New York State Geological Association

79th Annual Meeting
State University of New York at Cortland
September 28—30, 2007

Field Trip Guidebook
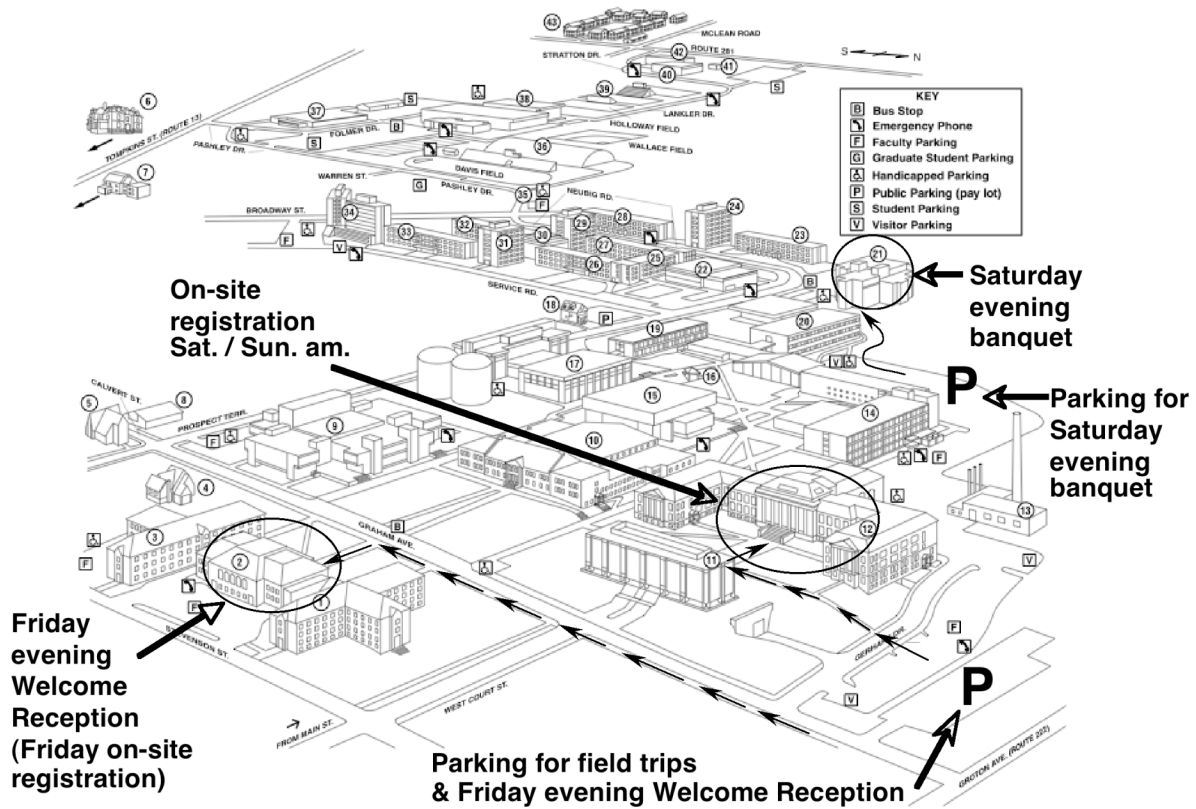
Christopher A. McRoberts, Editor

On-site registration Sat. / Sun. am.

Saturday evening banquet

P **Parking for Saturday evening banquet**

Friday evening Welcome Reception (Friday on-site registration)

Parking for field trips & Friday evening Welcome Reception

P

State University of New York College at Cortland

Bldg. Name	Bldg. No.	Bldg. Name	Bldg. No.	Bldg. Name	Bldg. No.
Alger Hall	29	Hayes Hall	32	President's Residence	4
Alumni House	6	Heating Plant	13	Randall Hall	28
Bishop Hall	27	Hendrick Hall	33	Service Group	42
Bowers Hall	14	Higgins Hall	31	Shea Hall	26
Brockway Hall	2	Interfaith Center	5	Smith Tower	34
Casey Tower	34	Leadership House	18	Sperry Center	15
Chemical Mgmt. Facility	41	Lusk Field House	36	Stadium Complex	39
Cheney Hall	1	McDonald Building	7	Studio West	37
Clark Hall	24	Memorial Library	17	Van Hoesen Hall	20
Commissary/Receiving	40	Miller Building	11	West Campus Apts.	43
Corey Union	21	Moffett Center	10	Whitaker Hall	35
Cornish Hall	19	Neubig Hall	22	Winchell Hall	30
DeGroat Hall	3	Newman Center	8		
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Cortland
State University of New York College at Cortland

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EDITOR'S PREFACE AND ACKNOWLEDGEMENTS

The Geology Department at SUNY Cortland is proud to host the 79th Annual Meeting of the New York State Geological Association. It has been 37 years since Cortland last hosted the NYSGA meeting, and in 1970 Graham Heaslip (at whose former desk I now type these words) served as both President of the NYSGA and editor of the 42nd field guide. Although much has changed in our department in the past three decades, we are still committed to providing an excellent undergraduate program of study involving students in coursework, independent research, and field experiences. Given that much of the current research of Cortland's geology faculty involves fieldwork far removed from central New York, we have relied much on the expertise of those who have made central New York geology their passion. In selecting trips for this meeting we have strived to focus, not only on what central New York is famous for (e.g., world-class Devonian strata and its contained fossils, and a beautifully glacially sculpted landscape), but also to encompass field trips appealing to other geoscientists and educators (be they hard-rockers, soft-rockers, or no-rockers). It is our hope that there is a trip of interest for everyone.

I wish to express my appreciation to the numerous individuals who have contributed to the content and production of this field guide. First and foremost to the 26 authors of the field trips who have devoted their time, energy, and expertise in offering a truly outstanding collection of excursions and, for the most part, getting their manuscripts to me in a timely fashion and in the correct format. I also acknowledge the many reviewers of manuscripts who have helped produce a more polished and error-free field guidebook and to the landowners who provided access to their property during the field trips.

The guidebook is only one facet of producing this yearly event and it is quite clear that the 79th NYSGA meeting would not have been possible without the members of SUNY Cortland's Organizing Committee. I wish to personally thank this committee comprised of Bob Darling (meeting coordinator), Gayle Gleason (logistics) and Dave Barclay (registration) for their countless hours in planning, organizing, and generally working together to 'pull this thing off'. Secretarial support was, as always, efficiently and flawlessly provided by Sue Nevins. We are also grateful for the Cortland undergraduate students who have volunteered their time to drive vehicles and help with the smooth running of the meeting. The Buffalo Geological Society graciously donated funds to help defray some of the printing costs of this field guide. We thank Alan Benimoff and others of the NYSGA executive for cajoling us into running this meeting, providing much needed guidance and answering our many questions. Finally, we are indeed fortunate to have Jim Bell for agreeing to speak at the Banquet on his work with the Mars Rover Project.

Enjoy the meeting and have a pleasant stay in central New York!

Christopher A. McRoberts

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Trip A-1

GLACIAL AND ENVIRONMENTAL GEOLOGY OF THE ONONDAGA TROUGH, CENTRAL NEW YORK

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INTRODUCTION

The Onondaga Trough valley is a vestige of the glacial history of central New York. The trough extends from the outlet of Onondaga Lake southward, through and south of the City of Syracuse, to and beyond the Valley Heads Moraine at Tully, N.Y. Geologic and hydrologic research focused on the Onondaga Trough has been ongoing for several decades through numerous academic institutions such as State University of New York at Cortland and Syracuse [School of Environmental Science and Forestry (ESF)], Syracuse University, Hamilton College, and Colgate University (among others). In addition, the USGS, in co-operation with USEPA-Region 2, Onondaga County, the Onondaga Lake Partnership, and the Onondaga Environmental Institute, has

conducted studies of the Onondaga Trough valley during the last two decades. On-going surficial and bedrock mapping associated with the USGS STATEMAP program have helped to construct a stratigraphic framework that has been used in the development of numerical ground-water-flow and brine-migration models. These models are being used to support central New York's major urban hydrogeology challenge – improving the water quality of Onondaga Lake (purportedly one of the most polluted lakes in the country) and fostering the urban renewal of the adjacent area. Participants on this NYSGA trip will see how glacial geology and hydrology, and engineering geology interact.

ROADSIDE STOPS OVERVIEW

This fieldtrip will be a 10-year retrospective assessment of activities that have occurred since the 1997 NYSGA 69th Annual Meeting at Hamilton College. This 2007 fieldtrip is co-authored by a number of individuals who are familiar with their subject area and each will provide a summary of their work in the following pages. The field-trip stops will commence at the Village of Homer well field (Stop 1), and then progress northward in the Onondaga Trough valley, with stops at the Tully Lakes area (kettle lakes) situated on the Valley Heads Moraine (Stop 2) where the headwaters of West Branch of Tioughnioga Creek (Susquehanna River Basin) will be viewed and its hydrogeology discussed, then further north over the Tully Moraine and down into the Onondaga Creek valley where the unique Tully Valley mudboils will be visited (Stop 3). The remaining two stops will be around Onondaga Lake in Syracuse. An introduction to the Onondaga Lake setting will take place at the Inner Harbor (Stop 4). Discussions will center on the interaction of glacial geology, engineering, anthropogenic contamination of nearshore and in-lake environments, and the brine-filled aquifer. We may also discuss urban redevelopment of shoreline areas, the unique hydrology of the Onondaga Lake Outlet. We'll then hear about the impact of wastebed deposits from the Solvay Process, the associated chemical wastes from the Allied Corporation, now Honeywell Corporation, and the current clean-up plan for Onondaga Lake while standing atop of these massive wastebeds (Stop 5).

Disclaimer

All of the field sites described in the guidebook article and road log are on private property. As such, it is imperative that we respect the rights and wishes of these land owners. Please do not visit these sites on your own, as it may jeopardize future field-trip opportunities. Over the past 15 years it has become increasingly difficult to maintain our access agreements to these sites as individuals and even groups of people have entered these properties without obtaining land-owner permission. We strive to maintain good relations with the land owners and do not want the inappropriate actions of a few to ruin the educational opportunities for many others who wish to enter these areas. The USGS can, and does act as the 'gate keeper' for the property owners and we would be happy to facilitate your future access to these sites. Please contact Bill Kappel for any questions as to access for you and your classes.

HYDROGEOLOGY OF THE ONONDAGA TROUGH

[This section prepared by W.M. Kappel, T.S. Miller, and D.L. Pair]

Onondaga Trough (Valley) Geologic Data Collection and Analyses

The primary sources of geologic data that contributes to our understanding of the Onondaga Trough (Figs. 1 and 2) are logs of test holes drilled at construction sites throughout the trough for buildings, roads, bridges, public utilities, and other projects. Many of these projects are concentrated in the urban area of Syracuse and along major highways. Test holes outside Syracuse are scant and the data are generally less detailed because only limited geologic data are required for most small-scale construction in rural areas. Test-hole logs (descriptions of materials penetrated) provide data on stratigraphy (layering of the glacial sediment), soil properties (permeability, compactness, texture, color), and ground-water levels encountered during drilling.

Information in rural areas was obtained mainly from local ground-water studies and from well records collected through the New York State Department of Environmental Conservation (NYSDEC) well permitting program established in 2000. Additional stratigraphic data for the valley-fill deposits were obtained from 12

deep test holes drilled for USGS investigations conducted between 2000 and 2004 in the central and northern parts of the Onondaga Trough (Fig. 2) and from wells drilled for the salt-dissolution operation on the backside of the Valley Heads moraine near Tully, N.Y.

A listing of USGS reports (including website addresses for these reports) is included as Appendix 1, while a bibliography of cited references is provided at the end of each of this fieldtrip. The USGS reports supplement the information contained herein and should be in-hand while reading the following sections or referenced during the field trip as the report figures (and colors) provide much greater detail.
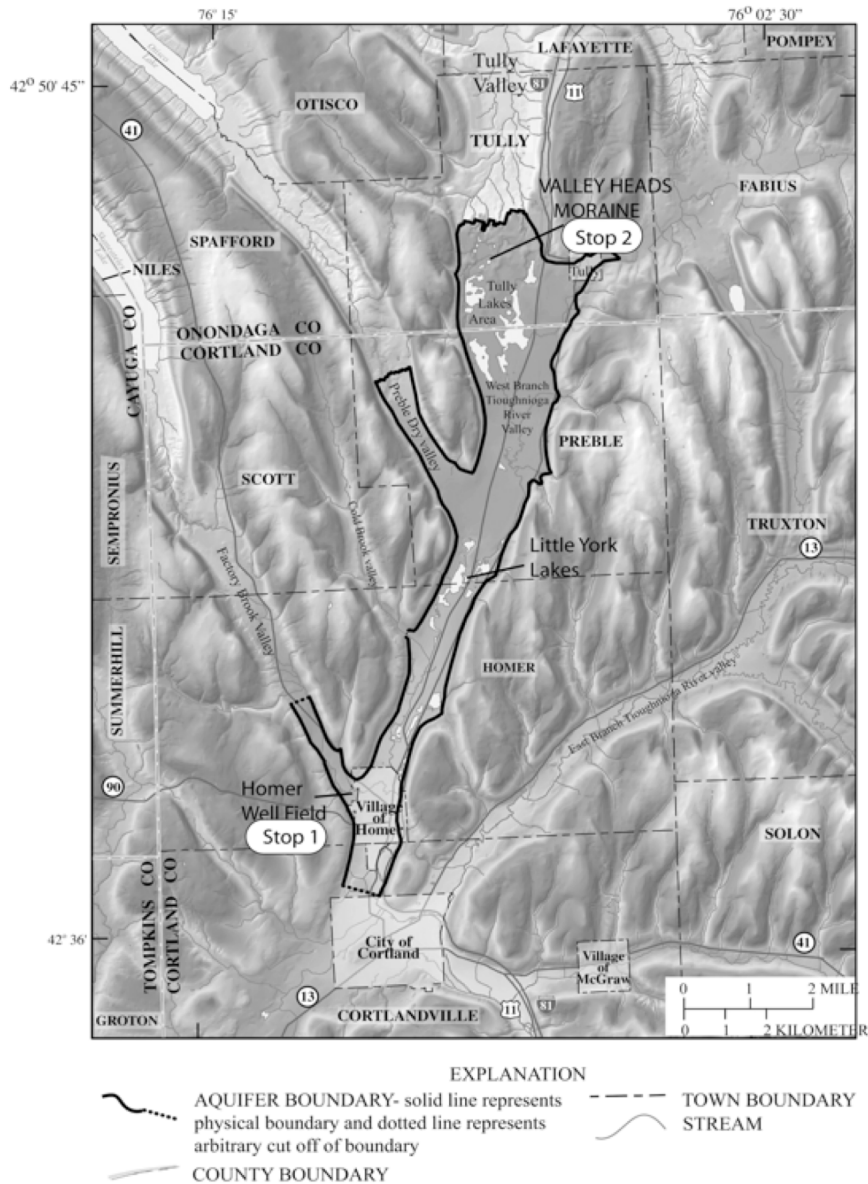
FIGURE 1—The southern part of the Onondaga Trough – West Branch of Tioughnioga Creek valley showing the aquifer boundary for the West Branch ground-water-flow model and location of key features.

The geologic section shown in Figure 3 is along the thalweg (deepest part) of the trough with hills projected in the background and depicts the general southward dip of the sedimentary bedrock units (40 to 50 feet per mile). The unconsolidated deposit geologic section shown in figure 4 is a longitudinal transect roughly following the axis of Onondaga Trough and extends from the Onondaga Lake outlet to and just south of the Valley Heads (Tully) Moraine. If the reader is interested in viewing more geologic sections, there are seventeen sections shown in the paper by Kappel and Miller (2005) that depict unconsolidated materials in the Onondaga trough and its tributary valleys. These sections were constructed to depict the configuration of the bedrock surface below the valley floor, as well as the generalized stratigraphy of the glacial deposits that overlie bedrock.
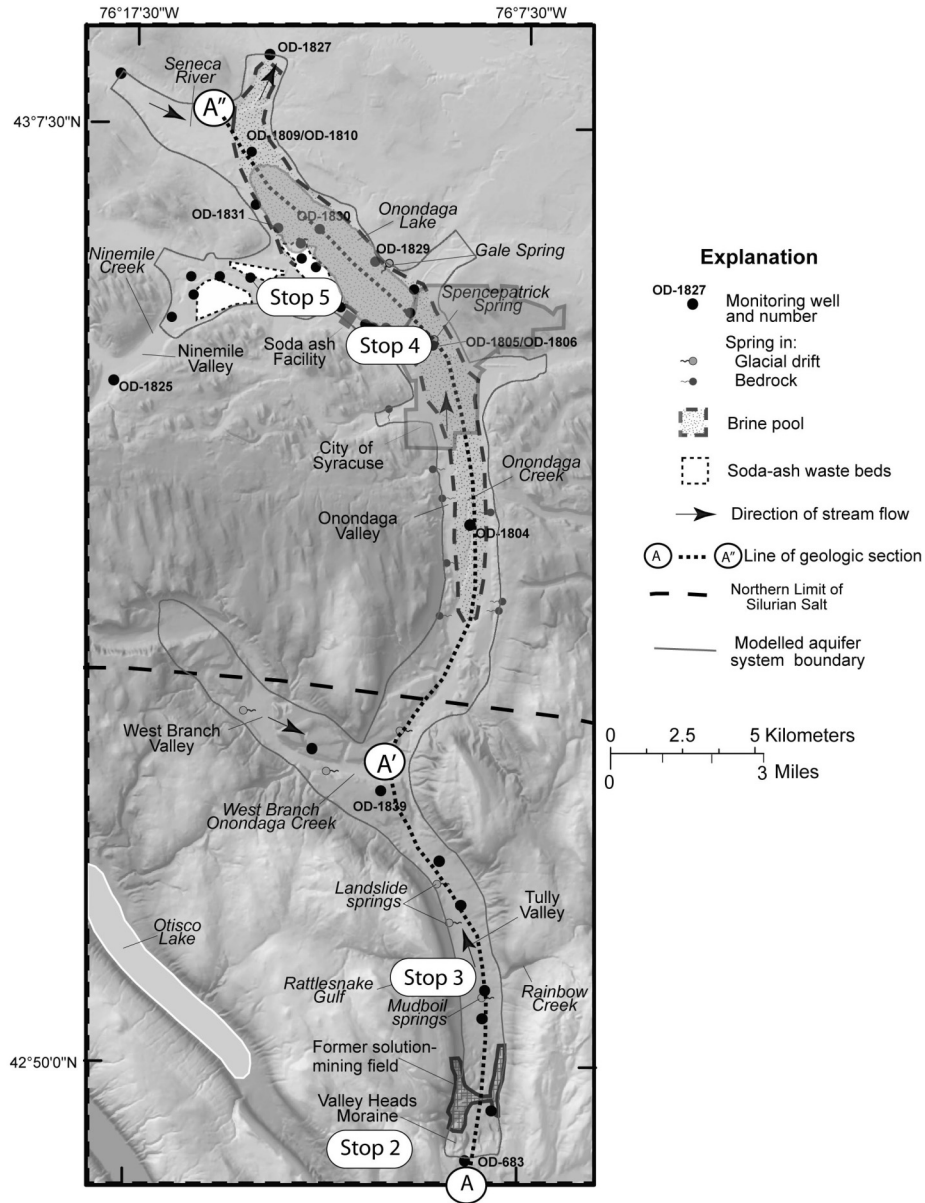
FIGURE 2—Geographic features of the central and northern Onondaga Trough including physical and man-made features, location of wells and borings, springs, the soda-ash facility and associated wastebeds, and location of the brine pool in the valley-fill sediments.

Additional products from the USGS study of the central and northern sections of the Onondaga trough (Tully Valley and Onondaga Valley) included geographic information system (GIS) based shaded-relief maps of (1) the land surface in the Syracuse area (fig. 1 in Kappel and Miller, 2005) and (2) the underlying bedrock surface in the Onondaga Trough (figs. 2 through 4 in Kappel and Miller, 2005). The elevation of bedrock surface map was generated by interpolating elevation data from test-hole logs and other data sources, and subtracting these values from the Digital Elevation Model (DEM). The resulting dataset was directly applied to the ground-water-flow model to generate elevation grids of the bedrock surface that established the hydrogeologic framework for model simulations of ground-water flow and brine migration in the Onondaga Valley unconsolidated aquifer.

The shaded-relief map in figure 2 of Kappel and Miller (2005) depicts an oblique "birds-eye-view" of the bedrock floor of the Onondaga Trough as if all the glacial sediments have been removed from the valley. However, there were little data available on bedrock elevation in the upland areas surrounding the Onondaga Creek, Otisco Lake, and Butternut Creek Valleys therefore, the land outside the Onondaga Trough represents land surface elevation, not bedrock elevation. A contoured representation of the elevation of bedrock surface in the Onondaga Trough is shown in figure 4 of Kappel and Miller (2005). The bedrock-surface elevations range from about -25 ft below mean sea in the deepest part of the trough (under the southern end of Onondaga Lake and the city of Syracuse) to greater than 550 feet along the edges of the trough. The bedrock elevations are greater than 1,400 feet along ridge tops on either side of the trough, but upland elevations higher than 550 feet are not depicted in that figure.

The bedrock floor of the trough rises southward from just south of Syracuse to at least Little York Lake in the West Branch Tioughnioga River valley (fig. 1). The bedrock floor levels out between Little York Lake and Cortland. The bedrock floor also rises to the north under Onondaga Lake (Fig. 2). The thickness of unconsolidated valley-fill deposits along the thalweg of the Onondaga Trough, from Onondaga Lake to the Tully Moraine, averages 420 feet, but near the moraine it exceeds 800 feet (Kappel and Miller, 2003). Valley-fill deposits in the southern part of the Onondaga Trough, in the West Branch valley of Tioughnioga Creek, range from 350 feet to the north, closer to the moraine, and thin to about 250 feet thick to the south, near Cortland.
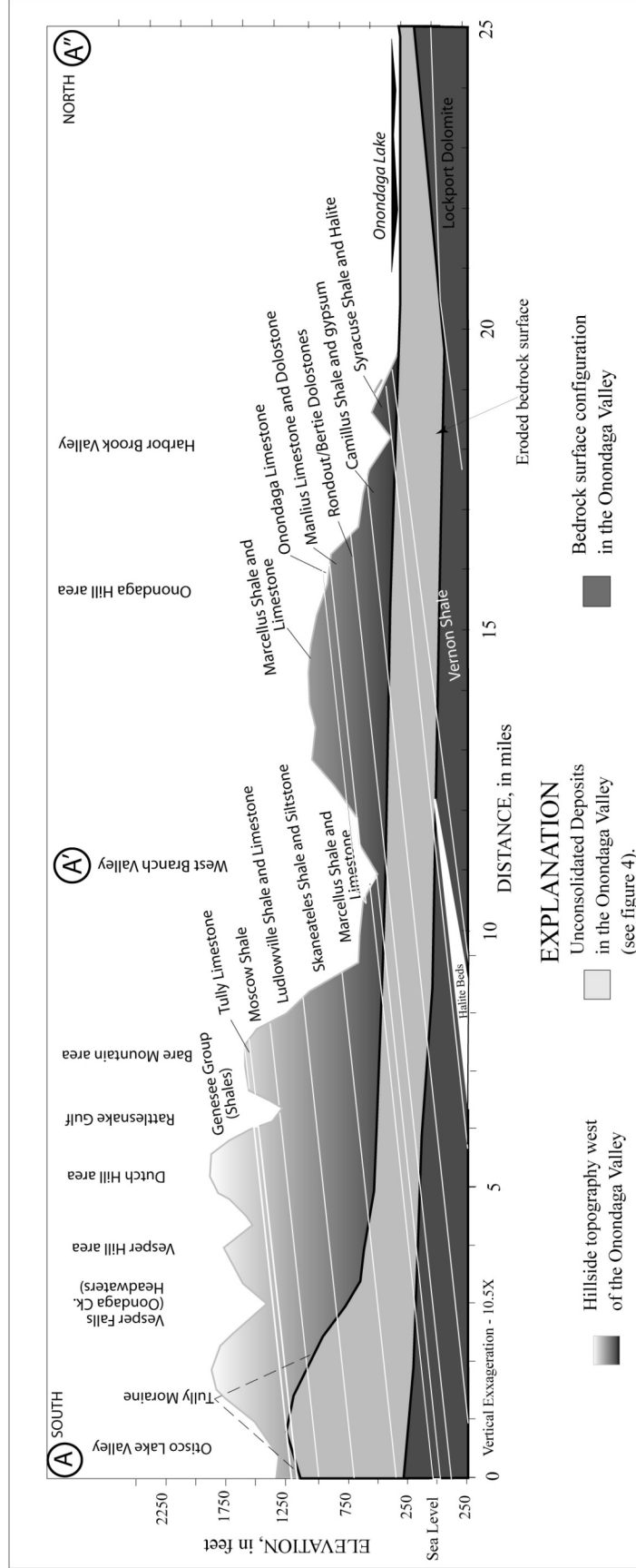
Bedrock Stratigraphy

Sedimentary rock units of Silurian and Devonian age gently dip 40 to 50 feet per mile to the south (Fig. 3) Halite units (salt) are probably exposed in the floor of the trough and thicken to the south (Fig. 3). Bedrock units in central New York appear as east-west trending bands, therefore the older (Silurian aged) units underlie the lowlands north of Syracuse, whereas the younger (Devonian aged) units underlie the southern part of the trough and form the hills bordering the trough (Fig. 3). The stratigraphic column (Fig. 5) summarizes the characteristics of the bedrock sequence from the Upper Silurian through Middle Devonian time period, and includes the group and formation names, the lithology, and relative thicknesses of the bedrock units.

Unconsolidated Stratigraphy of the Onondaga Trough

Geologic sections along and across the Onondaga Trough and some tributary valleys (fig. 7 and sections A-A' through Q-Q' in Kappel and Miller, 2005) were constructed using records from hundreds of test-hole logs drilled in the greater Syracuse area. These sections were assembled from data sources that span more than 50 years of varied drilling techniques and sediment descriptions. The geologic interpretation of driller and consultant logs have been modified and many times simplified, to conform to the New York State Geological Survey's lithologic characterization of glacial sediments (Pair, 1998a, b). The simplified longitudinal geologic section of unconsolidated sediments along the Onondaga Trough from Tully, N.Y. to the Onondaga Lake outlet is shown in Figure 4. This section depicts the layering of sediments deposited generally from south to north during the recession of the glacial ice from this part of New York (from 12,000 to 14,000 radiocarbon years ago). This period of recession included pauses and brief readvances of the ice margin in the Onondaga Trough.

GENERALIZED BEDROCK GEOLOGIC SECTION - VALLEY HEADS MORAINE AT TULLY TO ONONDAGA LAKE NEAR SYRACUSE

FIGURE 3—Geologic section along the thalweg of the Onondaga trough valley from the Valley Heads moraine to the Onondaga Lake outlet showing the generalized bedrock stratigraphic sequence.

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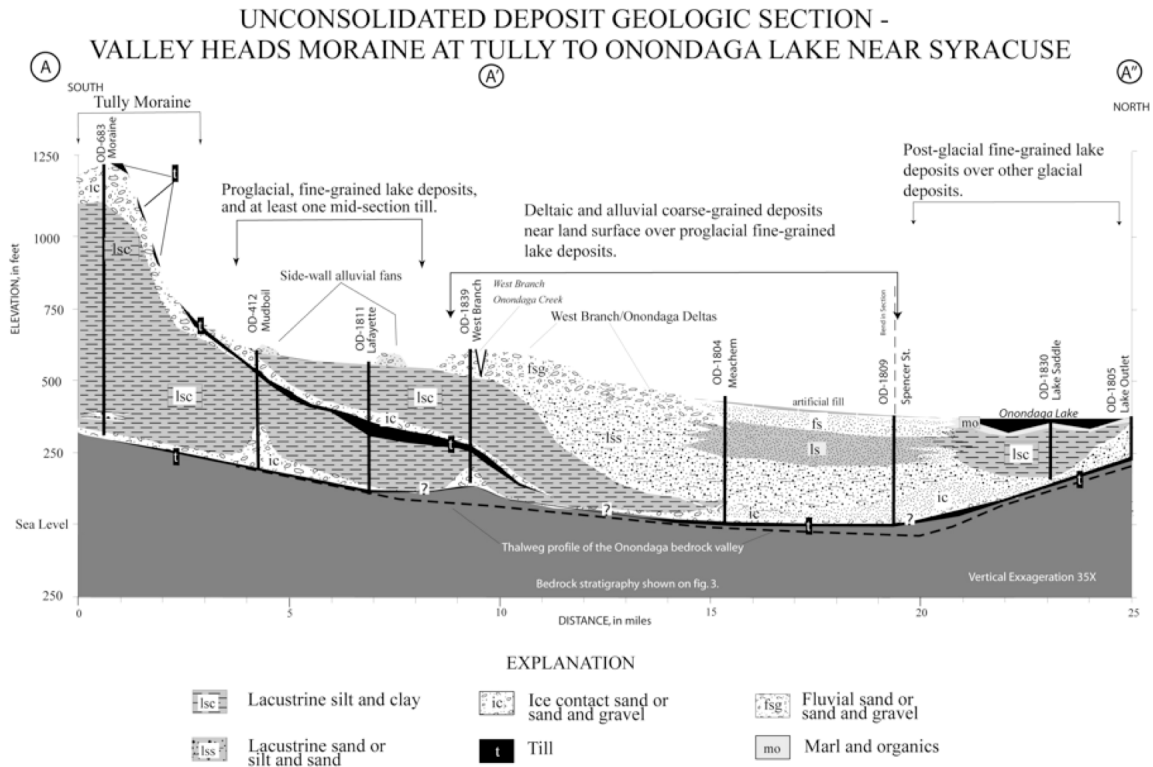
FIGURE 4—Geologic section along the thalweg of the Onondaga trough valley from the Tully Moraine to the Onondaga Lake outlet, showing the generalized layering of unconsolidated sediment in the valley-fill deposits.

Glacial History and Hydrogeology of the Valley Heads Moraine at Tully

The Valley Heads Moraine at Tully, N.Y. forms part of the largest end-moraine complex in central New York and forms the present surface-water divide between drainages that flow north to St. Lawrence River and drainage that flow south to Susquehanna River. The ice in the Onondaga trough came to its standstill position just west of Tully, N.Y. at about 17,000 Cal BP years ago (Ridge, 2003) where it formed the segment of the Valley Heads Moraine shown in figures 3 and 4. The surface-water divide is just south of New York State Route 80 where it crosses the moraine. The surface-water divide doesn't correspond to the ground-water divide as it is generally further to the south, and its position changes seasonally according to local ground-water conditions (Kappel and others, 2001).

The melting of the ice front and the glacial retreat northward from the Valley Heads Moraine resulted in the formation of a proglacial lake between the retreating ice front and the Valley Heads Moraine. Initially, this lake drained to the south through an outlet channel(s) across the Valley Heads Moraine, but as the ice front retreated northward, toward present-day Syracuse, successive outlet channels with progressively lower elevations were exposed along the east side of the valley about 10 miles to the north (Hand, 1978), and sequentially lowered the lake levels. As the ice retreated from the Tully valley segment, large amounts of coarse-grained sediment (sand and gravel) were transported southward through subglacial tunnels; some was probably deposited within ice tunnels as sinuous esker ridges, and the remainder was disgorged into the proglacial lake as subaquatic fans. The finer-grained sediment (fine sand, silt, and clay) was carried out into the lake where it settled to form lake-bottom deposits that buried the older, coarser-grained subglacial sediment. When the Valley Heads ice retreated past Syracuse (about 15 miles north of Tully), the lake drained entirely and glacial deposition in the trough ceased.

STRATIGRAPHIC COLUMN OF BEDROCK UNITS IN THE ONONDAGA TROUGH VALLEY

	Age, in millions of years	GROUP	FORMATION / Member	LITHOLGY	THICK-NESS
DEVONIAN — Middle	374		TULLY	Limestone	30
			MOSCOW / Windom	Shale and sandstone	180
		HAMILTON	LUDLOWVILLE / Portland Point ls. / Owasco / Spafford / Ivy Point / Otisco / Centerfield ls.	Shale and limestone	10 / 1-3 / 25 / 50-60 / 160-180 / 30
			SKANEATELES / Butternut / Pompey / Delphi Station / Mottville	Shale and Siltsone	100-200 / 60 / 100 / 45
			MARCELLUS / Cardiff / Chittenango / Cherry Valley / Union Springs	Black Shale and limestone	185-200 / 100 / 3 / 13-15
			ONONDAGA / Seneca / Moorehouse / Nedrow / Edgecliff	Limestone, grey, with some cherty layers and one or more bentonite layers	27 / 24 / 14 / 12
DEVONIAN — Lower	387	TRISTATES	ORISKANY	Sandstone, white	0-5
		HELDERBERG	MANLIUS / Pools Brook / Jamesville / Clarks Reservation / Elmwood / Olney	Limestone and Dolostone, grey/brown	5 / 16 / 15 / 12 / 40
			RONDOUT / Chrysler	Dolostone	50
SILURIAN — Upper	408		COBLESKILL	Limestone	25
		BERTIE	Oxbow / Forge Hollow / Fiddler s Green	Mostly Dolostone with Gypsum layers	9 / 55 / 30
		SALINA	CAMILLUS	Shale, tr. Gypsum, some Dolomite	250
			SYRACUSE	Shale, Halite and tr. Gypsum, black	275
			VERNON	Shale, Gypsum - green/red	500-600
	421	LOCKPORT	LOCKPORT	Dolomite, Grey	120-140

FIGURE 5—Stratigraphic column of bedrock found below and surrounding the central and northern sections of the Onondaga trough

At Tully, N.Y., large amounts of sediment accumulated at the ice front as a result of the meltout of englacial material that had been eroded and transported from the bedrock valley floor and walls. In addition, subglacial meltwater discharging from conduits at the base of the ice disgorged sediment as subaqueous fans at the ice front. Meltwater that flowed on top of the ice transported supraglacial material to the ice front in channels on or within the disintegrating ice front. A hummocky landscape developed at the ice margin as large amounts of sediment were deposited on the disintegrating ice to form kame, kettle, and glacial karst features. Kame mounds and ridges of glacial drift (mixed glacial sediments) formed across the Valley Heads Moraine; the drift generally consists mostly of coarse-grained sediment (sand and gravel) that settled to the land surface

as the ice melted. Kettle hollows sometimes formed between kame mounds as a result of the melting of buried ice masses that became separated from the retreating ice front. Glacial karst formed in areas of relatively thick glacial debris that covered and insulated residual ice masses. Internal drainage networks may then have developed beneath these masses and these drainage conduits slowly enlarged to the point that the overlying ice collapsed to form depressions and sinuous chains of craterlike depressions. Green and Tully Lakes together have a sinuous form that may reflect a former collapsed meltwater conduit(s) (see fig. 4, Kappel and Miller, 2003).

Two deep test wells (well OD-685, 560 feet deep; well OD-683, 830 feet deep, (fig. 5, Kappel and Miller, 2003) were drilled by the USGS on the Valley Heads Moraine and over the deepest part of the valley. Neither well reached bedrock, although the deeper well (OD-683) probably ends close to bedrock. The other well (OD-685) penetrated 135 feet of kame-moraine deposits (sand and gravel) overlying 30 feet of deltaic sand, which is underlain by 240 feet of glaciolacustrine sediments (fine sand, silt, and clay) interbedded with till, and then by 160 feet of silty sand and gravel. This silty sand and gravel unit may be part of a buried moraine of pre-Valley Heads origin, or it may represent an early-Valley Heads standstill position. The thick lacustrine unit between the Valley Heads Moraine and the underlying and presumably older moraine indicates the presence of a lake in the Tully Trough before the Valley Heads readvance. The lake apparently continued to exist or redeveloped, as indicated by fine-grained sediment that buries this early moraine as well as the older coarse deposits north and south of the moraine. Whether the 160-foot thick silty sand and gravel sequence extends to bedrock and continues within the subglacial deposits of sand and gravel (as shown in figure 5, Kappel and Miller, 2003) is uncertain.

Well OD 683 (see fig. 5, Kappel and Miller, 2003) was drilled to 830 ft and did not reach bedrock. The well penetrated, in descending order, 7 feet of till, 118 feet of kame moraine (mostly sand and gravel), 600 feet of fine-grained lacustrine sediment (fine sand, silt, and clay), 40 feet of sand and gravel, another 60 feet of fine-grained lacustrine sediment, and 3 ft of very coarse gravel which may be part of a subaquatic (buried) fan or moraine. Test hole data are insufficient to reveal the extent of the hypothesized buried fan or moraine, and do not indicate whether these deposits extend to bedrock, as implied by figure 5 of Kappel and Miller (2003).

The complex stratigraphy of the moraine results in an aquifer system with multiple-aquifer units. The surficial kame-moraine deposits (ice-contact and collapsed outwash) forms an extensive unconfined sand and gravel aquifer (70 to 120 feet thick) that is underlain by a thick sequence of fine-grained glaciolacustrine deposits that, in turn, confine one or more deep sand and gravel aquifers (presumably subglacial conduit deposits and subaquatic fans) of highly variable thickness (15 to 150 feet thick), that may be discontinuous. The moraine also contains large kettle lakes, such as Song, Crooked, Green, and Tully lakes (see fig. 4, Kappel and Miller, 2003). The water levels in some of the kettle lakes represent the local water table, whereas, the water level in others appear to be perched (Kappel and others, 2001). The bottoms of these perched lakes and ponds are probably lined with poorly-permeable sediment and decayed organic material (muck) and(or) till that impedes the downward movement of lake water.

The kame sand and gravel aquifer at the moraine is connected to the surficial outwash sand-and-gravel aquifer in the West Branch Tioughnioga River valley. Both aquifers typically yield 10 to 50 gallons per minute to open-ended domestic wells. The water quality is generally good, although the water is hard, as would be expected from the large limestone content within the gravel (Ku, and others, 1975, fig. 22; Denny and Lyford, 1963, pl. 3).

Ground water in the northern part of the kame sand and gravel aquifer flows northward and discharges from springs along the north side of the moraine. Some of the springs are perennial, including the springs along Route 11A which are used for a small public-water-supply system for residents in the Tully Valley, whereas others flow only during the wet season. Some domestic wells at the crest of the moraine are completed in a sand and gravel layer, about 15 feet thick that apparently is a thin lens within the upper part of the glaciolacustrine unit (see fig. 5, Kappel and Miller, 2003). Ground water in this unit probably also discharges to the springs along the northwestern side of the moraine (see fig. 4, Kappel and Miller, 2003).

Two domestic wells on the crest of the moraine, each about 400 feet deep (OD-675 in the east central part of the moraine and OD-674 on the western side), were drilled through the surficial kame moraine and the lacustrine deposits and completed in the thin basal aquifer that overlies bedrock. Water from the confined aquifer is typically turbid, moderately mineralized, and similar to water in the shale—with a hydrogen sulfide odor and enough iron to cause staining. The surficial sand and gravel aquifer near the crest of the moraine is

thinly saturated because springs on the north side of the moraine drain much of the water from this aquifer. USGS test well OD-683, near the intersection of NYS Route 80 and Gatehouse Road (see fig. 4, Kappel and Miller, 2003), penetrated an unconfined sand and gravel aquifer between depths of 10 and 85 feet, and a thin sand-and-gravel aquifer (lens?) between 107 and 118 feet which was confined within the upper part of the glaciolacustrine unit. Both aquifers yield more than 10 gallons per minute to domestic wells. Underlying the thick confining unit are two confined sand and gravel aquifers between depths of 730 and 770 feet and 827 and 830 feet. Refusal was encountered at depth 830 feet, but a well could only be installed in the aquifer at 730 feet. The driller estimated the yield from the 730 foot deep aquifer at several hundred gallons per minute (gpm). It is uncertain whether the well finished in this aquifer is connected to the basal confined aquifer in the West Branch Tioughnioga Creek valley and that found in the Tully valley.

The Tully Valley Mudboils

The Tully Valley mudboils are volcanolike cones of fine sand and silt that range from several inches to several feet high and from several inches to more than 30 feet in diameter. Active mudboils are dynamic ebb-and-flow features that can erupt and form a large cone in several days, then cease flowing, or they may discharge continuously for several years. Mudboils have been observed in the Tully Valley in Onondaga County, in central New York, since the late 1890's but probably have existed since the last proglacial lake drained from the valley. Mudboils have continuously discharged sediment-laden (turbid) water into nearby Onondaga Creek at least since the early 1950's. The discharge of sediment causes gradual land-surface subsidence that, in the past, necessitated rerouting a major petroleum pipeline and a buried telephone cable, and caused two road bridges to collapse. The water discharged from mudboils can be either fresh or brackish (salty).

