parameter for the curve fits was 0.935 for Montario and 0.999 for Southwick. These preliminary results from profiling into deeper water are encouraging, and we hope to continue repeating these profiles at least twice per year, in early summer and late summer, in order to assess seasonal and longer-term changes in profiles.

a



b



c

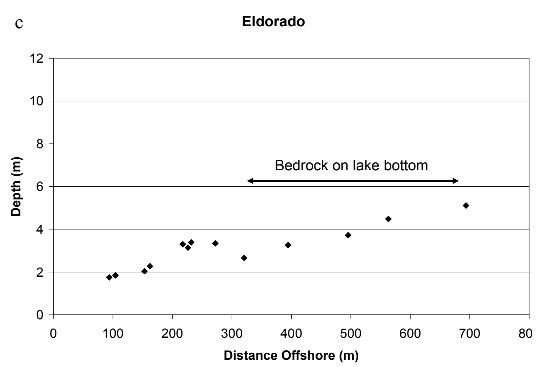


Figure 11 - Beach profiles measured from boat, August 14, 2004. a) Montario Point b) Southwick Beach State Park; c) Eldorado. Trendlines on the Montario Point and Southwick profiles are of the form $d = Ax^b$.

Sandy Pond Inlet

Sandy Pond Inlet connects North Pond to Lake Ontario (Fig. 12). Boaters use the sandy beach on the north side of the inlet for sunbathing and picnicking. The vegetated area and low dunes just south of the inlet channel is a bird sanctuary, and access is restricted. Farther south, there is a washover channel and fan just south of the line of low dunes closest to the inlet. This area is used for boaters to walk from the pond side to the lake side for swimming and sunbathing (Fig. 13). During late summer, declining lake level and deposition in the inlet channel make it increasingly difficult for larger boats to navigate through the channel as the season progresses. The area between the inlet and Carl Island becomes shallow and difficult to navigate as well. In 2004, a group of private citizens obtained permits to dredge the inlet in order to maintain channel navigability.

In order to study changes in bed topography, the inlet was mapped using an electronic distance meter (EDM or "total station"; Fig. 14) in July 2003 during high early-summer lake level, and at lower lake levels in September, 2003 and September 2004. A limited amount of dredging was done in late summer, 2004, before the September, 2004 map was made (E. Bernstein, pers. comm.). Survey data were loaded into Gocad ™ drafting software and contoured. Figure 15 shows topographic maps made from the EDM surveys of the channel in July 2003, September 2003, and September 2004. It should be noted that few areas on land were surveyed, so the topography depicted in land areas is poorly constrained.

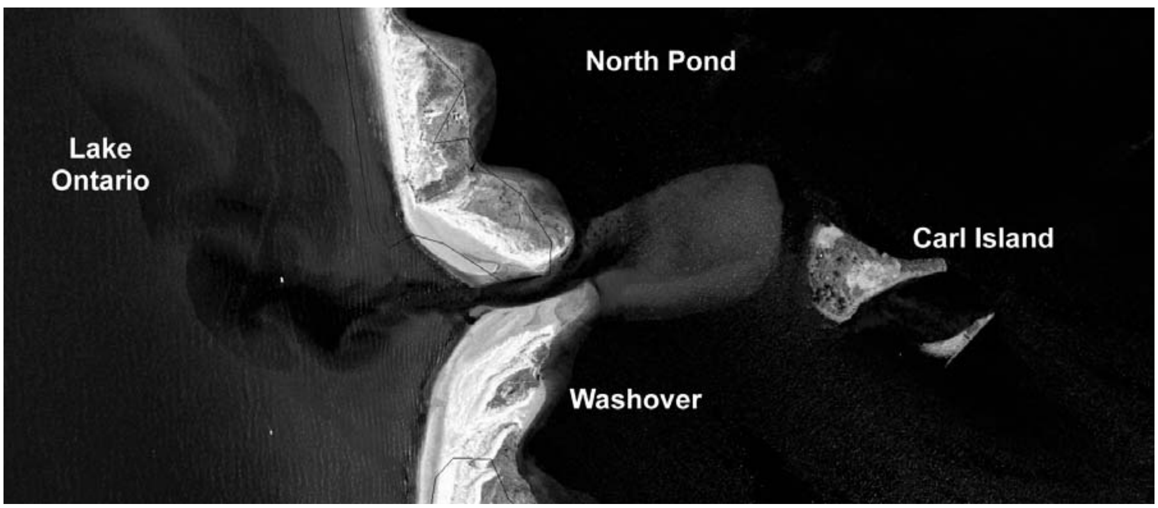


Figure 12 - Aerial photograph of Sandy Pond Inlet, April 2003, north to top of photo. Note sand shoal on pond side of inlet. Source: NY GIS Clearinghouse (n.d.).

Figure 13 - Photomosaic of washover channel just south of Sandy Pond Inlet, view toward North Pond.

Comparing the July and September, 2003 maps, one can see that the shoreline shifted lakeward and pondward as lake level dropped. Using Gocad ™, elevations from the July map were compared with September map elevations to identify areas of erosion and deposition (Fig. 16a). Bed elevations increased on the lake side and pond side of the inlet channel between July and September. Deposition on the pond side of the channel occurred on the north and south sides of the channel, but occurred over a slightly larger area on the south side. Water depths measured in the inlet channel exceeded 8 ft. in July, 2003, but the greatest depths measured in September were less than 6 ft.

Figure 14 - EDM used to survey Sandy Pond Inlet

Dramatic changes occurred in the inlet between the September 2003 and September 2004 surveys (Fig. 16b). The transition zone, where the shoreline angles toward the inlet channel, became widened and shifted landward. The south shore of the transition zone retreated to the edge of the low sand dunes, and on the north side the water edge shifted eastward by nearly 100m. Erosion lowered lakebed elevations on north and south portions of the transition zone that had been exposed in September 2003 by more than a 0.3m on the south side and by more than half a meter in many places on the north side. The inlet channel in 2004 was reduced to less than one-half its length the previous year. Deposition occurred where the lakeward end of the inlet channel had been in 2003. Greater buildups of sediment occurred on the sides of the inlet channel. Deposits up to 1m thick helped to build up the channel margin on the north side of the inlet channel. On the south side of the channel, an elongate deposit more than a meter thick extends lakeward of the inlet mouth, and deposition increased the beach area on the pond side of the inlet south of the inlet channel. Erosion lowered the bed in the inlet channel by more than half a meter, but depths in the channel were similar in 2003 and 2004.

The fall, 2003 wind storms caused many of the changes observed between September 2003 and September 2004. Photographs taken a day after the November 13-14, 2003 wind storm show that the inlet mouth had widened compared to earlier in the year (Fig 18). Photographs taken from approximately the same location on the south shore of the inlet channel in July, 2003 and November, 2003 show that widening of the transition zone and shortening of the inlet channel occurred during Fall, 2003 (Fig. 17). Modifications of the channel shoreline did occur between November, 2003 and September 2003, and it is not possible from photographs to determine the extent of deposition and erosion due to the storms. A local resident indicated that there was significant storm surge on North Pond during the October, 2003 wind storm (T. Jones, pers. comm.). Wave ripples were observed in the washover channel the day after the November, 2003 windstorm, indicating that water levels had been quite high during that storm as well.

Weir (1977) mapped two inlets that were open between North Pond and Lake Ontario at that time, one at the location of the present-day inlet and the other roughly 300m to the south. The inlet to the south had closed by 1978 (DelPrete, 1995). Weir's map shows sheet piling located on the eastern side of the barrier north of the north (i.e. present-day) inlet channel. The southern edge of the sheet piling is now located on the western side of the barrier approximately 180m north of the inlet channel. This indicates that the barrier has shifted landward at a rate of at least 0.93 m/y. The rate is likely higher as Weir's study indicated recession rates of about 1.3 m/y since the late 1800s.

Since 1829, the inlet between North Pond and Lake Ontario occupied at least 5 different locations along the barrier south of the present day inlet (Weir, 1977). Weir stated that inlets do not shift gradually along the shore in the direction of longshore drift. Instead they are stable for a period of time, and progressively lengthen. This decreases channel efficiency due to increased friction and favors inlet abandonment. When a new inlet is established during a large storm, for example, the older less efficient inlet is abandoned. Presumably then, the shortening of the inlet that occurred between the 2003 and 2004 surveys would increase the likelihood that the inlet will remain in its current location. However, if storms similar to the Fall, 2003 wind storms were to occur at times of higher lake levels, erosion may well create a new inlet. Candidate locations for a new inlet would be areas where the barrier is narrow, large dunes are absent, and vegetation is sparse. One such candidate is the washover area just south of the inlet, and another is an older washover to the north of the present inlet location.

Sediment transported by longshore drift and onshore transport during storm surge into Sandy Pond Inlet and other inlets in the eastern Lake Ontario barrier system may be stored in the inlet or pond for long periods of time. Although some sediment may return to the lake during spring snowmelt or other large runoff events, much of this sediment is essentially lost from the barrier system. The direction of longshore drift on the Lake Ontario coast is not precisely known at most locations, but recent work has indicated that Sandy Pond Inlet may receive sediment from both southward and northward longshore drift (Woodrow, et al., 2002). Additional study is needed to determine the quantities of sediment that are transported from adjacent beaches via longshore drift and deposited in pond and wetland areas.