Mudboil activity was first reported in the Syracuse (New York) Post Standard, in a short article dated October 19, 1899:

> "***Tully Valley—A Miniature Volcano—*** *Few people are aware of the existence of a volcano in this town. It is a small one, to be sure, but very interesting. In the 20-rod gorge where the crossroad leads by the Tully Valley grist mill the hard highway bed has been rising foot after foot till the apex of a cone which has been booming has broken open and quicksand and water flow down the miniature mountain sides. It is an ever increasing cone obliterating wagon tracks as soon as crossed. The nearby bluff is slowly sinking. Probably the highway must sometime be changed on account of the sand and water volcano, unless it ceases its eruption.*"

This newspaper article accurately describes Tully Valley mudboils and presages the collapse of the Otisco Road bridge 92 years later in 1991. The article indicates that land subsidence occurred nearby, but gives no indication that Onondaga Creek was turbid; this was either an oversight by the reporter or was not a concern.

Flow from a mudboil is driven by artesian pressure that forces water and sediment upward from two confined sand and gravel aquifers through a 60-foot-thick layer of dense silt and clay at land surface. The artesian pressure within the aquifers can lift water 20 feet above land surface along most of the valley floor and 30 feet above land surface near Onondaga Creek. The source of the artesian pressure is surface water entering the ground-water system along the valley walls — primarily at the southern end of the valley at the Valley Heads Moraine and from the alluvial fans at the mouth of Rattlesnake Gulf and Rainbow Creeks. Additional water may also enter the mudboil aquifers from the Tully brinefield area in the southern part of the valley where former solution mining of halite deposits has led to fracturing of bedrock and land-surface subsidence in the brine-mining area.

The flow of water from the mudboils changes seasonally in response to changes in artesian pressure in the two aquifers. In the spring, when ground-water recharge is greatest, the mudboils in the main mudboil/depression area (MDA) (see fig. 3, Kappel and McPherson, 1998) can discharge 400 gpm or more. As recharge to aquifers declines during the summer, artesian pressure in the aquifers also declines, and flow from mudboils typically decreases to 200 gpm or less. The rate of mudboil flow does not change in response to individual rainstorms but does respond to seasonal variations in precipitation and resulting changes in hydrostatic pressure within the mudboil aquifers.

Suspended-sediment discharged from the MDA to Onondaga Creek has been measured from October 1991 to present (September, 2007). The daily average suspended sediment load in the 1993 water year was approximately 30 tons per day (tons/d). Most of the suspended sediment is very fine clay and silt with a small fraction of very fine sand. Chemical analyses of mudboil discharge in the MDA indicate that the source of water can be either the confined fresher-water aquifer or an underlying brackish-water (salty) aquifer (Fig 4). Chloride concentrations in the upper, freshwater aquifer range from 37 to 430 milligrams per liter (mg/L) and from 2,000 to 7,100 mg/L in the lower, brackish-water aquifer. The difference in chloride concentration between these two aquifers is due partly to the greater density of the saltwater, which causes the brackish water to concentrate in the lower aquifer. Remedial efforts near the Tully Valley mudboils during the late 1990's included: (1) diverting flow from the tributary that feeds the MDA to an adjacent tributary; (2) installing depressurizing wells at several locations around the MDA and along Onondaga Creek to decrease the artesian pressure; and (3) constructing a dam and sediment-settling impoundment to detain mudboil sediment that would normally discharge to Onondaga Creek.

Surface-Water Diversion.—Flow from the upper 0.7 square miles of the mudboil tributary drainage was diverted south to an adjacent tributary drainage in June 1992. This diversion reduced total annual surface water inflow to the MDA by about two-thirds, which, in turn, reduced sediment loading to Onondaga Creek by half — from about 30 tons/d before diversion to about 15 tons/d thereafter.

Aquifer Depressurizing Wells.—Depressurizing wells were installed near the collapsed Otisco Road bridge during the winter of 1992-93 in an effort to reduce artesian pressures in the upper aquifer and subsequently slow nearby mudboil activity. The wells were drilled to the base of the fresher-water upper aquifer, and 10-foot-long well screens were installed to allow artesian-pressured water to flow out of the well while holding the fine-grained sand and silt in place.

These wells initially had a combined discharge of about 25 gpm of sediment-free water and have modestly reduced artesian pressure in the freshwater aquifer by about 1 pound per square inch, or about 2.5 feet of hydraulic head in the late 1990's. Currently these wells discharge about 15 gpm. While nearby mudboil activity has not increased since they were drilled, and no new mudboils have developed, these wells only influence the artesian pressure within about 100 feet of the well and do not preclude new mudboil activity in the area. Eight additional wells were installed in the aquifers underlying the MDA and Onondaga Creek in the summer of 1996 to further reduce artesian pressure and slow mudboil activity. Total ground-water discharge from all wells averaged about 350 gpm in 1997, but today that flow is averages only about 180 gpm. The chemical quality of water discharging from these wells varies with the position of the well in relation to the MDA and has varied over time. Most of the flows from depressurizing wells screened in the upper aquifer around and downgradient from the MDA are slightly brackish to salty, indicating that water from the lower basal aquifer is migrating upward into the base of the upper, fresher-water aquifer. The salinity of water upgradient (south) of the MDA is generally low. Water quality measurements between 1997 and today indicate a slow but steady increase in conductivity (salinity) in the discharged water. Discharge from individual wells ranged from less than 5 to as much as 100 gpm, depending on (1) the well location within the aquifer unit, (2) the aquifer material, and (3) the time of year In the 10 years of operation of these free-flowing wells, the flow rates have diminished even with well redevelopment in 2005. The decrease in flow is probably caused by the clogging of the fine-grained mudboil aquifer outside the well screen, and outside the influence of well-redevelopment techniques.

Impoundment Dam.—A temporary dam was constructed at the outlet of the MDA (see fig. 3, Kappel and McPherson, 1998) in July 1993 to reduce the average daily load of sediment discharging to Onondaga Creek. The impounded water covered several mudboils and allowed most of the silt and sand to settle out before flowing to Onondaga Creek. Also, the weight of water over active mudboils, and the additional weight of sediment settling on the mudboils likely decreased mudboil discharge. The impoundment, in conjunction with the depressurizing wells, has slowed mudboil activity in the MDA, and slowed land subsidence in this area as well. The impoundment reduced the average daily load of sediment discharged from the MDA to Onondaga Creek, from 15 tons/d in 1992 to about 1.5 tons/d during water years 1993 and 1994, but by 1995, the entire impounded area was filled with sediment. Consequently, sediment loading to Onondaga Creek increased from 1.8 tons/d in water year 1995 to 2.8 tons/d in water year 1996. The dam was reconstructed to allow the outflow elevation to be raised in the summer of 1996 and has kept the discharge of sediment from the MDA to

Onondaga Creek in the range of 1 ton/d or less until about 2004. At that point mudboil activity within the MDA had filled the impoundment with sediment. In the summer of 2005 a 'moat' was dug around the MDA to allow any mudboil-generated sediment to settle out prior to the water being discharged from the MDA. While this process did reduce mudboil sediment loads to less than a half ton/d, dredging of the moat will be required on a semi-annual basis to maintain any sediment-retention capability.

Future Remedial Activities.—Based on the past 10-years of remedial activity and monitoring at the main mudboil depression area, the results of these efforts have been mixed. While sediment loading has been reduced by more than 95 percent, the cost of operation and maintenance at the MDA cannot be sustained into the future due to diminishing sources of funding. Rather than treat the sediment problem at the discharge area, it has been proposed that reducing the amount of water which enters the aquifers which are hydraulically connected to the mudboils might be a better, longer-term solution to reduce mudboil sediment loading to Onondaga Creek.

A series of pilot projects have been proposed by the Onondaga Lake Partnership to be implemented in the Tully brinefield area as access to this area is fairly uncomplicated and the landowner has granted permission to implement these activities. One stream (Emerson Gulf) looses about 500 gal/min at the edge of the valley wall where numerous bedrock fractures, related to nearby sinkhole subsidence, are present. (See the Sunday Fieldtrip "B1" for further information.) A 300-foot long section of stream channel will be lined to prevent surface water infiltration. On the east side of the Tully valley, in one of the larger sinkholes, the outlet of the sinkhole will be lowered to remove as much water as possible and route it to the surface water stream course. Supplemental pumping of water from the sinkhole will also be attempted using wind- and solar-powered pumps to test the efficacy moving water to the surface-water system and further reducing the amount of water which enters the ground water system from surface- and ground-water sources which enter this sinkhole. Water levels in this sinkhole generally change by 10 to 15 feet in a normal year, with less water-level change in wet years. If these pilot projects can be shown to be successful, additional remedial activities could be implemented on the alluvial fans of Rattlesnake and Rainbow Creeks, as well as elsewhere near the brinefield. It is not the objective of these remedial activities to stop all mudboil activity, but to return it to the semi-seasonal activity noted in the late 1800's.

EXCAVATIONS IN AND AROUND ONONDAGA LAKE

[This section prepared by J. P. Stewart and C. W. Dickhut]

The firm of John P. Stopen Engineering has designed several large and deep excavations in the sediments surrounding Onondaga Lake in the last seven years. This section describes the special considerations that the local geology poses for this type of work and is illustrated by case histories.

The area around Onondaga Lake is a deep bedrock trough filled with glacial and soft recent sediments. The groundwater level is shallow with quality that ranges from fresh to highly saline. These conditions are often challenging for the deep excavations required for civil works.

A significant construction problem for excavations is managing groundwater. The deeper porous glacial deposits are confined below less permeable lacustrine soils, and often require differing dewatering considerations, as illustrated by the following case histories.

Case History 1 (Saline Groundwater in Porous Soil)

A below-grade pump station was constructed in 1930 in the lowland area of Syracuse on the south side of Kirkpatrick Street and on the banks of Onondaga Creek. The pump station was replaced around 1975 with a larger, deeper and more modern facility. The pump station was constructed where the porous glacial outwash sands are thick and have no cover of lacustrine sediments. The area had natural artesian salt brine springs and salt was produced commercially throughout the area in the 1800's. During that century, over 10 million cubic yards of salt were extracted from brine wells.
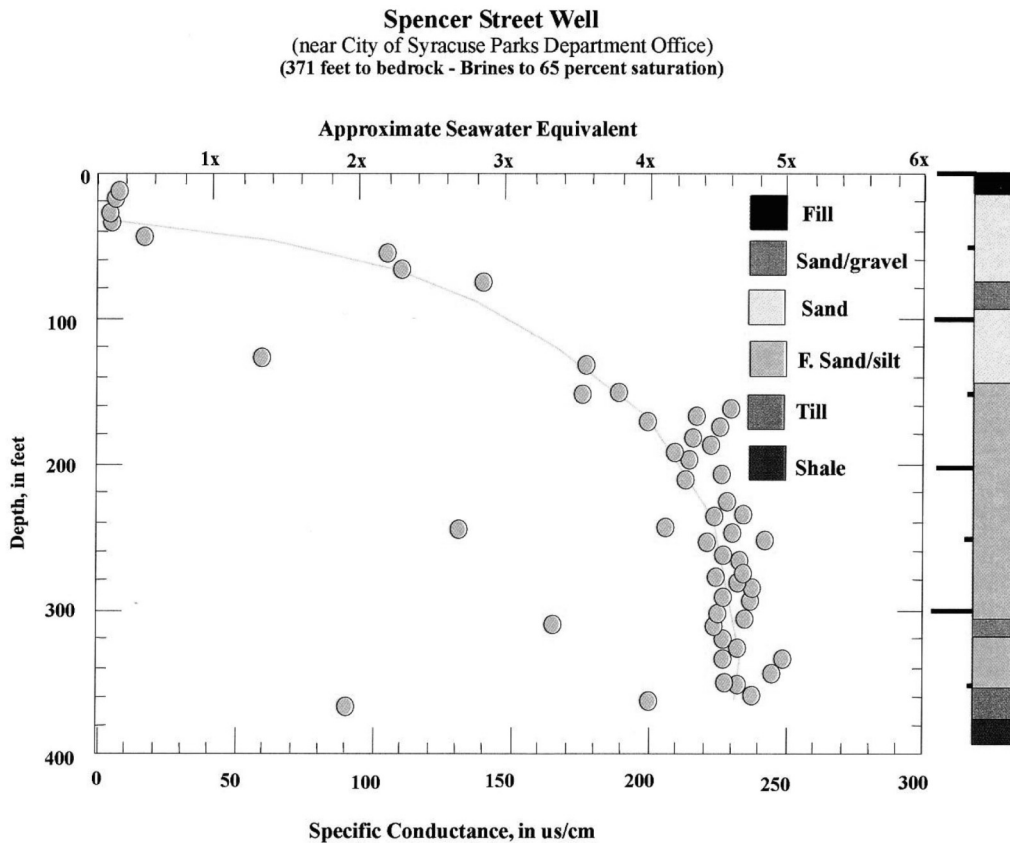
The 1975 construction required a major excavation dewatering effort to lower the water table 15 to 20 ft next to Onondaga Creek. Thousands of gallons per minute (gpm) were pumped from the deep dewatering wells installed inside the excavation. With that volume of water, the effluent was checked for suspended solids by burning off the

water and weighing the residue. Oddly enough, these tests showed significant solids in the essentially clear discharge. It was uncertain if significant fines were being removed from the ground that could presage a collapse. Consequently, the job was shut down for several months.

The job did not resume until it was determined that the groundwater had significant dissolved salt content. The solids in the dewatering discharge were not suspended solids, but dissolved solids that posed no threat for ground settlement or collapse.

After construction of the pump station addition began in 2002, the USGS drilled an exploratory boring within a few hundred feet of the site. The 6-inch-diameter vibro-sonic core boring retrieved a continuous sampling of the valley fill and the bedrock below (Kappel, personal communication, 2003). Figure 6 shows a log of the boring and Figure 7 shows a geologic cross-section through the valley at the pump station. Of particular interest is the salt content of the pore fluid that shows salt content increasing with depth, but also that the salt content of the deep soil is more than 4 times that of sea water.

Project borings and old test borings showed that the site was underlain by about 15 ft of old fill. The water table was about 10 ft below grade. The natural soils consisted of sands and gravels to depths of more than 60 ft. The USGS boring indicated that the deeper soils had seams with greater sand and silt content than the shallower soils.

Spencer Street Well
(near City of Syracuse Parks Department Office)
(371 feet to bedrock - Brines to 65 percent saturation)

FIGURE 6—Salt Concentration versus Depth near the Kirkpatrick Street site. (Kappel, 2005)

The project required excavation to about 35 ft depth and about 25 ft below the water table. The soils consisted of deep permeable soil that was most likely hydraulically connected to the adjacent creek and could not be easily cutoff with sheetpiling or a cut off wall.

To evaluate the required dewatering effort, a pump test was performed in a test well installed next to the pump station as shown in Figure 8. Back evaluation of the pump test (Continental Placer, 2002) indicated that dewatering efforts could require 10,000 to 15,000 gpm.

Removal of the estimated volume of groundwater had several undesirable effects, including the cost of pumping and the practicality of installing enough well capacity in the limited project area. Furthermore, it was uncertain if contamination on the site to the south might be drawn in.

The dewatering posed a significant environmental concern because the salt content was several times that of sea water. Scientists studying the lake during the 3-day pump test detected a significant spike in chlorides. Since the construction would require about 2 months of pumping at about 10 times the test rate, the dewatering was not acceptable. The owner subsequently issued a change order to construct the excavation without dewatering in excess of 100 gpm.

The most practical way to conform to the dewatering limits was to seal the excavation bottom and sides by jet grouting before beginning excavation. The seal needed to be thick enough to resist the buoyancy of the open excavation that would extend below the water table by about 20 ft. Relying on the existing structure to resist the uplift was not considered reliable because it was eccentrically located next to the excavation and it needed its own weight to resist its own buoyancy. It was viewed prudent to limit the risk of shifting the structure that might occur if load was transferred to it. As a consequence, the bottom seal needed to be stable without contribution from the surrounding structure. Evaluation showed the most economical seal would be attained with a 10-ft-thick bottom seal supplemented with 18 60-kip pressure-grouted ground anchors.
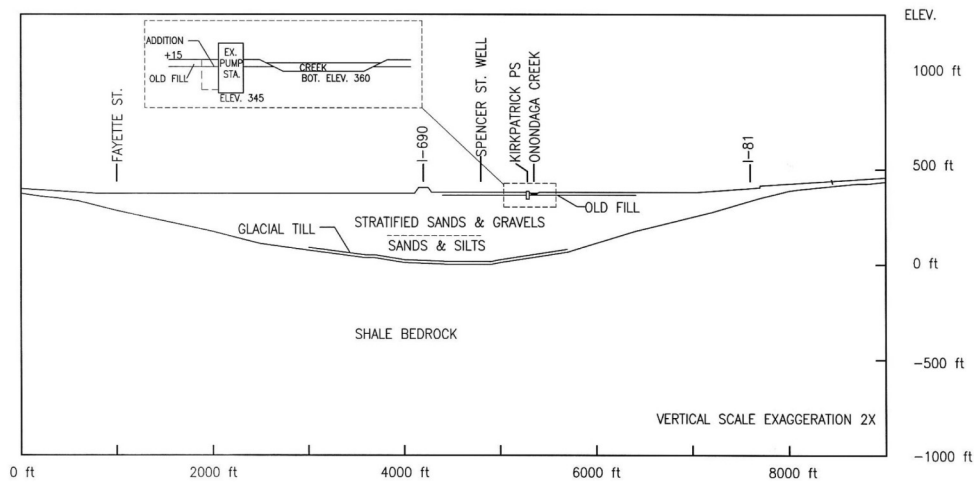

FIGURE 7—Geologic section at the Kirkpatrick Street pump station.
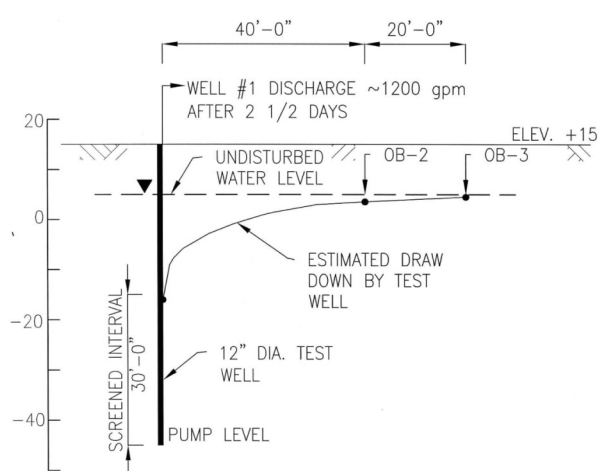
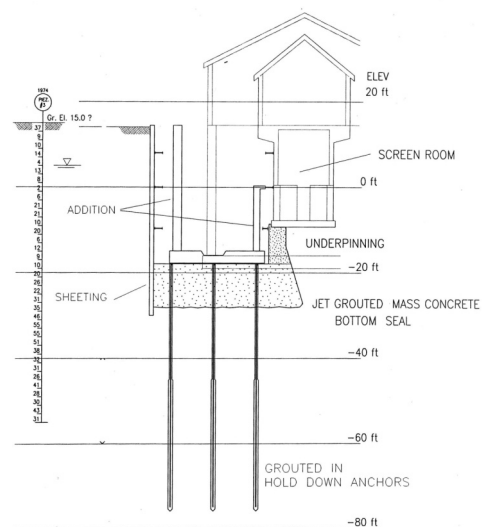

FIGURE 8—Results of aquifer test at Kirkpatrick St.

FIGURE 9—Excavation cross section with down-hole anchors and bottom seal

The first step of construction was to drive interlocked hot-rolled steel sheetpiling on the edges of the excavation that were accessible. Sheeting extended to the bottom of the proposed bottom seal. The next step was to construct a few jet grout test columns to confirm proper procedures and equipment were implemented for the site. The test columns were cored and demonstrated the grout was of good quality. The cores showed thin silty layers and significant proportion of large pieces of gravel and small cobbles. The third step consisted of underpinning the existing structure by jet grouting. Concurrently, jet grout columns were installed to construct the bottom seal, as shown in Figure 9.

After constructing the underpinning and bottom seal by jet grout columns, the fourth step was to install the ground anchors. Ground anchors were drilled from the ground surface through the bottom seal and approximately 30 ft below the bottom seal. After the last anchors were installed, excavation began inside the sheeting and bracing was installed. The work was completed and no movement of the existing structure was detected. The maximum dewatering effort was about 20 gpm and well within the requirements set by the Owner. Most of the water entered the excavation in one localized leak in the side seal between the sheeting and the jet grout. It is likely that the sheeting flexed under the earth pressure loads and separated slightly from the jet grout mass. Nevertheless the leakage was essentially insignificant.

Case History 2 (Thick Lacustrine Sediments Over Porous Glacial Sand)

A second deep excavation closer to Onondaga Lake was constructed about one year later to a similar depth. The site was in an area where the lacustrine soil overlying the glacial sands and gravel was much thicker than for Case History 1. There was a shallow groundwater table with about 10 ft of old fill overlying about 160 ft of interlayered soft clayey silt and loose sandy silt that overlaid the glacial sands.

Even though this excavation was of similar depth and over 250 times larger, the dewatering volume was several orders of magnitude less than would have been required for Case History 1 without the grout seal. As a consequence, the excavation was dewatered by confining the excavation within interlocked steel sheetpiling and pumping from widely-spaced well points around the perimeter. The structure was supported on steel H-piles driven to bear deep in the glacial sands almost 300 ft below the ground surface.

FIGURE 10—Excavation photograph of Case History #2

Case History 3 (Thin Lacustrine Sediments Over Porous Glacial Sand)

A third deep excavation was constructed near the north end of the lake in an area of shallow groundwater and where about 30 ft of low permeability lacustrine clayey silt and silty sand lie over more porous glacial sand. An excavation for a pump station in the 1960's was apparently constructed to a depth of about 20 ft with limited dewatering, although no records were available to confirm.

FIGURE 11—Case History #3 photographs. A) Construction of hold-down anchors. B) Inside excavation.

Pumping test results indicated that the excavation could be dewatered only if the deep sands were depressurized by pumping several thousand gpm from deep wells. Dewatering at this rate posed risks to a nearby gas transmission line and existing structures. It was subsequently decided to construct a bottom seal inside interlocked steel sheetpiling before making the excavation.

The bottom seal was constructed by jet grouting a 10-ft-thick plug of soilcrete and installing 48 hold down anchors that extended about 70 ft below grade. After completing the bottom seal by jet grouting and installing the hold down anchors, the excavation was completed by dewatering inside at a rate of less than 2 gpm.

These case histories describe how the local geology influences dewatering efforts required to construct civil works in the Onondaga Lakefront area.

REMEDIATION AND ENGINEERING CHALLENGES AT ONONDAGA LAKE

[This section prepared by G. Swenson and T. Johnson]

This portion of the trip is focused on the geology and hydrogeology of the western side of Onondaga Lake and the Ninemile Creek Valley (Figure 12). Due to the complexities associated with this system, various remediation and engineering challenges have been encountered as part of the ongoing cleanup efforts conducted by Honeywell adjacent to Onondaga Lake. A summary of these topics are presented below and will be discussed in more detail during the field program.

Geology of Onondaga Lake and the Ninemile Creek Valley

The geology of Onondaga Lake and the Ninemile Creek Valley is dominated by unconsolidated deposits associated with glaciofluvial and glaciolacustrine deposition. Onondaga Lake occupies a glacial trough through the Silurian shales and evaporites. The Ninemile Creek valley also occupies a bedrock low developed by fluvial and glacial activity. The overburden deposits associated with Onondaga Lake are represented by the following sequence beginning at the top of bedrock (Figure 13):

- till;

- ice contact basal sand and gravel layer;

- glaciolacustrine silt and fine sand layer – this layer is relatively massive and tends to have a coarser texture toward the lake outlet;

- glaciolacustrine and lacustrine silt and clay layer – this layer tends to be thicker toward southern end of lake;

- lacustrine marl - some of these deposits have been dated between 5,000 – 7,000 years B.P. (Kappel 2006); and

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- recent fill/settling basins along shore and lake sediments and in-lake waste materials.

The overburden deposits associated with the Ninemile Creek Valley are represented by the following sequence beginning at the top of bedrock (Figure 14):

- glaciofluvial and deltaic deposits of sand and gravel, sand, and silt, which are generally coarser in texture to the southwest and become finer in texture near Onondaga Lake;

- recent fluvial and floodplain deposits of sand, silt and clay;

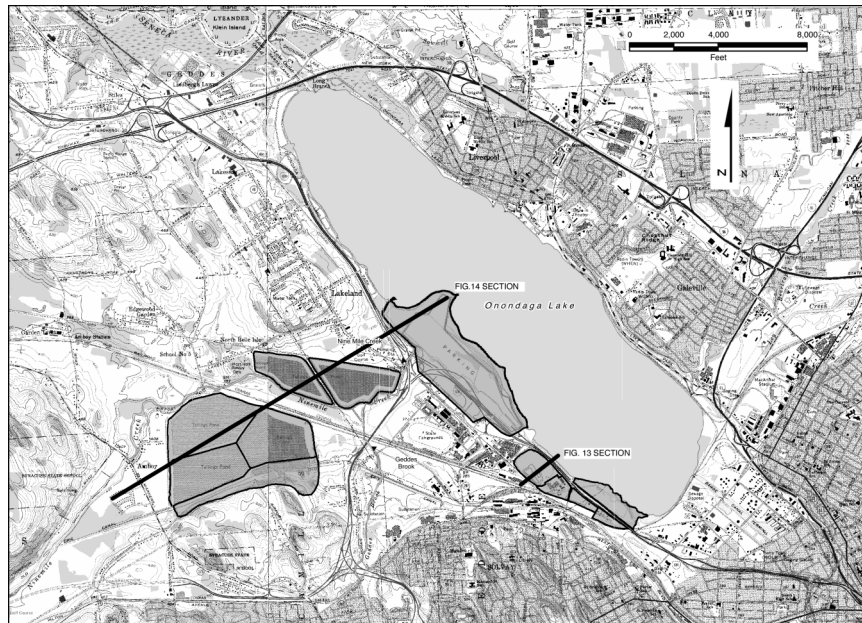- recent deltaic deposits; and

- fill/settling basins.

FIGURE 12—Location of settling basins and cross sections.
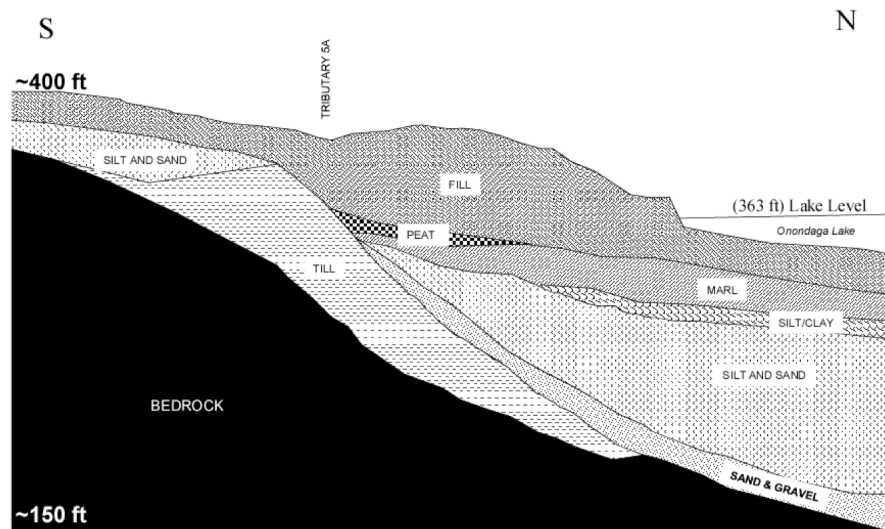

FIGURE 13—Representative cross section of the Willis/Semet site

The deposits in Ninemile Creek Valley reflect fluvial and deltaic deposition. These materials merge with the Onondaga Lake sequence to reflect fluvial discharge to historic Lake Iroquois. Much of the unconsolidated deposits under Onondaga Lake reflect the more recent lacustrine deposition of the sediments associated with glaciofluvial discharges to the lake.
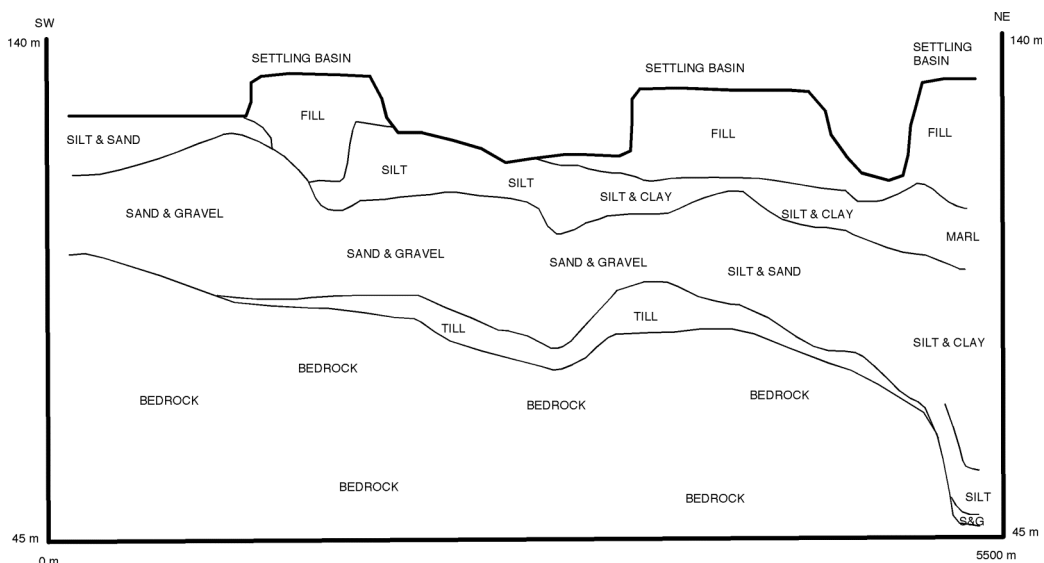

FIGURE 14—*General Ninemile Creek cross section.*

Groundwater of Onondaga Lake and the Ninemile Creek Valley

   Background.—Groundwater associated with the Onondaga Lake and Ninemile Creek Valley consists of three principle types of water: naturally occurring halite brine, fresh water, and a leachate associated with the settling basins. The halite brine saturates the overburden deposits under the lake and has a density of up to 1.16 g/cm^3, chloride concentrations over 120,000 mg/l, and halite saturations of up to 80%. The halite brine was formed over 16,000 years ago (Yager and others, 2007) and is confined by the silt and clay layer present beneath the lake. Discharge of the brine through the silt and clay confining layer to Onondaga Lake is limited, estimated to be very low (Effler 1996) (Kappel, 2006). Some discharge also occurs through the shallow units along the shore of the lake, where the confining layer pinches out, and historic discharges initiated the Syracuse salt industry in the late 1700's (Kappel, 2000).

   Leachate associated with the settling basins is present in both the shallow and deep overburden deposits. This leachate has a calcium, sodium, chloride composition and is differentiated from the halite brine by the higher percentage of calcium. The density of this leachate can measure up to 1.06 g/cm3. During active use of the settling basins during the 1900's, high volumes of process water were discharged to the settling basins. Currently, recharge of the settling basins only occurs via precipitation.

   Groundwater Flow.—Characterization of groundwater flow in the vicinity of Onondaga Lake is complicated by the presence of the dense halite brine and the leachate. In general, the regional direction of groundwater flow is toward Onondaga Lake. Because of the presence of the dense brine and leachate, which the fresh water can not displace, much of the fresh water is forced to discharge to streams surrounding the lake, such as Ninemile Creek. However, there is small amount of localized fresh water discharge to the lake through the shallow sediments.

   The distribution and flow of the settling basin leachate is governed by historic settling basin operations and the interaction with native brines and fresh water. The current hypothesis is that historic loading of the settling basins resulted in a dense plume of leachate that migrated downward to the basal sand and gravel unit

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displacing the halite brine along the lakeshore. Once the settling basin loading ceased, the deep leachate plume remained trapped beneath the confining layer along with the halite brine. Current leachate created by natural recharge to the settling basins discharges to the lake and streams though the shallow deposits.

Due to the complexities associated with evaluating groundwater flow with multiple densities, a groundwater model was developed to support multiple evaluations. This model uses the USGS code SEAWAT 2000 (Langevin et al. 2003) to simulate the movement of the fresh and dense groundwater in the Ninemile Creek Valley and around Onondaga Lake. This model is currently being used to evaluate groundwater flow and various evaluations associated with the remediation of Onondaga Lake and associated Honeywell sites.

Remediation and Engineering Challenges.—Historic industrial practices, in addition to geologic and hydrogeologic conditions around Onondaga Lake, present a variety of remediation and engineering challenges. Much of the overburden deposits around Onondaga Lake are low density. Weight of hammer or low blow counts are common when using conventional soil boring sampling techniques. These low density materials present several construction challenges for buildings and other structures. For example, Carousel Mall was designed to "float" on the soft sediments. The proposed expansion of the mall and the Interstate 690 bridge to the east of this area are using piles driven to depths of 30 m or more to provide an engineering base for construction.

As part of the remediation program for several sites adjacent to Onondaga Lake, Interim Remedial Measures (IRMs) are being implemented to stop the flow of contaminated groundwater and DNAPL to Onondaga Lake. The IRMs will consist of a sheet pile barrier wall and groundwater collection trench with wick drains to capture and pump the water to a treatment plant on the Willis Avenue site. DNAPL recovery wells will also be installed along the shoreline to extract this material from behind the barrier wall. The sheet pile wall will be keyed into the silt/clay layer approximately 25-45 ft below the surface, which is a confining layer to groundwater flow. The low density soils, presence of the silt/clay confining layer, potential need for lake dredging near the wall, and the close proximity of Interstate 690 and utilities present multiple geotechnical design challenges for the construction of the wall and trench.

A portion of the IRMs noted above are focused on the deeper sand and gravel units. The groundwater in the deep zone does not appear to be migrating to Onondaga Lake, however, leachate and other organic constituents in the deep basal sand and gravel layer may require some type of remediation. The deep unit is highly permeable and is under artesian conditions (up to 10 ft above lake level) due to the presence of the silt/clay confining layer. Design of a remedy for this unit is complicated since recovery wells in the deep unit could produce high volumes of water and potentially cause preferential migration of the brines in this unit.

Discharges to Onondaga Lake from historic industrial practices have resulted in sediments and other deposits on the bottom of the lake that will require remediation. Isolation capping has been selected by the NYSDEC as a key component of the remedy for the impacted sediments within the lake. In some areas of the lake, the isolation cap will consist of 1-2 m of sand carefully placed over the soft sediments. Stability of these sediments is a key component of the ongoing design evaluations for the remedy. The amount of groundwater upwelling through the shallow sediments and ultimately into the isolation cap is also a critical portion of the design. The low upwelling velocities are difficult to measure and therefore multiple methods are being used to document this information. Data is currently being collected as part of ongoing pre-design investigation activities on the lake to evaluate stability and groundwater upwelling estimates.

REFERENCES CITED

Continental Placer, 2001, Kirkpatrick Street Dewatering Project (unpublished Consultant's Report).

Denny, C.S., and Lyford, W.H., 1963, Surficial geology and soils of the Elmira-Williamsport region, New York—Pennsylvania: U.S. Geological Survey Professional Paper 379, 60 p.

Effler, S.W., 1996, Limnological and Engineering Analysis of a Polluted Urban Lake: Springer, New York, 832 p.

Hand, B.M., 1978, The Syracuse meltwater channels: *in,* Merriam, D.F., ed., New York State Geological Association Fieldtrip Guidebook, p. 225-249.

Kappel, W.M., 2000, Salt production in Syracuse, New York ("the salt city") and the hydrogeology of the Onondaga Creek valley: U.S. Geological Survey Fact Sheet FS 139-00, 8 p.

Kappel, W.M., and Miller, T.S., 2005, Hydrogeology of the Valley-Fill Aquifer in the Onondaga Trough, Onondaga County, New York, U.S. Geological Survey Scientific Investigations Report 2005-5007, 13 p.

Kappel, W.M., and Miller, T.S., 2003, Hydrogeology of the Tully Trough -- Southern Onondaga County and Northern Cortland County, New York, U.S. Geological Survey Water-Resources Investigations Report 03-4112, 16 p.

Kappel, W.M., Miller, T.S., and Hetcher, K.K., 2001, Hydrogeology of the Tully Lakes area in southern Onondaga and northern Cortland Counties, New York, U.S. Geological Survey Water-Resources Investigations Report 01-4166, 16 p.

Kappel, W.M., and McPherson, W.S., 1998, Remediation of mudboil discharges in the Tully Valley of Central New York, U.S. Geological Survey Fact Sheet FS 143-97, 4 p.

Ku, F.K.H., Randall, A.D., and MacNish, R.D., 1975, Streamflow in the New York part of the Susquehanna River basin: New York State Environmental Conservation Bulletin 71, 130 p.

Langevin, C.D., Shoemaker, W.B, and Guo, W., 2003, MODFLOW-2000, the U.S. Geological Survey Modular Ground-Water Model – Documentation of the SEAWAT-2000 Version with the Variable-Density Flow Process (VDF) and the Integrated MT3DMS Transport Process (IMT): USGS Open-File Report 03-426. Tallahassee, Florida.

Leonards, G.A., ed., 1962, *Foundation Engineering*, McGraw-Hill Book Company, New York. 1146 p.

Pair, D.L., 1998a, The surficial geology of Otisco Valley, NY Quadrangle (7.5 minute): United States Geological Survey–New York State Geological Survey Map no. 2gk1234, scale 1:24:000.

Pair, D.L., 1998b, The surficial geology of Tully, NY Quadrangle (7.5 minute): United States Geological Survey–New York State Geological Survey Map no. 2gk1236, scale 1:24:000.

Ridge, J.C., 2003, The last deglaciation of the northeastern United States—a combined varve, paleomagnetic, and calibrated 14C chronology, *in,* Cremeens, D.L., and Hart, J.P., (eds.), Geoarchaeology of landscapes in the glaciated northeast—Proceedings of a symposium held at the New York Natural History Conference VI: Albany, NY, New York State Museum, Bulletin 497, p. 15-45.

Stewart, G.W., 1937, "Subsurface Conditions in and Around Syracuse, NY," M.S. Thesis, Dept. of Geology, Syracuse University, NY.

Stewart, J.P., Peterson, A.L., Begley, T.J., and Falter, D.C., 2003, Kirkpatrick Street Pump Station Addition Excavation – Small Project, Big Twist: Earth retention Systems 2003 – A Joint Conference, American Society of Civil Engineers Metropolitan Section Geotechnical Group Meeting, New York, New York, 16 p.

Yager, R.M. Kappel, W.M., and Plummer, L.N., 2007, Origin of halite brine in the Onondaga Trough near Syracuse, New York, USA: modeling geochemistry and variable-density flow: Hydrogeology Journal, v. 15, no.5, p. 713-728

APPENDIX

Selected U.S. Geological Survey Reports Relative to the Onondaga Trough and the surrounding region, with web addresses.