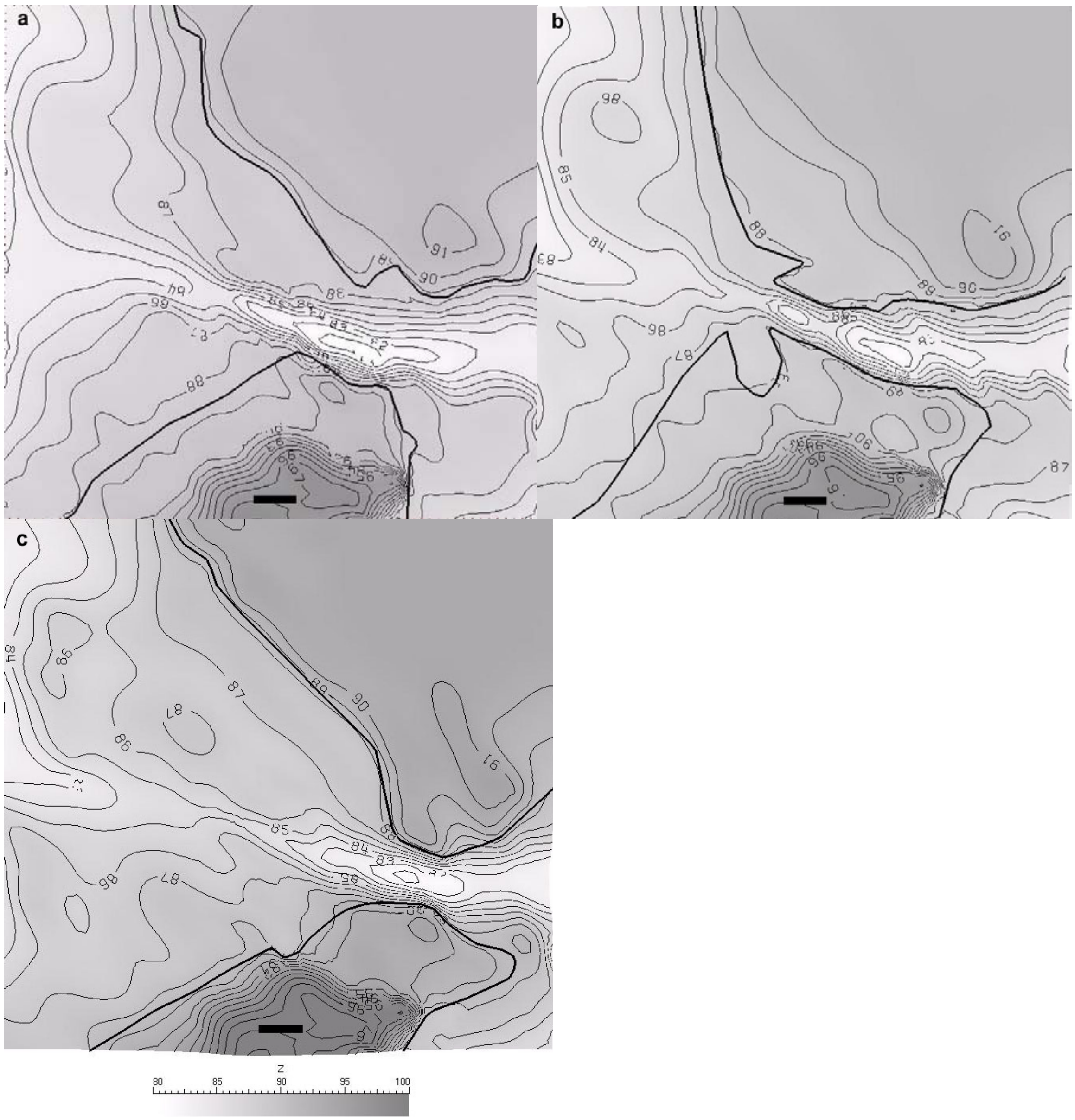


Figure 15 - Topographic maps of Sandy Pond Inlet on a) July 21, 2003, b) September 12, 2003, and c) September 21, 2004. Contour interval is 1 ft. Scale bar in each map represents 100 ft.

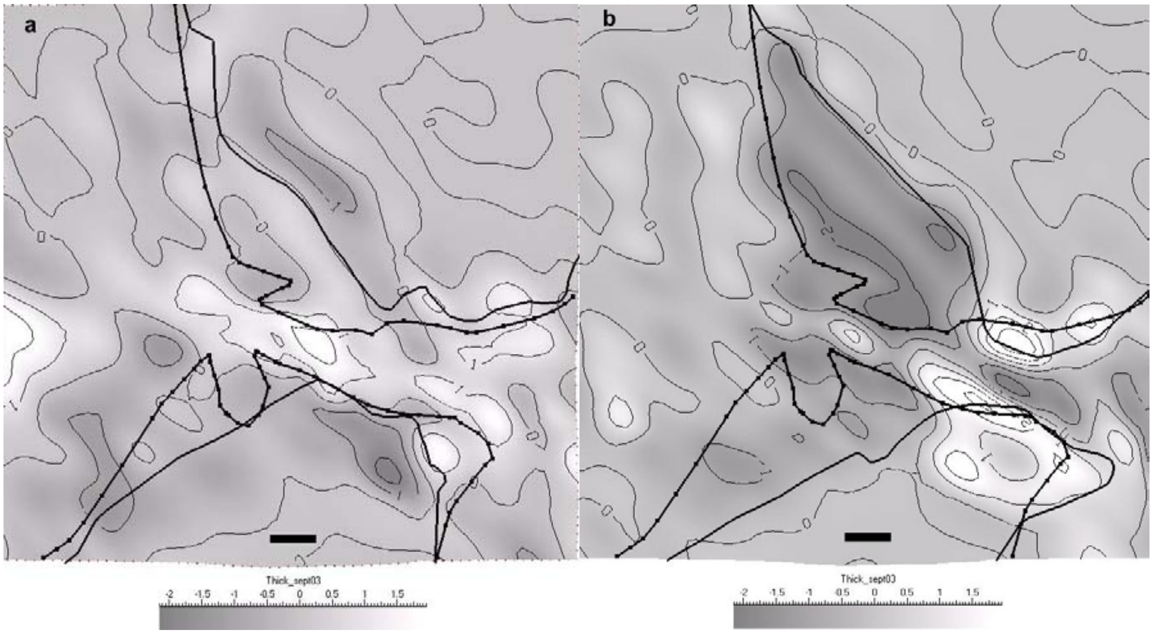


Figure 16 - Shaded relief map showing change in topography and position of water edge a) from July, 2003 (plain solid line) to September, 2003 (solid line with dots) and b) from September 2003 (solid line with dots) to September 2004 (plain solid line). Contours represent change in elevation in feet. Scale bar in each map represents 100 ft.

Figure 17 - Views of the transition zone of Sandy Pond Inlet in a) September, 2003, b) the day after the wind storm in November, 2003, and c) September, 2004. View to north.