Kappel, W.M., and Teece, M.A., 2007, Paleoenvironmental Assessment and Deglacial Chronology of the Onondaga Trough, Onondaga County, New York: U.S. Geological Survey Open-File Report 2007-1060, 12 p., http://pubs.usgs.gov/of/2007/1060/

Kappel, W.M. and Miller, T.S., 2005, Hydrogeology of the Valley-Fill Aquifer in the Onondaga Trough, Onondaga County, New York, U.S. Geological Survey Scientific Investigations Report 2005-5007, 13 p., http://ny.water.usgs.gov/pubs/wri/sir055007/

Kappel, W.M. and Miller, T.S., 2003, Hydrogeology of the Tully Trough -- Southern Onondaga County and Northern Cortland County, New York, U.S. Geological Survey Water-Resources Investigations Report 03-4112, 16 p. http://ny.water.usgs.gov/pubs/wri/wri034112/

Kappel, W.M., Miller, T.S., and Hetcher, K.K., 2001, Hydrogeology of the Tully Lakes area in southern Onondaga and northern Cortland Counties, New York, U.S. Geological Survey Water-Resources Investigations Report 01-4166, 16 p. http://ny.water.usgs.gov/pubs/wri/wri014166/

Kappel, W.M. and Landre, B.F., 2000 (Revised 2002), Managing the water resources of the Oswego River Basin in central New York, U.S. Geological Survey Fact Sheet FS 180-99, 6 p., http://ny.water.usgs.gov/pubs/fs/fs18099/

Kappel, W.M., 2000, Salt production in Syracuse, New York ("The Salt City") and the hydrogeology of the Onondaga Creek Valley, U.S. Geological Survey Fact Sheet FS 139-00, 8 p., http://ny.water.usgs.gov/pubs/fs/fs13900/

Kappel, W.M. and McPherson, W.S., 1998, Remediation of mudboil discharges in the Tully Valley of Central New York, U.S. Geological Survey Fact Sheet FS 143-97, 4 p. http://ny.water.usgs.gov/pubs/fs/fs14397/

Kappel, W.M., Sherwood, D.A., and Johnston, W.H., 1996, Hydrogeology of the Tully Valley and characterization of mudboil activity, Onondaga County, New York, U.S. Geological Survey Water-Resources Investigations Report 96-4043, 71 p., http://ny.water.usgs.gov/pubs/wri/wri964043/

McKenna, J.E., Chiotti, T.L., and Kappel, W.M., 1999, Ecological status of Onondaga Creek in Tully Valley, New York -- Summer 1998, U.S. Geological Survey Fact Sheet FS 141-99, 6 p., http://ny.water.usgs.gov/pubs/fs/fs14199/

Pair, D.L., Kappel, W.M., and Walker, M.S., 2000, History of landslides at the base of Bare Mountain, Tully Valley, Onondaga County, New York, U.S. Geological Survey Fact Sheet FS 0190-99, 6 p., http://ny.water.usgs.gov/pubs/fs/fs19099/

Wieczorek, G.F., Negussey, Dawit, and Kappel, W.M., 1998. Landslide hazards in glacial lake clays-Tully Valley, New York., http://pubs.er.usgs.gov/pubs/fs/fs01398

Yager, R.M., Kappel, W.M., and Plummer, L.N., 2007a, Origin of halite brine in the Onondaga Trough near Syracuse, New York State, USA: modeling geochemistry and variable-density flow: Hydrogeology Journal, v. 15, n.5, p. 713-728., http://www.springerlink.com/content/008m772313n73413/

Yager, R.M., Kappel, W.M., and Plummer, L.N., 2007b, Halite brine in the Onondaga Trough near Syracuse NY: Characterization and simulation of variable-density flow: U.S. Geological Survey Scientific Investigations Report 2007-5058. http://ny.water.usgs.gov/pubs/ (IN PROCESS)

Yanosky, T.M. and Kappel, W.M.,1997, Tree rings record 100 years of hydrologic change within a wetland. U.S. Geological Survey Fact Sheet FS 0057-97, 4 p. http://ny.water.usgs.gov/pubs/fs/fs05797/

ROAD LOG FOR TRIP -1

GLACIAL AND ENVIRONMENTAL GEOLOGY OF THE ONONDAGA TROUGH, CENTRAL NEW YORK

All of the field sites described in the road log are on private property. As such, it is imperative that we respect the rights and wishes of these land owners. Please do not visit these sites on your own, as it may jeopardize future field-trip opportunities. Over the past 15 years it has become increasingly difficult to maintain our access agreements to these sites as individuals and even groups of people have entered these properties without obtaining land-owner permission. We strive to maintain good relations with the land owners and do not want the inappropriate actions of a few to ruin the educational opportunities for many others who wish to enter these areas. The USGS can, and does act as the 'point of contact' for the property owners and we would be happy to facilitate your future access to these sites. Please contact Bill Kappel for any questions as to access for you and your classes.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Start at parking lot north of Old Main building on SUNY Cortland campus. From parking lot turn left (E) on Gerhart Dr., Yield sign turn left (N) onto Graham Ave. At bottom of hill (traffic light) turn left (W) onto Groton Ave, (Route 222).
1.1	1.1	CORTLAND CAMPUS TO NYS-ROUTE 281. At intersection with NYS-Route 281, turn right (N) and proceed through the outskirts of Homer. Once past the Homer High School, and crossing NYS-Route 90, on your left is the Homer waterworks about 0.7 mi from Rte. 90. Turn left into waterworks
4.4	3.3	ROUTE 281 TO HOMER WELLFIELD

STOP 1. HOMER WATERWORKS—SITE OF HANGING VALLEY ENTERING THE SOUTHERN END OF THE ONONDAGA TROUGH

At this location the Village of Homer has two municipal supply wells (75 and 83 ft deep) that tap an unconfined outwash sand and gravel aquifer that yield 500 to greater than 1,000 gal/min of potable water. Seismic refraction was used to determine the depth to bedrock within the Factory Brook valley where the water works are located. About 1 mi up-valley from the well field the depth to bedrock was only about 110 ft (elevation. of 1,110 ft) – a subsequent test well ground-truthed the seismic data. These results indicated that the Factory Brook V. is a hanging valley to the W. Br. Tioughnioga River V. where the elevation of bedrock in the center of the valley is at about 900 ft or 200 ft deeper than the Factory Brook Valley. We believe the Homer wells are at the lip of the hanging valley and starting at Rt 281 (a stone throws away) the buried valley wall plunges down into the very southern extent of the Onondaga Trough.

Leave waterworks and turn left, back on to Route 281 (N). Follow Route 281 North, cross over I-81 near Preble, NY, and continue to the intersection with Route 80 just past the Tully exit of I-81. *As you travel this route, you are moving in and across the southern end of the Onondaga Trough. The bridge crossing of Route 281 at I-81 is the location of geologic section B-B' (Fig. 6, Kappel and Miller, 2003). The intersection of NYS-Route 80 and I-81 is the location of geologic section A-A' of figure 6, cited above.*

| 16.6 | 12.2 | HOMER WELLFIELD TO NYS-ROUTE 80 AT TULLY I-81 INTERCHANGE |
| 17.6 | 1.0 | TULLY I-81 INTERCHANGE TO LAKE RD. AND GREEN LAKE |

STOP 2. TULLY/VALLEY HEADS MORAINE AT GREEN LAKE

At this location we are on the Valleys Heads Moraine at Tully. This moraine 'plugs' the Onondaga Trough. The term 'trough' stems from a definition by W.M. Davis – *"Along a number of valleys it is possible to pass from one drainage system to the other through open valleys in which the present divides are determined not by rock, but by drift deposits"*. Although Tarr (1905) provided the definition, the term "through valley" is attributed to W.M. Davis, who coined the phrase during the discussion of Tarr's paper at the 1904 Geological Society meeting (Tarr, 1905, p. 233). The moraine forms the surface water divide between St. Lawrence and Susquehanna River Basins.

The location of the Valley Heads Moraine in central New York was thought to closely coincide with a preglacial bedrock divide. However, the study of the moraine has shown that during the last glacial episode, there is no bedrock ridge or high point high beneath the moraine. Rather, the trough continues a gradual slope upward starting near Syracuse, goes beneath the moraine, and continues south for several more miles to about Homer where it eventually flattens out and then begins to slope southward in the Tioughnioga River valley.

The surface-water/ground-water interaction in the Tully Lakes area is very complex and the surface-water divide often does not coincide with the ground-water divide. Ground water generally flows from the edges of the valley toward the center, where it discharges to lakes, ponds, wetlands, springs, and the West Branch Tioughnioga River. Most ground water north of the ground-water divide discharges to springs at an elevation of about 1,125 feet on the northern slope of the moraine; the rest flows northward as underflow to deeper zones of valley-fill deposits in Tully/Onondaga Creek valley. Ground water south of the divide flows southward eventually entering the Tioughnioga River.

A round of synoptic water-level measurements indicated that water levels in the western Tully Lakes are generally above those in the surrounding aquifer; therefore, water seeps from the lakes to the aquifer most of the time. In contrast, the eastern lakes generally receive water from the aquifer. Green Lake commonly receives ground-water flow along its east side and discharges to the aquifer along its west side—while at the same time surface water drains southward out of the lake through a channel at the south end and empties into Tully Lake. Tully Lake receives ground-water flow from all sides during the spring recharge conditions; then, as water levels decline from summer through early winter, it discharges to the aquifer along its southern edge.

		stop sign you will intersect with Route 80 again. *See geologic section F-F' (Kappel, Miller, 2005) for the unconsolidated stratigraphic section of this area.*
19.8	2.2	LAKE RD.GRAVEL PIT TO NYS-ROUTE 80

Proceed north across Route 80 **(***BE VERY CAREFUL – cars zip along Route 80 and there is limited sight-distance***)**. After you have safely crossed Route 80 you are now on Tully Farms Road, headed down the north side of the Tully (Valley Heads) Moraine. *Note the steepness of the road, the many springs about half way down the back side of the Moraine.*

Continue North on Tully Farms Road. *The road has been affected by subsidence where it crosses Onondaga Creek, due to brine mining.* As you pass through a dog-leg in the road, and through the middle of a farm (*watch for loose cows, cats, kids, and farm machinery*) you will go up a slight incline (*Rattlesnake Gulf alluvial fan*) to the intersection with Otisco Road. *See geologic section E-E' (Kappel, Miller, 2005) for the unconsolidated stratigraphic section of this area.*

At this intersection, turn right (E) and follow the road to the dead end. *Again be careful of small kids in the cluster of houses, and when you park, do so to the south side of the road, and do not block the vehicle turn-around area on the right.*

24.1	4.3	NYS ROUTE 80 TO OTISCO ROAD – DEAD END

STOP 3. TULLY VALLEY MUDBOILS

At the parking area you can view several different features. Going around the barrier, and walking toward what used to be the bridge over Onondaga Creek you will see the remains of the foundation for the bridge. The downstream side wing walls a tipped back to the south and have dropped about a foot. The upstream wingwalls are just at or under the water surface. The mudboil was located about 30 feet upstream of the former bridge.

Adjacent to Otisco Road on the south side was a formerly active mudboil area (circa 1970's). The only vegetation present now is water-tolerant grassy species and willow trees. The road way experienced minor subsidence.

Walking back up Otisco Road (W) and on your left there will be a red-pipe farm gate. Walk around the gate (watch for poison ivy) and proceed south, down the farm lane. Where the field on the left ends at the tree line, continue south about 300 feet along the farm lane, and there will be a spur road on your left, that drops off the lane (SE) to the Rogue mudboil area. A former depressurizing well is now marked by a mudboil 'pool' and you should be able to note subsidence cracks as you come down this roadway. Walk around the large pond (former mudboil, now diked) where you should find a 6-inch PVC pipe discharging into the pond from another depressurizing well. This well and surrounding area has subsided about 3+ feet. With extreme caution you should be able to walk across the volcano-like cone of the mudboil sediments at the base of this well. Note the fine-grained nature of these sediments, and if you lightly tap these sediments with your foot, you might get them to liquefy. *DO NOT OVERDO THIS LIQUIFACTION!* Back in the woods are other mudboils – *BE VERY CAREFUL IF YOU DECIDE TO TAKE THIS WALK IN THE WOODS, THE MUDBOILS ARE QUICKSAND-LIKE.*

The main mudboil area (MDA) is further south down the farm lane, but the area is covered with phragmities so it is difficult to see into the MDA. If you venture in that direction, there is a moat around the MDA, and if you carefully walk on the dike you can find some active mudboils in the southwest quadrant, and associated land subsidence further uphill. At the outlet of the MDA there is a Parshall flume with associated discharge measuring equipment.

Once you have safely returned to Otisco Road and your vehicle, head back to Tully Farms Road (W) and at the intersection turn right (N) and take Tully Farms Road until it intersects with US Route 20. *As you travel down Tully Farms Road (passing Nichols Road on your right) you will travel through the remains of the 1993 landslide area. This is not a scheduled stop, but if you pull off the by the white cinderblock house, you can see the results of the landslide. Please refer to Wieczorek and others, 1998; Pair and others, 2000 for further information on this slide. Please do not walk into this slide area without permission or venture into the tumble-down house – the house is unsafe and both are private property!*

27.4	3.3	OTISCO ROAD DEAD-END TO US ROUTE 20

At Route 20 you have two options – if you would like to see an overview of the Tully Valley, the directions follow. If you want to continue the 'official' tour route, skip the following VALLEY OVERVIEW section and follow the CONTINUATION directions.

SIDE TRIP FOR VALLEY OVERVIEW

At the intersection of Tully Farms Road and Route 20, take a right on Route 20 (E), (*Careful! Dangerous intersection – cars moving quickly and you have limited sight distance*) proceed up the hill, driving toward I-81. Near the top of the hill take a left hand turn on to Webb Road (~1.7 miles) – careful, as it is almost a U-turn onto Webb and Route 20 traffic is moving quickly. Follow Webb for ~0.4 miles then turn left on to Amidon Road. Follow Amidon Road until it bends to the right (uphill) and turns into Summer Ridge Road. Take the road to the top and then park on the edge of the road where it takes a right-hand turn. Get out and look to the south and see a panorama of the southern Onondaga Trough --the Tully Valley, Tully Moraine in the distance, and Song Mountain ski area in the far distance. Follow the reverse route back down to Tully Farms Road at Route 20 – take a right on to Tully Farms Road Extension (N) from Route 20.

0.0	*3.2*	SIDE TRIP -- ROUTE 20 TO TULLY VALLEY OVERVIEW AREA

CONTINUATION OF FIELDTRIP
See geologic section D-D' in Kappel, Miller, 2005 for the unconsolidated stratigraphic section near US Route 20

Drive across Route 20 (*Careful! Dangerous intersection – cars moving quickly and you have limited sight distance*). Follow the Extension Road north to the intersection with NYS-Route 11A. Take a left on to Route 11A (N) and on your left after a few miles will be the Onondaga Creek flood control dam. *See geologic section I-I' (Kappel, Miller, 2005) for the unconsolidated stratigraphic section of this area. This is near the intersection of the main and West Branch valleys of Onondaga Creek. The high bluffs to the west side of the dam are over the thalweg of the bedrock valley, and these bluffs are fluvial deposits discharged from the West Branch valley when the ice front was still located further to the north during the*

development of the Syracuse Channels (Hand, 1978). If you stop here, you can see the spillway for the dam cut into the local limestone bedrock and the coarse sand and gravel of the fluvial deposits. This dam is on Onondaga Nation property, so don't wander and don't stay too long as the Onondagas are protective of their land and don't like the dam being here.

Continue North on Route 11A, through the Onondaga Nation for and at the 'T' intersection with US Route 11, take a right and go east toward I-81. Just under I-81, take an immediate left for the entrance ramp for I-81 North.

37.7	7.1	ROUTE 20 TO I-81 INTERCHANGE AT NEDROW
44.4	6.7	I-81 AT NEDROW TO I-690 WEST INTERCHANGE

Follow I-81 North to the intersection with I-690 West [~6.7 mi]– this can be a difficult right lane merge with another entrance ramp to I-81—that is, you will move over one lane to the right to get into the I-690 West lane (toward Baldwinsville). Follow this lane and then merge on to I-690 West – right hand merge. Go past the West St. exit and then again move right to get off at Geddes Street. At the foot of the exit ramp, take a right and drive to the traffic light intersection with Kirkpatrick Street. At this light, take a right onto Kirkpatrick St. (NE) and cross Van Rensselaer St. Just ahead, on the left is the parking lot for the Inner Harbor amphitheater (do not cross the bridge over Onondaga Creek). Turn left, park, and go over the knoll to the Inner Harbors tent structure (white sail-like roof).

46.7	1.8	I-690 WEST TO INNER HARBOR

STOP 4. THE SYRACUSE INNER HARBOR – FUTURE COMMERCIAL AND RESIDENTIAL REDEVELOPMENT OF THE FORMER OIL CITY AREA – ENGINEERING GEOLOGY

The Inner Harbor is at the southern end of extensive urban redevelopment for the greater Syracuse area. The expanse of land and water surrounding the Inner Harbor site is slated for massive construction in the near future as is the intervening area northeastward toward the Carousel Mall (green topped buildings) with the development of an expanded Carousel Mall, the potential development of DestiNY USA, and the supporting infrastructure. Discussions here will center around the difficulties and innovations needed to cope with the glacial sediments and salty water that underlie this entire area.

Leave the parking lot and turn right (SW) and back down Kirkpatrick St. to the traffic light. Go straight (a slight right turn) ahead on Spencer Street to the Bear Street traffic light. Just ahead on the left is the on-ramp for I-690 West, get on I-690 (there is a lot of construction on I-690 in 2007) and the next exit is for State Fair Blvd. Exit right, but at the end of the ramp do not follow the road under I-690; go straight ahead and up on toward the Wastebed Parking lot. There is a gate that is normally locked but will be open for the fieldtrip. Those that follow can park off to the side of the road and walk out on the wastebeds, but again this is private property – you'll need permission to do so or go there when the State Fair is ongoing. *There are plans to develop the area with a walking trail as the wastebeds and associated wastes in the lake are to be remediated which we will hear about at Stop 5.*

| 48.8 | 2.6 | INNER HARBOR TO WASTE BEDS |

STOP 5. THE WASTEBED DEPOSITS ADJACENT TO ONONDAGA LAKE AND THE REMEDIATION OF GROUNDWATER, IN-LAKE WASTE DEPOSITS, AND OTHER ENGINEERING CHALLENGES

This portion of the trip is focused on the geology and hydrogeology of the western side of Onondaga Lake and the Ninemile Creek Valley. Meeting on top of one of the wastebeds gives a view of the lake, and allows one to see where various remedial activities are and will occur in the future. Due to the complexities associated with this hydrogeologic system, various remediation and engineering challenges have been encountered as part of the ongoing cleanup efforts conducted by Honeywell adjacent to Onondaga Lake. A summary of these topics are presented and discussed in more detail during the field program.

To return to Cortland follow the signs to Route I-690 East (presently there is a detour back toward the State Fairgrounds) and follow I-690 east and watch for the I-81 South (Binghamton) exit ramp. (This is another kamikaze entrance ramp) As you leave I-690 (right exit off of I-690) and approach I-81 you will need to get over to the left lane of I-81 as it narrows to one lane temporarily – watch for traffic entering from behind, and then join I-81 South (two lanes). Travel south on I-81 back to the Homer exit. Exit at Homer, and at the 'T' (NYS-Route 281) intersection, turn left on to Route 281 South. Follow 281 until it intersects with Route 222 – get in left hand turning lane at the light. Turn left on to 222 (Groton Road) and go to Graham Road, then up the hill to the Cortland Campus

| 84.8 | 36.0 | WASTEBEDS TO CORTLAND CAMPUS |

END OF FIELD TRIP

Trip A-2

PLEISTOCENE GLACIATION OF THE TIOUGHNIOGA RIVER AND FALL CREEK VALLEYS

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INTRODUCTION

The Pleistocene Epoch in eastern North America was characterized by a series of glaciations, the most recent of which was the Wisconsinan (from about 117 to 11.9 ka B.P.). Advance and retreat of glacial ice resulted in erosion of the landscape and the formation of landforms and deposition of glacial and related deposits. At the peak of the Wisconsinan glacial period, the Laurentide Ice Sheet extended from Canada and covered most of eastern North America, with its margin located in Ohio, northern Pennsylvania, northern New Jersey, and extending through Long Island into the Atlantic off the southern coasts of Connecticut, Rhode Island, and Massachusetts. Deposits of the previous Illinoian glaciation can be found to the south of the Laurentide margin, but are mostly lacking to the north of the margin as a result of erosion and reworking of sediments and deposition during the Wisconsinan glaciation (Randall 2001).

In New York State, glacial ice flowed south/southeast from Canada, as indicated by glacial striations deposits/landforms. Valleys parallel to ice flow were typically eroded into U-shaped glacial valleys, whereas valleys perpendicular to flow were infilled with glacial deposits and, upon ice retreat, eroded by meltwater and rivers. Over successive glaciations, through valleys such as the valleys of the Finger Lakes were carved out, and glacial landforms and deposits, including the drumlin fields of the Ontario Lowlands in western New York and several moraine systems were deposited / formed.

In the Late Wisconsinan Last Glacial Maximum (LGM), following ice advance to the position of the terminal moraine around 20,000 (radiocarbon) years ago, ice began to retreat back to the north. The retreat phase in the Appalachian Plateau of New York State was characterized by a relatively rapid retreat of ice in the uplands, while ice tongues remained longer in glacially eroded U-shaped valleys. Valley ice tongues retreated through the processes of back- and downwasting (Fleisher 1993). Backwasting involved active ice retreat and typically occurred in through valleys. Downwasting was characterized by stagnant ice retreat and took place in smaller valleys (Fleisher 1991).

Ice retreat from the LGM was accompanied by formation of lakes in front of the retreating ice margin, especially where previously deposited glacially derived sediment blocked southward drainage. At least one period of stagnation and readvance of ice resulted in the deposition of the Valley Heads Moraine (Stops 1, 3, and 4) within several valleys of Central and Western New York (Fig.. 1) (Miller et al. 1998). The Valley Heads Moraine is a characteristic ablation or dead ice moraine that formed as a result of ice stagnation, as shown by the large kettle ponds and lakes in the Tully region (Stop 1) confined between the moraine to the north and the valley train to the south (Fleisher 1991). Deposition of the Valley Heads moraine resulted in the formation of large proglacial valley train sequences confined to valleys south and east of the moraine (such as the Otisco, Factory Brook, West and East Branch Tioughnioga, Otter/Dry Creek, Trout Brook, and Main Stem Tioughnioga valleys, which are the subject of this field trip). Retreat of ice north from the Valley Heads position also led to development of large proglacial lakes.

Ice re-advance and the resulting deposition of the Valley Heads Moraine reorganized pre-LGM drainage divides in several of the through valleys in central New York. Two segments of the Valley Heads Moraine provide drainage divides and boundaries for the Tioughnioga and its associated drainage. The Tully Moraine (Stop 1) forms the drainage divide at the northern end of the West Branch Tioughnioga Valley. Drainage to the north of the moraine, in the Onondaga Trough, flows north towards the Seneca River and Lake Ontario, while drainage to the south of the moraine flows south through the West Branch and Main Stem Tioughnioga River to the Chenango River and eventually the Susquehanna. To the southeast of Cortland, the South Cortland Moraine (Stops 3 and 4) forms another drainage divide. Flow to the west of the moraine is west-southwest through the Fall Creek Valley to Cayuga Lake. To the east of this moraine, Dry Creek and Otter Creek flow east-northeast into the West Branch Tioughnioga River, which then turns southwest down the Main Stem Tioughnioga.

This trip focuses principally on the deglaciation of the Tioughnioga Valley between Tully and Whitney Point, NY, with additional examination of the Fall Creek Valley between Cortland and Ithaca, NY. The goal is to illustrate most of the key morphologic and sedimentologic features that characterize active and dead-ice retreat of the LGM ice through this area of central New York. In addition, exposures in Fall Creek Valley allow us to examine some pre-LGM deposits.
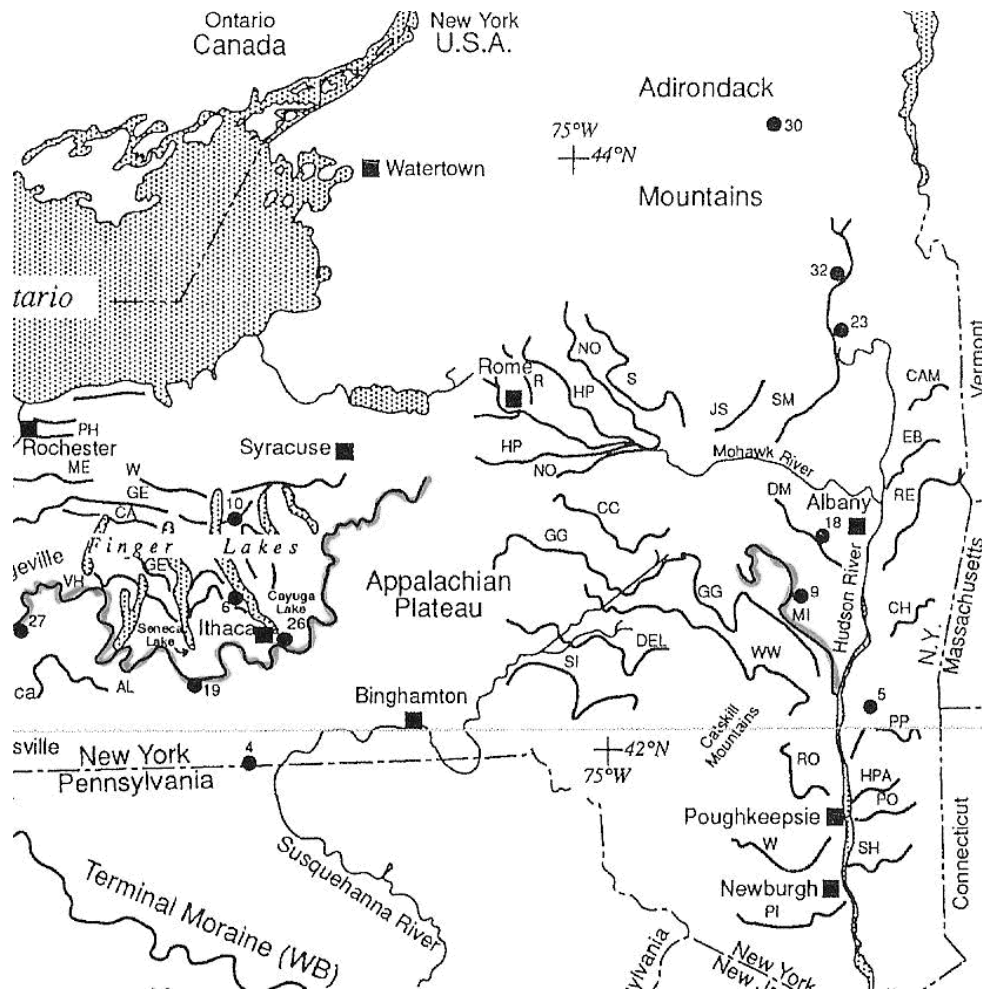

FIGURE 1—Valley Heads Moraine in Central New York (Muller and Calkin 1993).

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Glacial Deposition

The advance and retreat of glacial ice both in the uplands and within valleys resulted in the deposition of till over most of Central New York. Glacial till, also known as hardpan or fragipan, consists of a poorly sorted mix of sediments ranging in size from clay to boulders. Erratics, sediment clasts that have origins far from the region of deposition, are often found within tills, as well as other glacial deposits. In the field-trip area, these erratics are commonly metamorphic rocks, igneous rocks (granite), and red sandstones.

Ice-contact sediment (till) was deposited as lodgement, meltout or ablation, flow, or deformation till. Lodgement tills consist of debris deposited at the base of active glacial ice, and are characterized by compacted, unsorted, unstratified sediment with faceted and striated clasts. Meltout or ablation tills can be deposited subglacially as basal ice melts and deposits sediment, or supraglacially as the ice surface melts and sediments within the ice are released. Both sub- and supraglacial meltout tills are less compact than lodgement tills, but may be interbedded with meltwater deposits and may show some stratification. Supraglacial meltout tills, deposited from the retreating glacial terminus are more common than subglacial meltout tills. Flow tills result from supraglacial sediment flowing off the surface of glacial ice and depositing in front of the terminus. Clasts in these tills are usually more rounded, sorted, and the deposits are more stratified than meltout tills.

As glacial ice retreated from Central New York, sediment-laden braided rivers of glacial meltwater flowed down valleys that were previously filled with ice, resulting in the deposition of outwash sediments as valley train sequences (outwash confined to valleys). Sands and gravels were deposited on top of previously deposited sediments, including lacustrine deposits, tills, and previously deposited outwash, as well as along, above, or within/below stagnant ice remaining in the valley. The formation of several glacial valley landforms resulted from the interaction of meltwater with stagnant and/or active ice. Sources of this meltwater could have originated from up valley where active ice was retreating or stagnant ice was melting, or from upland meltwater sources flowing into the valley onto or up-valley from active or stagnant ice.

Stagnant ice blocks that were buried by sands and gravels produced kettle holes, lakes, and ponds when the ice melted and outwash sediments collapsed downward. Meltwater that flowed within or below stagnant ice deposited sands and gravels within a tunnel that was later exposed, after ice melted, as a ridge of outwash or an esker. Kame deposits, including isolated kames, kame terraces, and kame deltas, form from glacial meltwater flowing over the surface of glacial ice or between the ice margin and a valley wall. In instances when the ice surface contained depressions or holes, outwash sediments can infill these regions as meltwater enters them, slows, and deposits its sediment load. When the ice melts, these outwash deposits would be lowered to the land surface as isolated mounds or kames. Due to the curved surface of valley ice tongues, meltwater often flows over the surface of active and/or stagnant ice in the region along the valley wall, where the ice elevation is at its lowest. Outwash would be deposited in the region between the glacial ice and the valley wall. Upon melting of the ice, this outwash would lose its ice-side support and thus collapse downward in the direction of the valley. The upper surface of this kame terrace deposit would remain nearly flat and at a higher elevation than the valley floor, while the side facing the valley typically would contain collapse features and faults oriented in the direction of the ice support. Kame deltas result from ice-contact meltwater flowing from the front edge of the ice into a proglacial lake. As the meltwater slows, it deposits sediment as a deltaic deposit.

Proglacial lakes formed in glacial valleys throughout central New York and are identified by their associated silt and clay lacustrine deposits, which are often varved, and/or by preserved tributary deltas that now "hang" topographically above valley bottoms. Proglacial lake formation results when drainage within the valley is blocked, either by active or stagnant ice, by glacial deposits including moraines and valley plugs, or by higher elevations and bedrock knolls down-valley that result from erosion and over-deepening of the valley upstream by glacial erosion. These lakes may drain catastrophically as the ice blocking their drainage retreats or as the down-valley outlet is eroded, producing hanging shorelines and terraces. The lakes may also be infilled with coarser-grained outwash sediments.

FIELD TRIP AREA

The field trip area is located in Cortland, Broome, and Tompkins Counties, New York, within the valleys of the West Branch Tioughnioga River from Tully to Cortland, Fall Creek from Cortland to Ithaca, and the Main Stem Tioughnioga River from Cortland to Whitney Point (Fig. 2). This area is located in the Eastern

Appalachian Plateau region of Central New York, also known as the Allegheny Plateau. The Appalachian Plateau is an upland region with elevations ranging between 450 and 670 m. Valleys within the plateau have been deepened an average of about 325 m, and widened, with major valleys ranging from 300 up to 2500 m wide, through multiple glacial episodes. Quaternary sediments within these plateau valleys are the result of past glaciations and consist of outwash sands and gravels or valley train landforms. Upland tributary valleys contain deposits of alluvium that overlie till and bedrock (Randall 2001).

The northern rim of the Appalachian Plateau and the Erie-Ontario Plain, with its associated northward drainage, is located just north of the field trip region. The Finger Lakes and associated drainage are located to the west, and although they are part of the Appalachian Plateau, their present-day drainage is also to the north. The southward draining Chenango Valley is located to the east of the Tioughnioga Valley. Confluence between the Chenango and Tioughnioga valleys occurs at Chenango Forks, and the Chenango Valley continues southward (Coates 1963).
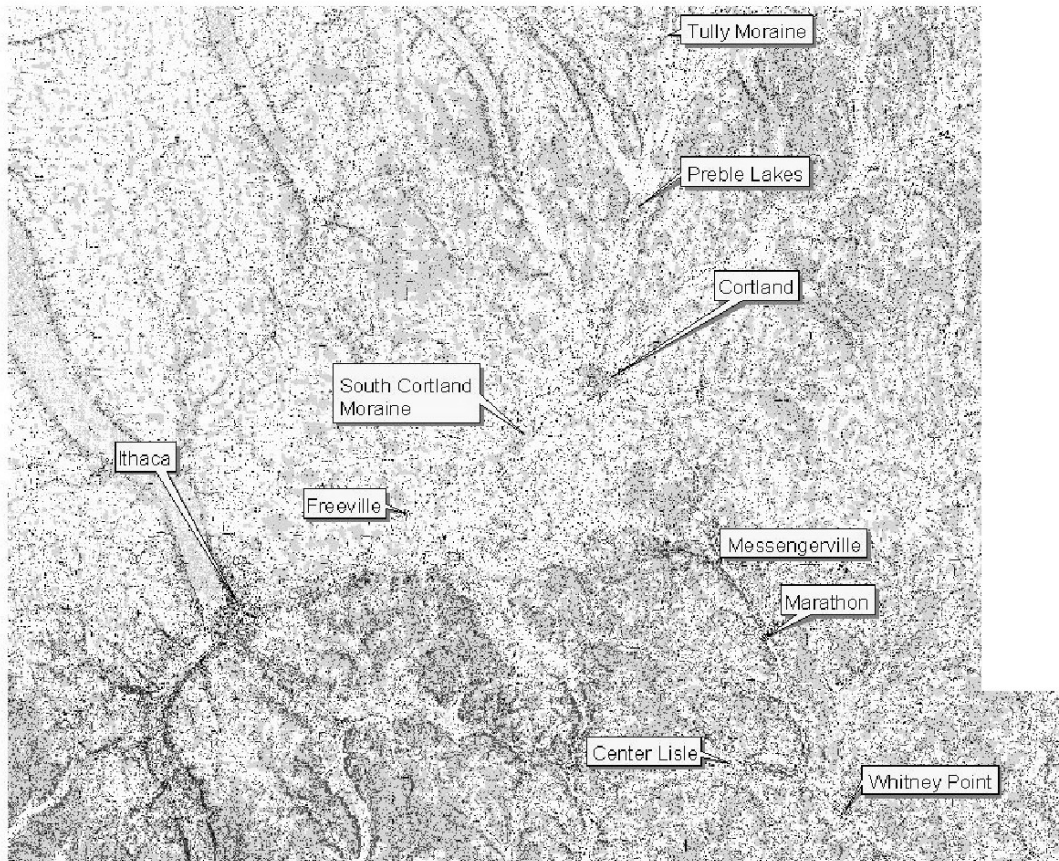
FIGURE 2—Map of field trip area; West Branch, Main Stem, and Fall Creek Valley locations in Cortland, Broome, and Tompkins County, Central New York.

West Branch and Main Stem Tioughnioga Valleys

The West Branch Tioughnioga Valley, located between Tully and Cortland NY, is a through valley that extends southward from the Tully Lakes region. This valley extends northward beyond the Tully Lakes (Stop 1) to the Tully (Valley Heads) moraine and is known as the Tully Trough. To the north of the moraine, the through valley continues northward towards Syracuse as the Onondaga Trough and contains the northward flowing Onondaga Creek. The West Branch Tioughnioga drains Tully Lake and flows south through the Preble area (Stop 2; Preble Swamp and Preble Lakes Goodale, Upper Little York, and Lower Little York), to Homer and Cortland. Major tributaries to the West Branch Tioughnioga include the Otisco Valley, from which there is no

present-day drainage, Cold Brook, and Factory Brook. To the north of the city of Cortland, the West Branch turns to the east and is joined by Otter and Dry Creeks before confluence with the East Branch Tioughnioga.

Upon confluence of the East and West Banches, the Main Stem Tioughnioga resumes a south-southeast course. To the southeast of Cortland (east of Polkville), the Trout Brook Valley and its west-flowing drainage merges with the Main Stem Tioughnioga. Tributaries to the Main Stem between Trout Brook and the next major confluence at Whitney Point, where the Tioughnioga and Otselic rivers join, include Hoxie Gorge Creek, Gridley Creek, Hunts Creek, Jennings Creek, and Dudley Creek, as well as some small unnamed tributaries. To the south of the Otselic-Tioughnioga confluence, tributaries to the Tioughnioga include Bull Creek, Halfway Brook, and several smaller unnamed tributaries. At Chenango Forks, the Tioughnioga converges with the Chenango River, flowing south to the Susquehanna.

The Main Stem Tioughnioga Valley retains a glacial U-shape, with steep valley walls and a broad, flat valley bottom, although the width of the valley decreases greatly south of the East and West Branch confluence. In the Cortland region, a juncture of four valleys occurs (West Branch Tioughnioga, East Branch Tioughnioga, Main Stem Tioughnioga, and Otter/Dry Creek valleys), producing an extremely large overall valley width of around 4500 m at the widest point. At the Trout Brook – Main Stem confluence, the Tioughnioga Valley is around 1100 m across. The valley width averages around 500 m at Blodgett Mills, and width continues to decrease southward, reaching around 200 m at Messengerville. From Messengerville onward, valley width fluctuates, with wider areas of around 400 to 600 m (1000 m at Whitney Point) and narrower areas of around 200 and 300 m (Fig. 2). According to Coates (1981), this fluctuation of valley width south of Cortland resulted from the Tioughnioga Valley having been a meltwater conduit or sluiceway during glacial retreat.

Bedrock in the Tully, NY, region consists of the Middle Devonian Hamilton Group (Kappel and Miller 2003). Bedrock in the valley near Marathon is of the Ithaca Formation, whereas bedrock in the uplands consists of Enfield Formation siltstone (HydroSource Associates Inc. 1999). In the region of Whitney Point, bedrock consists of Late Devonian Genessee Formation and Sonyea Group sandstones and shales (Rickard and Fisher 1970). Sandstone and shale of the Genesee Formation is found in the valleys at Cortland and within the Fall Creek Basin (Karig and Elkins 1986). Bedrock within the West Branch and Main Stem Tioughnioga valleys is buried by thick surficial deposits of glacial and alluvial origin. Deposits include glacial outwash sands and gravels, deltaic deposits of medium and fine sand, glaciolacustrine silts and clays, till, and alluvial silt, sand, and gravel (Muller and Cadwell 1986).
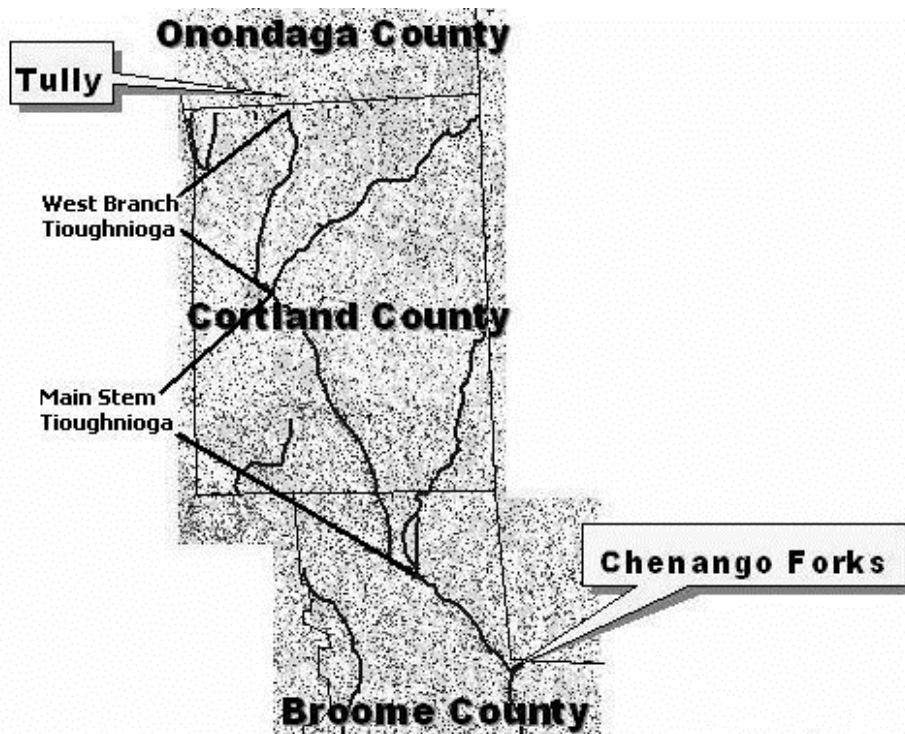
Fall Creek Valley

The Fall Creek Valley is located in the Finger Lakes Region of Central New York west of Cortland and extends from the western side of the South Cortland / Fish Hatchery (Valley Heads) Moraine (Stops 3 and 4) to Cayuga Lake at Ithaca. Tributaries to the valley include Lake Como Outlet, Webster Brook, Mud Creek, Mill Creek, Virgil Creek, and several smaller tributaries. The creek flows in an east–southeast direction. The valley turns northward at Ithaca, and Fall Creek flows into Cayuga Lake. Due to its east-west orientation, the valley was perpendicular to glacial ice flow and thus filled with glacial deposits during both glacial advance and retreat (Stop 5). Following glacial retreat, Fall Creek excavated its valley, with incision to bedrock in the lower reaches beyond the Cornell Arboretum. Upstream of this region, surface deposits consist of alluvium, and terraces record the incision of Fall Creek into its channel.

GLACIAL HISTORY OF THE WEST BRANCH AND MAIN STEM TIOUGHNIOGA VALLEY

The West Branch and Main Stem Tioughnioga valleys (Fig. 3) are oriented parallel to the direction of ice flow, and as a result, were widened and deepened, forming U-shaped glacial valleys. Retreat methods were defined using Fleisher's (1991, 1993) through and non-through valley glacial assemblage model. Fleisher proposes that kame moraines with associated valley train, deltaic gravel terraces, lacustrine plains, kame fields, and dead-ice sinks are associated with backwasting (active retreat) in through valleys, whereas kames, kame fields, and discontinuous gravel plain remnants with dead-ice sinks and eskers are characteristic of downwasting in non-through valleys. In the case of downwasting, dead-ice sinks are bounded up- and down-valley by landforms such as terraces, kame moraines and kame fields, and valley trains (Fleisher 1991). Both

the West Branch and Main Stem Tioughnioga valleys are identified as through valleys, in which retreat occurred through backwasting with some episodes of detached margins and downwasting.

A glacial plug deposit of sand and gravel is found in the Tioghnioga Valley at Chenango Valley State Park just upstream of Chenango Forks, where the Tioughnioga River flows into the Chenango River. This deposit formed as a block of ice detached from the actively retreating ice margin and meltwater deposition increased (Cadwell 1972). As the Tioughnioga Valley was deglaciated, flow eroded through the valley plug deposits (Cadwell 1972). Between Chenango Forks and Whitney Point ice was actively retreating at this time. Outwash was deposited on the floor of the valley, and kame terraces and deltas were deposited along the valley walls. A kame terrace/delta is found where Halfway Brook enters the valley from the east, and a kame terrace is found along the western valley wall at Istaka. Well logs collected from the NY State Department of Environmental Conservation, Department of Transportation, county health departments, and the U.S. Geological Survey show a buried deposit of clay and silt, which we interpret as a glaciolacustrine deposit, beneath outwash sediments at and south of Whitney Point. The southward extent of this lake is not known, due to a lack of subsurface information in that region. The lake may have formed by impoundment of meltwater between the retreating ice margin and valley plug deposits downvalley. A large kame terrace and delta deposit is found along the eastern valley wall to the north of Whitney Point, within the Town of Lisle. Sediments of this deposit are recorded in well logs within the terrace and exposed at the Taylor Ready Mix sand and gravel pit. Kame delta structure is preserved in a deposit along Watts Road.

FIGURE 3—West Branch and Main Stem Tioughnioga valleys oriented parallel to ice flow direction.

Stagnant ice within Dudley Creek Valley west of Lisle led to the deposition of associated non-through valley deposits (Fig. 4). Kame terraces are found along both sides of the valley wall along Dudley Creek near the Dudley Creek–Tioughnioga junction. The terrace deposits on the north side of the valley are exposed along Dudley Creek behind the Town of Lisle firehouse (Stop 7) and above the valley in a road cut along Owen Hill Road (Stop 8). On the southern side of the valley, kame terrace deposits are exposed in the valley across the road from the firehouse and above the valley in gravel pits along Smith Hill Road. A large subglacial channel deposit known as the Lisle Esker (Stop 6) is found between Richford and Lisle. In some areas, the esker is

attached to the valley wall, whereas it is detached in other locations. It is thought to have formed when a distributary tongue of ice extended through a gap in the Owego Creek Valley to the west and stagnated during deglaciation, when it was cut off from the main ice tongue to the west (von Engeln 1961).
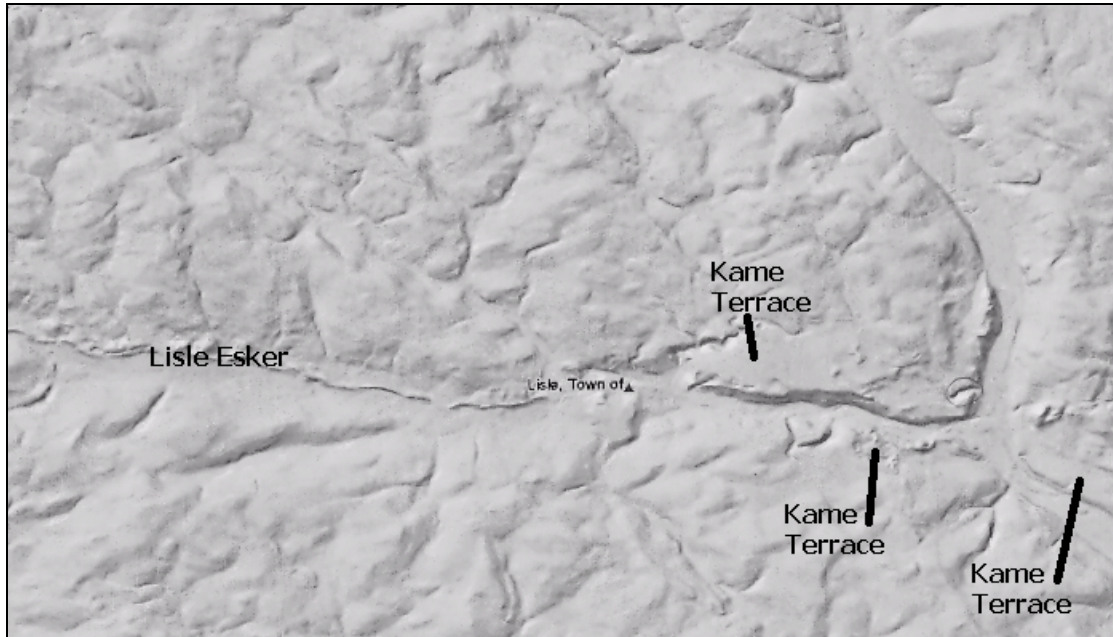
FIGURE 4—Dudley Creek Valley stagnant ice landforms.

Ice retreat from Whitney Point to Marathon was characterized by active ice retreat. Stratigraphy in the Tioughnioga Valley north of Lisle, as interpreted from well logs, includes glacial till overlying outwash and/or bedrock that is overlain by outwash. Surface deposits in the valley are composed of alluvium. Kame terraces are present along the valley walls east of Killawog and both east and west of Marathon (Optional Stop). The lack of glaciolacustrine deposits in this area shows that drainage to the south was not blocked by glacial deposits downvalley.

A preglacial drainage divide probably was located north of Marathon, in the region of Messengerville, based on the orientation of tributary valleys. Flow to the south of this divide was south toward Marathon, while flow to the north was toward Cortland, and down the Otter-Dry Creek valley southwest toward the Fall Creek Valley (Fig. 5) (Miller 2000). The dendritic pattern formed by valleys in the region of Cortland shows that flow from the Main Stem Tioughnioga and Trout Brook north of Blodgett Mills was toward Cortland. Bedrock in the Main Stem Tioughnioga Valley from Messengerville to Cortland slopes northward, providing additional support for preglacial flow in this direction (Fig. 6). As a result of glacial processes, drainage at Cortland was diverted southward, down the Main Stem Tioughnioga Valley and across the former drainage divide, forming the present-day drainage patterns. The narrowing of the through valley southeast of Cortland is a result of drainage diversion from a large valley system at Cortland, where valleys and their associated drainage converged, to the smaller Main Stem Tioughnioga Valley, where preglacial drainage was to the north via a small stream valley. The present day Otter Creek Valley is underfit, in which a small stream occupies a much larger valley (Miller et al. 1998).

Map labels:

76°15' 76°10' 76°5'

1500

Tully

1000

ONONDAGA CO
CORTLAND CO

1500

42°45'

1500

CAYUGA CO

1500

1500

City of
Cortland

1500

TOMPKINS CO

1500

Tioughnioga River Flow

1500

Preglacial drainage
flowed southwestward
through Fall Creek valley

1500

42°30'

1500

1500
Messengerville

Tioughnioga River
diverted southward
through preglacial
divide

CONTOUR INTERVAL 500 FEET
DATUM IS SEA LEVEL

Marathon

Base from U.S. Geological Survey,
1954, 1:250,000

Geology modified from
Miller and others, 1998

B. PRE- AND POSTGLACIAL PERIOD

PRESENT-DAY STREAM

AQUIFER BOUNDARY

PREGLACIAL-DRAINAGE
COURSES- arrow indicates
direction of flow

FIGURE 5—Pre- and Post-glacial drainage in the Cortland area (Miller 2000).
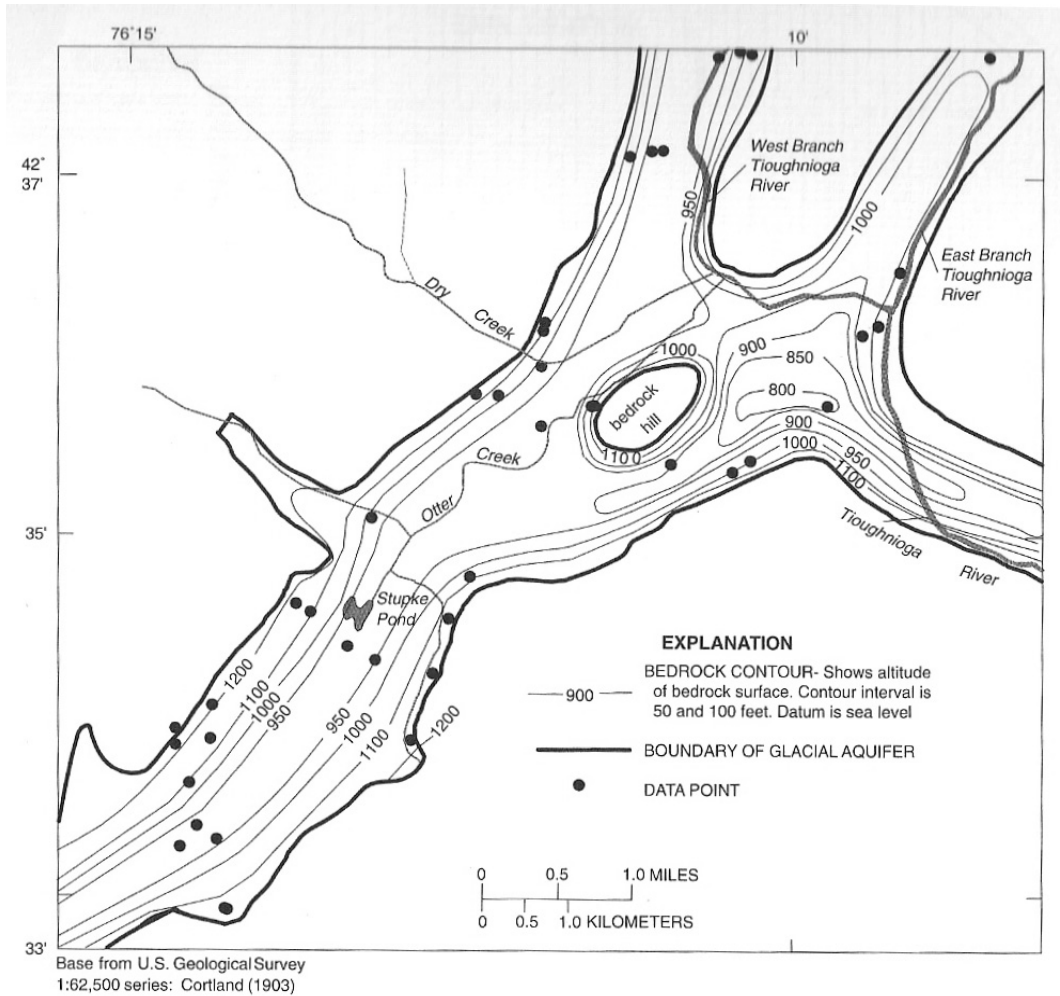
FIGURE 6—Map of depth to bedrock, Cortland area (Miller et al. 1998).

Glacial retreat and ice in the Cortland region led to the deposition of large kame deposits that formed as meltwater from upland regions flowed into the valley. A large kame terrace is found on the southern valley wall at Cortland, with elevations reaching up to 385 m. Terrace deposits are also found along the northwest valley wall, between the West Branch and Dry/Otter Creek valleys. Mapping of glacial deposits based on soil parent materials suggests that the terraces might have been deposited as a moraine (Fig. 7). Glacial stratigraphy as interpreted from well logs and subsurface information shows that outwash and possible buried kame deposits overlie bedrock. These outwash deposits are overlain by a thick layer of glaciolacustrine silts, clays, and very fine sands that were deposited in a proglacial lake. Surficial outwash deposits overlie the glaciolacustrine deposits. Uplands and the uppermost reaches of the valleys contain deposits of till over bedrock. The South Cortland moraine is found to the southwest, in the Dry/Otter Creek Valley. In this region, bedrock is overlain by till, outwash (kame) deposits, and kame moraine (Miller et al. 1998). A kame deposit is found in the Otter/Dry Creek valley along Route 28 just north of Webb Road. Glaciolacustrine deposits in the valley extend south of Cortland to Blodgett Mills. Ice retreat is interpreted as having been active, following Fleisher's (1991) through valley assemblages.
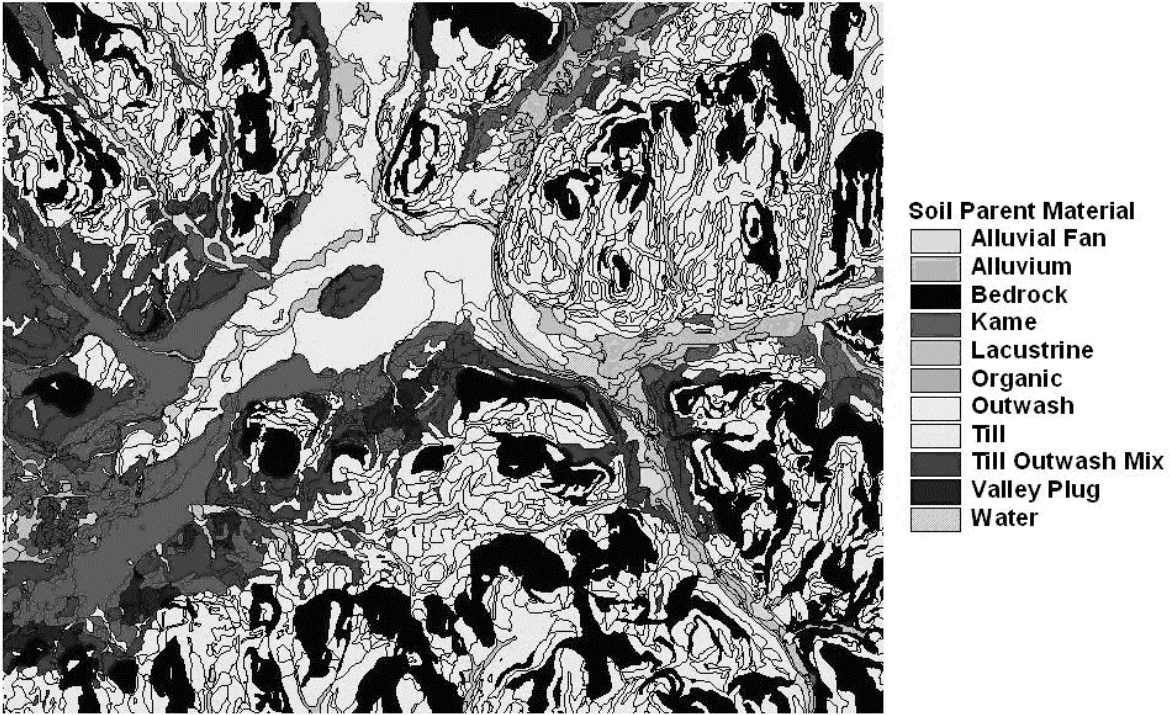
FIGURE 7—Surface deposits based on soil parent materials (base map from USDA, NRCS, SSURGO for Cortland County).

The West Branch Tioughnioga flows southward through the Homer-Preble Valley, north of Cortland. Glacial retreat through this valley led to a stratigraphy consisting of till and/or outwash overlying bedrock and overlain by glaciolacustrine deposits, with some silt and clay units over 30 m thick. In some areas, glaciolacustrine deposits directly overlie bedrock. Glaciolacustrine deposits are buried beneath deltaic sands and outwash composed of sand and gravel. At the Factory Brook junction in Homer, glacial till at the surface may be an indication that a glacial readvance down the Factory Brook Valley reached as far as the West Branch Tioughnioga–Factory Brook valley confluence. Large kame terraces are found within the Factory Brook region north through Cortland and the West Branch Tioughnioga River Valley south of the Skaneateles Valley Heads Moraine (Fig. 8). These most likely were produced as upland meltwaters flowed into the valley, depositing outwash between the ice and the valley walls. Glaciolacustrine deposits extend northward through the Preble Lakes area to the Tully Lakes and Tully (Valley Heads) Moraine (Buller 1978; Kappel and Miller 2003).

Bedrock in the region slopes northward toward the moraine, and an actively retreating ice tongue was present within this valley (Buller 1978). The Preble Lakes just south of the Otisco Valley confluence, and the Preble swamp just north of the confluence, are large kettles that may be interpreted as dead ice sinks (Fig. 9). Valley train deposits are found both up and down valley from the lakes and swamp. We interpret that the Preble Lakes and swamp formed as part of the retreating ice margin detached, resulting in large, stagnant ice blocks within the valley that were buried beneath outwash deposits. Upon melting of the ice, overlying sediments collapsed downward, forming depressions in the valley floor that filled with water and resulted in kettle lakes. The Preble swamp may have been a shallow lake that was gradually infilled, eventually producing a swamp. A kame terrace is found along the eastern valley wall just north of the Preble swamp.

FIGURE 8—Factory Brook Kame Terrace Deposits

FIGURE 9—Preble Lakes and Swamp

The Tully Lakes and Tully Moraine (Stop 1) are found north of the Preble Swamp (Fig. 10). The Tully Moraine formed during the Valley Heads episode as a result of ice stagnation. The uppermost surface of the moraine is at around 365 m, around 200 m above the valley floor. The northernmost edge of the moraine in the Onondaga Trough has an elevation of around 210 m. From this region north, the elevation of the valley floor continues to decrease, reaching around 170 m in Lafayette and around 90 m at Onondaga Lake in Syracuse. To the south of the moraine, elevations remain at around 365 m, with the Tully Lakes occupying lower elevation basins. Kames are found in association with the kettles, and valley train deposits extend downvalley from the lakes. Valley-floor elevations are higher to the south of the moraine than those to the north due to glacial readvance, due to the scouring and deepening of the Onondaga Trough and the deposition of the Tully Moraine (Kappel and Miller 2003). As ice stagnated at the Valley Heads moraine within the Tioughnioga and associated Otisco, Skaneateles, and Fall Creek valleys, drainage to the south dammed between the moraine and the drainage divide in the Main Stem Tioughnioga Valley. An outlet to the lake was located at Blodgett Mills, and drainage diversion occurred at Cortland, with flow to the southeast down the Main Stem Tioughnioga, rather than down the Otter Creek Valley (Miller et al. 1998). As the Tioughnioga River established its channel, glacial and alluvial deposits and drainage into the valley controlled the location of the river. This is evident at the confluence of the Otisco Valley at Preble and the Factory Brook Valley at Homer, where large glaciofluvial fan deposits were produced as meltwater from the retreating ice margins in the Otisco and Skaneateles Lakes troughs flowed into the West Branch Tioughnioga Valley. In these areas, flow in the West Branch Tioughnioga River is confined along the eastern valley wall.
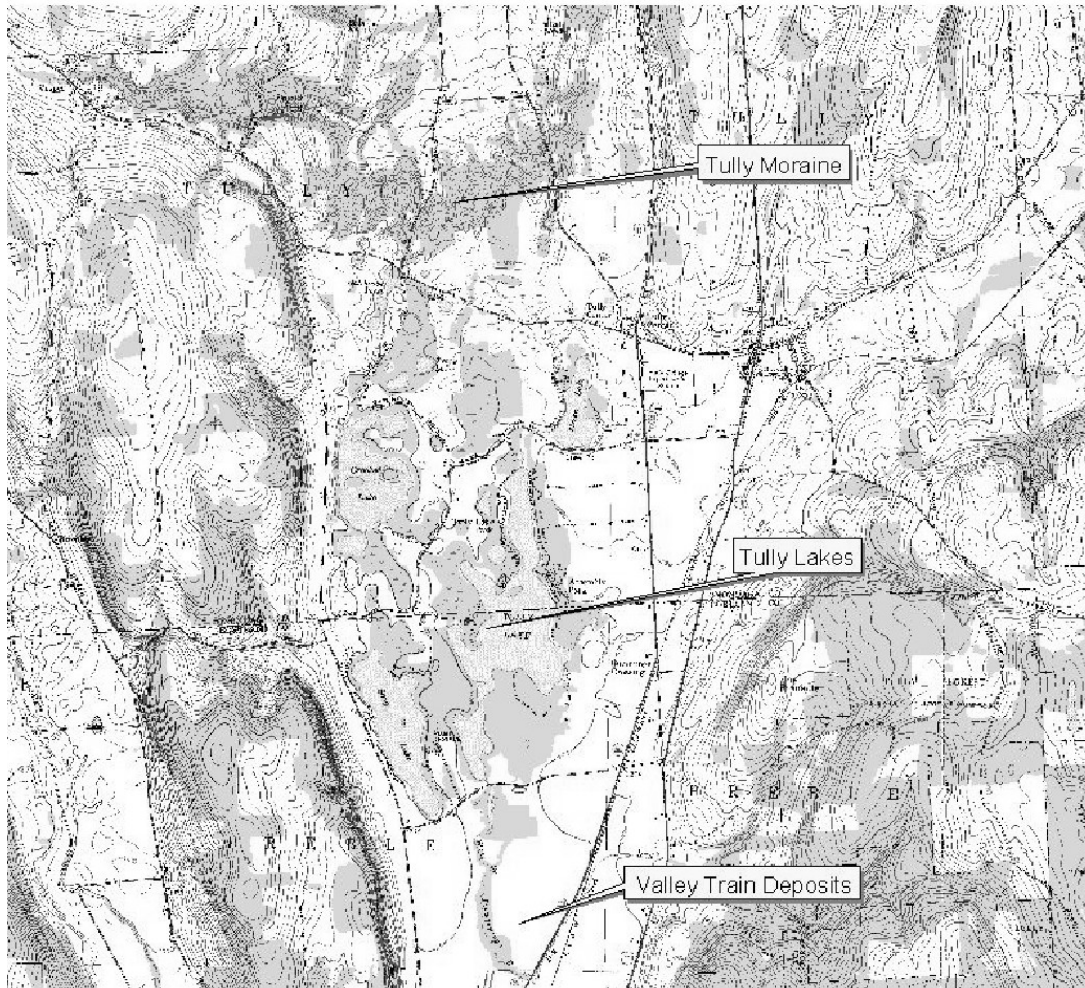
FIGURE 10—Tully Lakes and Moraine Region.

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GLACIAL HISTORY OF THE FALL CREEK VALLEY

Drumlins located in northwestern central New York to the north of the Finger Lakes indicate that ice, possibly laden with subglacial meltwater, flowed fast across that region during late stages of the LGM (Ridky and Bindschadler 1990; Shaw and Gilbert 1990). Erosion resulted in the low elevation of the Ontario Lowlands region. The Finger Lakes were produced over successive glacial episodes as ice flowed into river valleys oriented parallel to the direction of ice flow. Subglacial erosion widened and deepened these valleys, producing overdeepening of the Finger Lakes troughs (Bloom 1986), with erosion in Seneca and Cayuga Lakes below sea level (Mullins and Hinchey 1989). The Finger Lakes gorges and hanging valleys were produced as glacial troughs were widened and deepened (Bloom 1986). Much of the overdeepening was associated with the readvance to the Valley Heads Moraine position (Mullins and Hinchey 1989), and the modern gorges and hanging valleys were rapidly incised as ice retreated back to the north and pro-glacial lakes dropped to near their modern level, in as little as 500 years (Knuepfer and Lowenstein 1998; Knuepfer and Hensler 2000).

Successive periods of glaciation in the region resulted in expansion of the Cayuga Trough far enough to intersect with the lower reaches of the Fall Creek Valley. A change in the valley floor material from alluvium to bedrock in the region of the Cornell Arboretum reflects the southward expansion of the trough and glacial erosion of the lower Fall Creek Valley. During the LGM, Fall Creek Valley, oriented perpendicular to ice flow direction, was infilled with glacial debris, burying earlier Wisconsinan glaciolacustrine sediment and till (Bloom 1986). Glacial ice from the Cayuga Trough flowed up the Fall Creek Valley, resulting in the erosion of unconsolidated deposits and bedrock and the deposition of lodgement till in the valley (Cline and Bloom 1965). As ice retreated from the valley, the retreating ice tongue and glacial deposits blocked drainage flowing from the Willseyville area, resulting in the formation of Freeville-Dryden Lake. A well sorted, stratified layer of glaciolacustrine sediments composed of sand and silt was deposited within this lake and overlies Fall Creek Valley lodgement till (Cline and Bloom 1965). A period of ice readvance in the Fall Creek basin is recorded by a deposit of glacial lodgement till and supraglacial meltout till that overlies moraine, outwash, and glaciolacustrine deposits. These deposits are overlain by fill (Miller et al. 1998). Glacial Lake Ithaca formed as Freeville-Dryden and other proglacial lakes in the region drained to a lower elevation and merged together. Glaciolacustrine silts and clays were deposited within this lake (Cline and Bloom 1965). As ice retreated and Freeville-Dryden Lake drained from the Fall Creek Valley, Fall Creek began to re-excavate its channel, and a fan-delta formed where Fall Creek flowed into Glacial Lake Ithaca. Following ice retreat and drainage of Glacial Lake Ithaca, hanging valleys, terraces, and deltas formed as Fall Creek re-excavated its valley in response to base-level lowering of the Cayuga trough (Bloom 1986).

The Varna high banks exposures (Stop 5; Fig. 11) provide a record of glaciation within the Fall Creek Valley. Basal deposits consist of laminated glaciolacustrine silts and clays, sometimes exposed at the base of the bluff. Elsewhere in the valley these varved sediments, which overlie strongly weathered till (Bloom 1986), have yielded radiocarbon ages in excess of 35,000 ^{14}C yr B.P. (Muller and Cadwell 1986). The high banks expose 30 m of poorly sorted till, composed of sand and gravel of local origin with larger crystalline clasts, above these lacustrine sediments. The till contains interbeds of stratified sand and gravel that were deposited as water from the ice-free headwaters of Fall Creek flowed south and was blocked by the ice tongue and its associated deposits (Bloom 1986). This till is overlain by a blue-gray matrix-dominated lodgement till. The majority of clasts are local shales and sandstones, but a greater number of crystalline rocks are present in this upper till than in the lower till. The lowermost part of the upper till contains clasts of limestone and dolostone, while the uppermost 3-6 m lacks carbonate clasts and shows evidence of oxidation and water reworking. The blue-gray lodgement till is overlain by a 3-m-thick, well-sorted, stratified glaciolacustrine deposit composed of silts, sands, and some clays deposited within Freeville-Dryden Lake (Cline and Bloom 1965).

The South Cortland moraine (Fig. 12) is an ablation moraine with hummocky topography and ice-contact meltwater deposits. Kames, kettles and eskers are found at the Malloryville esker, bog, and wetland complex (Stop 4; Fig. 13) and the Lime Hollow Nature Center (Stop 3), with kettle holes occupied by ponds and wetlands. Gravel pits contain subrounded to rounded clasts, some of which are faceted. The majority of clasts are composed of local rocks, but exotics, including granite, gneiss, and red sandstone, are also found. The deposits are massive and matrix dominated, with the majority of clasts ranging between sand and gravel size.
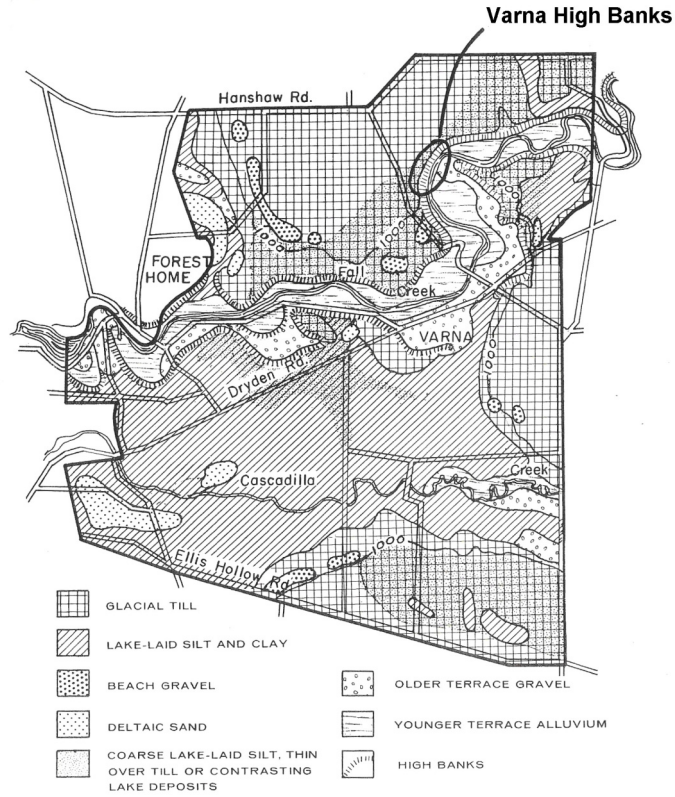
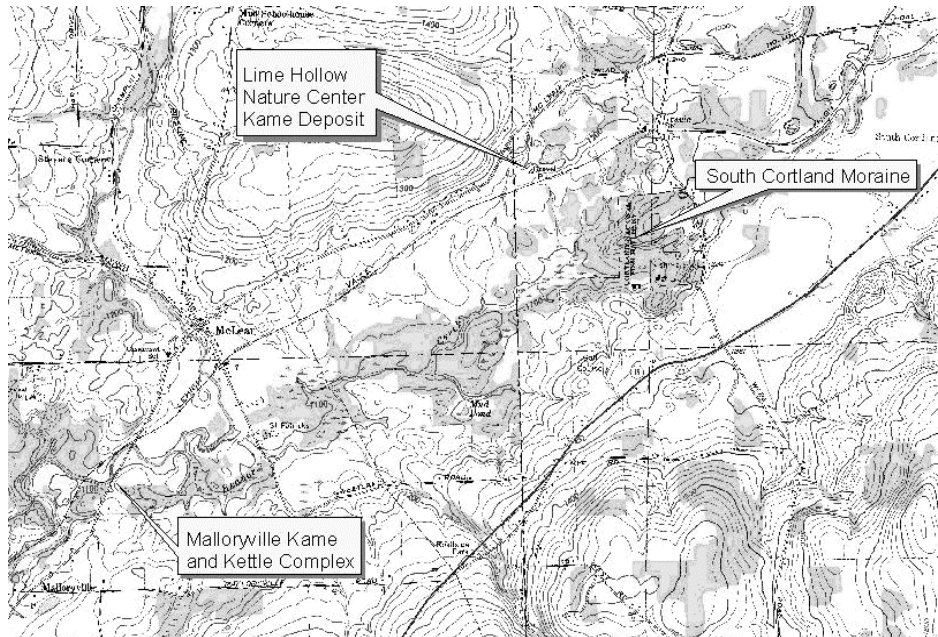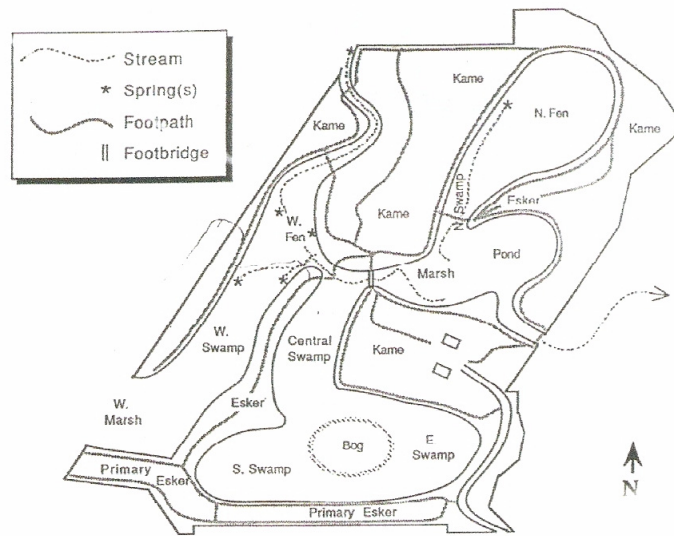**FIGURE 11**—Fall Creek glacial deposits and Varna High Banks area (Cline and Bloom 1965).

FIGURE 12—South Cortland Moraine and kame, kettle, and esker locations.

FIGURE 13—Malloryville Kame and Kettle Complex: Glacial landform locations (McNamara et al. 1992).

REFERENCES CITED

Bloom, A.L., 1986, Geomorphology of the Cayuga Lake Basin, in Field Trip Guide for the 58th Annual Meeting of the New York State Geological Association: Cornell University, Ithaca, NY, ,p.261-279.

Buller, W., 1978, Hydrologic appraisal of the water resources of the Homer-Preble Valley, New York: U.S. Geological Survey Water Resources Investigation WRI 78-94, 31 p.

Cadwell, D.H., 1972, Late Wisconsinan deglaciation chronology of the Chenango River Valley and vicinity, New York [PhD Thesis]: Binghamton, SUNY Binghamton, 102 p.

Cline, M.G., and Bloom, A.L., 1965, Soil survey of Cornell University property and adjacent areas: Cornell Miscellaneous Bulletin 68, New York State College of Agriculture,.31 p,

Coates, D.R., 1963, General geology of South-Central New York, in Coates, D.R., ed., Field Trip Guide for the 35th Annual Meeting of the New York State Geological Association : State University of New York at Binghamton, Binghamton, NY, p.21-58.

Coates, D. R., 1981, Geomorphology of south-central New York, in Enos, P., ed., Field Trip Guide for the 53rd Annual Meeting of the New York State Geological Association : State University of New York at Binghamton, Binghamton, NY, p. 171-200.

Fleisher, P.J., 1991. Active and stagnant ice retreat: deglaciation of Central New York in Rayne, T.W., Bailey, Da.G., and Tewksbury, B.J., eds., Field Trip Guide for the 63rd Annual Meeting of the New York State Geological Association: Hamilton College, Clinton, NY, p.307-371.

Fleisher, P. J., 1993, Pleistocene sediment sources, debris transport mechanisms, and depositional environments: a Bering Glacier model applied to northeastern Appalachian Plateau deglaciation, central New York: Geomorphology, v. 6, p. 331-355.

HydroSource Associates, Inc., 1999, Groundwater Exploration Phase I Report: Village of Marathon, New York: unpublished consultants' report..

Kappel, W.M., and Miller, T.S., 2003, Hydrogeology of the Tully trough, southern Onondaga County, and northern Cortland County, New York: U.S. Geological Survey Water-Resources Investigations Report 03-4112, 16 p.

Karig, D.E., and Elkins, L.T., 1986, Geological summary of the Cayuga Region in Field Trip Guide for the 70th Annual Meeting of the New York Geological Association: Cornell University, Ithaca, NY, , p.23-42.

Knuepfer, PLK., and Hensler, S.M., 2000, Proglacial lakes and development of stream gorges in the Cayuga and Seneca Valleys, in McKinney, D.B., ed., Field Trip Guide for the 72nd Annual Meeting of the New York State Geological Association:.Hobart and William Smith Colleges, Geneva, NY, p. 75-90.

Knuepfer, P.L.K., and Lowenstein, T.K., 1998, Finger Lakes gorges revisited, in Naslund, H.R., ed., Field Trip Guide for the 70th Annual Meeting of the New York State Geological Association: New York State Geological Association, p. 23-42.

McNamara, J.P., Siegel, D.I., Glaser, P.H. and Beck, R.M., 1992, Hydrogeologic controls on peatland development in the Malloryville Wetland, New York (USA): Journal of Hydrology, v. 140, p. 279-296.

Miller, T.S., 2000, Simulation of ground-water flow in an unconfined sand and gravel aquifer at Marathon, Cortland County, New York: U.S. Geological Survey Water-Resources Investigation Report 00-4026, 24 p.

Miller, T.S., Sherwood, D.A., Jeffers, P.M., and Mueller, N., 1998, Hydrology, water-quality, and simulation of ground-water flow in a glacial aquifer system, Cortland County: U.S. Geological Survey Water Resources Investigations Report 96-4255, 84 p.

Muller, E.H., and Cadwell, D.H., 1986, Surficial geologic map of New York; Finger Lakes sheet: New York State Museum, Geological Survey, Map and Chart Series #40, scale 1:250,000.

Muller, E.H., and Calkin, P.E., 1993, Timing of Pleistocene glacial events in New York State: Canadian Journal of Earth Sciences, v. 30, p.1829-1845.

Mullins, H.T., and Hinchey, E.J., 1989, Erosion and infill of New York Finger Lakes: Implications for Laurentide ice sheet deglaciation: Geology, v. 17, p. 622-625.

Randall, A.D., 2001, Hydrogeologic framework of stratified-drift aquifers in the glaciated Northeastern United States: U.S. Geological Survey Professional Paper 1415-B, 179 p.

Rickard, L.V., and Fisher, D.W., 1970, Geologic map of New York: Finger Lakes sheet: New York State Museum and Science Service Map and Chart Series No. 15, scale 1:250,000.

Ridky, R.W., and Bindschadler, R.A., 1990, Reconstruction and dynamics of the Late Wisconsin "Ontario" ice dome in the Finger Lakes region, New York: Geological Society of America Bulletin, v. 102, p. 1055-1064.

Shaw, J., and Gilbert, R., 1990, Evidence for large-scale subglacial meltwater flood events in southern Ontario and northern New York State: Geology, v. 18, p. 1169-1172.

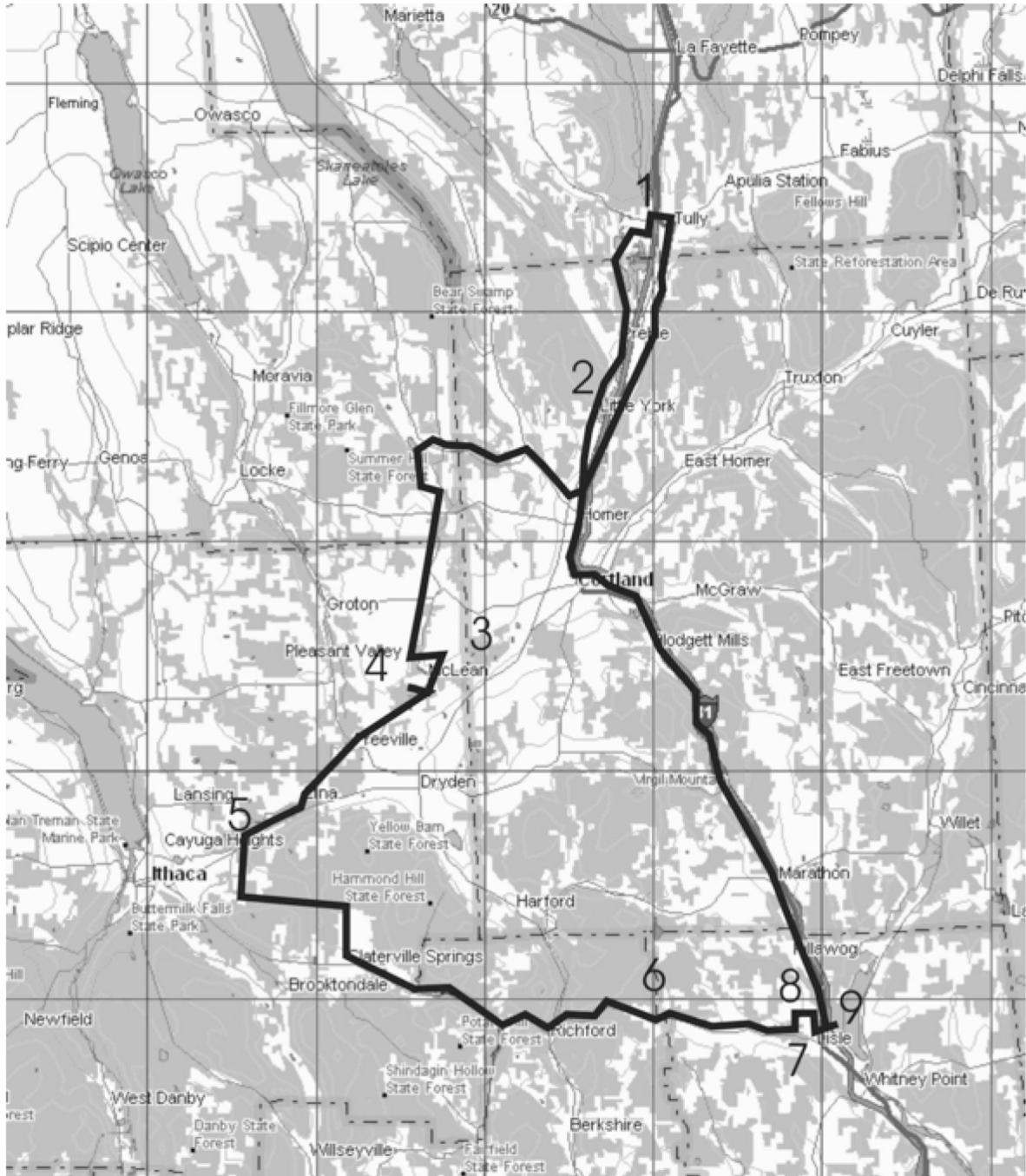von Engeln, O.D., 1961, The Finger Lakes region: Its origin and Nature: Cornell University Press, Ithaca, NY, 156 p.