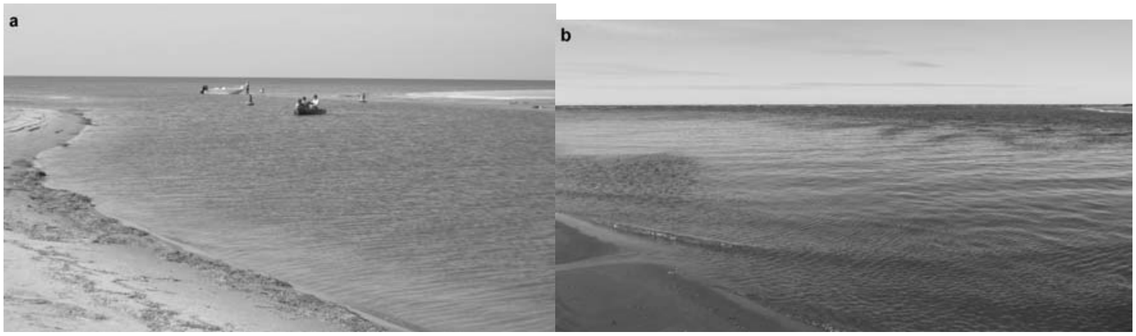


Figure 18 - View of inlet channel, looking lakeward in a) September, 2003 and b) the day after the windstorm of November, 2003.

SUMMARY

Monthly beach profiling revealed short-term, seasonal and longer-term variations in beach profiles along the eastern Lake Ontario coast. Profiles are eroded in the shallow nearshore zone up to the beach berm as lake level rises in the spring. As lake level drops during the summer and into autumn, the beach widens as the beach face progrades lakeward. The elevation of the beach face declines as the beach face progrades. In most locations nearshore bars were observed during the summer during at least some years, and in some cases profiles revealed that bars were eroded as the beach face prograded. This supports Bruun's (1983) contention lake level drops may lead to bar formation as sediment in the nearshore zone is mobilized, and that deposition of that sediment would occur in landward areas.

Storms had a significant impact on some beaches and little to no impact on others. Erosion of the beach and base of the dune at Montario Point during the November, 2003 wind storm was the most significant alteration observed. It took nearly a year and a half for the beach to recover.

Only minor longer-term changes in beach profiles were observed during the Fall, 2002-Spring, 2005 study period. Some beaches (Montario Point and Eldorado) showed larger fluctuations in profile shape than others, but these fluctuations showed no overall trend. The only consistent trend observed was accumulation of sediment in vegetated portions of the beach profiling transects at Sandy Island, Southwick and Eldorado. This shows that efforts to maintain vegetative cover by planting beach grasses and/or erecting fencing to keep people off vegetated areas are having a positive effect.

Beach profiles measured from a boat extend into deeper water show that the profiles at Montario Point and Southwick have the shape expected for equilibrium profiles. Best fit power functions to the profiles measured at these two locations yielded results similar to other studies on Great Lakes coasts. The exponent in the power function at Southwick was close to the value predicted for equilibrium profiles, but the Montario Point profile was steeper than the equilibrium profile. This may be due to modification of the profile due to the Fall, 2003 windstorms. More profiling in deeper water is needed to fully understand the nature of erosion and deposition on the eastern Lake Ontario shore and whether equilibrium profiles are maintained in this area.

Sandy Pond Inlet provides a vital navigational link between North Pond and Lake Ontario. Reduced water depth in the inlet channel between the inlet and Carl Island hinders navigation. Detailed mapping revealed that as lake level drops, channel depths are reduced by deposition on the lake side of the channel as well as lowered water level. Sediment deposited in the channel is derived from erosion of the shallow water transition zone on the lake side of the inlet during falling water levels. Wind storms during Fall, 2003 enlarged the transition zone and shortened the inlet channel to less than one-half its previous length as the transition zone shifted nearly 100m landward. Contributions of sediment from longshore drift from areas of the coast north and south of the transition zone could not be determined from the present study. Such information is needed to determine the role that sedimentation in inlets and ponds plays in the beach budget for eastern Lake Ontario.

ACKNOWLEDGEMENTS

Funds from a Student – Faculty Collaboration Challenge Grant from SUNY Oswego, and a State Water Quality Improvement Project Grant from the New York State Department of Environmental Conservation helped fund this project. Dawn Schmidt, Natural Resources Conservation Service, provided the equipment and expertise running the total station surveys of Sandy Pond Inlet. Tom and Greg Jones provided boat transportation and, with Sarah Green and Natalie Menielly, assisted with mapping Sandy Pond Inlet. Brian Willis, Texas A&M University, provided equipment and ran the total station for our boat beach profiles, and provided training on Gocad. The following people ably assisted in beach profiling: Audra Crocetti, William Deerfield, Jill Draper, students in Gabel's Great Lakes Environmental Issues course and Introductory Geology labs, and Raymond and Elizabeth Willis. We thank Randall Korwin, Dan Leary, and Dan Whitehead for kindly allowing us access to their property.

0.0	Parking lot near Hewitt Union and Culkin Hall, SUNY Oswego. Proceed south out of lot, turn right onto West End Ave. then left onto Sweet Road.
0.2	Turn left (east) onto State Rt. 104
1.1	Bear right at 5-way intersection (traffic light) to stay on Rt. 104 (AKA Bridge Street)
10.1	Turn left onto Rt. 104B
15.6	Rt. 104B becomes State Rt. 3
36.3	Turn left onto Eldorado Road
36.5	Stop sign; go straight. Road becomes gravel road.
37.5	Yield sign; turn left.
37.8	Bear left
38.2	STOP 1 Black Pond Wildlife Management Area (gravel parking lot)

We will walk from the parking lot along the boardwalk that takes us through the wetland adjacent to Black Pond onto the barrier. Note the tall dunes at this location. The boardwalk takes us over the dunes to the beach. At this location we will examine the wetland and pond environments of the barrier system as well as the beach and dunes. Zebra mussel and quagga mussel shells are abundant here, particularly in the spring or after storms. A short hike on the beach to the north will take us to the inlet that connects Black Pond with Lake Ontario.

40.1	Turn right (south) onto State Rt. 3
42.0	Intersection with Rt. 197 and Southwick Beach State Park access road. Turn right onto Southwick Beach State Park access road.
42.9	STOP 2 Southwick Beach State Park picnic area adjacent to beach.

We will walk along the beach to our beach profiling locality. Here we can observe some of the efforts that have been made to reduce human impacts on the beach and dunes. This will also be our lunch stop.

43.8	Turn right (south) onto State Rt. 3
55.7	Turn right (west) onto Rainbow Shores road.
57.4	Turn left (south) onto dirt road that parallels the shoreline.
58.4	STOP 3 Deer Creek Marsh (grass parking lot)

A short trail will lead us to an viewing platform overlooking Deer Creek Marsh where we can discuss some of the human impacts on coastal wetlands. We will then proceed to the cobble beach.

61.6	Turn left (north) onto State Rt. 3
63.5	Turn left (west) onto Oswego County Rt. 15.
65.8	STOP 4 Sandy Island Beach State Park

Sandy Island Beach is the location of another beach profiling transect. We will discuss the many changes in management this beach has undergone over the decades. Tremendous efforts have been made by government agencies and conservation groups to improve conditions at this beach. One such effort involved moving a vast quantity of sand that had blown from the tall aeolian dune on the eastern margin of the park into North Pond. The sand was trucked to the lakeward side of the dune. Beach grasses were planted by conservation groups to help stabilize the dune. Cars are no longer permitted on the beach, except for access to residences on the barrier. Fencing was erected to help stabilize sand and keep people off the dunes. Our beach profiling indicates that these efforts have helped protect the dunes.

STOP 5 (Optional) We will walk along the beach approximately 2 miles north to Sandy Pond Inlet. There we will examine the beach, inlet, pond and washover.

68.1	Turn right (south) onto State Rt. 3
83.5	Turn right (west) onto State Rt. 104
93.4	Sweet Road (main entrance to SUNY Oswego campus)
93.6	Hewitt Union, SUNY Oswego

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Trip B-4

EURYPTERIDS AND FACIES CHANGES WITHIN SILURIAN/DEVONIAN 'EURYPTERID BEDS' OF NEW YORK STATE

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Perhaps nowhere else in the world are eurypterids found in the abundance they are within the Silurodevonian strata of New York State. More importantly, little-understood cyclic sequences, mostly within an evaporite basin, preserved a multitude of distinctive eurypterid faunas. Many of these are currently known, and several now known, need to be better understood. This is especially true of Akron-Cobleskill horizons (see Ciurca 1978, 1994) that are providing abundant remains of successor faunas to those of the well-known Bertie faunas, viz. *Eurypterus remipes* (older) and *Eurypterus lacustris* Faunas.

The purpose of this paper is to comment on Bertie Group and overlying Akron-Cobleskill occurrences – their distribution geographically and stratigraphically of microbialites and eurypterids – and suggest that much more research is required if we are to truly understand the paleoenvironment not only of the sedimentary rocks in which we find specimens, but the environment in which the animals actually lived. The association of many of the eurypterid horizons with stromatolites/thrombolites is particularly emphasized. This is an exciting new area of observation and research.

The evolution of fish, and other organisms, has taught us that given enough time, many animals evolve to occupy disparate environments within the marine realm and often within the fresh water realm (rivers and lakes). Many authors have already suggested that the various eurypterid morphologies we encounter can be attributed to their existence in a variety of environments. Yet, it has proved difficult to provide specific environments for specific eurypterids – especially faunas that contain a great variety of species (forms) co-mingled like those of the Bertie Group <u>assemblages</u>.

The discovery of numerous post Akron-Cobleskill eurypterid horizons (Ciurca 1973, 1994) and horizons within the Cobleskill Limestone (itself) of eastern New York suggested lithologically and faunally that the Bertie Group needed redefinition. Redefinition was proposed previously (Ciurca 1980) and provided for practical recognition of the Silurian-Devonian boundary across New York State from Albany westward to Buffalo and into the Niagara Peninsula of Ontario, Canada based upon the abrupt appearance of *Erieopterus*.

In addition, examination of the typical Bertie Group lithologies and faunas can help us look for and understand newly discovered eurypterid horizons. An example of this, discussed below, is the occurrence of numerous stromatolite horizons within most sequences (See section on Microbialites within the Bertie Group).

The eurypterid-bearing horizons mostly begin in basal Salina Group units that are more argillaceous than Bertie Group (and higher) horizons. While not much is known about facies relationships within these lower horizons (e.g. Pittsford and other members of the Vernon Formation), some observations have been made (Ciurca in NYSGA 1990). Of course, during the deposition of the Salina sequence, the thick redbeds of the Vernon-Bloomsburg to the south and east were important to the paleoenvironmental setting in which we find the eurypterids. To the west of the basin were the 'Niagaran' reefs, some of which are believed to have continued growth into late Salina and Bertie time.