FIGURE 14—Map of field trip stops

ROAD LOG FOR TRIP A-2

PLEISTOCENE GLACIATION OF THE TIOUGHNIOGA RIVER AND FALL CREEK VALLEYS

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	SUNY Cortland Gerhart Drive Parking Lot, Bowers Hall. Departure
0.1	0.1	Exit SUNY Cortland Parking Lot: Turn Left onto Graham Avenue, proceed to Groton Ave
0.2	0.1	Turn Right onto Groton Ave (route 222), proceed on Groton Ave to North Main Street.
0.6	0.4	Turn left onto North Main Street (route 11)
1.2	0.6	Bear right onto Homer Ave, route 11. Continue on route 11 through Homer to Preble Lakes region. Note the gravel pits in outwash deposits along route 11 in Homer.
12.0	10.8	At the Preble Swamp area, note the gravel exposure in the kame terrace to the east. The Preble swamp is a kettle, possibly a dead ice sink, formed as an ice block broke off from the actively retreating ice margin and was buried beneath glacial outwash. As the ice melted, outwash deposits collapsed downward, forming a kettle at the surface. To the east of the kettle, a kame terrace formed as glacial meltwater deposited sediment between the ice block and the valley wall. As the ice melted, the sediments collapsed and a kame terrace resulted. An exposure in this terrace is located just north of the swamp. Continue North on route 11 to Tully Lakes region.
15.3	3.3	Turn left onto route 80/11 at Tully
15.9	0.6	Continue straight on route 80 and go under I-81 bridge.
17.4	0.1	Turn left onto Lake Road
18.2	0.8	STOP 1 at Green Lake, parking area on left side of road.

STOP 1. TULLY LAKES KETTLES AND KAMES

Green Lakes and the rest of the Tully Lakes, ponds, and depressions, are kettles. These lakes formed as ice stagnated and blocks of ice were buried beneath outwash deposits. As ice melted, outwash deposits collapsed downward, forming depressions and kettles. Some of these depressions were below the water table and thus filled with water, forming kettle ponds and lakes. In areas where depressions were found on the ice surface, sand and gravel was deposited by glacial meltwater flowing over the ice surface. As the ice melted, these deposits were lowered to the valley floor, forming mounds known as kames. The gravel pit to the south of Green Lakes is excavated into such a kame deposit. Overall, the deposit is poorly sorted, with larger clasts of gravel in a sand matrix. Clasts are mostly of local origin, with some exotics, including granites, metamorphics, red sandstones, and limestones transported to the region from areas to the north. Gravels commonly contain calcium carbonate rims, and large post-Pleistocene conglomerates composed of sands, gravels, and cobbles, cemented with calcium carbonate, are found at the site. This Tully Lakes moraine complex commonly is considered part of the Valley Heads Moraines, although the cross-section of Kappel and Miller (2003) implies that this Valley Heads morphology may be developed on a relatively shallow till deposit that in turn overlies laminated silt and clay over a buried till deposit. Thus, whether this area represents the "true" Valley Heads Moraine—or a slightly younger recessional moraine formed during retreat from the Valley Heads position—is unclear.

Return to vehicles, turn left onto Lake Road.

18.9	0.7	Bear left, staying on Lake Road. Continue south and southeast to Song Lake Road.
20.2	1.3	At intersection of Lake Road and Song Lake Road, turn left, traveling south toward Homer. Note Song Lake, another kettle lake, to the left.
22.8	2.6	Song Lake Road merges onto route 281.
23.5	0.7	Cross through the Otsico-West Branch Tioughnioga Valley confluence. Glacial and alluvial fan deposits from the Otisco Valley control the location of the West Branch Tioughnioga River at this junction, with the river confined along the eastern valley wall. Valley Heads Moraine deposits and Otisco Lake (one of the Finger Lakes) are found up the Otisco Valley to the northwest. Continue south on route 281 to the Preble Lakes region.
25.2	1.7	Turn left onto Little York Lake road, toward Dwyer Memorial Park
25.3	0.1	Turn left onto New Road.
26.1	0.8	Proceed to end of New Road, pull over along Little York Lake. STOP 2 is here at Preble Lakes

STOP 2. PREBLE LAKES

These lakes are large kettles that occupy most of the valley width, and are possible dead-ice sink deposits. They were formed in association with the Preble Swamp to the north, as ice blocks detached from the retreating ice margin and were buried beneath glacial outwash deposits. As ice melted, outwash at the surface collapsed, forming depressions that intersect the water table. Kettle lakes and ponds of the Preble area resulted.

Return to vehicles, turn around and proceed back to route 281.

26.9	0.8	Turn left on route 281, continue south to Homer.
31.9	5.0	At Clinton Street (route 41) in Homer, turn right.
33.6	1.7	Proceed up the Factory Brook Valley along route 41. Note large active EZ Acres gravel pit to the left. This pit is in a kame terrace, deposited as glacial meltwater flowed between the ice surface and the valley wall. Although this is an excellent exposure through the entire kame terrace deposit, the height of the quarry wall at the time of this writing presupposes a safe visit. Continue north along route 41.
34.5	0.9	Turn left onto route 41A. Note exposure of bedrock as you enter Homer Gulf; this would be the equivalent of the valley-wall side of the kame terrace.
37.8	3.3	Turn left onto Cutler School Road. Note that road isn't marked, but cross-road to the north (right) is Atwood Road.
38.2	1.4	Turn right onto Creech Road.
38.9	0.7	Turn left onto Champlin Road.
46.1	7.2	Hamlet of McLean. Turn left at stop sign onto McLean Road.
47.8	1.7	Pull into parking area of Lime Hollow Trail Head for STOP 3.

STOP 3. LIME HOLLOW NATURE CENTER TRAILS

This area contains a pitted, hummocky kame and kettle landscape, with wetlands located in low lying depressions and kettles. Walk along the trail, heading southeast. Pass the trail sign, continuing straight. Cross the field, heading towards a mostly reclaimed gravel exposure. This is an exposure in a kame deposit. The unit overall is poorly sorted and matrix dominated, with grain sizes ranging from sand and silt to cobbles. Clasts include local sandstones and shales, as well as limestones, granites, red sandstones, and other exotics transported to the region from the north.

Return to vehicles, exit Lime Hollow trailhead, turning right (northeast) onto McLean Road.

48.5	0.7	Drive to Lime Hollow Visitor Center for LUNCH STOP. After lunch, reverse course back to McLean
50.0	1.5	McLean Road becomes Fall Creek road. Continue southwest.
51.5	1.5	Turn right onto W. Malloryville Road.
52.0	0.5	Turn right at sign on right into parking area for O.D. von Engeln Malloryville Preserve of the Nature Conservancy.

STOP 4. O.D.VON ENGELN PRESERVE AT MALLORYVILLE

This preserve is named for Otto D. von Engeln, late Professor of Geology at Cornell, who bequeathed much of the site in order to preserve the glacial geology and wetlands ecosystem; von Engeln was keenly interested in the glacial history of New York. The preserve is located within glacial deposits associated with deposition of the South Cortland (Valley Heads) Moraine. The area contained stagnant ice that resulted in the production of kame and kettle topography, as well as an esker. The site also contains abundant wetland types, including marshes, bogs, fens, and swamps, within depressions and low, poorly drained areas. Wetlands are in low-lying areas and depressions that formed as kettles at as buried ice blocks melted and overlying deposits collapsed downward. The esker was deposited as a subglacial meltwater channel beneath stagnant ice.

Return to vehicles, exit the Malloryville Reserve, and proceed northeast along Fall Creek Road.

52.5	0.5	Return to Fall Creek Road and turn right.
55.2	2.7	Freeville: Continue southwest on route 366. Fall Creek Road becomes Main Street.
58.7	3.5	Bear right onto route 13 / route 366 (Dryden Road).
60.0	1.3	Route 13 and route 366 split. Turn left onto route 366 (Dryden Road) toward Varna.
61.1	1.1	Turn right onto Monkey Run Road Proceed to end of road, parking where possible. Walk down to Fall Creek.

STOP 5. VARNA HIGH BANKS.

In this section of Fall Creek Valley, a large meander loop of Fall Creek cuts into blue-gray till and associated deposits. Basal deposits, exposed occasionally along the base of the exposures on either side of the creek, consist of laminated glaciolacustrine silts and clays. Elsewhere in the valley these varved sediments, which overlie strongly weathered till (Bloom 1986), have yielded radiocarbon ages in excess of 35,000 14C yr B.P. (Muller and Cadwell 1986). Lodgement till is typically exposed at the base of the southeastern bank of Fall Creek. This unit is highly compact, poorly sorted and matrix dominated, with clasts ranging from angular to rounded in shape. Some clasts contain glacial striations. Grain size ranges from clay to cobble, with an assortment of clast lithologies, including sandstone, granite, limestone, quartzite, and others. The lower layer of

this till are contains a higher percentage of gravel and is better sorted, with some stratification. Till in this bank has been eroded by Fall Creek and is overlain by alluvial deposits.

Along the northwestern side of Fall Creek, the Varna High Banks are exposed. The lower section of this deposit is poorly sorted and matrix dominated, with grain size ranging from clay to cobbles. Clasts are predominantly of local origin, with some exotics. Precipitation of calcium carbonate has led to the cementation of the lower unit. Some layering and stratification is present in the unit, with interbeds of unsorted material and sorted, layered sands and gravels. The lower unit of the Varna High Banks is overlain by a blue-gray glacial lodgement till. This is the same till as that exposed in the eastern bank. The uppermost section of the till shows evidence of being reworked by flowing water. Glaciolacustrine deposits from Freeville-Dryden Lake overlie the till. This unit has been eroded from the eastern bank. Grains consist primarily of silt and fine sand.

Return to vehicles. Return to route 366, turn right toward Varna and Ithaca.

61.7	0.6	Turn left onto Mt. Pleasant Road.
62.1	0.4	Turn right onto Turkey Hill Road.
63.7	1.6	Turn left onto Ellis Hollow Road.
68.6	4.9	Turn left onto route 79, head toward Lisle.
71.8	3.2	Caroline area. Note truncated spur ridges ahead that dominate the narrowing valley through which route 79 passes. The morphology of the valley-facing slopes is typical of east-west valleys in this part of New York: a steep north-facing slope on the south side of the valley, a gentler south-facing slope on the north side. Coates (1966) attributed this morphology to till shadows, formed because south-flowing ice dropped its basal load at the stress reduction caused by the north valley wall and had increased erosional capacity with increased basal stress at the south valley wall.
78.1	6.3	Richford. This is the through valley that connects Owego Creek with Dryden and, ultimately, the Fall Creek Valley.
81.2	3.1	Michigan Hill Road. This is the approximate location of the west end of the Lisle / Dudley Creek esker, which continues east for approximately 5 miles (8 km). The esker is intermittent at this west end, but becomes more continuous to the east (except where it has been quarried away).
83.4	2.2	Turn left into Broome-Tioga Sport Center (motocross) for STOP 6

STOP 6. LISLE ESKER.

Most of the deposit has been removed at this location during construction of the speedway parking area, but part of the esker is exposed to the west of the parking area. This esker was deposited within a fluvial channel beneath (likely) stagnant ice. Clasts range from sand to cobble in size, and are rounded. The unit is stratified and contains well-sorted lenses of sand and gravel. Clast lithologies are predominantly of local origin, with a small percentage of exotics.

Return to vehicles, turn left onto route 79, continue toward Lisle.

85.8	2.4	Center Lisle; east end of Lisle Esker. Continuing to the east, the valley margins are marked by a series of kame terrace deposits.
88.8	3.0	Turn left into Lisle Fire Station parking lot. Walk down to Dudley Creek for STOP 7.

STOP 7. KAME TERRACE WITHIN NORTHERN CUTBANK OF DUDLEY CREEK.

Dudley Creek has been undercutting its north bank at a number of locations through this narrow valley. This has produced a number of large slumps, the best exposed of which is here. Most of the slumps expose stratified sands and gravels. The deposits at this location also are stratified, with a massive, well-sorted sand unit in the lower section overlain by a unit of interbedded sand and gravel. The lower sand unit contains cross beds and collapse structures. Layers within the interbedded unit dip upstream and contain large scale cross beds.

Return to vehicles, turn right onto route 79 to retrace your steps to the west end of the narrower part of Dudley Creek Valley.

91.6	2.8	Turn right onto Tony Road.
91.7	0.1	Turn right onto Owen Hill Road. Owen Hill Road climbs onto broad kame terrace surface. This is the same kame that is exposed in the bank of Dudley Creek (Stop 7).
93.2	1.5	Crest of Owen Hill Road onto kame terrace surface.
94.5	1.3	Descend onto lower kame terrace surface.
95.0	0.5	Begin descent from lower kame surface.
95.1	0.1	Pull over to left at exposure on north side of Owen Hill Road for STOP 8.

STOP 8. KAME TERRACE

This terrace is a continuation of the deposit along the northern wall of the Dudley Creek Valley. Clasts within this deposit are rounded and range from fine sand to boulders. The lithologies are mostly of local origin, but exotics, including red sandstone and metamorphics are also found. Clasts are oriented into the deposit, indicating meltwater flowing south toward Owen Hill Road and parallel to the Tioughnioga Valley. The sand and gravel near the top of this deposit is cemented by calcium carbonate.

Return to vehicles, turn left onto Owen Hill Road continuing downhill.

95.8	0.7	Turn right onto Walker Road.
96.3	0.5	Turn left onto route 79.
96.7	0.4	Turn left onto route 11 and take an immediate right onto Johnson Hill Road
97.0	0.3	Turn left onto Watts Road.
97.3	0.3	Turn left along dirt road into exposure for STOP 9.

STOP 9. KAME DELTA

This deposit is above the Tioughnioga Valley on the eastern valley wall. A large kame terrace with a working sand and gravel operation is found below this deposit. Here, the kame deposits are deltaic. The delta deposit is stratified, with topsets and forsets exposed. A channel deposit is present in the upper left section of the deposit, with a channel bar to the right. A fault offsets channel sand beds, and gravels along the fault have slumped and rotated. Overall, the deposit is poorly sorted, but individual well-sorted layers with imbricated pebbles are present.

Return to vehicles, turn right onto Watts Road.

97.6	0.3	Turn right onto Johnson Hill Road.
97.9	0.3	Turn right onto route 11.
98.3	0.4	Note the working gravel pit and kame terrace exposure to the right, along the east valley wall. This is the gravel pit and kame surface west of and below Stop 9. Continue north up the Tioughnioga Valley.
104.8	6.5	In Marathon, optional side trip to an overview of the valley. For side trip, turn right onto East Main Street / route 221. Proceed 0.1 mile (under I-81), and turn left onto Galatia Street . Continue 0.3 miles to Alboro Road and turn left. Proceed 0.8 miles to Appleby School (Marathon Elementary) for Marathon Glacial Landscape Overview

OPTIONAL STOP. MARATHON GLACIAL LANDSCAPE OVERVIEW

The valley walls at Marathon contain large kame terrace deposits, including a deposit at the location of the graveyard just north of the Interstate 81 onramp. A terrace is present to the southwest, across the valley from the high school. Note the relatively flat topography of the terrace surface. Kame terraces are also found to the south of Marathon, just north of the Cortland – Broome County boundary.

Return to vehicles, leave the Marathon school, and retrace route to route 11 in Marathon. Turn north.

104.8	0.0	Continue on route 11 from Marathon. Head north along the east valley wall. Note the narrowing of the Tioughnioga Valley and the bedrock outcrops along the east valley wall.
109.0	4.2	Just north of Messengerville is the likely location of a pre-glacial upland and valley drainage divide. Drainage north was toward Cortland and drainage south was toward Marathon. As a result of glaciation, this drainage divide was eroded and the Tioughnioga River now flows south from Cortland toward Marathon and Whitney Point.
111.3	2.3	Abandoned gravel pit just east of road. Pit exposes more than 30 m of stratified sand and gravel; most is cemented by calcium carbonate. This deposit may be part of a recessional moraine that was a valley plug deposit, impounding a lake that deposited thick glaciolacustrine sediments now buried beneath the Tioughnioga Valley to the north.
113.1	1.8	Hoxie Gorge Road. Spectacular bridges take I-81 across this deep gorge, which is incised into glacial deposits and (mostly) bedrock.
116.9	3.8	Continue north to Polkville. This area is within the valley of the Main Stem Tioughnioga. The Trout Brook Valley merges with the Main Stem Valley from the east in this area. Pre-glacial drainage in this area was to the northwest, as shown by the orientation of the Trout Brook and Tioughnioga valleys, the change in valley width from northwest (wide valley) to southeast (narrow), and the northwestward sloping longitudinal bedrock profile within the Tioughnioga Valley. A sand and gravel mining operation in glacial outwash deposits is located within the valley, just southwest of the highway. Beneath the outwash deposits, a thick layer of glaciolacustrine silt and clay is present. This silt and clay was deposited in a large glacial lake that extended to the Valley Heads Moraine deposits. The lake outlet was located south of Blodgett Mills, quite possibly impounded by the valley plug deposit described above at mile 111.3. Continue north on route 11.

118.2	1.3	Bear left onto Port Watson Street, the continuation of route 11.
119.0	0.8	Pendleton Street. A kame surface is located to the left (south) along the southern valley wall here at Cortland. The terrace deposits, as shown by soil parent materials, extend southwest along the Otter Creek Valley. Additional kame terraces are found along the northern valley wall of Otter Creek. These kame terraces may be part of a moraine that was deposited as ice retreated northward from the region, possibly pinned against the high topography south of Cortland. Within the valley itself, outwash deposits overlie glaciolacustrine deposits, buried outwash, and till.
119.4	0.4	Turn left at Tompkins Street (Route 13).
119.5	0.1	Turn right at Prospect Ave.
119.6	0.1	Bear left on Prospect Ave.
119.8	0.2	Turn right at Graham Ave.
120.0	0.2	Turn left onto Gerhart Ave into SUNY Cortland.
120.1	0.1	Return to Bowers building. END OF TRIP.

END OF FIELD TRIP

Trip A-3

KIMBERLITIC ROCKS OF CENTRAL NEW YORK

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INTRODUCTION

The term "kimberlite" was first formally used to refer to the diamond-bearing mica peridotites of South Africa by Henry Lewis in 1887 at a meeting of the British Association for the Advancement of Science (Lewis,1888; Mitchell, 1986). Interestingly, similar mica peridotites were first discovered in upstate New York fifty years earlier. In his first annual report on the geology of the third district of New York, Lardner Vanuxem reported the presence of "serpentine" veins cutting through the Paleozoic sediments on the banks of East Canada Creek (approximately 10km east of the city of Little Falls) (Vanuxem, 1837). In his final geological report for the Third District, Vanuxem (1842) described four dikes of uncommon igneous rocks in a ravine east of Ludlowville (Tompkins County), the dikes on East Canada Creek (Herkimer/Montgomery County), as well as a "serpentine body" in the city of Syracuse (Onondaga County). According to Williams (1887a,b), the Syracuse serpentinite was discovered in 1837 by Oren Root who brought the rock to the attention of Vanuxem. Thus, 1837 appears to be the year when kimberlitic rocks were first observed and described anywhere in the world.

The New York dikes are part of a larger north-south belt of kimberlitic intrusions on the western flank of the Appalachian mountain belt that extends from Tennessee to Quebec. These dikes are of particular interest to North American geologists because they provide the only direct information on the nature of the mantle and lower crust in the Appalachian interior, and they are the only expressions of Mesozoic magmatism in the region. Their age, origin, and relationship to plate tectonic processes are still poorly known or understood.

DISTRIBUTION OF DIKES

To date more than 80 distinct kimberlitic dikes and irregular intrusive bodies have been found in New York (Table 1). Most exist in clusters within an elongate NNE – SSW area between Syracuse and Ithaca, with the vast majority in ravines that feed into Cayuga Lake from the east, south, and west (Figure 1). Two somewhat isolated dikes in the Ithaca area are one at Filmore Glen State Park, and the other in a small quarry northwest of Ithaca (McDougal Rd dike) (Figure 2). The Ogdensburg (Eel Weir) dike in St. Lawrence County is the northernmost and most isolated dike in the state (Newland, 1931). Unfortunately, this dike is no longer exposed
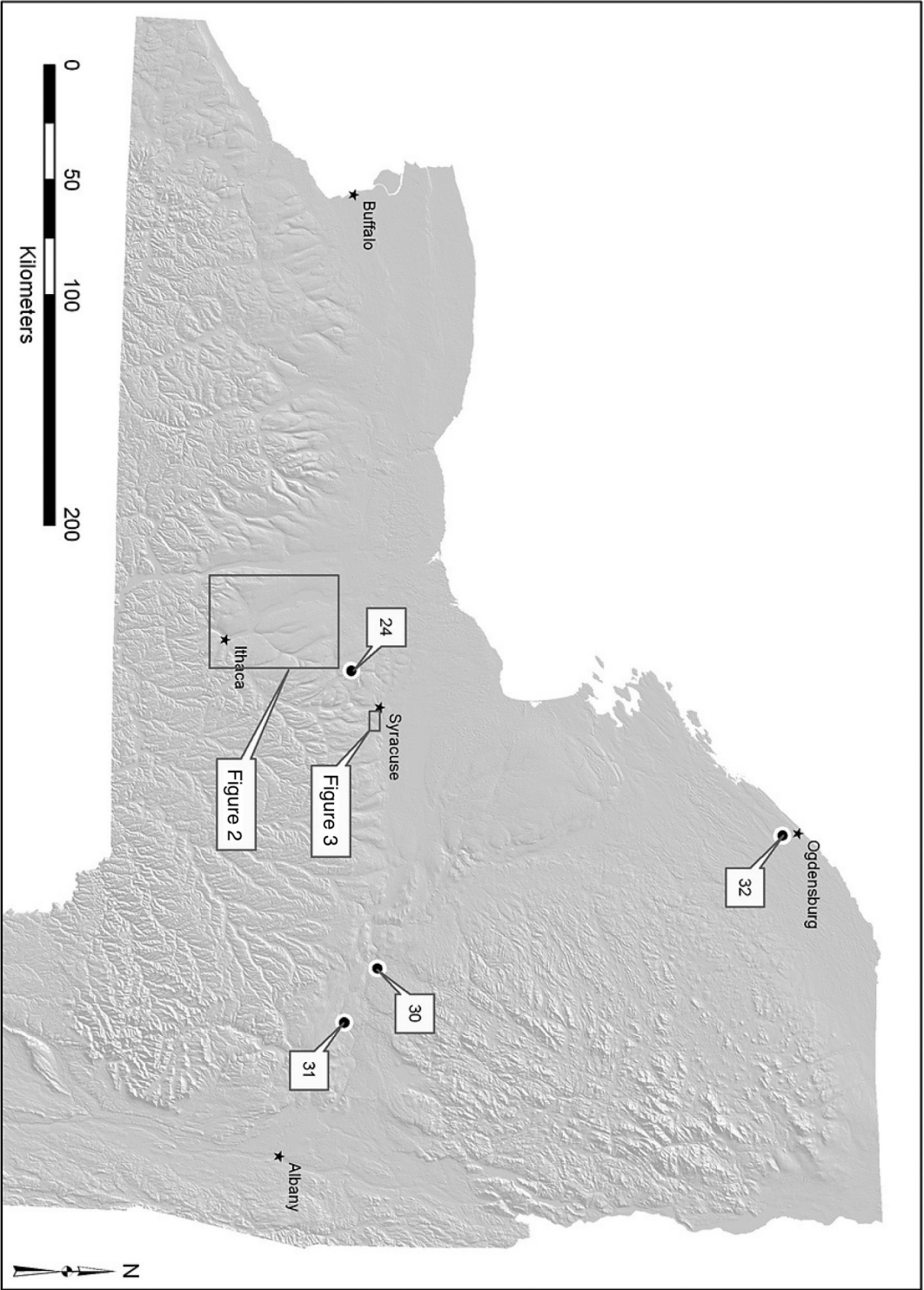
TABLE 1—List of known kimberlitic intrusions in New York State (modified after Martens (1924), Foster (1970), and Kay & Foster (1986)). Numbers in first column correspond to location numbers displayed on the following maps

Map #	Locality Name	Location	Width (cm)	Strike	References	2007 Status
1	Macdougal Rd Quarry	Small quarry 400m west of intersection of NY96 and NY336. Dike in quarry floor.	40	N35W	Wells (1961); Foster (1970)	Unknown
2	Lively Run	Lively Run; 2.1km northeast of Interlaken; 425m east of N-S road. Two dikes, 6m apart.	2.5 4.5	N10W	Martens (1924)	Unknown
3	Frontenac Creek	Frontenac Creek; 460m west of NY89. Cluster of three dikes. One additional dike 150m upstream.	5 - 60	N10W	Foster (1970)	Unknown
4	Taughannock Creek	Five dikes at foot of Taughannock Falls.	2 - 10	~N-S	Matson (1905); Martens (1924)	Unknown
5	Taughannock Creek	Taughannock Creek; 1km east of NY96 bridge; ten dikes over distance of 500m.	250	N10W	Martens (1924); Foster (1970)	Cluster of three dikes (Foster's dikes 6-8) exposed on north side of creek. 18T 0366828 UTM 4709908
6	Glenwood Creek	Glenwood Creek; ~400m west of RR; 365m east of Duboise Rd; one dike on south side of creek.	270	N6E	Matson (1905); Martens (1924); Foster (1970)	Unknown
7	Glenwood Heights Rd	Unnamed creek north of Glenwood Creek, where creek meets bend in Glenwood Heights Rd.	45		Filmer (1939)	Unknown
8	Poyer Orchard 1	1.2km south of Glenwood Ck dike; diatreme(?).	6000		Barnett (1905); Martens (1924); Filmer (1939); Foster (1970)	Unknown
	Poyer Orchard 2 & 3	1.2km south of Glenwood Ck dike; two dikes, one 150m north of diatreme, one 180m north.	90 45			
9	Indian Creek	Indian Creek; north of old hospital along road to old heating plant; three dikes over 20m	4 - 60	N8W	Martens (1924); Foster (1970)	Unknown
10	Williams Brook	60m west of NY96; dike on north side of creek.	370	N3W	Filmer (1939); Foster (1970)	Well exposed on north bank. 18T 0374573 UTM 4701668
11	Six Mile Creek 1	Six Mile Creek; 1.2 km south of dam at Green Tree Falls.	5	N5E	Martens (1924); Foster (1970)	Unknown
12	Six Mile Creek 2	Six Mile Creek; 90m north of dam at Green Tree Falls.	5	N2W	Martens (1924); Foster (1970)	Unknown
13	Six Mile Creek 3	Six Mile Creek; 80m south of pumping station; two dikes 10m apart.	25 8	~N-S	Martens (1924); Foster (1970)	Unknown
14	Six Mile Creek 4	Six Mile Creek; 180m north of old pumping station; two dikes 10m apart.	25 10	N3E	Martens (1924); Foster (1970)	Unknown
15	Ithaca Reservoir	Near Six Mile Creek; 60m up ravine near small lake (now covered by Ithaca reservoir). Weathered diatreme?	300 by 240		Filmer (1939)	Unknown
16	Brandon Place Quarry	Small quarry in southeast portion of Ithaca (no longer accessible). Four dikes.	2 - 20	N2E	Martens (1924)	Unknown
17	Cascadilla Creek	Gorge below Central Ave; Cornell campus (next to Snee hall); Six dikes over 100m east and west of bridge.	5 - 100	N10W to N2E	Martens (1924); Foster (1970)	Unknown

TABLE 1 (continued)

Map #	Locality Name	Location	Width (cm)	Strike	References	2007 Status
18	Willard Ravine	Ravine, Cornell campus; south of Willard Straight Union.	35		Sheldon (1927); Filmer (1939); Foster (1970)	Unknown
19	Fall Creek	Fall Creek Gorge; south side of gorge in first deep notch east of Stewart Ave.	10	~N-S	Martens (1924); Foster (1970)	Unknown
20	Portland Point	Portland Point; two dikes in east wall of Cayuga crushed stone quarry; one in underground Cargill salt mine.	25 - 70	N1W to N14W	Sheldon (1921); Martens (1924); Broughton (1950); Foster (1970)	Unknown
21	Townley Creek	Townley Creek east of Ludlowville; Seven dikes between 70 and 340m east of falls.	10 - 20	N2W to N10E	Vanuxem (1839); Matson (1905); Martens (1924); Foster (1970)	Unknown
22	Ludlowville	First ravine south of Townley's Creek; Eleven dikes between 300m and 700m east of falls.	2 - 18	N6W to N10E	Schneider (1903a); Martens (1924); Foster (1970)	Unknown
23	Fillmore Glen	South of Moravia; Fillmore Glen State Park, 120m above Pinnacle lookout.	40	N12E	Wells (1961); Foster (1970)	Could not be located.
24	Clintonville	Clintonville, north of Otisco Lake; first stream north of US20 that flows east to Nine Mile Creek; 366m upstream in south wall, 2 clusters of 3 dikes each.	4 - 33	N6E to N12E	Smith (1909), (1931); Hopkins (1914)	Only three dikes found on south bank of creek. 18T 0390385 UTM 4754434
25	Salt Springs Rd	Near intersection of Salt Springs Road and Genesee Street.	130	~N-S, 82E	Maynard & Ploger (1946)	Covered.
26	Butternut St./ Highland St./ Griffiths St.	Exposed during 1902 excavation for trunk sewer and subsequent excavations.	1000	N5E	Smyth (1902); Schneider (1902,1903b); Kraus (1904)	Covered.
27	Foot St./ Green Street/ James St.	Excavations exposed many poorly defined bodies of peridotite.	150 - 1200		Williams (1887a,b), (1890a,b); Schneider (1903b); Smyth (1902); Hogeboom (1958)	Loose blocks in vacant lot on Green St. 18T 0407359 UTM 4767653
28	DeWitt	0.5mi S of Dewitt Center; 3 mi east of Syracuse; under reservoir excavated in (1895)	~600	N80E	Darton & Kemp (1895a,b); Schneider (1903b)	Abundant blocks on banks /slopes of LeMoyne reservoir. 18T 0411980 UTM 4766854
29	Euclid Ave	Dike exposed in excavated hill south of Euclid Ave.	60	N35W 80E	Van Tyne (1958)	Covered.
30	East Canada Creek	East bank of ECC, 25m upstream of hydroelectric plant, two dikes 30 - 50m south of falls	20-30	N35E	Vanuxem (1842); Smyth (1892), (1893), (1896), (1898)	Well exposed on east and west banks. 18T 0521072 UTM 4763037
31	Big Nose	Dike exposed in road outcrop at "Big Nose" along the Mohawk River, north of NY Rt. 5.	4		Allers (pers. comm..)	Covered.
32	Ogdensburg	Dike exposed during construction of Eel Weir dam.	700(?)	NE, vertical	Newland (1931)	Covered.

FIGURE 1—Locations of kimberlitic intrusions in New York State. Numbers correspond to localities listed in Table 1. Boxes outline areas covered by detailed maps (Figs. 2 and 3).

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AGE OF THE INTRUSIONS

In general, kimberlitic rocks are very difficult to date because the xenoliths, xenocrysts, macrocrysts, and groundmass phases they contain can each have a different age. In addition, even if ages are determined on individual minerals, it is not always possible to determine if the minerals crystallized from the kimberlitic fluid or are xenocrystic. Finally, extensive post-emplacement metasomatism, alteration, and weathering make it difficult to find and extract suitable material for dating.

With the above caveats in mind, the published isotopic data on the kimberlitic rocks in New York suggest a Late Jurassic / Early Cretaceous emplacement age. The initial dates by Zartman et al. (1967) on coarse grained, phlogopite macrocrysts from two dikes, (one at Portland Point and the other one at Manheim), yielded K-Ar ages of 439 ± 22 Ma and 371 ± 19 Ma, respectively. While the age of the Manheim dike is geologically possible (the dike cuts the Upper Cambrian Little Falls dolostone), the age for the Portland Point dike is not (the dike intrudes a mid-Devonian limestone). Zartman et al. concluded that the K-Ar ages were affected by excess radiogenic argon. Rb-Sr isotopes on the same phlogopite samples yielded ages of 136 ± 8 Ma (Manheim dike) and 118 ± 15 Ma (Portland Point), suggesting crystallization (and emplacement?) during the Cretaceous.

Basu et al. (1984) reported whole-rock K-Ar ages of 139 ± 7 Ma (Williams Brook), 140 ± 8 Ma (Frontenac dike), 146 ± 8 Ma (Cascadilla Gorge), 121 ± 23 Ma (Taughannock Creek) and 113 ± 11 Ma (Portland Point). Kay and Foster (1986) suggested that these data indicated at least two distinct intrusion events. The large uncertainties associated with these dates, and the unresolved question of what exactly a bulk rock age on serpentinized peridotites represents, make the interpretation of distinct intrusive events highly speculative.

Miller and Duddy (1989) reported an apatite fission-track age data from one of the Ithaca dikes of 104 ± 22 Ma. They interpreted this age as the time of cooling when the region was exhumed and eroded.

Heaman and Kjarsgaard (2000) reported high precision U-Pb perovskite ages of 144 ± 8 and 146 ± 7 Ma for the Williams Brook dike, and 147 ± 5 Ma for the Glenwood Creek dike. These are probably the dates that most accurately reflect the emplacement age of the dikes because the perovskite crystals are ubiquitous in the groundmass of many of the dikes and are clearly of magmatic origin.

MINERALOGY AND PETROGRAPHY

Relatively large (2-15 mm) macrocrysts "floating" in a very fine-grained matrix form the usual texture of these rocks. The most common macrocrysts are olivine, phlogopite, garnet, clinopyroxene, and spinel; the groundmass contains phlogopite, calcite, serpentine, perovskite, and magnetite along with minor (and localized) clinopyroxene, clinoamphibole, epidote, chlorite, barite, celestite, spinels, ilmenite, pyrrhotite, pentlandite, and pyrite. Despite many extensive searches for diamonds, none have ever been found. The shallow emplacement level of the dikes, and the chemical compositions of the garnet and pyroxene macrocrysts indicate that diamonds are unlikely to exist in these intrusions. A brief overview of the features of the major mineral phases follows:

TABLE 2—Ranges of major element oxide contents (wt. %) in the silicate macrocrysts. Data by electron microprobe. (nd = not detected)

Oxide	Olivine	Garnet	Diopside	Phlogopite
SiO_2	41.07-41.63	38.96-46.49	55.02-56.59	37.58-41.25
TiO_2	nd	0.08-2.11	0.00-0.17	0.84-3.41
Al_2O_3	0.00-0.03	11.25-22.44	0.45-2.65	6.62-16.93
FeO	8.42-12.49	6.06-24.25	1.75-6.66	4.00-13.34
MnO	0.09-0.20	0.05-0.45	0.09-0.36	0.01-0.14
MgO	47.56-50.63	8.60-20.55	13.46-17.77	19.70-24.51
CaO	0.03-0.10	4.61-22.89	19.35-25.78	0.01-0.42
Cr_2O_3	nd	0.00-5.77	0.00-1.60	0.00-2.00
V_2O_3	nd	0.02-0.10	nd	nd
NiO	0.29-0.57	0.00-0.08	nd	0.01-0.07
BaO	nd	nd	nd	0.06-3.24
Na_2O	nd	nd	0.00-1.91	0.00-0.45
K_2O	nd	nd	nd	7.88-10.04
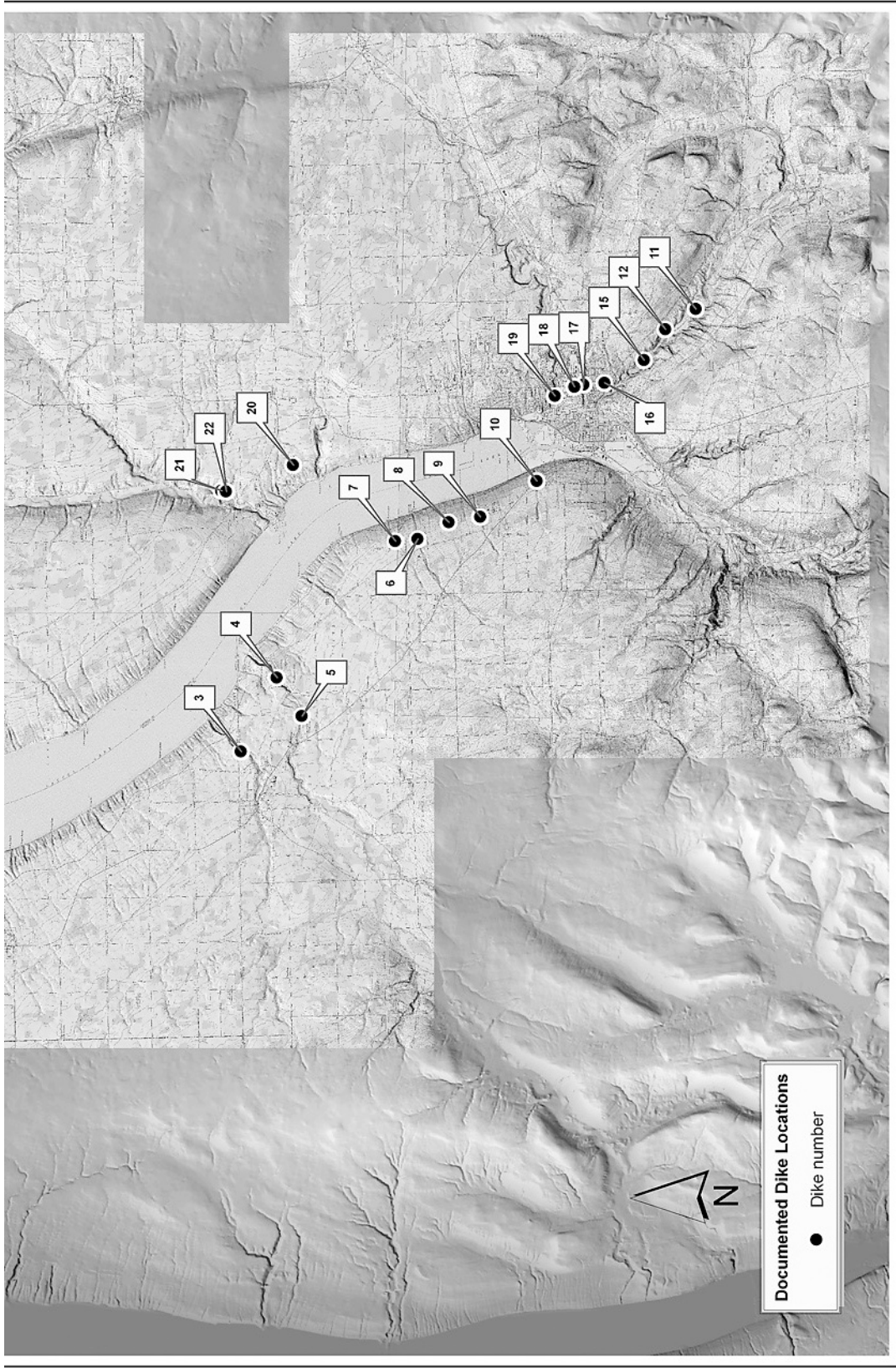
FIGURE 2—Locations of kimberlitic rocks in the Ithaca area. See Table 1 for descriptions
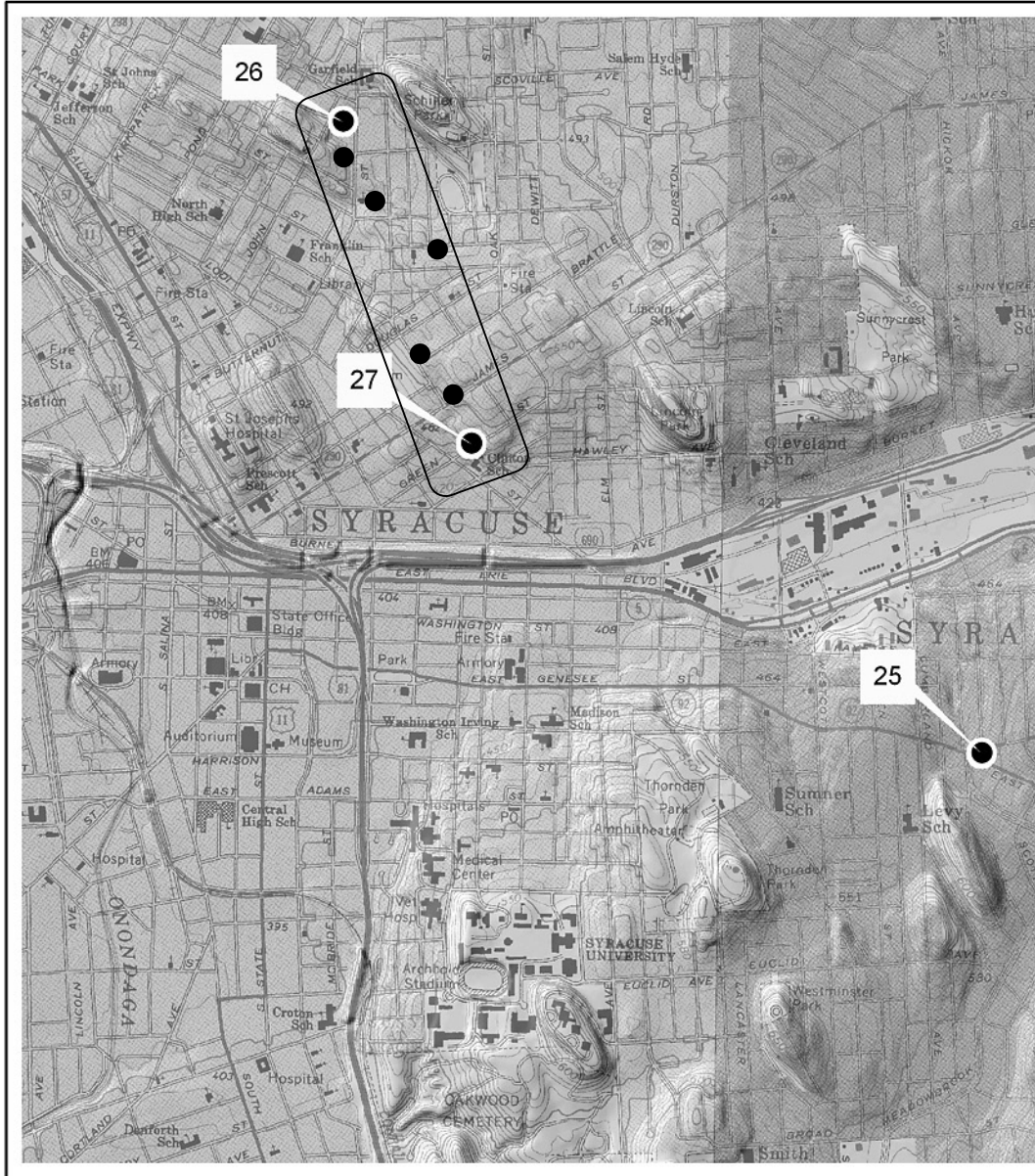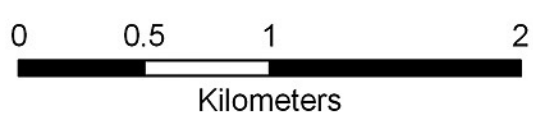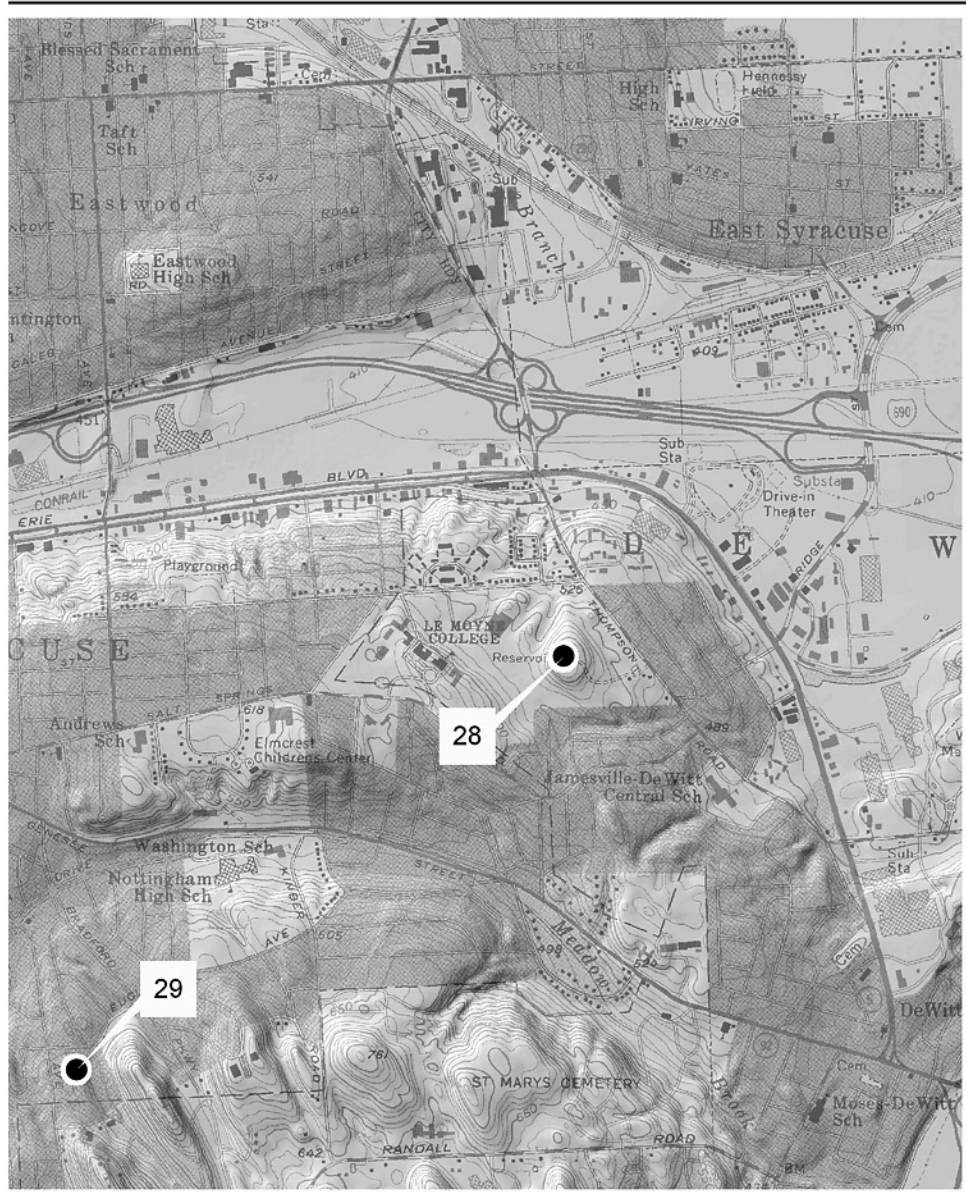
FIGURE 3—Locations of kimberlitic rocks in the Syracuse area. Box encloses zone of Green St. - Griffith St. intrusions. All of these intrusions were discovered during excavations. In 2007, none of the dikes are exposed; large blocks from previous excavations can still be found at Green St (#27) and at Dewitt Reservoir (#28). See Table 1 for descriptions.

Olivine—Olivine is, for the most part, completely serpentinized, although some dikes from the Syracuse, Ithaca and Manheim groups contain macrocrysts with fresh cores of olivine. Most of the olivine macrocrysts range in size from a few millimeters to about one centimeter, although one macrocryst in the Euclid Ave. dike is 3.6 cm in length. Generally, large and small olivine crystals coexist in the same rock.

In some dikes two generations of olivine crystals can be recognized: one has an elongate lamellar habit and the other has a rounded habit and is often partially deformed. Remnant fresh cores are more common in the somewhat larger, rounded macrocrysts. The serpentine replacing the olivine macrocrysts is fibrous to scaly, or an aggregate of extremely fine-grained and nearly isotropic scales (chlorophaite). Color varies from a pale yellow to a very dark olive brown. Other products of the alteration of olivine are calcite, millerite, chlorite and spinels (predominantly magnetite).

Chemically, the olivine macrocrysts are relatively homogeneous (Fo_{87-91}), and contain moderate NiO contents (0.29 to 0.57 wt. %), and relatively low concentrations of CaO (0.03 to 0.10 wt. %) (Table 2). The Fo_{87-91} compositions are in the range of olivines from mantle peridotites (Dawson, 1980), so it is possible that they are coming from disaggregated xenoliths. They also are similar to the compositions of phenocrysts in ultrabasic magmas, and therefore, could also be of primary magmatic origin. Macrocrysts with distinct deformation lamellae are clearly of xenocrystic origin, but these are rare.

Garnet—Garnet was found only in the Taughannock Creek, Portland Point, and Dewitt Reservoir dikes, although the initial reports of the Green Street / Foot Rd dike in Syracuse describe the local children collecting large "rubies" from the excavated material (Williams, 1887a). In all occurrences, the garnet macrocrysts are rounded and have well-developed kelyphitic coronae (Figure 4). In contrast to the uniform nature of the olivine macrocrysts, the garnet macrocrysts are variable in color (pink to orange to yellow), and composition (Table 2).

Electron microprobe data reveal four groups of garnets: i) chrome-rich pyrope, ii) pyrope, iii) magnesian almandine, and iv) high-titanium ferro-magnesian grossular (Dawson & Stephens, 1975). The Cr-pyropes (which contain 3.67 to 5.77 wt. % Cr_2O_3), are compositionally similar to garnets in garnet lherzolites, one of the most common mantle xenoliths in kimberlites (Dawson & Stephens, 1975). Unfortunately, there is considerable compositional overlap with garnets in garnet peridotites and garnet pyroxenites, so these sources cannot be excluded on the basis of bulk composition. Trace element compositions obtained through laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) reveal high concentrations of the HREE (Lu_{cn} = 9.65 to 11.14) and extremely low concentrations of the LREE (La/Nb_{cn} = 0.03) (Figure 5). The Sc/Yb and Ti/Sc ratios of the garnet macrocrysts are close to those of C1-chondrites. These features strongly suggest that the pyrope macrocrysts are derived from a garnet-lherzolite source.

The pyrope (1.96 to 2.14 wt. % Cr_2O_3), magnesian almandine (8.6 wr. % MgO) and grossular (2.11 wt. % TiO, 6.53 wt. % FeO) macrocrysts are most likely derived from disaggregated xenoliths that come from shallower mantle or crustal sources (eclogites?, grosspydites?).
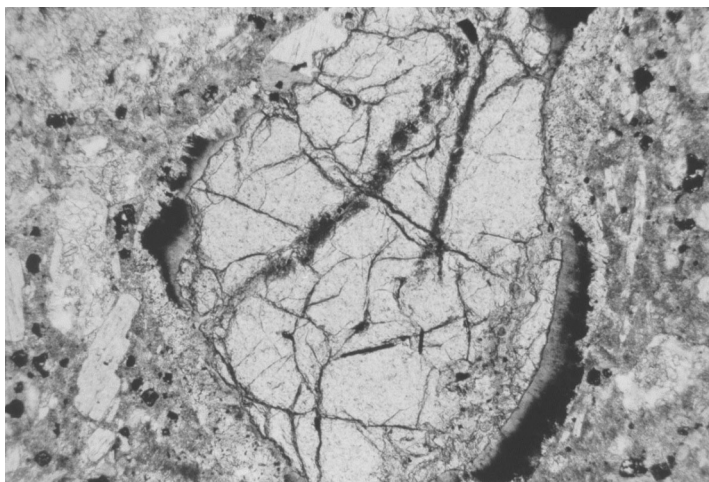
FIGURE 4—Fractured garnet macrocryst with kelyphitic corona, Taughannock Creek, Tompkins County. Width of photo = 3mm.
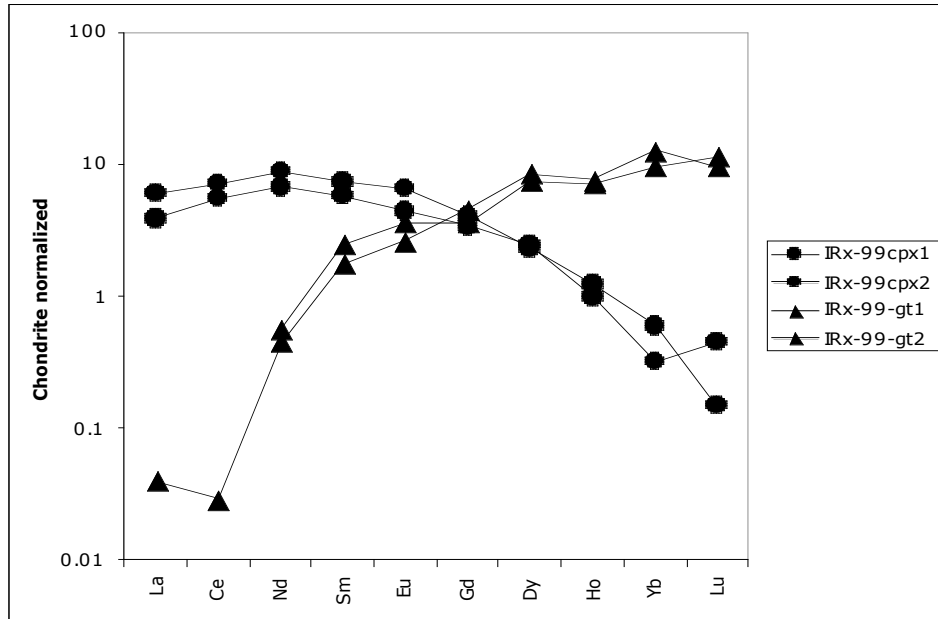
FIGURE 5—Chondrite normalized REE pattern for pyrope (IRx-99-gt) and low-Cr diopside (IRx-99-cpx) macrocrysts.

Diopside—Diopside is a fairly common macrocryst in many of the dikes in the Ithaca area but is quite scarce in dikes from other regions. It ranges in color from light to dark emerald green, to a pale yellow-green color. Similar to the garnet macrocrysts, the clinopyroxenes also exhibit kelyphitic coronae in some occurrences (Figure 6). The diopside crystals are no larger than a few millimeters in diameter. All are anhedral and extensively fractured.

Microprobe data show two groups of clinopyroxenes: low Cr-diopsides (with 0 to 0.05 wt. % Cr_2O_3) and high Cr-diopsides (with 1.52 to 1.60 wt. % Cr_2O_3). The high Cr-diopsides have 1.81 to 1.91 wt. % Na_2O, 2.54 to 2.65 wt. % Al_2O_3 and 2.83 to 2.89 wt. % FeO, showing affinities with clinopyroxenes in garnet lherzolites. The low Cr-diopsides have low Na_2O (0 to 0.37 wt. %) and low Al_2O_3 (0.45 to 1.04 wt. %). Kay et al. (1983) reported clinopyroxene compositions matching those of pyroxenes in garnet lherzolites and in garnet (or spinel) peridotites.
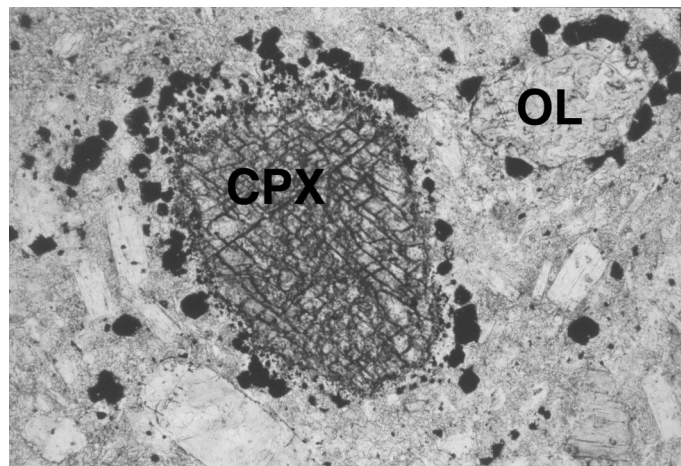

FIGURE 6—Diopside with a reaction rim of very small and coarse spinels. Olivine with a spinel rim in the upper right corner and phlogopite in the groundmass. Width of photo = 3.3mm.

The low-Cr clinopyroxene macrocrysts are relatively enriched in the LREE (La/Yb_N = 10.19 to – 12.16) and depleted in the HREE (Lu_N = 0.15 to 0.45) leading to subchondritic ratios of Sm/Nd_N (0.83 to 0.86) and Lu/Hf_N (0.02 to 0.14) (Figure 6). The Sc/Yb and Ti/Sc ratios of the clinopyroxene macrocrysts are suprachondritic. The trace element data are insufficient to characterize the source of the low-Cr diopside macrocrysts.

Phlogopite—The modal proportion of phlogopite to the other macrocrysts varies from dike to dike. Phlogopite is most abundant in the two East Canada Creek dikes, where flakes can reach 1.5 cm in diameter (Figure 5b). It is also abundant in the groundmass of many dikes. Rounding and strong deformation characterize all of the phlogopite macrocrysts. Chlorite, calcite, and Fe-Ti oxides replace some crystals along the (001) cleavage planes.
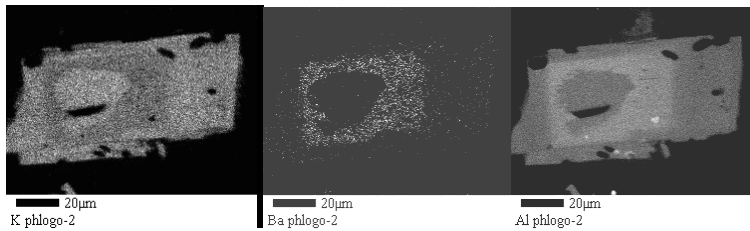

FIGURE 7a—X-ray elemental distribution maps on a zoned groundmass phlogopite (dike from the Syracuse group).
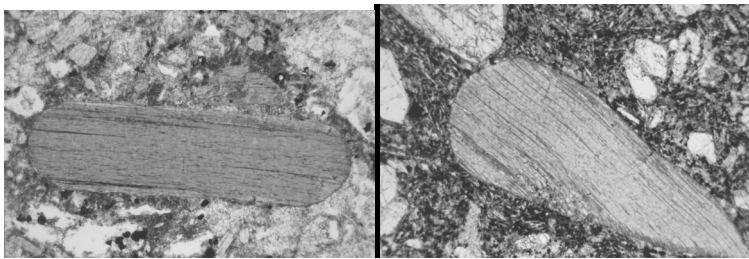
FIGURE 7b—Rounded and deformed phlogopite macrocrysts, East Canada Creek. Width of photo = 4mm.

Microprobe studies to date have not revealed any significant compositional differences between macrocryst and groundmass grains. Some of the phlogopite grains display zoning (Figure 7a) having a Ba-rich core (3.24 wt. % BaO) and a Ba-depleted rim (0.20 wt. % BaO). There also are compositional differences between dikes; for example, the phlogopites in the Williams Brook dike contain 1.45 wt. % Cr_2O_3 whereas the grains in the Portland Point dike have Cr_2O_3 contents around 2 %. TiO_2 contents also vary (between 0.84 and 3.41 wt. %), and NiO contents are very low (0.01 to 0.06 wt. %). These compositions are all within the range of mica compositions observed in other kimberlitic rocks (Mitchell, 1986).

Perovskite—Perovskite is found in all the dikes, but it is abundant in only a few of them (e.g. Williams Brook and Manheim dikes). It has a characteristic square cross section, is yellow-brown in color, and occurs as small (<0.25mm), isolated octahedral crystals in the serpentine and calcite dominated matrix. Clusters of small crystals also can be found as rims on magnetite grains (Figure 8). These two occurrences represent two different generations of perovskite: primary perovskite that crystallized from the kimberlitic fluid and secondary perovskite that formed as a post-magmatic reaction rim on magnetite. The chemical composition of both is in the range of perovskites from other kimberlitic rocks (Mitchell, 1986).
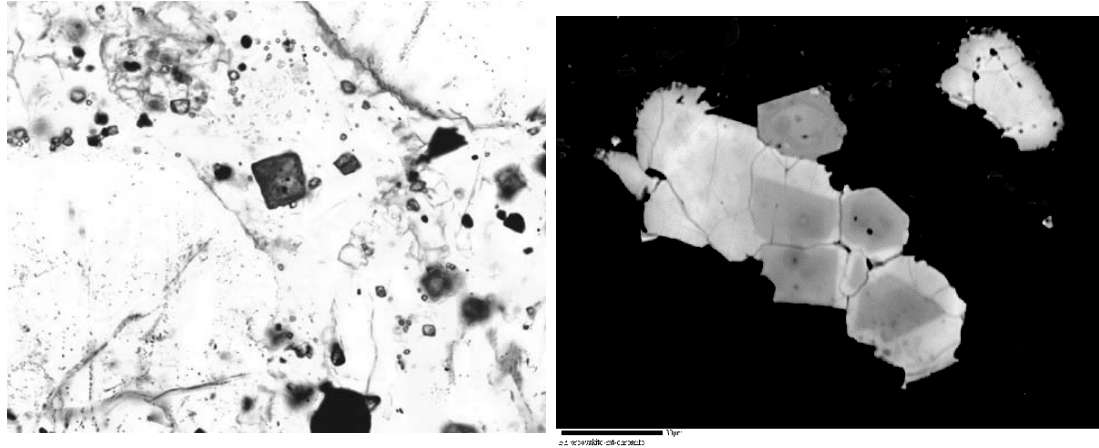
FIGURE 8—Photomicrograph (left) of primary <50μm perovskite crystals with characteristic square cross section, and back-scattered electron image (right) of 5-10μm perovskite rims (light gray) on magnetite (gray).

Spinel—Spinels are found with different sizes, textures and compositions (Figure 9). Grains range in size from micron scale grains in the groundmass (usually magnetite), to large (up to 5mm) macrocrysts (usually with Cr and Al-rich cores). They occur as isolated grains or as clusters of crystals, and most grains display compositional zoning. Three groups of spinels can be recognized according to their Cr, Al and Fe content. The first group contains the Cr- (> 23.49 wt. % Cr_2O_3) and Al-rich (> 11.99 wt. % Al_2O_3) spinels. These spinels are zoned with a Cr- and Al-rich cores and Fe-rich rims.
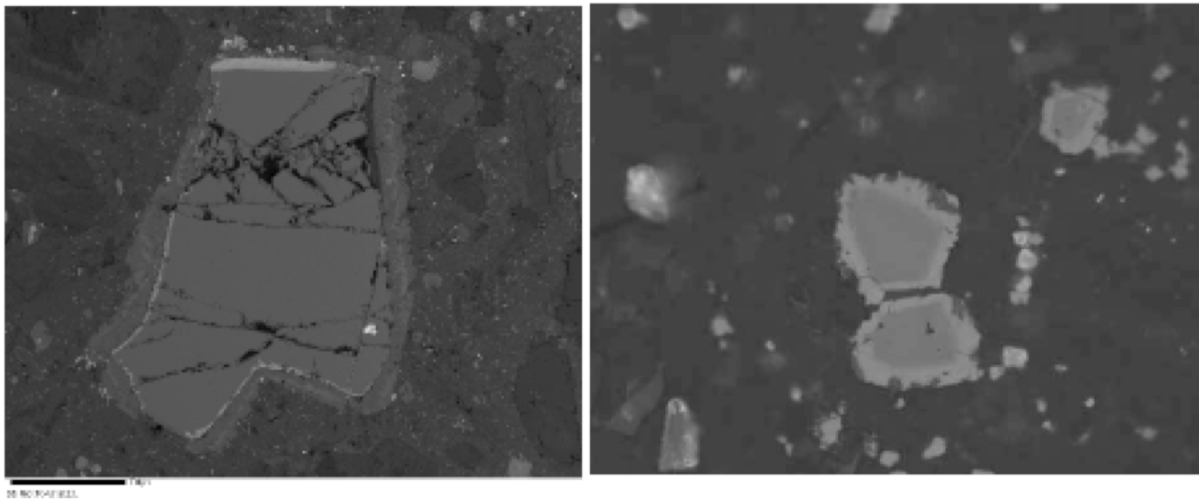
FIGURE 9—Left: Back-scattered electron (BSE) image of a spinel macrocryst (Taughannock Creek dike). The core is chromite and the fine whiter rim is magnetite. Both are enveloped by an unknown, non-stoichiometric Fe-rich compound. Below: Reflected light image of chromite (darker) with magnetite rim (lighter) in the Green Street dike.

The second group contains the low Al (< 5.15wt. % Al_2O_3) Fe-rich (> 70 wt. % FeO as total iron) spinels that plot into the magnetite compositional range. The TiO_2 (7.94 to 9.98 wt. %) and MgO (6.07 to 8.08 wt. %) contents confer a titanian and magnesian character. The third group makes the compositional transition from chromite to magnetite (magnesioferrite, magnesiochromite).

The wide range of spinel compositions observed is typical of kimberlitic rocks (Mitchell, 1986) and reflects their complex nature and history.

Ilmenite—Ilmenite is fairly common in the groundmass as irregular and tabular grains. Unlike the large MgO-rich macrocrysts commonly found in Group I kimberlites (Mitchell, 1986) these ilmenites are small and have relatively low MgO contents (0.90 to 6.85 wt. %).

Calcite—Calcite is a major component of the groundmass; in places, it also replaces the cores of olivine macrocrysts. Calcite could come from two sources: primary calcite derived from the CO_2-rich kimberlitic fluid, and secondary calcite derived largely from the surrounding calcareous shales and limestones.

WHOLE-ROCK CHEMISTRY

While there have been nearly one hundred scientific reports on the kimberlitic rocks of New York since they were discovered 170 years ago, only twenty bulk chemical analyses have been published. The first was a partial analysis by T. S. Hunt in 1858, and the most recent were three analyses by Foster in 1970. The paucity of chemical data is a consequence of the fact that the rocks are difficult to analyze, and the resulting data are difficult to interpret.

Kimberlites are, by nature, hybrid rocks containing complex mixtures of mantle and crustal derived materials, and almost all have experienced extensive post-emplacement hydrothermal and/or groundwater alteration. Because of these complications, whole-rock compositions almost certainly do not represent, or even approximate, magmatic liquid compositions. This limits our ability to understand the mineralogical and chemical nature of the mantle source of kimberlitic magmas, and their subsequent evolution. Nevertheless, whole-rock chemistry does provide important information that allows us to categorize and classify these unusual rocks, and to constrain the geological processes involved in their formation.

Thirty-two samples were analyzed by XRF and ICP-MS spectrometry for major and trace element compositions. Representative analyses are presented in Table 3. Major element oxide proportions reflect the fact that these rocks are composed primarily of variable proportions of serpentine, calcite, and magnetite (Figure 10).
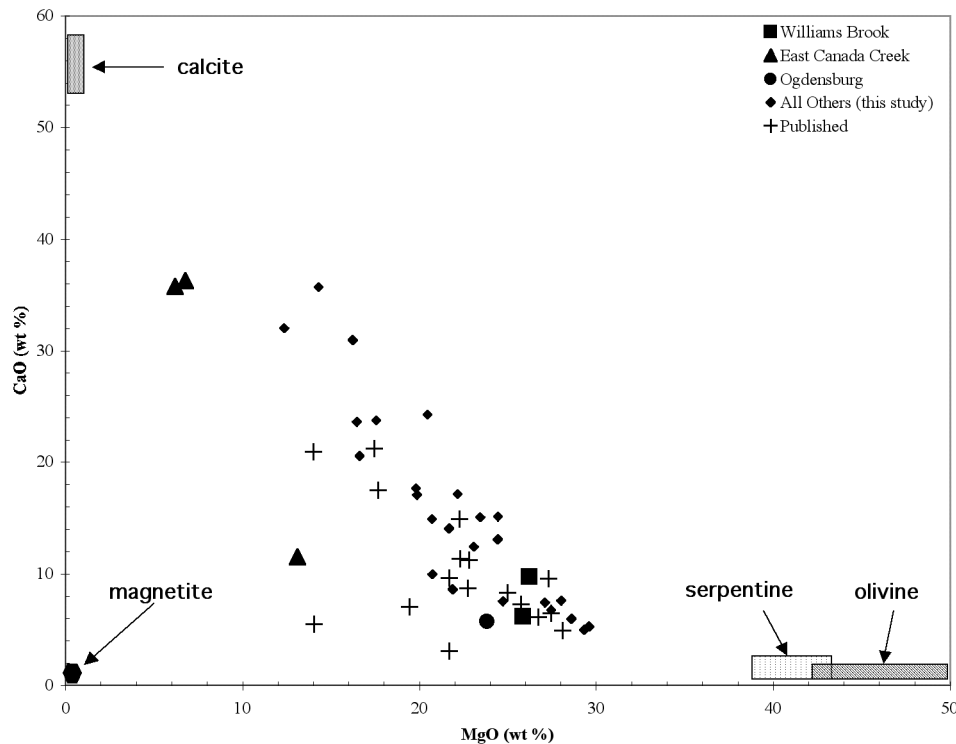
FIGURE 10—CaO vs. MgO in kimberlitic rocks of New York State. Major element oxide compositions reflect the variable modal mineral composition within and between dikes.

TABLE 2—Representative whole-rock analyses of kimberlitic rocks from New York State.

Map #:	5	10	17	20	24	25
Location Name:	Taughannock	Williams Br.	Cascadilla	Portland Pt.	Clintonville	Salt Spgs Rd
Sample#:	T1	W2	10069	IRx-99	IRx-100	10065
Major Element Oxides (XRF wt.%)						
SiO_2	27.34	32.05	34.40	27.55	34.55	37.70
TiO_2	1.53	2.42	1.55	1.37	1.58	1.39
Al_2O_3	2.89	3.09	3.26	2.59	6.37	4.33
FeO*	8.42	9.14	8.75	8.33	7.86	8.28
MnO	0.21	0.17	0.21	0.18	0.22	0.14
MgO	17.81	26.38	28.52	22.62	21.47	26.86
CaO	15.29	9.81	5.95	13.97	8.46	6.79
Na_2O	0.16	0.11	0.08	0.14	0.07	0.11
K_2O	1.37	1.94	1.50	0.93	2.09	2.88
P_2O_5	1.13	0.41	0.57	0.97	0.77	0.40
SO_3 >/=	0.34	0.22	0.30	0.09	0.10	0.30
Major oxides	76.50	85.74	85.09	78.73	83.53	89.19
Trace oxides	0.87	0.75	1.07	0.98	1.03	0.83
LOI	21.28	12.73	13.68	19.38	15.59	9.68
TOTAL	98.65	99.22	99.84	99.09	100.15	99.70
Trace Elements (XRF ppm)						
Ni	988	1013	1195	986	693	1083
Cr	1388	1875	1737	1340	1686	1755
Sc	24	17	19	19	26	16
V	262	227	262	216	323	147
Ba	1436	840	2861	1674	2795	1583
Rb	50	80	68	38	102	101
Sr	1473	511	926	1981	1152	806
Zr	287	142	189	246	225	147
Y	22	11	17	20	19	13
Nb	183	110	193	157	189	111
Ga	9	8	8	6	12	8
Cu	68	63	64	83	69	48
Zn	74	66	96	127	93	90
Pb	10	8	12	13	12	6
La	160	81	206	132	152	77
Ce	292	152	300	240	272	140
Th	21	13	27	20	25	14
Nd	112	59	98	93	95	50

TABLE 2 (Cont.)—Representative whole-rock analyses of kimberlitic rocks from New York State.

| Map #:	27	28	29	30	31	32
Location Name:	Green St.	Dewitt	Euclid Ave	East Canada	Big Nose	Ogdensburg
Sample #:	GS1	15333	EA1	EC4f	BN1	11430
Major Element Oxides (XRF wt.%)						
SiO_2	34.25	37.41	36.99	22.24	30.63	31.62
TiO_2	1.27	1.72	1.13	2.55	1.19	2.33
Al_2O_3	3.46	4.96	3.71	4.66	3.50	7.67
FeO^*	8.87	8.45	8.13	3.22	7.61	11.44
MnO	0.17	0.15	0.16	0.09	0.12	0.30
MgO	29.48	26.46	28.16	5.72	22.90	23.70
CaO	5.25	8.03	7.68	30.77	12.27	5.73
Na_2O	0.07	0.17	0.12	0.08	0.08	0.07
K_2O	1.43	3.20	1.86	2.43	0.79	1.37
P_2O_5	0.39	0.44	0.34	0.33	0.62	0.40
$SO_3 >/=$	0.08	0.22	0.23	0.34	0.34	0.03
Major oxides	84.73	91.21	88.51	72.43	80.04	84.66
Trace oxides	0.76	0.83	0.81	1.10	1.82	0.90
LOI	13.75	8.04	10.87	25.29	16.82	15.13
TOTAL	99.24	100.08	100.19	98.82	98.69	100.69
Trace Elements (XRF ppm)						
Ni	1160	1021	1226	943	993	535
Cr	1943	1523	2000	1382	2338	902
Sc	16	16	17	17	21	28
V	170	130	157	361	168	107
Ba	959	1432	1089	4351	9304	3360
Rb	66	99	74	112	34	85
Sr	402	1175	658	1016	1104	575
Zr	100	199	101	118	213	147
Y	12	12	9	11	13	25
Nb	105	132	86	77	135	218
Ga	7	11	7	9	7	16
Cu	48	42	40	99	57	23
Zn	75	125	70	161	18	64
Pb	6	17	5	6	7	7
La	81	89	68	58	93	207
Ce	145	148	124	102	172	328
Th	13	11	14	16	15	26
Nd	53	54	48	42	60	121

Because of the complex, hybrid nature of kimberlitic rocks, a number of chemical filters have been proposed to screen whole-rock analyses for the chemical effects of crustal contamination and/or weathering. Two that have been found to be useful are the contamination index [C.I. = $(SiO_2+Al_2O_3+Na_2O)$ / $(MgO+2*K_2O)$] (Ilupin and Lutz, 1971), and the molar Si/Mg ratio (Clement, 1982). Both of these indices are based upon the assumption that contamination by crustal rocks usually results in the addition of SiO_2, Al_2O_3, and Na_2O to the ultrabasic kimberlitic magma, and upon the fact that weathering usually leads to the removal of Mg and K cations and the relative enrichment in Si and Al. According to Mitchell (1986), contaminated rocks are considered to have Si/Mg > 0.88 and C.I. >1.5. Unfortunately, these indices are not universally applicable because of the diverse nature of potential crustal contaminants within and between kimberlite fields. In New York, for example, the kimberlite xenolith assemblage includes syenitic gneisses and amphibolites from the Grenvillian basement, as well as shale, dolostone, and limestone xenoliths from the Paleozoic cover. In one extreme example, the dike from the Cargill salt mine (sample BG 30) shows clear chemical evidence of contamination by the surrounding rock salt (Cl >1.3 wt. % and Na_2O > 2.2 wt. %). Considering the general nature and origin of kimberlitic rocks, probably all are chemically modified to some degree by the crustal rocks that they intrude.

Figure 11 is a plot of contamination index (CI) versus molar Si/Mg ratio; the boundaries between the fields for uncontaminated, moderately contaminated, and highly contaminated rock compositions are set at 1.25 and 1.75 for the CI, and at 0.9 and 1.2 for the Si/Mg ratio. While these values are somewhat arbitrary, they are consistent with the values adopted by Mitchell (1986) and are generally consistent with the macroscopic evidence for different degrees of contamination (i.e. abundance of crustal xenoliths). The diagram suggests that virtually all of the NY State kimberlites have been moderately to extensively chemically modified by weathering and/or crustal contamination. Somewhat surprisingly, the East Canada Creek dikes in the town of Manheim plot as the most contaminated samples, even though they contain the largest, most abundant and freshest phlogopite macrocrysts.
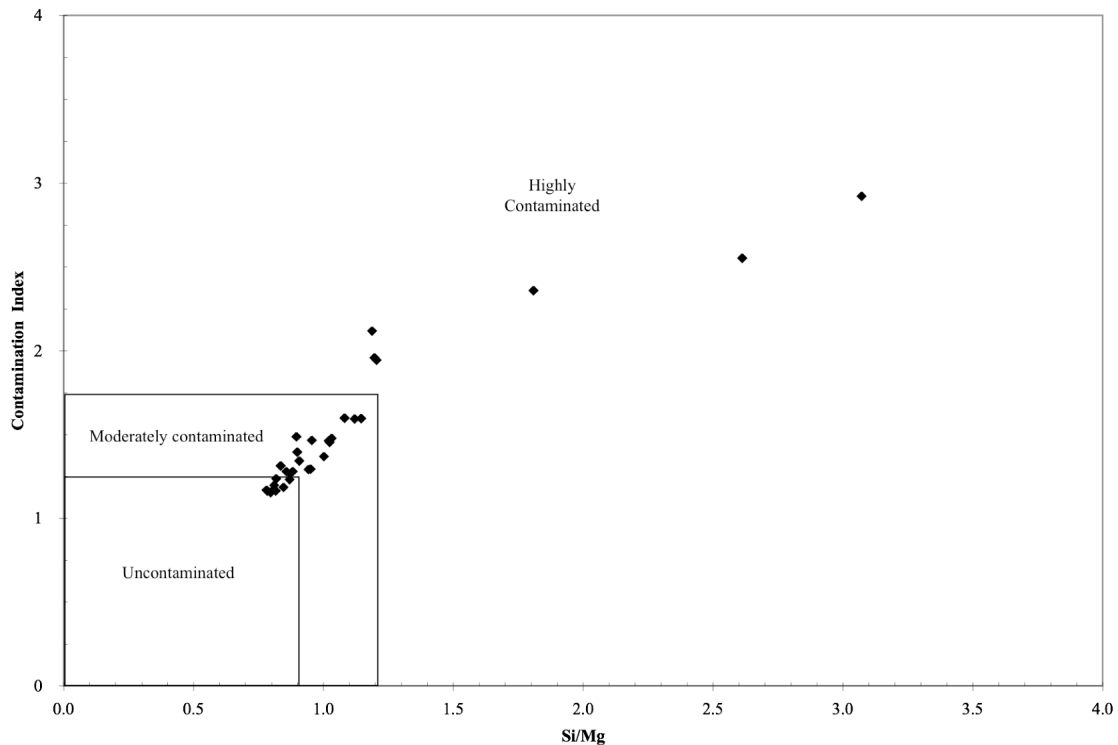
FIGURE 11—Plot of contamination index vs. molar Si/Mg ratio for New York State kimberlite-like rocks. [C.I.= $(SiO_2+Al_2O_3+Na_2O)$ / $(MgO+2*K_2O)$].

Despite the extensive alteration and contamination, the new whole-rock data allow us to identify individual dikes and/or clusters of dikes that have distinct geochemical signatures, particularly in terms of the ratios of relatively immobile, high field strength (HFS) minor and trace elements (e.g. Nb /TiO$_2$ - Figure 10). It is clear from these diagrams that the Williams Brook dike northwest of Ithaca, and the two dikes on East Canada Creek are chemically distinct intrusions and quite unlike all of the other dikes in central NY. While the significance of this is not yet fully understood, this is the first time that such systematic compositional differences have been observed in the kimberlitic rocks of New York. Additional radiometric dating and mineral trace element studies are planned to investigate these differences in more detail.
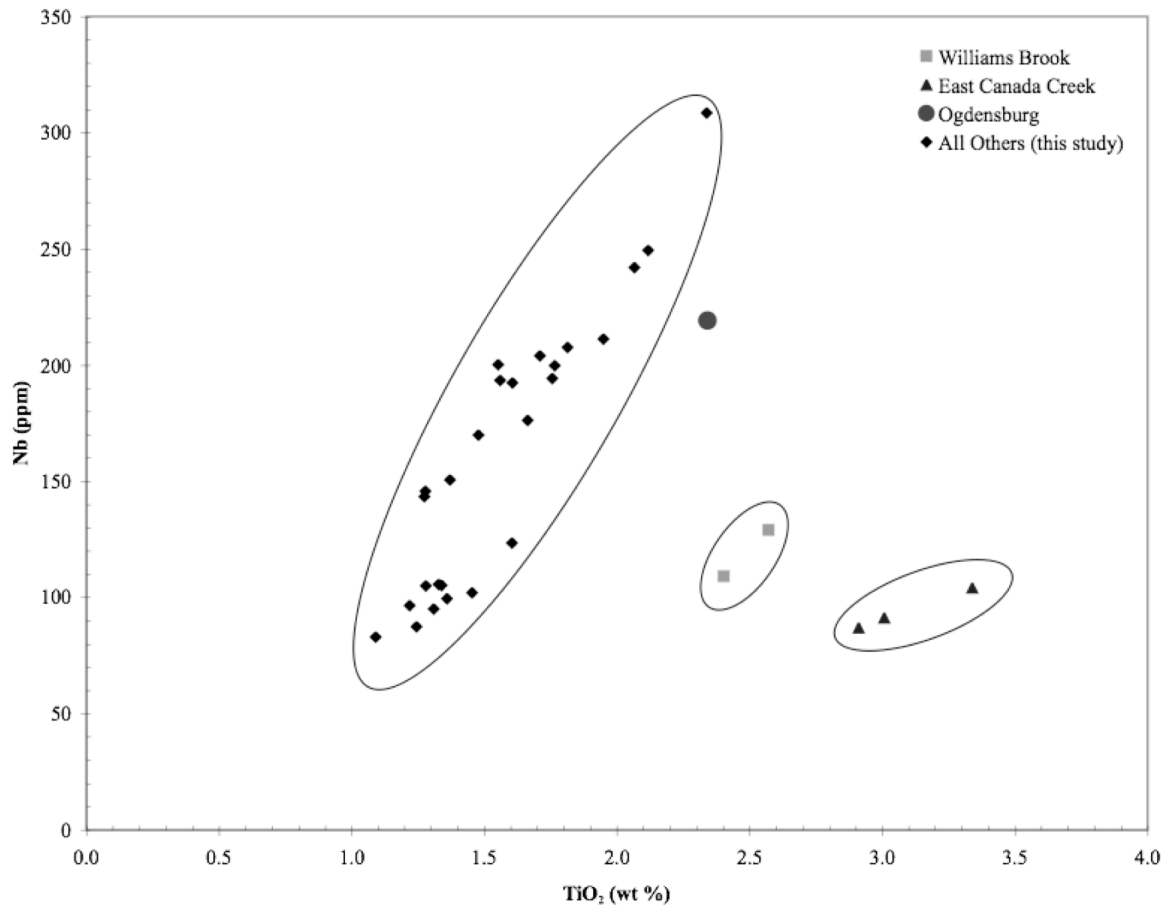
FIGURE 12—Nb (ppm) vs TiO$_2$ (wt. %) concentrations in kimberlitic rocks from New York State. Dikes from East Canada Creek and Williams Brook exhibit distinct HFS element ratios.