<u>All</u> of the Late Silurian and Early Devonian eurypterid 'deposits' indicate formation in shallow water and most are associated with features of desiccation represented by the numerous (cyclical) horizons bearing mudcracks, occasional ripple marks, rip-up clasts and probable strandline deposits. Uniquely, the Yale Peabody Museum ory is now the repository of a great collection of the sedimentary structures and sedimentological features, including a suite of associated lithological samples personally observed and collected over a period of about 40 years.

AKRON-COBLESKILL (*EURYPTERUS*) FACIES CHANGES

The type Cobleskill Limestone occurs in eastern New York while the type Akron Dolostone occurs in western New York. Between the two regions, important and little-understood facies changes take place. These changes are important to understand if we wish to gain insight into the evolution of *Eurypterus* and associated fauna. Part of the problem exists because various authors have interpreted, especially the Cobleskill Limestone, differently. The limestone is not one continuous bed or lithology. Westward from the type section, the interval is replaced by numerous *Eurypterus*-bearing beds that are useful in delineating the Silurian-Devonian boundary in the northeastern United States. Above these lateral and vertical changes, *Erieopterus* makes its appearance during the Helderbergian Transgression.

If anything characterizes the Cobleskill Limestone, it is its brachiopod fauna. While associated forms including horn corals, stromatoporoids, favositids and cephalopods are often important and typically marine forms; the brachiopods cross facies and are even associated with *Eurypterus* in some areas. The abundance of *Eurypterus* and associated eurypterids within post-Bertie units has generally gone unrecognized for over 100 years.

Until a redefinition of the eurypterid-bearing Akron/Cobleskill suite of litho- and biofacies is completed, it is difficult to discuss thoroughly the distribution of the eurypterid faunas, the emphasis of this work on a complete survey of the distribution of eurypterid faunas above the faunas of the Bertie Group. However, several observations can be recorded. The Moran Corner Waterlime (Ciurca 1973, 1994) is currently one of the most important of the post-Akron/Cobleskill units only because so much material has been recovered already and most eurypterid types have been recognized, viz. *Eurypterus* (2 species), *Acutiramus*, *Dolichopterus* (probably 2 species of dolichopterids) and an associated fauna not unlike that of typical Bertie Group associations

Figure 1 - Stromatoporoid biostrome, generally regarded as Cobleskill Limestone, within Owasco Outlet near Auburn, New York. Unnamed overlying beds are rich in eurypterid remains and may represent a channel through the Akron-Cobleskill sequence in this area. Presumably, the stromatoporoid beds are the same as those found on Frontenac Island in Cayuga Lake where the Cobleskill fauna is well known and consists of many types of brachiopods, rugose and colonial corals, pelecypods and cephalopods.

The critical area, probably more central in the basin, is Auburn-Syracuse. *Eurypterus* occurs above stromatoporoid biostromes that would be interpreted, as Cobleskill Limestone and that are also present on Frontenac Island in Cayuga Lake. Above this is a sequence with still other *Eurypterus*-bearing units including a brachiopod fauna that would also be interpreted as "Cobleskill," but which is a chert-bearing dolostone. It is possible that at least one of these *Eurypterus*-bearing units represents a large channel cutting through the complex of Akron/Cobleskill sediments in a relatively north-south orientation. Stromatoporoids extend westward at least to the Honeoye Falls area and corals to Buffalo and Fort Erie, Ontario, Canada. The zonal brachiopod *Eccentricosta jerseyensis*, occurs sporadically throughout the extent of this complex of post-Bertie units but also occurs within the (Bertie) Williamsville Formation in western New York and Ontario, Canada.

It is likely that the *Eurypterus lacustris* Biofacies developed behind Akron/Cobleskill deposition of stromatoporoid biostromes and probably small patch reefs. There is a distinct change in fauna within the Williamsville Waterlime itself as we go from central to western New York. The west is dominated by the *Eurypterus lacustris* Fauna, and the east by the *Paracarcinosma scorpionis* Fauna (presumably a deeper water regime). At the same time, the ubiquitous *Lingula* seems to disappear almost completely in the Fort Erie, Ontario region.

MARTISCO REEF COMPLEX

A road cut near Marcellus Falls, New York, exposes massive beds of cherty dolostone that are termed "Cobleskill Formation" in the literature. However, lithologically there is no similarity between the beds exposed here and the type Cobleskill Limestone of eastern New York. Additionally, near the top of the beds exposed at Marcellus falls, and just below the overlying Chrysler Formation, occurs a small patch reef. This reef and the associated cherty dolostone are referred to as the "Martisco Reef Complex." (from an Abstract presented before the Rochester Academy of Science Fall Paper Session at the Rochester Institute of Technology, Rochester, New York, November 15, 2003).

Irregular bedding and stylolitic contacts characterize the massive reef beds exposed here. Fossils are difficult to extract and identify due to recrystallization. However, silicified stromatoporoids preserved here provide evidence as to the nature of the mound occurring near the top of this exposure. Brecciation is prominent in certain beds and indicates the high-energy environment in which these beds formed.

The Martisco Reef Complex is capped by stromatolites and a bed of brecciated dolostone. An important eurypterid bed occurs near the top. The basal beds are not obvious at this site, but a nearby ravine reveals the expected Williamsville (or Oxbow) Formation with *Lingula* as seen at many outcrops of the unit.

A similar complex occurs in the Rock Cut Gorge near Syracuse and is herein termed the Rock Cut Bioherm. The sequence is nearly the same as that at Marcellus Falls, but the dark chert is less prominent. Large ostracods, a conulariid, and eurypterid remains have been observed above the irregular biohermal beds.