CLASSIFICATION: KIMBERLITES, ORANGEITES, OR LAMPROITES?

Are these unusal rocks really kimberlites? Over the years they have been referred to as "serpentine bodies" (Vanuxem, 1842), peridotites (Williams, 1887a), alnoites (Smyth, 1893), and kimberlites (Matson, 1905). For most igneous rocks, classification is now straightforward, based primarily on modal mineralogy, rock texture, and/or rock chemistry (Le Maitre et al., 2002). Unfortunately, due to the mineralogical complexity of "kimberlites", a simple definition does not exist; they are, in fact, a clan of complexly related rocks. The situation is nicely summarized by Winter (2001) who states: "The confusion (in classification) is most evident in the highly potassic lamprophyre-lamproite-kimberlite group, a diverse array of mafic to ultramafic rocks with high volatile contents. The numerous intertwined petrographic and genetic similarities and differences in this broad group present a classification nightmare." (p.362)

Kimberlites are currently divided (somewhat arbitrarily) into two groups (Smith et al., 1985; Skinner, 1989). Group I kimberlites are the analogue of the rocks originally found and described at Kimberley, South Africa (the "basaltic kimberlites" of Wagner, 1914). Group II kimberlites are the equivalent of the "micaceous kimberlites" of the Orange Free State, South Africa (or the "lamprophyric kimberlites" of Wagner (1914)). The two groups of kimberlites display subtle differences in their mineralogical composition (Smith et al., 1985; Skinner, 1989; Mitchell, 1995; Tainton & Browning, 1991).

The current formal definition for Group I kimberlites states that they "are volatile-rich (dominantly carbon dioxide) potassic, ultrabasic rocks commonly exhibiting a distinctive inequigranular texture resulting from the presence of macrocrysts (large crystals, typically 0.5–10 mm diameter) and, in some cases, megacrysts (larger crystals, typically 1–20 cm) set in a fine grained matrix" (Le Maitre et al., 2002). Some of the minerals from the macrocryst–megacryst association are mantle and crustal xenocrysts that were sampled and carried up by the kimberlitic fluid.

There is no formal definition at this moment for Group II kimberlites because they are less well studied. They were originally named "micaceous kimberlites" by Wagner (1914) and later "orangeites" (Wagner, 1928). According to Le Maitre et al. (2002), Group II kimberlites "belong to a clan of ultrapotassic, peralkaline volatile-rich (dominantly H_2O) rocks, characterized by phlogopite macrocrysts and microphenocrysts, together with groundmass micas." Mitchell (1986, 1995) argues that Group II kimberlites are not true kimberlites, but a distinct group of rocks. He suggests that they should be called "orangeites" to recognize their fundamentally different mineralogical character. According to Mitchell (1995; p.14) orangeites can be distinguished from kimberlites by "the absence of monticellite, magnesian ulvospinel, and Ba-rich micas belonging to the barian phlogopite-kinoshitalite series." In addition, orangeites can be distinguished from alnoites, lamprophyres and many other alkaline rocks by their lack of "melilite, alkali feldspar, plagioclase, kalsilite, or nepheline" (Mitchell, 1995; p.14). If these criteria are applied, then the central New York dikes would not be considered true (Group I) kimberlites; they would be classified as orangeiites. However, in terms of bulk chemical composition, the dikes often plot between the fields of kimberlites and orangeites (as defined by South African intrusions), with slightly more overlap with the Group I kimberlite field for most elements (Figure 13).

Lacking any widely accepted criteria for identifying and classifying this complex clan of igneous rocks, and the ambiguous petrographic and chemical features of the central New York dikes, we have chosen to refer to them as "kimberlitic" or "kimberlite-like", rather to call them kimberlites and imply that they are true analogues of group I kimberlites.
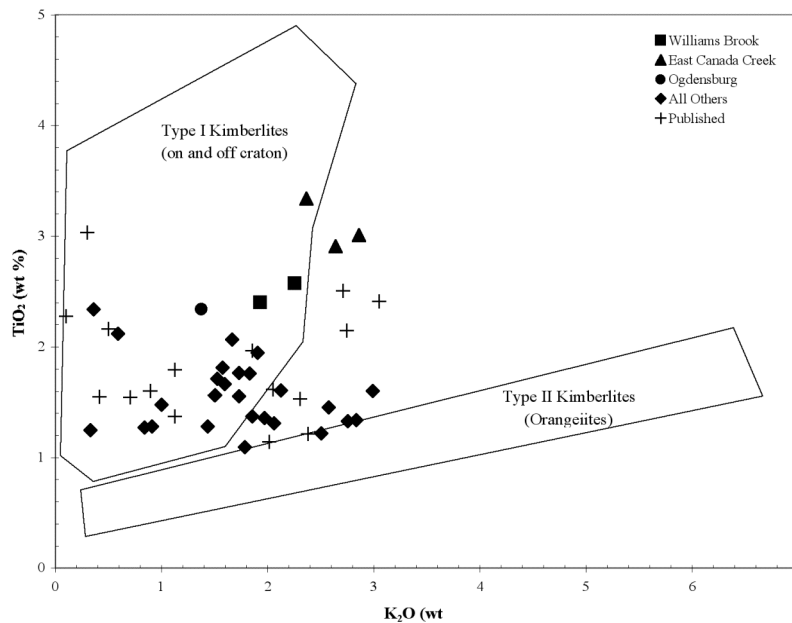
FIGURE 13—TiO_2 vs. K_2O concentrations in New York kimberlitic rocks. Fields for Group I and Group II kimberlites are from Mitchell (1995).

GEOGRAPHIC DISTRIBUTION AND AGES OF KIMBERLITIC ROCKS IN EASTERN NORTH AMERICA

Kimberlitic rocks in the eastern part of North America have been described from Tennessee (Safford, 1869; Gordon, 1927; Hall and Amick, 1944; Meyer, 1976), Kentucky (Diller 1885; Crandall 1887), Virginia (Sears and Gilbert, 1973), Pennsylvania (Kemp and Ross, 1907) and New York. These rocks are poorly exposed and occur as dikes, small diatremes, and pipe-like structures (Kentucky). The largest dike currently known in eastern North America is in Masontown, PA where it can be traced along strike for over 4 km.

These rocks were originally referred to as "mica peridotites", but mineralogically and texturally they can all be broadly classified as kimberlites. Dikes from all of the areas contain macrocrysts of olivine (usually highly serpentinized), phlogopite, pyrope garnet, and Cr-rich diopside, set in a strongly altered and calcite-rich groundmass. Distinct, mantle-derived ultramafic xenoliths are very rare; compositions of the observed macrocryst phases suggest many grains are from disaggregated garnet lherzolite (and possibly spinel lherzolite) xenoliths.

Even though there are many difficulties and uncertainties related to establishing the real age of the kimberlitic rocks of eastern North America, a compilation of the data published by Hall and Amick (1944), Zartman et al. (1967), Sears and Gilbert (1973), Pimentel et al. (1975) and Heaman and Kjarsgaard (2000) suggest a general trend toward progressively younger ages from south to north: Tennessee 225 to 350 Ma; Kentucky 257 to 275 Ma; Pennsylvania ~185 Ma; New York ~145 Ma. If valid, this apparent progressive age trend might be the continental expression of the northward migration of the opening of the Atlantic basin during this time (Taylor, 1984). A much more comprehensive study of all of the eastern North American dikes involving modern radiometric dating techniques is needed to test this hypothesis.

The kimberlitic rocks of New York are an important part of the record of Mesozoic tectonic and magmatic activity in eastern North America. They have fascinated and intrigued geologists since their discovery in 1837. Poor exposures along with extensive alteration and contamination have made unraveling their story an extremely difficult task. This report presents new whole-rock and mineral data that have made it possible to discern subtle differences between the clusters of dikes in central New York. At the completion of this study, we hope to be able to say a bit more about the age and origin of these unusual rocks.

ACKNOWLEDGEMENTS

REFERENCES CITED

Barnett, V.H., 1905, Notice of the discovery of a new dike at Ithaca, New York: American Journal of Science, v. 19, p. 210.

Basu, A.R., Rubury, E., Mehnert, H.H., and Tatsumoto, M., 1984, Sm-Nd, K-Ar and petrologic study of some kimberlites from eastern United States and their implication for mantle evolution: Contributions to Mineralogy and Petrology, v. 86, p. 35-44.

Broughton, J.G., 1950, Observations on the intrusion of rock salt by peridotite: Eos, Transactions American Geophysical Union, v. 31, p. 229-233.

Clement, C.R., 1982, A comparative geological study of some major kimberlite pipes in the Northern Cape and Orange Free State. University of Cape town, Cape Town.

Crandall, A.R., 1887, Report on the geology of Elliot County; also, notes on the trap dikes of Elliot County, Kentucky: Kentucky Geological Survey, v.13, 198 p.

Darton, N.H., and Kemp, J.F., 1895a, A new intrusive rock near Syracuse: Geological Society of America Bulletin, v. 6, p. 477-478.

—, 1895b, A newly discovered dike at DeWitt, near Syracuse, New York: American Journal of Science, v. 49, p. 456-462.

Dawson, J.B., 1980, Kimberlites and their xenoliths: New York, Springer-Verlag, 252 p.

Dawson, J.B., and Stephens, W.E., 1975, Statistical classification of garnets from kimberlites and associated xenoliths: Journal of Geology, v. 83, p. 589-607.

Diller, J.S., 1885, Dikes of peridotite cutting the Carboniferous rocks of Kentucky: Science, v.5, p,65.

Filmer, E.A., 1939, New peridotite dikes of Ithaca: Pan-American Geologist, v. 72, p. 207-214.

Foster, B.P., 1970, Study of the kimberlite-alnoite dikes in central New York (Finger Lakes region). SUNY at Buffalo, Buffalo, NY (Master's),

Gordon, C.H., 1927, Mica-peridotite dike in Union County, Tennessee: Geological Society of America Bulletin, v.28, p.125-216.

Hall, G.M., and Amick, H.C., 1944, Igneous rocks areas in the Norris Region, Tennessee: Journal of Geology, v.52, P.622-625.

Heaman, L.M., and Kjarsgaard, B.A., 2000, Timing of eastern North American kimberlite magmatism: continental extension of the Great Meteor hotspot track?: Earth and Planetary Science Letters, v.178, p.253-268.

Hogeboom, W.L., 1958, The petrology of the Green Street peridotite dikes in Syracuse, New York. [M.Sc. thesis]: Syracuse University, Syracuse, NY. 95p.

Hopkins, T.C., 1914, The geology of the Syracuse Quadrangle [New York]: Bulletin - New York State Museum, v. 171.

Hunt, T.S., 1858, Contributions to the History of Ophiolites. Part II.: American Journal of Science, v. 26, p. 234-240.

Ilupin, I.P., and Lutz, B.G., 1971, The chemical composition of kimberlite and questions on the origin of kimberlite magmas: Sovietskaya Geol., v. 6, p. 61-73.

Kay, S.M., and Foster, B.P., 1986, Kimberlites of the Finger Lakes region: New York State Geological Association, Field Trip Guidebook, v. 58, p. 219-238.

Kay, S.M., Snedden, W.T., Foster, B.P., and Kay, R.W., 1983, Upper mantle and crustal fragments in the Ithaca kimberlites: Journal of Geology, v. 91, p. 277-290.

Kemp, J.F., 1891, Peridotite dikes in the Portage sandstones near Ithaca, New York: American Journal of Science, v. 42, p. 410-412.

Kemp, J.F., and Ross, C.G., 1907, A peridotite dike in coal measures of southwestern Pennsylvania: New York Academy of Science Annals, v.17, p.509-518.

Kraus, E.H., 1904, A new exposure of serpentine at Syracuse, New York: American Geologist, v. 33, p. 330-332.

Le Maitre, R.W., Streckeisen, A., Zanettin, B., Le Bas, M.J., Bonin, B., Bateman, P., Bellieni, G., Dudek, A., Efremova, S., Keller, J., Lameyre, J., Sabine, P.A., Schmid, R., Sorensen, H., and Woolley, A.R., 2002,

Igneous rocks; a classification and glossary of terms; recommendations of the International Union of Geological Sciences Subcommission on the Systematics of Igneous Rocks Edition: 2: Cambridge, Cambridge University Press, 236 p.

Lewis, H.C., 1887, On diamantiferous peridotite and the genesis of diamond: Geological Magazine, v. 4, p. 22-24.

—, 1888, The matrix of diamond: Geological Magazine, v. 5, p. 129-131.

Martens, J.H.C., 1924, Igneous Rocks of Ithaca, New York, and Vicinity: Geological Society of America Bulletin, v. 35, p. 305-320.

Matson, G.C., 1905, Peridotite dikes near Ithaca, New York: Journal of Geology, v. 13, p. 264-275.

Maynard, J.E., and Ploger, L.W., 1946, A study of the Salt Springs Road peridotite dike in Syracuse, New York: American Mineralogist, v. 31, p. 471-485.

Meyer, H.O.A., 1976, Kimberlites of the continental United States: A review: Journal of Geology, v. 84, p. 377-402.

Miller, D.S., and Duddy, I.R., 1989, Early Cretaceous uplift and erosion of the northern Appalachian Basin, New York, based on apatite fission track analysis: Earth and Planetary Science Letters, v. 93, p. 35-49.

Mitchell, R.H., 1986, Kimberlites: Mineralogy, Geochemistry, and Petrology: New York, Plenum Press, 442 p.

—, 1995, Kimberlites, Orangeites, and Related Tocks: New York, Plenum Press, 410 p.

Newland, D.H., 1931, An occurrence of peridotite near Ogdensburg, New York: Bulletin - New York State Museum, v. 286, p. 113-117.

Pimentel, N., Bickerman, M., and Flint, N.K., 1975, A new K-Ar date on the Masontown dike, southwestern Pennsylvania: Pennsylvania Geologist, v.6, p.5-7.

Safford, J. M. (1869) Geology of Tennessee; Mercier, Printer to the State, Nashville.

Schneider, P.F., 1902, New exposures of eruptive dikes in Syracuse, N Y: American Journal of Science, v. 14, p. 24-25.

Schneider, P.F., 1903a, Notes on some eruptive dikes near Ithaca [New York]: Proceedings of the Onondaga Academy of Sciences, v. 1, p. 130-136.

—, 1903b, The geology of the serpentines of central New York: Proceedings of the Onondaga Academy of Sciences, v. 1, p. 110-117.

Sears, C.E., and Gilbert, M.C., 1973, Petrography of the Mt. Horeb Virginia kimberlite: Geological Society of America Program with Abstracts, v.5, p.434.

Sheldon, P.G., 1921, A new dike near Ithaca, New York: Science, New Series, v. 53, p. 20-21.

—, 1927, On the association of faulting with dike intrusion: Journal of Geology, v. 35, p. 353-367.

Skinner, E.M.W., 1989, Contrasting group I and group II kimberlite petrology; towards a genetic model for kimberlites, *in* Ross, J., Jaques, A.L., Ferguson, J., Green, D.H., O'Reilly, S.Y., Danchin, R.V., and Janse, A.J.A., eds., Fourth international kimberlite conference: Perth, Australia, Geological Society of Australia, Sydney, N.S.W., Australia, p. 528-544.

Smith, B., 1909, Dikes in the Hamilton shale near Clintonville Onondaga County, New York: Science, v. 30, p. 724.

Smith, B., 1931, Notes on the Clintonville dikes, Onondaga County, New York: Bulletin - New York State Museum, v. 286, p. 119-122.

Smith, C.B., Gurney, J.J., Skinner, E.M.W., Clement, C.R., and Ebrahim, N., 1985, Geochemical character of southern African kimberlites: a new approach based upon isotopic constraints: Transactions Geological Society of South Africa, v. 88, p. 267-280.

Smyth, C.H., 1892, A third occurrence of peridotite in central New York: American Journal of Science, v. 43, p. 322-327.

—, 1893, Alnoite containing an uncommon variety of melilite: American Journal of Science, v. 46, p. 104-107.

—, 1896, Note on recently discovered dikes of alnoite at Manheim, New York: American Journal of Science, v. 2, p. 290-292.

—, 1898, Weathering of alnoite in Manheim, New York: Geological Society of America Bulletin, v. 9, p. 257-268.

—, 1902, Petrography of recently discovered dikes in Syracuse, New York, with note on the presence of melilite in the Green Street dike: American Journal of Science, v. 14, p. 26-30.

Tainton, K., and Browning, P., 1991, The relationship between Group-2 kimberlites and lamproites; an example from the northern Cape Province, South Africa: European Union of Geosciences Terra Abstracts, v. 3, p. 17-18.

Taylor, L.A., 1984, Kimberlitic magmatism in the eastern United States: Relationship to mid-Atlantic tectonism: Proceedings of the Third International Kimberlite Conference, v.1, p.417-424.

Van Tyne, A.M., 1958, Petrology of the Euclid Avenue peridotite intrusive. [M.Sc. thesis]: Syracuse University, Syracuse, NY. 144p.

Vanuxem, L., 1837, First annual report of the geological survey of the third district, New York: Albany, 265 p.

—, 1839, Third annual report of the geological survey of the third district, New York, 283 p.

—, 1842, Geology of New York; Part III, Comprising the survey of the third geological district: Albany, NY, W. and A. White and J. Visscher, 306 p.

Wagner, P.A., 1914, The diamond fields of South Africa: Johannesburg.

—, 1928, The evidence of the kimberlite pipes on the constitution of the outer part of the Earth: South African Journal of Science, v. 25, p. 127-148.

Wells, J. W., 1961, Unpublished New York State bedrock geology maps: in Broughton, J. G., Fisher, D. W., Isachsen, Y. W., and Richard, L. V., eds., Geologic Map of New York, Finger Lakes Sheet: New York State Museum and Science Service Map and Chart Series #5.

Williams, G.H., 1887a, On the serpentine (peridotite) occurring in the Onondaga salt group at Syracuse, New York: American Journal of Science, v. 34, p. 137-145.

—, 1887b, On the serpentine of Syracuse, N.Y.: Science, v. 9, p. 232-233.

—, 1890a, Note on the eruptive origin of the Syracuse serpentine (with discussion by James Hall and J. F. Kemp): Geological Society of America Bulletin, v. 1, p. 533-534.

—, 1890b, Origin of serpentine of Syracuse, New York: American Geologist, v. 5, p. 118.

Winter, J.D., 2001, An Introduction to Igneous and Metamorphic Petrology: Upper Saddle River, NJ, Prentice Hall, 697 p.

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

ROAD LOG FOR FIELD TRIP A-3

KIMBERLITIC ROCKS OF NEW YORK

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Start at McDonald's parking lot on RT 13 just south of intersection with RT 281 on SW outskirts of Cortland.
		Follow Route 13 to Ithaca.
7.4	7.4	Turn right at light in village of Dryden. Stay on RT 13.
13.4	6.0	Turn left at light onto RT 366 (Dryden Rd).
17.4	4.0	Continue straight on Hoy Rd onto Cornell campus (RT 366 bears off to left)
17.7	0.3	Turn left onto Campus Rd.
18.2	0.5	Bear left onto Stewart Ave. Continue downhill on Stewart.
18.4	0.2	Turn right onto E. Buffalo St.
19.6	1.2	Intersection of RT 89 and RT 96. Continue straight on RT 96 north (Cliff St.)
28.7	9.1	Turn right onto Taughannock Park Rd. (Sign for Taughannock Falls State Park)
29.6	0.9	Stop 1. Pull off and park in gravel turn off on right side of road next to creek. Follow foot-path south down to stream bed.

STOP 1. TAUGHANNOCK FALLS STATE PARK

 The entire stream valley is part of Taughannock Falls State Park. No hammering or collecting is allowed

At least three small (2-8cm wide) dikes are exposed on the north bank of the creek underneath a small poplar tree, slightly west of the parking area. Foster (1970) identified and mapped out 10 dikes in the streambed over a distance of ~ 800m. Erosion and sampling over the past 30+ years has made it increasingly difficult to locate half of these dikes.

The dikes intrude the Devonian West River Shale (Genesee Group), and follow the prominent N-S joints. The tendencies of the dikes to occur in clusters, to contain xenoliths of the local country rock, and to pinch out along strike, are all displayed nicely here.

Petrographically, the Taughannock Creek dikes contain the best preserved macrocryst assemblage of all the NY kimberlites, with unaltered olivine, phlogopite, pyrope garnet, Cr-bearing diopside, and various spinels being relatively common. The groundmass of the dikes is composed of varying proportions of serpentine, calcite, phlogopite, perovskite, apatite and magnetite. Relative to the other central NY dikes, the groundmass of the Taughannock Creek dikes is relatively calcite-rich.

The bulk chemistry of the Taughannock dikes is quite variable, but they tend to have relatively low SiO_2 contents and relatively high CaO, P_2O_5, Sr and Zr concentrations compared to other NY state kimberlites (Table 2).

29.6	0.0	Turn around and return to RT 96.
30.5	0.9	Turn left onto RT 96 S (Cliff St.)
38.2	7.3	Stop 2. Pull off in parking area on right side of road just north of intersection with Hopkins Rd. Hike up stream ~ 50m. Dike is well exposed on north wall of creek bed.

STOP 2. Williams Brook

** Unusual and nicely exposed dike. Please do not hammer on outcrop. Collect only loose pieces found in streambed**

As noted by Kay & Foster (1986), the Williams Brook dike is the largest dike in the Cayuga lake region, with a width of ~3.7m. The contacts with the surrounding shale are not exposed along the eastern margin, and only poorly exposed on the western margin. Little to no thermal effects are visible.

The dike is petrographically and chemically distinct from all of the other Cayuga region dikes. It is very dark colored and dense, with abundant large (2-12mm) black serpentine pseudomorphs after olivine macrocrysts. Large phlogopite macrocrysts are also common, but in contrast to the Taughannock Creek dikes, no garnet, pyroxene, or spinel macroscrysts are observed. Some samples contain up to 10% unaltered olivine. The groundmass is similar to the other Cayuga area dikes, although perovskite and phlogopite are somewhat more abundant.

Chemically, the Williams Brook dike is unlike any of the other Ithaca region dikes (Table 2). It has a relatively high TiO_2 content, and relatively low Ba, Sr, and Zr concentrations. The bulk composition of the Williams Brook dike is most similar to the northernmost Ogdensburg dike.

38.2	0.0	Continue south on RT 96.
39.8	1.6	Intersection with RT 13N and RT 34 N. Turn left onto RT 34N (Meadow St.)
41.4	1.6	Take exit for RT 34N (Stewart Park / Auburn). Turn left at stop sign onto RT 34N.
47.1	5.7	Turn left onto RT 34B toward King Ferry.
47.7	0.6	Turn left onto Portland Point Rd.
48.1	0.4	Stop 3. Turn left into Hanson (South Lansing) rock quarry (just past rock conveyor that crosses over the roadway).

STOP 3. PORTLAND POINT QUARRY

Private Property. Contact Hanson Aggregates New York for permission to enter quarry

Two dikes were discovered in the Hanson Aggregates (former Portland Point) limestone quarry. One dike (5 to 7 cm wide) was found cutting the floor, and another, (12 to 20 cm wide) is exposed in the west wall of the quarry. Both strike N-S, have near vertical dips, and are strongly altered. They contain upper crustal xenoliths of both limestone and shale. Large serpentine pseudomorphs after olivine are common in some portions of the

dikes, but are absent in others. Pyrope garnets are very rare, but when present they contain unusually high amounts of chromium (5.77% Cr_2O_3). Calcite and serpentine are the most common minerals in the groundmass.

   In 1947 a dike cutting the halite in the underground works of the Cayuga Rock Salt Company (just below the Portland Point limestone quarry) was discovered. The dike, studied by Broughton (1950), occurred at a depth of 0.4 mile, and was exposed underground for 300 feet. The dike is an extension of one of the dikes exposed in the overlying limestone quarry, and had been predicted by the local mine geologists.

48.1	0.0	Leave quarry. Turn right onto Portland Point Rd.
48.5	0.4	Turn right onto RT 34B.
49.6	0.7	Junction with RT 34. Continue straight on RT 34 toward Genoa.
55.4	5.8	Turn right onto Locke Rd.
56.5	1.1	Bear left; stay on Locke Rd.
59.3	2.8	Turn right onto RT 90 at stop sign.
60.8	1.5	Village of Locke. Turn left onto RT 38 (Main St).
64.5	3.7	Village of Moravia. Turn right onto RT 38A north (E Cayuga St).
76.0	11.5	Continue straight on RT 359.
77.7	1.7	Turn left onto RT 41A (West Lake Rd.)
83.5	5.8	Turn right onto US 20 (West Genesee St.)
83.6	0.3	Village of Skaneateles. Continue east on US 20.
88.7	5.2	Turn left onto Williams Rd.
89.0	0.3	Stop 4. Park along side of road near Melody Meadows Farm

STOP 4. CLINTONVILLE

** Private property. Please ask permission to access ravine from property owner. Please do not hammer on outcrop. Collect only loose pieces found in streambed**

   The Clintonville dikes were discovered nearly 100 years ago (Smith, 1909). While only two dikes were originally identified, subsequent workers (Hopkins, 1914; Smith 1931) described a total of six dikes exposed over a distance of ~75m in the small ravine.

   The dikes range in width from 2 to 30 cm, and all are nearly vertical with a N6-12°E strike. The dikes cut through shales of the Devonian Skaneateles Formation (Hamilton Group). What is most striking about the Clintonville dikes is the dramatically different degree of weathering and alteration exhibited by immediately

adjacent dikes. Two of the largest dikes exposed on the southern wall of the ravine are only centimeters apart, but one is completely altered to a soft yellow clay/hydroxide mixture while the other remains fairly coherent and is composed primarily of serpentine and calcite.

The fresher dikes contain macrocrysts of phlogopite, and nice euhedral pseudomorphs of olivine, along with scarce macrocrysts of diopside and spinel.

The bulk composition of the least altered Clintonville dikes closely approximates the "average" composition of all of the NY kimberlites.

89.0	0.0	Turn around and return to US 20.
89.4	0.4	Turn left onto US 20 (east).
89.7	0.3	Turn left onto RT 174 (Sevier Rd.)
92.0	2.3	Turn right onto RT 175E / RT 174
93.5	1.5	Bear right and stay on RT 175E (W Seneca Turnpike)
101.7	8.2	Turn left and stay on RT 175E (South Avenue)
103.6	1.9	Turn right onto W. Kennedy St. / RT 175
104.2	0.6	Turn left onto S. Salina St / US 11
105.7	1.5	Turn right onto James St / RT 5
		(Note: City fountain and park will be on your left; large white Post-Standard Building will be in front of you just past the intersection).
106.4	0.7	Turn right onto Lodi St. (Note: Large Regency Towers Apt. building will be at on your right).
106.5	0.1	Turn left onto Green Street.
106.6	350 ft.	Stop 5. Park in parking lot on your left. Walk uphill to vacant lot immediately adjacent to parking lot. Blocks of kimberlitic material are exposed in the slope to the north.

STOP 5. GREEN STREET

Only a few blocks remain of what was probably the very first kimberlite discovered in New York State. Please do not hammer on, or remove, any of the remaining large pieces

The Green Street, or "Foot-street road serpentinite", was first described by Vanuxem in 1839, although it was reportedly discovered by Oren Root in 1837 (Williams, 1887a). The Green Street dike is part of a cluster of relatively small to moderate sized (5 cm up to 10 m wide) dikes (and sill offshoots?) that run NNW from Green Street in the south to Griffith Street in the north. Excavations in this area over the years encountered the serpentinite in at least five separate localities (see Figure 1b). The actual number, size, and orientation of intrusions are not known. All intrude Silurian dolostones and shales of the Syracuse Formation (Salina Group), and many contain abundant crustal xenoliths. Most xenoliths are from the local country rock, although lower crustal metamorphic rocks (gneisses and amphibolites) are also found.

As with all the kimberlitic rocks, samples of the Green St.–Griffith St. intrusions are quite variable in color, texture, and mineralogy. During the early excavations, it was reported that the local children collected "rubies" (pyrope garnets) and "emeralds" (Cr-bearing diopsides) up to 1 cm in diameter (Williams, 1887a). Other macrocrysts reported include phlogopite, olivine, orthopyroxene, clinopyroxene, feldspar, garnet and spinel (Hogeboom, 1958). We have only observed phlogopite, olivine and spinel macrocrysts in the samples currently available for study. The matrix is composed of phlogopite, serpentine, calcite, magnetite, apatite, and perovskite. Smyth (1902) reported the presence of melilite, but this has not been confirmed.

106.6	0.0	Turn right out of parking lot onto Green St.
106.65	0.05	Turn left onto Lodi St.
107.1	0.45	Turn left onto Erie Blvd East.
109.6	2.5	Turn right onto Thompson Rd.
109.9	0.3	Turn right onto Springfield Rd.
110.0	0.1	Stop 6. Turn left into LeMoyne College Physical Plant parking lot. Hike up hill to Dewitt reservoir. Blocks of kimberlitic material are common along the slopes of the reservoir.

STOP 6. DEWITT RESERVOIR

Private property. Contact LeMoyne College security for permission to hike on trails up to reservoir. Abundant material is available for collecting on the banks of the reservoir, although please leave the larger blocks intact for future field trips*

Other than the few small blocks on Green St., this is the only other kimberlitic material in the Syracuse region that is still exposed and accessible. The "dike" was discovered in 1894 by P. F. Schneider, and first described by Darton & Kemp (1895a.b). According to the contractor engaged in the initial excavation of the reservoir, blocks of the kimberlitic rock "occurred in masses imbedded in a greenish-yellow earth which underlaid the entire area of the excavation, which was about 200 by 250 feet" (Darton and Kemp, 1895a,b; p.456). Some of the individual blocks were reported to be 20' x 50', indicating that the original intrusion was a very large dike or possibly a diatreme or sill.

What is most striking about the DeWitt material is the abundance of dark, dense, relatively hard rocks. In fact, some of the "surplus" blocks excavated from the reservoir were "crushed and used for road material in DeWitt with satisfactory results" (Schneider, 1903b).

The freshest samples contain abundant, well-defined pseudomorphs of serpentine after olivine and rare macrocrysts of phlogopite and pyrope garnet. The matrix consists largely of phlogopite, serpentine, calcite, magnetite, and perovskite.

		To get onto the NY Thruway (I-90)
110.0	0.0	Leave parking lot. Turn right (east) onto Springfield Rd.
110.1	0.1	Turn left onto Thompson Rd.

112.7	2.6	At Carrier traffic circle, take 2nd exit for I-90 tollbooths (Exit 35).

To return to Cortland / SUNY Cortland

110.0	0.0	Leave parking lot. Turn right (east) onto Springfield Rd.
110.1	0.1	Turn left onto Thompson Rd.
110.6	0.5	Take the ramp onto I-690 W
113.6	3.0	Take the exit for I-81 S toward Binghamton. (Note: Exit will be on your left). Stay on I-81 S
144.3	30.7	Take Exit 12 (Homer / Cortland) off of I-81 S.
144.8	0.5	Follow signs for US-11 S / RT 41 S and merge onto US-11 / RT 41. Stay on US-11 / RT 41 into downtown Cortland.
146.9	2.1	Turn right onto Groton Ave / RT 222
147.3	0.4	SUNY Cortland Campus will be on your left (south).

END OF FIELD TRIP

Trip A-4

RE-EXAMINATION OF THE TYPE ITHACA FORMATION: CORRELATIONS WITH SECTIONS IN WESTERN NEW YORK

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INTRODUCTION

Stratigraphic correlation of the early Frasnian (Genesee Group) in New York has been ongoing for over 150 years (see Kirchgasser 1985 and Kirchgasser et al. 1994 for summary). While recent work has recognized key temporal horizons (e.g., sequence boundaries, condensed sections, and flooding surfaces) in western New York sections, correlation of such horizons into the vicinity of Ithaca, New York is currently ongoing (Kirchgasser 2000, Baird et al. 2006). This study focuses on detailed re-examination of the Ithaca Formation of east-central New York in its type area, which has not been studied in any detail since the early 1980's (see Kirchgasser 1985). We use sequence stratigraphic approach to describe and correlate these strata; a model which was previously unavailable to past researchers. Herein we summarize progress in characterizing small-scale sedimentary cycles within the Ithaca Formation, as well as the identification of horizons and/or beds that may be correlative with those previously recognized in western New York sections. We present a working hypothesis of high-resolution correlation within the Ithaca Formation. These include fossil-rich calcareous siltstone and shell-rich limestone beds that appear to record sediment starved intervals associated with both small- and large-scale transgressions. The latter may at least correlate with condensed, styliolinid-rich pelagic limestones in western New York.

Given that certain Genesee divisional units recognized in western New York (Schumacher Bed of the Penn Yan Formation, Genundewa Formation, "Huddle Bed" of conodonts and *Styliolina* in the basal West River Formation) have not yet been correlated with any certainty into the Ithaca region, and, given that certain Ithaca area units (Renwick Formation, Beebe Limestone, Williams Brook Limestone, as well as some newly described condensed intervals in the lower Ithaca Formation) have not been confidently linked into the stable stratigraphic succession of the Genesee Valley-Erie County region, the results presented here are a "work in progress". By establishing a "cat's cradle" of bed-to-bed matches between the various, large Ithaca gorge sections, we will be able to construct a hierarchy of units and beds that reflect eustatic changes and important bioevents. This linkage of local sections will also allow us to out-correlate marker beds both to the Genesee Valley area and to the east.

GEOLOGIC SETTING

The Devonian Appalachian Basin deposits of New York have long been known as a global reference section. The presence of a relatively complete stratigraphic succession, as well as an onshore-offshore gradient from red beds to black shales, is ideal for studies of stratigraphy, paleoecology, and sedimentology. These strata were deposited within a foreland basin resulting from the Acadian Orogeny; during which oblique convergence occurred between the Laurentia and Avalon terranes (Ettensohn 1985; Ver Straeten and Brett 1995, Ettensohn et al. 1988). Erosion of the collisional highlands produced the classic progradational complex known widely as the "Catskill Delta", which advanced in a generally westward direction and largely filled the foreland basin by the early Mississippian. More specifically, this trip focuses on a portion of the delta progradation (Ithaca Fm.) resulting from the creation of accommodation space following the third collisional tectophase of the Acadian orogeny which involved cratonward emplacement of overthrusts which depressed the lithosphere to enhance the foreland basin (see Ettensohn 1998).

This tectophase drastically changed both the geological and biological character of the basin. Following deposition of the Tully Limestone, orogenic activity caused major basin subsidence which was coincident with eustatic sea level rise (Johnson et al. 1985). This resulted initially in the deposition of the black Geneseo Shale, and its equivalents, over much of the Appalachian Basin. Furthermore, these basinal changes (and probably associated global warming) may have caused what is called the global Taghanic Bioevent, which resulted in the demise of Middle Devonian faunas worldwide (Aboussalam et al. 2001, Aboussalam 2003). Past observations suggest that this fauna returns anachronistically in the Appalachian Basin within the Ithaca Formation, and subsequently, in younger strata (Williams 1913). During the time of the deposition of the Geneseo Shale, the foreland basin was for the most part devoid of benthos, except for organisms adapted to a dysaerobic sea floor (*Pterochaenia, Buchiola*, rare chonetid brachiopods). This was episodically punctuated by times of slightly better bottom-water conditions, resulting in the deposition of the Fir Tree and the Lodi Limestones and their associated faunas (Baird et al. 1988). During and following this interval, progradation outpaced the rise of sea level, resulting in the deposition of the coarser-grained Sherburne and Ithaca Formations. Between the time of deposition of these formations, which are primarily comprised of turbiditic sequences, another sea-level deepening resulted in the deposition of the Renwick Shale. It is during the shallowing phase after the initial sea level rise associated with the Renwick Shale that the 'recurrent Hamilton Fauna' is first observed.

LOWER GENESEE GROUP STRATIGRAPHY

Background

Units seen on this trip were deposited as sediment along the delta-front during early Frasnian time. Correlation from the thin, basinal deposits of western New York into the delta-front area around Cayuga Lake has long been plagued by a variety of factors, including: misunderstanding of facies relationships (intertonguing), under-estimation of the thickening rate of these strata and within bed lithological variation, and also a case of biostratigraphic misidentification (see Kirchgasser 1985 and Williams 1951 for a complete summary). In addition, eastern sections are physically prohibitive; the 50-fold eastward thickening of the Genesee succession from Lake Erie to Ithaca and the sheer size of waterfalls and cliff walls in the eastern sections renders any attempts at regional section matching difficult, and access to the entire succession somewhat problematic.

Chadwick (1933, 1935) was the first to realize that the stratigraphic sequence of Genesee, Portage, and Chemung was also seen in the complex intertonguing of facies representing an offshore to onshore gradient (Genesee to Portage to Chemung facies). Around this same time, Caster (1933, 1933a) focusing on Ithaca area sections, attempted to divide the Ithaca Formation into traceable units as will be explained below. G.Q. Williams (1951) further expanded upon this work. Subsequently, the tracing of the Middlesex and Rhinestreet black shales into the area of Cayuga Lake by Sutton (1959, 1963), Sutton et al. (1962), and deWitt and Colton (1959, 1978) correctly placed both the Sherburne and Ithaca Formations within the Genesee Group (also see Rickard, 1964, 1975, 1981). Following this, examination and revision of Genesee Group ammonoids by House and Kirchgasser (see Kirchgasser and House 1981, Kirchgasser 1985, House and Kirchgasser 1993, Kirchgasser 2000, and references therein) began the process of building a high-resolution biostratigraphic correlational framework for the Genesee Group. Specifically, the location of the "Linden Horizon" yielding the ammonoid

Koenenites styliophilus in the Cayuga Lake section provides a datum from which other marker beds in the western section can be located in the Cayuga Valley (Kirchgasser, 1985). Also, Baird et al. (1988) correlated the Fir Tree and Lodi horizons from the west into the vicinity of Cayuga Lake and beyond, further aiding in high-resolution correlation of the Genesee Group across New York. Huddle (1981) attempted biostratigraphic correlation using conodonts, but his efforts were focused mostly in western New York and likely need some taxonomic revision in order to be applied to the more recent zonations provided by Klapper and Johnson (1990, also, see references therein). Most recent summaries of the information provided above can be found in Kirchgasser (2000), Baird et al. (2006), and other references therein.

Pre-Ithaca Formation Strata

Pre-Renwick Units.—Underlying the Renwick Formation in the study area are the Geneseo and Sherburne Formations. The Geneseo Formation as described above is an anoxic black shale, deposited as part of the Taghanic Onlap event representing a substantial eustatic sea level rise (Johnson 1970, Baird and Brett 2003). Brief excursions of more oxic conditions resulted in the deposition of the Auloporid-rich Lodi and Fir Tree Limestone submembers (Baird et al. 1988). Beginning around Seneca Lake, and continuing eastward, the upper portion of the Geneseo Formation (including the Fir Tree and Lodi Limestone submembers) begins to interfinger with the turbiditic siltstones facies of the Sherburne Formation (Vanuxem 1840), and in the vicinity of Ithaca, the shales of the Penn Yan Formation have completely lost their identity to such a facies transition. Recent mapping has identified a condensed horizon above the Fir Tree and Lodi Limestone submembers within the Sherburne Formation (Figs. 1 and 2), but further work is needed to determine if this bed is even traceable. If this bed can be traced, this would allow for differentiation of the Sherburne Formation into smaller units. In this trip we will give a brief overview of sub-Ithaca units, and then focus on the strata above the Sherburne Formation, beginning with the first known traceable unit above the Lodi Limestone submember, the *Warrenella* Zone of the basal Renwick Formation.

Renwick Formation.—The name Renwick was originally proposed by Caster (1933) in an abstract and later in a fieldtrip guidebook (1933a), but no specific type locality was listed. Included in Caster's (1933a) Renwick is the "Ithaca *Lingula* Shale" of Williams (1906), which is also the "*Lingula complanatum* Zone" of Williams et al. (1909). G.Q. Williams (1951) described this interval as having a stratigraphic range of about 30 meters, but gave only an arbitrary lithological upper boundary. As a lower boundary, Williams (1951) used the base of the *Warrenella* Zone of Williams (1884), Kindle (1896, 1906), and Williams et al. (1909), thereby including it in the Ithaca Formation. Previously this was equivalent to the Cornell Member of the Sherburne Formation as proposed by Smith (1935). The Renwick was later named for exposures along Renwick Brook, to the northeast of Ithaca (deWitt and Colton 1959, 1978). In doing so, deWitt and Colton (1959, 1978) also removed the Renwick from the Ithaca, giving it 'equal' stratigraphic status. As defined here, the Renwick Formation begins at an abrupt contact of dark shale with the underlying siltstones of the Sherburne Formation; near the base of the *Warrenella* (*Spirifer laevis*) Zone. While the faunal boundary is likely to be facies-related and therefore diachronous, further detailed mapping of the Sherburne-Renwick contact interval is necessary so that a suitable, isochronous boundary can be given for the base of the Renwick.