MICROBIALITES (STROMATOLITES & THROMBOLITES) WITHIN THE BERTIE GROUP

Stromatolites form extensive biostromes in some sections. Within the Bertie Group, occurrences have been uncovered from the highly eurypterid-bearing localities at Litchfield (CIURCA Locality 56) and Passage Gulf (CIURCA 57) westward into Ontario, Canada. They are common structures in many eurypterid horizons and it has taken many years to recognize just how widely distributed they are. Current research at one site, the Neid Road Quarry northeast of LeRoy, New York, has revealed 'sheets' of stromatolite mounds packed within a bed of waterlime in the upper Fiddlers Green Formation (i.e. the Ellicott Creek Breccia).

Figure 2 - A portion of the 'Martisco Reef Complex' at Marcellus Falls. Note white mineralized zones associated with silicified stromatoporoids and irregular bedding. Bands of dark chert overly the reef.

There are a variety of morphotypes recognized within the organosedimentary structures referred to here as simply stromatolites. Recently, a large ramus of the giant pterygotid, *Pterygotus cobbi*, was recovered from the waterlime just above a bed of stromatolite mounds at the Neid Road Quarry in 2005. Thus far, four eurypterid species have been observed within this sequence of uppermost Fiddlers Green Formation in this region. The most common feature observed within the strata here are bedding planes replete with what appear to be countless ripped up 'algal' clasts. This is part of the *Eurypterus remipes* Biofacies occurring across New York State.

Stromatolitic structures have now been observed within the Fiddlers Green Formation from the Niagara Peninsula of Ontario, Canada eastward all the way to the *Eurypterus remipes* localities at Litchfield and Passage Gulf in eastern New York. Enormous thrombolites occur within the Victor Member in Ontario, Canada.

Figure 3 - Stromatolite mounds at the Neid Road Quarry. The intermound areas are light-colored, eurypterid-bearing waterlimes typical of the Bertie Group. Black (presumably carbonaceous) material easily weathers off the surface of the mounds. Excavation is a work in progress.

Figure 4 - "*Pterygotus cobbi*" – Part of a very large ramus (about 20 cm) found just above a bed of stromatolite mounds at the Neid Road Quarry northeast of LeRoy, New York in the Summer of 2005. This must have been one hell-of-a pterygotid. The sediments around are replete with unusually large salt hoppers.

Figure 5 - '*Eurypterus' laculatus*, preserved in the waterlimes of the Ellicott Creek Breccia, Fiddlers Green Formation, Bertie Group. This specimen was retrieved from intermound waterlime at the Neid Road Quarry. See Ciurca (1973) for a description of the quarry. Specimen now in the Yale Peabody Museum collections.

ROAD LOG FOR MICROBIALITE-BEARING EURYPTERID SITES
SILURIAN AND DEVONIAN OF UPSTATE NEW YORK
Part 1

CUMULATIVE MILES FROM	MILEAGE LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Exit 30 Toll booth, NYST I-90 at Herkimer
0.1	0.1	Turn left at intersection. Follow NY 28 south through Mohawk to Illion
0.7	0.6	NY 28 (continue straight through town, W. Main St.
1.5	0.8	Entering Illion
2.2	0.7	Remington Arms Factory on left
2.5	0.3	Turn left (south NY-51)
4.0	1.5	Entering Illion Gorge (Steele Creek) – Entering Ordovician Time Zone
5.7	1.7	Town of Litchfield
6.1	0.4	Entering Silurian Time Zone
8.1	2.0	Vernon Red Shale
8.9	0.8	Syracuse Formation on left
10.2	1.3	Cedarville
10.6	0.4	Turn right at stop sign onto Cedarville Road
11.3	0.7	Early Devonian limestones on left
13.3	2.0	STOP 1 at Litchfield Town Hall (on right)

STOP 1: EXPOSURES OF THE LATE SILURIAN FIDDLERS GREEN FORMATION

Approximately the upper half of the Fiddlers Green Fm. (Bertie Group) is exposed along the road directly opposite the Litchfield Town Hall (Ciurca Locality 56, Cedarville at Jerusalem Hill Roads). In 1966, the exposure was a glacially polished surface on the Phelps Waterlime Member and very resistant to excavation. However, during this and many subsequent years, hundreds of specimens of eurypterids and associated fauna were obtained from this relatively small outcrop. This material is now in the Yale Peabody Museum in New Haven, Connecticut. The eurypterid-bearing units here are the Phelps Waterlime at the top of the exposure and the underlying finely-crystalline Victor Member. *Eurypterus remipes* is the common element of the eurypterid fauna here and has drawn collectors from all over to this area trying to find one.

13.3		Turn back on Cedarville Rd. to NY-51. Turn left and head back through the gorge to Illion.
24.2	10.9	Turn right (Clark St., McDonald's on right). Continue towards I-90.
25.0	0.8	Village of Mohawk.
25.5	0.5	NY-28 (continue forward).
26.4	0.9	POW/MIA Remembrance Bridge. Follow sign to I-90. Head west on Thruway to Canastota Exit 34.
69.4	43.0	Toll booth – Canastota.
69.6	0.2	Turn right on NY-13 and stop at McDonalds.
69.7	0.1	Leave McDonalds, turn right and follow NY-13 south.
70.3	0.6	Erie Canal Museum on right.
71.0	0.7	NY-5, continue south on Oxbow Road.
72.9	1.9	Cotton Road, Clockville (continue south).
73.7	0.8	STOP 2 at road cut on east side of Oxbow Road – Oxbow Falls on right.

STOP 2: EXPOSURES OF THE LATE SILURIAN AND EARLY DEVONIAN STRATA

Stratigraphically, this is an important section. At the base is the uppermost Fiddlers Green Fm. (Bertie Group) with eurypterids and salt hoppers. Overlying strata include the Forge Hollow, Cobleskill, Chrysler, Thacher and part of the Olney Formations. There are at least three eurypterid horizons present in the section. The Silurian/Devonian boundary lies within the lower part of the Chrysler Fm. here – this lower portion is rich in strontium (celestite) at Chittenango Falls. This section, again, exhibits the importance of microbialites in sections preserving eurypterid remains. In the upper portions of this section, within the 'Thacher' here, are more than two thick beds (more than 2.5 m) of stromatolites (including digitate forms) and thrombolites intimately associated with the Early Devonian eurypterid, *Erieopterus*.

73.7		Turn back to NY-5.
76.9	3.2	NY-5, Village of Canastota. Continue north to I-90.
78.2	1.3	Thruway toll booth (end of Part 1).

Part 2

0.0	0.0	Start at Exit 34A toll booth (from the east) to I-481. Follow I-481 south towards Jamesville.
6.6	6.6	Jamesville exit (follow road towards Jamesville).
7.5	0.9	Turn right (Jamesville Toll Road).
8.2	0.7	STOP 3 near road cut in Late Silurian dolostones.

STOP 3: ROCK CUT BIOHERM (NEW NAME)

This road cut exposes what is here interpreted as another example of a patch reef within the Akron/Cobleskill sequence like that proposed for the Marcellus Falls area (Martisco Reef Complex). Above the sequence here is the lower portion of the overlying Crysler Formation. While chert is a minor component there, as compared to the Marcellus Falls exposures, it allows for a direct correlation with the units exposed at Marcellus Falls. Both localities record the same post-Bertie eurypterid facies.

		Continue forward to Brighton Avenue, turn left up the hill
10.7	2.5	Turn right on NY-173, down the hill.
11.4	0.7	US-11 (continue west on NY-173.
13.3	1.9	NY-175, NY-173, etc. Follow NY-175 west towards Marcellus.
21.4	8.1	NY-174 (continue into Marcellus village).
21.8	0.4	Turn right (follow NY-174 north).
23.0	1.2	STOP 4 at the Martisco Reef Complex and associated facies.

STOP 4: MARTISCO REEF COMPLEX, MARCELLUS FALLS, AND OVERLYING CHRYSLER FORMATION

Interpreted herein (and associated website) to be a Late Silurian patch reef associated with the Akron/Cobleskill sedimentation in upstate New York in this area, and eastward to the Rock Cut Gorge and also westward to the Auburn area. Eurypterid paleoenvironments can only legitimately be interpreted if we take into account the known distribution of stromatoporoids biohermal and biostromal deposits and understand the impact of the variety of morphotypes exhibited within the Bertie Group (and overlying sequences) of microbialites as we come to recognize them within the eurypterid-bearing Late Silurian and Early Devonian sequences.

The most notable feature here is the dark chert facies which is seen to laterally replace the reef but also overlies what looks like the core of the reef – a highly dolomitized massive section with vugs, dolomite crystals, and silicified stromatoporoids. This lithology resembles parts of the Lockport Dolostone of western New York.

		Continue north on NY-174 (winding road)
23.6	0.6	Ninemile Creek Fishing Area on right
26.3	2.7	Entering Camillus
26.8	0.5	NY-5 (turn left and head for Auburn)
42.0	15.2	Entering City of Auburn, continue west on NY-5
45.5	3.5	NY-326
46.0	0.5	Turn right on Clark St. (bear right), note shopping mall
46.4	0.4	Turn left on Beech Tree Rd. (head north)
46.9	0.5	Turn right on Canoga Road
47.4	0.5	STOP 5 at Owasco Outlet and Cobleskill (stromatoporoidal) Limestone

STOP 5: EXPOSURES OF THE COBLESKILL LIMESTONE STROMATOPOROID BIOSTROME

The intimate contact of an eurypterid horizon with deeper water facies is illustrated at this site. This may represent an extreme regression as eurypterid beds lie upon a massive stromatoporoid biostrome. And to complicate matters more, there are higher *Eurypterus*-bearing, pre-*Erieopterus*, units preserved upstream. This is the most complex area of eurypterid stratigraphy in upstate New York and is especially important because the units are all post-Bertie Group, i.e. post *Eurypterus lacustris* Biofacies. Thus, there are numerous eurypterid horizons exposed in the Owasco Outlet and we need to know why they are preserved here and how this important section relates to those to the west, south and east.

		Leave parking lot and head back to NY-5.
47.9	0.5	Turn left (onto Beech Tree Road).
48.4	0.5	Turn right on Clark St. to NY-5 and US-20. Head west to get to the New York State Thruway.
56.3	7.9	NY-90, continue west across Montezuma National Wildlife Refuge
58.3	2.0	Bear right onto NY-318 to NYST (I-90).
69.3	11.0	Entrance to the NYST, I-90.
		END OF TRIP.

ACKNOWLEDGEMENTS

The road log to the stops (circuitous as it is) benefited greatly from the assistance of Linda Heffron. Thank you! Pamella Lawry read part of the manuscript and I thank her for her comments. Special thanks go to Dr. Susan Butts of the Yale Peabody Museum for reviewing the section on Early Devonian Facies Relationships. Her suggestions were more than welcome. I also wish to thank Dr. O. Erik Tetlie (now at the Yale Peabody Museum) for the many interesting exchanges over the past couple of years – I have benefited from them greatly.

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