The top of the Renwick has been defined previously, first, as the contact with the Sixmile Creek member of Caster (1933); however this name is not used in Caster (1933a) or subsequently by other researchers, and, second lithologically, at the top of the youngest siltstone-filled channel in the sequence of dark shale and intercalated siltstone that composes the Renwick, by Williams (1951), deWitt and Colton (1978), and Grasso et al. (1986). We place the top of the Renwick Formation much higher, at the base of the Ithaca Falls limestone submember of the Ithaca Formation (Fig. 1, see below), and thereby abandoning any use of the name Sixmile Creek member. Within the Ithaca area, the Renwick Formation varies in thickness from 30 to 45 meters. Recent measurement of the section at Renwick Brook failed to locate the Ithaca Falls limestone with confidence; this gully has many covered intervals due to the numerous roads that cross it as well as a number of man-made retaining walls. Further work is necessary in order to find the Ithaca Falls limestone at this locality with certainty; or determine a new type section for the Renwick Formation as defined herein.

Further mapping of the Renwick Formation should allow division of this unit into different members. Some possible divisions include: the Cornell Shale of Smith (1935) (*Warrenella* Zone of Williams (1884), Kindle (1896, 1906), and Williams et al. (1909); the "Ithaca *Lingula* Shale" of Williams (1906) (also the "*Lingula complanatum* Zone" of Williams et al. (1909)); and a separate division for the upper, more siltstone-rich unit.
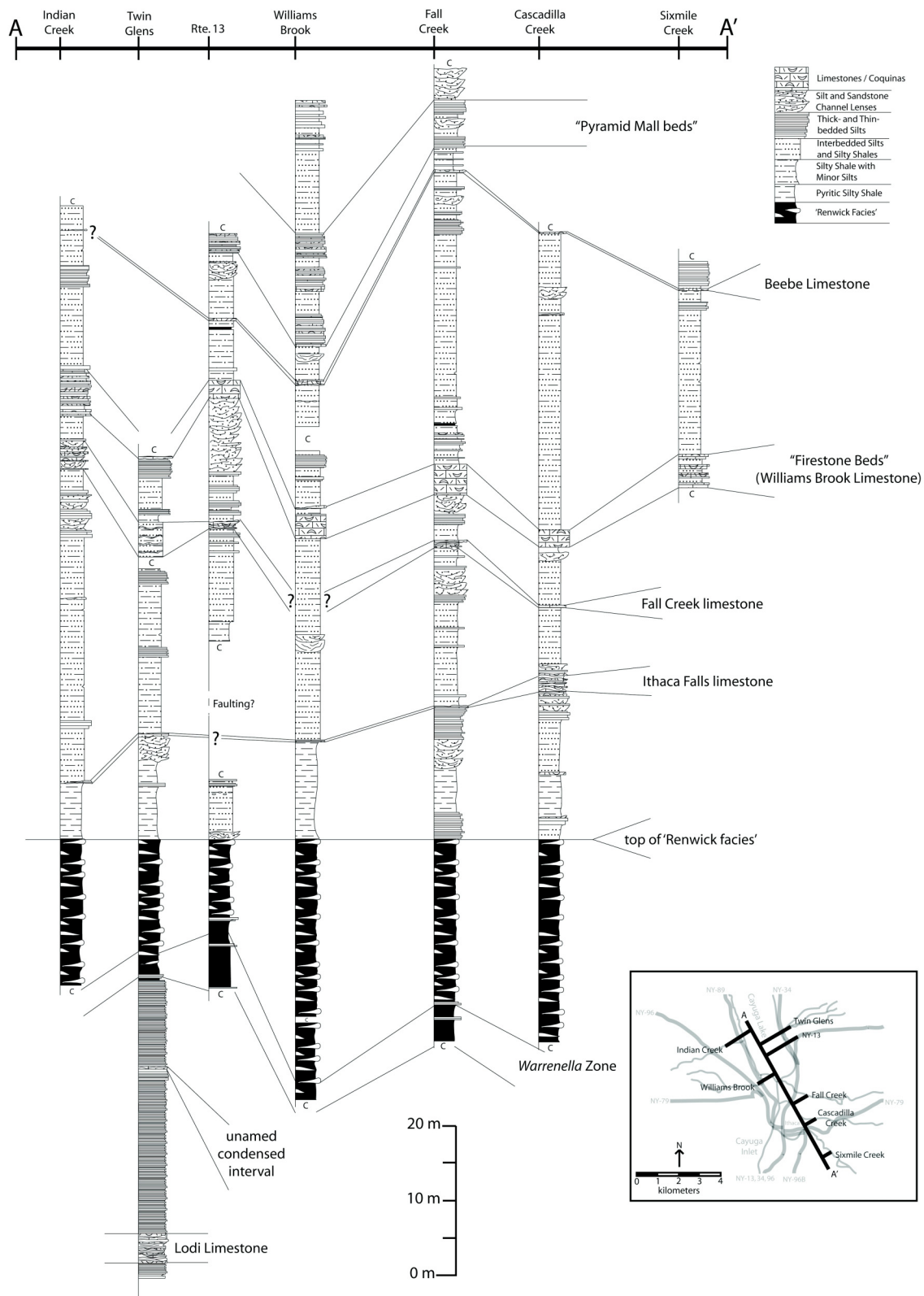
FIGURE 1—Stratigraphic correlation of various sections in the vicinity of Ithaca, New York; transect follows inset map. Stratigraphic nomenclature is discussed in text.
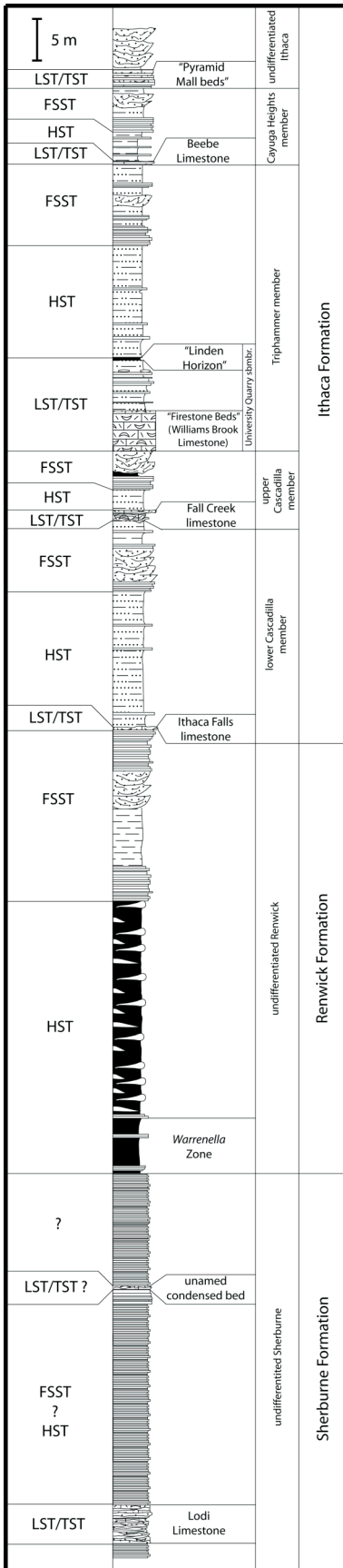
The term 'Renwick facies', is used herein to refer to the dark, thin-bedded shales and silty shales with interbedded siltstone channel lenses, which overlies the *Warrenella* Zone. This consists of the lowest portion of the Renwick Formation, comprised of the "*Lingula* Shales" of Williams (1906) and the numerous siltstone channel fills that locally cap many small cascades, and will be visible at STOPS 1B, 2, 3, and the optional stop. The 'Renwick facies' varies in thickness within the study area from 20 to 35 meters. This term is useful because it refers to the interval of the Renwick Formation (as defined here) that is equivalent to the Renwick of past researchers (Williams 1951, deWitt and Colton 1978, and Grasso et al. 1986). These channel fills commonly have basal lag deposits and exhibit soft-sediment deformation. These deformation features are typically seen at the base of siltstone channel complexes. In order for such channels to form, it is presumed that there was initial scouring into the sea floor, such that the base of the channel would be in a semi-compacted, or firm, mud. If this was the case, then load casts and ball and pillow structures by density inversion alone may not be sufficient to form these features. Some form of liquefaction of the firm, muddy base of the channels must have occurred; seismic activity would be one such possible trigger for liquefaction of thixotropic mud.

The Renwick Formation has a number of distinct faunal zones. The fauna of the *Warrenella* Zone includes: *Warrenella laevis* (with *Vermiformichnus* borings)*, Cyrtina hamiltonensis, Arcuaminetes scitulus, Sinochonetes lepidus, Lunulicardium ornatum, Paleoneilo filosa, P. constricta, Pseudoaviculopecten* sp., *Modiomorpha sublata, Grammysioidea subarcuata, Nuculoidea* sp., *Glyptotomaria capillaria, Eutaxocrinus* sp.*, Ponticeras perlatum,* Orthocones, plant and wood fragments. *Warrenella laevis* is an immigrant brachiopod from the Old World Realm fauna of western North America. See Williams (1884) and Kindle (1896) for a more complete species list. This interval will be accessible for collection at STOP 2 and the optional stop.

←

FIGURE 2—Sequence stratigraphic interpretation of the strata as discussed in text. This is a composite section of strata measured at the Fall Creek, Twin Glens, and NY-13 localities.

The fauna of the 'Renwick facies' is rather similar to the shales below, with the addition of *Plumulina plumaria, Cupularostrum* sp.*, Phthonia* sp., *Cypricardella* sp., *Paleozygopleura* sp., various species of *Lingula* (hence the name "*Lingula* shales"), *Productella* sp., 'Leiorhynchids', and the absence of *W. laevis*. From the top of the 'Renwick facies' upward to the base of the Ithaca Falls limestone, the remainder of the Renwick Formation is a shallowing upward succession of rusty-weathering pyritic silty shales, thin- and thick-bedded siltstones, and silt and sandstone channel lenses displaying deformation and possibly altered volcanic ash beds (bentonites). This is the position of the 'Recurrent Hamilton' fauna as proposed originally by Williams (1884) and subsequently defined in greater detail (Williams 1913 and references therein). The fauna of the uppermost Renwick Formation is similar to the '*Lingula*' shales below, with the addition of *Ambocoelia umbonata, Elyta fimbriata*, "*Mediospirifer*" *angusta, Tylothyris mesacostales,* and *Rhipidomella* sp. (Williams 1884). Although more work is necessary, this may be interpreted as a slightly more aerobic version of the fauna of the '*Lingula*' shales and both faunas may be considered 'Hamilton-like'; the recurrence of the Middle Devonian 'Hamilton fauna' within the Frasnian deposits of the Finger Lakes Region is currently being investigated as part of the dissertation research of JZ.

Stratigraphy of the lower Ithaca Formation in the vicinity of Ithaca

Accounts in the past literature of the Ithaca Formation contain observations of a prograding delta into the basin, the introduction of the Ithaca fauna into the area of Cayuga Valley, and the observation of a variety of coquinitic lenses that were thought to be only local, and therefore could not be correlated between sections (Williams 1884, Kindle 1896, deWitt and Colton 1978, W.H. Hass in Huddle 1981). Recent mapping during the last two field seasons, however, has identified a number of condensed horizons that can be traced around the area of Ithaca. We present the current status of our correlations below and propose an informal stratigraphic nomenclature based on our correlations using a sequence stratigraphic approach (Figures 1 and 2). The benefit of using this approach to delineate boundaries between sedimentary sequences is that it places unit boundaries at the contacts of traceable, and presumably isochronous beds, rather than at an arbitrary lithological change. In particular, we have traced these condensed, fossiliferous units around the Ithaca area. Such beds are interpreted as being deposited during transgressions, when sediment supply to the basin is presumed to be sequestered onshore, allowing for the accumulation of time-rich, fossiliferous beds. Although most sediment was sequestered in such a way during these times, variation within the amount of condensation seen within traceable beds argues that sediment influx was still occurring and may have been locally sporadic (see below). It is believed by the current authors that this variability, probably in conjunction with shifting depocenters, is what led to problems with attempts of earlier workers to produce a high-resolution stratigraphic correlation. It was the recognition of this variability that allows us to produce such correlations within the Cayuga Valley. Our correlations are further supported by stratigraphic, faunal, and biostratigraphic evidence.

We intend for these names to be informal and serve as a guide for correlating mappable units into the sections of western New York, whereby possible correlations with western marker beds would allow for correct labeling of the tentative, 'working-hypothesis', marker beds presented here. The list of stratigraphic nomenclature within the Ithaca Formation has a long and varied history. Whenever possible we attempt to use the first name given to a unit, provided 1) the name represents a meaningful geographic location, and 2) the unit fits reasonably well into a sequence stratigraphic framework that allows high-resolution correlation of marker beds. We also use the informal subdivision of "submember" to denote intervals, up to several meters in thickness, of distinctive lithology. This is particularly advantageous in the current study, in that condensed units (which are the units that are traceable) are typically intervals of strata, rather than discrete "beds". The following will proceed from the oldest to the youngest beds found in the lower Ithaca Formation.

Lower Cascadilla Member.—The name Cascadilla was originally used by Caster (1933, 1933a) presumably for the exposures along Cascadilla Creek, but no boundaries for this unit or a type section were given. G.Q. Williams (1951) subsequently defined this unit from the top of his Renwick to a level above the Williams Brook Limestone (see below).The lower Cascadilla member as defined herein commences with the Ithaca Falls limestone submember and is marked at the top by the base of the Fall Creek limestone submember. This interval is well exposed along the trail at Cascadilla Creek between Linn Street and College Avenue. This sequence is similar to the Renwick Formation, but lacks the 'Renwick facies'. This is also the *Paracyclas lirata* Zone of Williams et al. (1909), another 'Hamilton' fossil. The Ithaca Falls limestone has been traced around Cayuga Lake and serves as the first distinctive marker bed above the *Warrenella* Zone (Figure 1). This bed is best developed in the cliff face above the lip of Ithaca Falls, for which it is named, and also in the southern-most

gully of Twin Glens (STOP 4). This bed is typically a composite of fossil hash lenses, displaying a sharp erosional base, cross-bedding, interbedded turbiditic siltstones, reworked turbiditic siltstones; varying in thickness from a few to 34 cm. At Twin Glens these hash lenses are even seen to erode down through, and laterally truncate, a 10 cm thick turbidite. In more proximal (southeastern) localities, this bed is dominated by *Paleoneilo* sp., Nuculid clams, *Stictoptera meeki, T. mesacostales*, and abundant crinoid material. This bed commonly yields undeformed clams and gastropods with calcite replaced shells. At Twin Glens and Fall Creek there is a thin crinoidal hash layer found approximately at 40 and 20 cm, respectively, above this bed that has yielded articulated crinoids at Fall Creek. At more distal localities, in particular the section at Indian Glen, this bed yields an abundance of rhynchonellid brachiopods including forms exhibiting the characteristics of *Cupularostrum sappho, C. exima, C. contracta, Leiorhynchus quadricostatus,* cf. *Cherryvalleyrostrum limitare* (*"Leiorhynchus" limitare*)*, C. mesacostalis,* and *Caryorhynchus globuliformes.* The overlying portions of the lower Cascadilla member include shallowing upward interbedded silty shales and thin siltstones, lithologically and faunally similar to the upper portion of the Renwick Formation. This interval ranges in thickness from 20 to 45 meters.

Upper Cascadilla Member.—The interval herein termed upper Cascadilla member was included within the previously recognized Cascadilla member. The observation of a mappable fossiliferous interval within the Cascadilla member of older terminology, probably representing the transgressive interval of sea level change as discussed below, enables refinement and defines the lower/upper Cascadilla member boundary. Like the lower Cascadilla member, the upper Cascadilla member is well exposed along the trail at Cascadilla Creek downstream of the College Ave. Bridge. The base of this member is marked by the Fall Creek limestone submember, a newly recognized mappable unit within the Ithaca vicinity. This unit has been observed so far to be best developed on Fall Creek at the lip of the falls downstream from the old Cornell University Power Plant. These beds can be traced to slightly below the strata on which the Power Plant foundation is built. The top of this member is marked by the base of the "Firestone Beds" (see below). On this trip, these beds can be seen at STOP 5. The upper Cascadilla member ranges in thickness from 8 to 17 meters.

The Fall Creek limestone shows more variation in thickness than the Ithaca Falls limestone between localities. Faunally, these beds are extremely similar. These beds, when best developed as seen at Fall Creek, and at STOP 4, are a series of closely spaced coquinitic lenses. While not as sharp a contrast to the surrounding lithology as seen with the Ithaca Falls limestone, the Fall Creek limestone does stands out well relative to the relatively unfossiliferous strata found between the condensed marker beds presented here.

Faunally, the strata of the upper Cascadilla member are very similar to those below, with the addition of a small morph of ?*Orthospirifer.* The sequence of the upper Cascadilla member is typically similar to that of the lower Cascadilla member: silty shales with few interbedded siltstones coarsening upward to more abundant silts and finally, where present, packages of siltstone and sandstone channel lenses.

Triphammer Member.—The term Triphammer member was first presented by Caster (1933, 1933a) presumably for exposures at Triphammer Falls on Fall Creek, yet no upper or lower bounds were specified. G.Q. Williams (1951) places this interval from a level above the Williams Brook Limestone (see below) to the base of the Enfield; an older stratigaphic name that included what is not only currently part of the Ithaca Formation but also the Middlesex Formation, and also other lower units within the Sonyea Group. This is the most studied and referred to unit in the Ithaca Formation, primarily due to this being the first incursion of the 'Ithaca fauna' into the area of study. Herein the Triphammer member is designated as beginning at the base of the University Quarry submember, an interval first described by Williams (1884). This interval was named such because it was formerly quarried by Cornell University for building stone (circa 1890); this abandoned quarry is thought to be somewhere along the edge of the Fall Creek Gorge (W.T. Kirchgasser, pers. comm., 2006). Williams et al. (1909) also identified the interval of the University Quarry at: Fall Creek (behind the Cornell University Power Plant), Cascadilla Creek (approximately half-way up the gorge), Williams Brook (between 600 and 650 feet elevation), as well as at the quarries along Quarry St., and the old McCormick Quarry. At Williams Brook, the Williams Brook Limestone of Caster (1933, 1933a; see also deWitt and Colton 1978) occurs at this elevation. The upper portion of the University Quarry interval, Caster (1933, 1933a) had called the Marathon Sandstone, presumably for an outcrop near Marathon, east of Ithaca. This is also likely the '*Spirifer mesastrialis* Zone' and the '*mesastrialis* sands' of Williams et al. (1909) and Williams (1913). The lower portion of the University Quarry interval was termed the "Firestone Beds" by Kirchgasser (1985), for an interval that is more fossiliferous, and therefore more calcareous, than the rest of the University Quarry interval. This term was used informally by Williams and Kindle to describe these beds and others like them, as the

calcareous nature of these layers made them resistant to breakage during heating (Williams 1884, Kindle, 1896). Kirchgasser also identified the "Linden Horizon" near the top of the University Quarry interval. This is described as a sharp contact (flooding surface) on which the goniatite *Koenenites styliophilus* is abundantly concentrated. The University Quarry interval shows the first appearance of this goniatite in this succession and a key biostratigraphic marker bed, allowing correlation between the Cayuga and western sections including the type section of the horizon at Linden, NY, south of the City of Batavia in Genesee County. At its westernmost localities, the "Linden Horizon" is a condensed bed of styliolinid limestone abounding in the goniatite *Koenenites styliophilus*; it splays eastward across the Wyoming Valley-Honeoye Valley region into a thicker bundle of thin limestone beds yielding variable amounts of *Styliolina* and *Koenenites*.

Subsequently, Baird et al. (2006) mapped the Genesee Group in the area surrounding Linden, NY, and his observations suggest the possibility that Kirchgasser's " Linden Horizon" on Fall Creek may correspond to the erosive/corrosive flooding surface discontinuity that both caps and oversteps the condensed "Linden Interval" in the type Linden area (see also Kirchgasser and House 1981). Given that the "Linden Interval" thickens eastward, it appears likely that the University Quarry submember, including the "Firestone Beds" succession may be a greatly thickened eastern equivalent of the "Linden Interval" itself. Discovery of a goniatite specimen within the Caster Collections at the University of Cincinnati by one of the authors (JZ), which was subsequently identified as *Koenenites styliophilus* (Kirchgasser, pers. comm., 2007), further substantiates the possibility that *Koenenites* enters the succession during the time represented by the condensed interval ("Firestone Beds"), later to become concentrated on the sediment-starved flooding surface known as the "Linden Horizon". This specimen was labeled as "cascade at Power House Falls", which would place it below Kirchgasser's "Linden Horizon" on Fall Creek. This would indicate that the styliolinid grainstones commonly observed within the condensed "Linden Interval" in western sections may be correlative with the "Firestone Beds" in the Cayuga Lake section. We herein use the term "Linden Horizon" to represent the overlying, related flooding surface discontinuity (see below). The University Quarry submember is therefore a key marker interval that helps to put the strata above and below into the proper relationship to possibly correlative beds in western sections.

While the entire transgressive interval ("Firestone Beds" and "Linden Horizon") is included within the University Quarry interval, we find it useful to keep these names as mappable marker units. This is particularly important in westernmost New York where the post-"Linden Interval" corrosional discontinuity surface, roughly corresponding to Kirchgasser's "Linden Horizon" on Fall Creek, cuts out the "Linden Interval" (= "Firestone Beds"-equivalent succession) at all localities west of Murder Creek (Baird et al. 2006). Conversely, in eastern sections near Ithaca, the "Firestone Beds" appear to be more mappable than the "Linden Horizon" (Fig. 1). Furthermore, though the Williams Brook Limestone, which correlates to the "Firestone Beds" interval at other Ithaca area localities, is an impressive, 3+ meters-thick, deposit of shell-rich, pelmatozoan encrinite that shows internal channelization and cross-bedding, it is atypical compared to all other area sections. Hence, the name Williams Brook Limestone is herein used as a local name to represent this unique facies of the "Firestone Beds". Typically, the "Firestone Beds" are comprised of a complex of up to 4 meters of fossil-rich siltstones and coquina lenses.

In summary, we mark the base of the Triphammer member with the University Quarry submember (comprised of the "Firestone Beds" and "Linden Horizon"), as the name University Quarry apparently supersedes Marathon Sandstone, *S. mesastrialis* zone, '*Mesastrialis Sands*', "Linden Horizon", and "Firestone Beds". We mark the top of the Triphammer member with the base of the Beebe Limestone (Caster 1933, 1933a; deWitt and Colton 1978; see below). This bed was in the upper half of Caster's Triphammer member as originally defined. The thickness of the Triphammer member, as described herein, varies greatly from 10 to 40 meters.

The University Quarry interval marks the first appearance of the 'Ithaca Fauna' in this succession, characterized by such forms as *Orthospirifer mesastrialis*, *Cranaena eudora, Productella speciosa, Schizophoria impressa, Pseudoatrypa* cf. *devoniana*, "*Pugnoides*", Heterophrentid corals, *Platyceras* sp., *Koenenites* sp., and abundant disarticulated crinoid material. Some more 'Hamilton fauna' taxa also enter at these beds, including *Schuchertella* (cf. *Protoleptostrohpia*) and *Spinatrypa spinosa*. Fossils commonly show some pyritization in these beds. The strata from the University Quarry interval to the top of the Triphammer member, herein delineated by the base of the Beebe Limestone, are thickest in the section in the southeast of the study area, which agrees with deWitt and Colton (1978) and indicates that the source of sediment was from that direction (Fig. 1). At the section in Fall Creek, this interval is dominated by taxa of the 'Ithaca fauna' and

shallows upward from silty shales and interbedded siltstones and silty shales, to heavier-bedded siltstones and eventually silt and sandstone channel deposits.

Cayuga Heights Member.—The Cayuga Heights member, so named for the exposures along the cloverleaf interchange of NY-13 and Cayuga Heights Rd. to the northeast of Ithaca, is a relatively thin sequence compared to those described previously. This interval will be observed at STOPS 5 and 7A. Although not as thick as lower members, the same succession from condensed, fossiliferous strata through silty shales, to thick-bedded siltstones and sandstones is still observed. The base of this unit is the Beebe Limestone of Caster (1933, 1933a), which has been correlated across the study area. The Beebe Limestone contains elements of the 'Ithaca fauna', similar to the "Firestone Beds". Like the "Firestone" Beds, the Beebe Limestone also displays dramatic thickness and lithological variations over small distances (discussed below). This bed was stated as occurring at about 40 to 45 meters below the Enfield in the Ithaca succession by Caster (1933, 1933a). Interestingly, based on Caster's description, the base of the Enfield is the equivalent of our "Pyramid Mall beds" (see below); 40 to 45 meters below this is about the level of the "Firestone Beds". The Beebe Limestone was not located by Williams (1951) in his study of the type Ithaca Formation, even after talking with Caster himself. deWitt and Colton (1978) describe the Beebe Limestone as occurring in the cliffs of Fall Creek west of Beebe Lake. Huddle (1981) describes the bed as occurring in the top of the Fall Creek Gorge near a bridge (possibly the Thurston Avenue bridge or the nearby foot bridge), as well as in float blocks at the bottom of the gorge. Both of these bridges are at least 35 meters above the "Firestone Beds". The present authors have located the float blocks, but have been unable to locate the Beebe Limestone in place west of Beebe Lake. There is no visible coquina near the footbridge, which is stratigraphically lower than the Beebe Limestone as we have defined it (see below). The base of the Thurston Avenue bridge is currently inaccessible due to ongoing construction, but its elevation is about where we would project the Beebe Limestone based on upstream information.

We have located what we understand to be the Beebe Limestone on the inlet gorge to Beebe Lake (northeast side) in place (STOP 6). While the float pieces within the gorge are a dense coquina with reworked phosphate pebbles and rip-up clasts, the Beebe Limestone on the far side of the Lake is a fossiliferous calcisiltite. We have observed both of these lithofacies, as well as intermediates, at similar stratigraphic levels throughout the study area. We discuss this variation, and that of the "Firestone Beds", below. Above the Beebe Limestone, *Warrenella laevis* has been reported by Kindle (1894) around Forest Home, on Fall Creek. Concurrent with the reappearance of *Warrenella* is a short return to deeper water deposits ('Renwick facies'), yet the 'Ithaca fauna' appears to persist above this. Above this, the section again shallows upward to siltstone and sandstone channel deposits (base Enfield of older literature). The Cayuga Heights member extends from the Beebe Limestone to the base of the "Pyramid Mall beds" (see below), attaining a thickness of approximately 5 meters.

Undifferentiated Upper Ithaca Formation.—The "Pyramid Mall beds", named for the exposure along NY-13 northbound next to the exit sign for Pyramid Mall and Triphammer Road, marks the base of the next heretofore undifferentiated member (STOP 5). This series of beds has also been located on Fall Creek upstream of the Caldwell Road Bridge in Forest Home, and at approximately 25 meters above the Williams Brook Limestone ("Firestone Beds") at Williams Brook. This bed consists of thin- to thick-bedded fossiliferous siltstones and sandstones, with a thickness of up to 8 meters (up to about 15 meters at Williams Brook). Crinoid material, *Schuchertella* sp., cf. *Protoleptostrohpia* sp., *Pseudoatrypa* cf. *devoniana,* and *Productella* sp. dominate the fauna of these beds. Based on observations at Fall Creek and Williams Brook (Fig. 1), what we herein call the "Pyramid Mall beds" may be more than one cycle. Further work and identification at additional sections should aid in sorting this out.

SEQUENCE STRATIGRAPHIC INTERPRETATION

Sedimentary Sequences at the Delta-front

Similarity observed in the sedimentary sequences presented above suggests some over-arching control on the amount of sediment being supplied to the basin; we interpret this as an affect of sea-level oscillation, assuming an approximate balance between basin subsidence and a relatively constant sediment supply. As sea-level initially rises, progradation of the delta slows and may stop completely depending on the rate and severity of transgression. During this time, sediment becomes sequestered in bays and estuaries, prohibiting the majority of the sediment supply, if not all, from reaching the basin. The initial onset of sea-level rise and the beginning

of sediment sequestration is called the Lowstand Systems Tract (LST). The fastest rate of sea level rise, would result in condensed intervals, generally expressed as fossil rich limestones and calcareous silt and sandstones (Transgressive Systems Tracts, or TST). LST deposits are commonly reworked by overlying TST deposits such that these units may form a composite bed. Owing to the proximity of the study area to the sediment source, and the likely position along the delta-front, sediment supply may become reduced during transgressions but is not completely shut off, as one would expect in more basinal settings. In essence, in the overall Ithaca constructional (progradational) regime, even relatively "sediment-starved" TST intervals could effectively have been characterized by sediment bypass conditions at the delta front. This is evidenced by the 'splaying' open of coquinas with interbedded siltstones, as well as the variability observed within condensed intervals over a short distance as seen in the "Firestone Beds" and the Beebe Limestone. The "Linden Horizon" serves as an example of a flooding surface (FS), associated with the highest sea-level, and therefore the least amount of sediment input into the basin.

The character of the "Firestone Beds" at various localities is presented in Figure 3. At the Indian Creek and NY-13 localities, the "Firestone Beds" occur as highly fossiliferous interbedded silts and coquinas. At Williams Brook, this same interval is over 3 meters of channelized and cross-bedded thick-bedded limestone lenses (Williams Brook Limestone). At Fall Creek and Cascadilla Creek, this interval is again seen as coquinitic lenses with interbedded silt and sandstones. At Sixmile Creek this interval has a similar lithological character as Fall and Cascadilla Creeks, however the "Firestone Beds" here are dominated by *Heterophrentis* corals. One possible interpretation, for the thickening of sedimentary sequences towards the southeast following on the suggestion by deWitt and Colton (1978) that the source of sediment was coming from this direction, is that this transect represents the slope of the delta-front. The presence of abundant corals at Sixmile Creek and the observation of *Stromatopora* sp. in this interval near Fall Creek (Williams 1884) suggests that the shallowest portion of the transect presented is in the southeast. The Williams Brook Limestone can be interpreted as a submarine channel deposit along the delta-front slope. Relative to the "Firestone Beds" at other localities, the Williams Brook Limestone has thicker coquinitic lenses with a lower amount of siliciclastics within the calcareous portions. It would be expected that within a submarine channel there would be increased winnowing, as well as increased accumulation of fossils as channel lag deposits. A critical test of this interpretation would be to examine Williams Brook Limestone corals and brachiopods for abrasional degradation and biofacies incongruity relative to synjacent facies (= downslope transport). This is prima facie evidence for downslope displacement of large corals over a submarine escarpment in the Ludlowville Formation (see Bartholomew, et al., this volume). At Indian Creek and NY-13 the coquina lenses of the "Firestone Beds" are again splayed-open, possibly due to the increased sediment, winnowed from the area of the Williams Brook Limestone. Although not as clear-cut, the Ithaca Falls Bed may also show a similar gradient. The Ithaca Falls Bed is dominated by clams and bryozoans at the Fall Creek section, while 'Leiorhynchids' are only found in this bed at Indian Creek.

The Beebe Limestone submember also shows a good deal of variation between and within localities (Fig. 3). Following the same transect as the "Firestone Beds" above, the Beebe Limestone is a fossiliferous silty shale to calcisiltite at Indian Creek. This same bed is seen better developed at NY-13 were it is a calcareous pack- to grainstone. At Williams Brook this bed is again a fossiliferous silty shale to calcisiltite. At Fall Creek the Beebe Limestone shows the most variation, as a well developed coquina containing reworked phosphate pebbles and rip-up clasts in float blocks within the gorge, and as a pack- to grainstone upstream of Beebe Lake. At Cascadilla Creek this bed is again seen as a fossiliferous calcisiltite, and finally, at Sixmile Creek this bed is once again a well-developed coquina containing reworked pyrite-coated concretions. Similar concretions are observed approximately 2 meters below the Beebe Limestone at Williams Brook. Likewise, we attribute such variation to the presence of small submarine channels and variations in sediment supply.

Besides such variation in sediment supply during a systems tract, present correlations suggest shifting depocenters as well, as seen in Figure 1. The Renwick Formation is thickest at the southeast end of the transect, thinning toward the northwest. The lower Cascadilla member displays an opposite thickness trend along the transect. The upper Cascadilla member is relatively constant in thickness across the transect. The Triphammer member, like the Renwick, apparently thins from southeast to northwest. More work is necessary to understand the relationships of, and thickness between, the Beebe Limestone and the "Pyramid Mall beds" (Cayuga Heights member). Although likely shifting, the overall trend agrees with deWitt and Colton (1978) that the majority of the sediment supply was from the southeast.
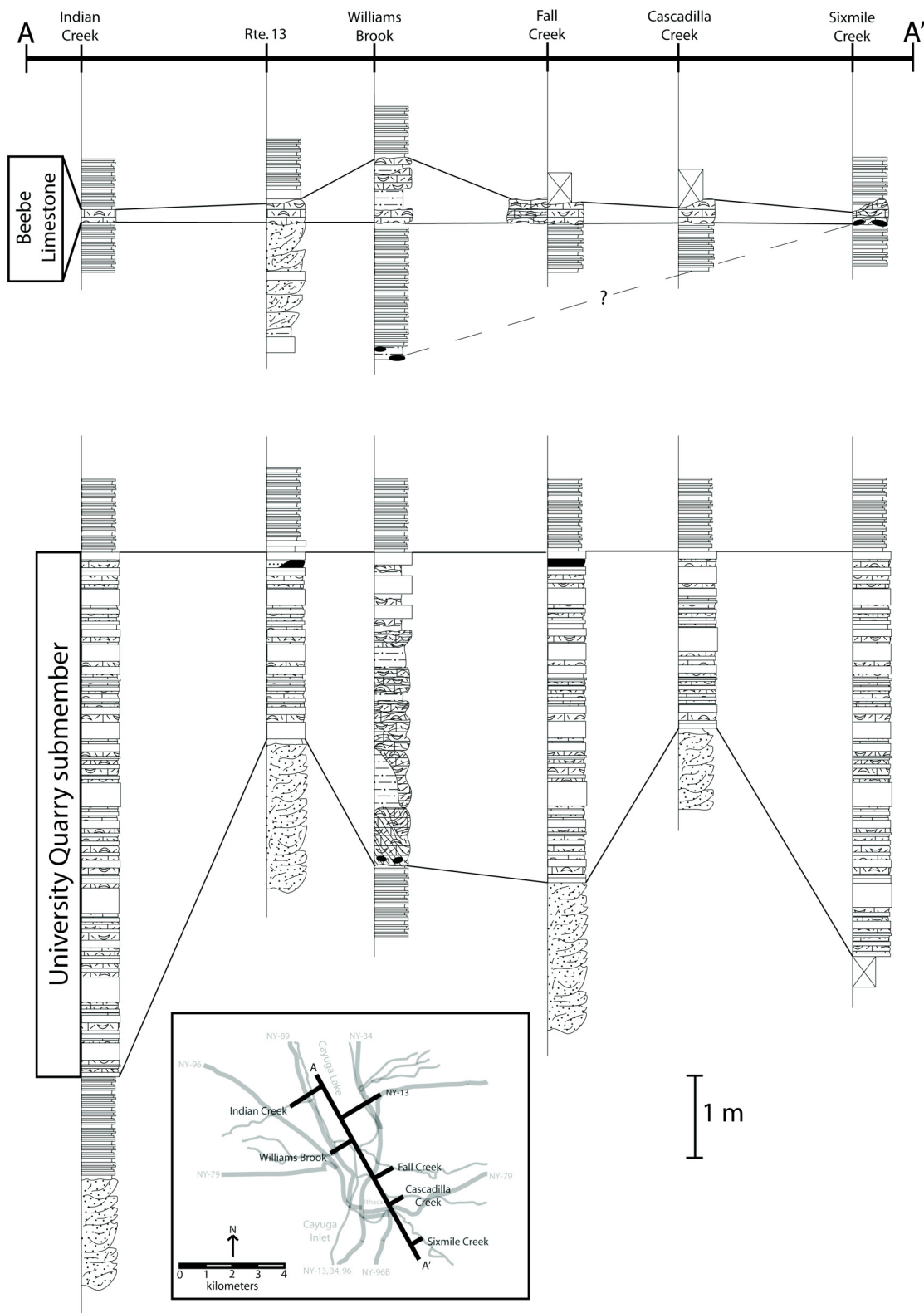
FIGURE 3—Lithological variation observed within the Beebe Limestone and University Quarry submembers at various localities in the vicinity of Ithaca, New York. Such variation is attributed to differences in sediment supply and shifting depocenters at the delta-front.

The rest of the sea-level oscillation is represented by strata that are relatively time poor and can be rather unfossiliferous (i.e. sediment diluted). This includes silty shales and interbedded thin siltstones representing deposition after progradation exceeds the rate of sea-level rise (Highstand Systems Tract, or HST). During sea-level fall, the sequence is seen as the thick-bedded silt and sandstone channel lenses observed (Falling Stage Systems Tract, or FSST). This is the shallowing (coarsening) upward pattern observed in each of the members described in this paper. The lowest point of sea-level, the sequence boundary (SB), would be observed at the base of the condensed interval of the next sea-level cycle and is used by us as the boundary between the described members of the Ithaca Formation. This boundary represents the lowest point of sea-level drop, and therefore when correlated between localities, is an isochronous surface. This surface is commonly composited (i.e. reworked) with the overlying TST, and can therefore be taken as the base of the condensed interval.

A summary diagram of the sequence stratigraphic interpretation of the lower Ithaca Formation is given in Figure 2.

CORRELATIONS WITH WESTERN SECTIONS

Now that we have a working hypothesis for sea-level cycle interpretation of the lower Ithaca and have identified a number of key marker beds within the Cayuga Lake area, it is possible to suggest tentative correlations with marker beds in western, more basinal sections. Because we interpret these condensed beds as forming during transgressions, it would follow that these should be traceable into discontinuity surfaces in the basin, as this is where a condensed interval should begin during a transgression. We outline such correlations in Figure 4.

Both the Fir Tree and Lodi Limestones have been well documented across western and central New York State (Baird et al. 1988). The Lodi Limestone is seen at the base of Twin Glens, just above the level of Cayuga Lake. Between the Lodi Limestone and the "Linden Horizon", which is herein understood to cap the University Quarry interval as discussed above, correlations are more uncertain. Kirchgasser (2000) suggests that the Renwick may be correlative to the Schumacher bed; a thin, conodont-rich black shale above the Lodi level, but this still has yet to be confirmed. Two fossil-bearing grey shale bands, informally termed the "Abbey Beds" (unpublished data) are observed to overlie the Schumacher Bed at the Rochester meridian; these may then correlate with the Ithaca Falls and Fall Creek limestones in the Ithaca sections presented here. Further fieldwork is needed to locate these beds in sections between Cayuga Lake and Genesee Valley.

The "Crosby Sandstone" (Fox 1932, Torrey et al. 1932) is a bed of resistant, sparsely fossiliferous, but bioturbated silt- to sandstone known from sections in the Keuka and Seneca valleys (deWitt and Colton, 1978). At the section at Mill Creek in Seneca Valley, this bed can be found above the "Linden Horizon" (University Quarry submember) correlatives (Kirchgasser and House 1981). It has not yet been confidently linked to Genesee Valley or Ithaca area localities (see discussion below), though it appears to be correlative with a bed of partially exhumed (truncated) concretions which is slightly below the Genundewa Formation in the Canandaigua Valley (Baird, 1976). In this region, Crosby Sandstone and correlative septarial concretions are overstepped westward across the Canandaigua Valley such that sub-Genundewa (upper Penn Yan Formation) dark grey shale rests unconformably on lower Penn Yan black shale deposits north of Seneca Point Glen. De Witt and Colton (1978) place the Crosby Sandstone at 70 to 80 feet above the Williams Brook Limestone at the section at Williams Brook. This corresponds well with the lower portion of a unit we have traced, and called the "Pyramid Mall beds". Huddle (1981) sampled the "Pyramid Mall beds" at Fall Creek for conodonts. The lack of *Ancyrodella rugosa*, unless due to facies restriction, suggests that this bed is older than the Upper Genundewa Limestone. Based on these observations, both the "Pyramid Malls beds" and the Beebe Limestone should be below the Genundewa Limestone in western sections. This is particularly troubling because the only known traceable, condensed unit between the "Linden Horizon" and the Genundewa Limestone is the Crosby Sandstone. The Crosby either represents both of these beds as a composite unit in western sections, or there is an, as yet, unrecognized discontinuity in the western sections as is suggested by the aforementioned dark shale-roofed discontinuity observed by Baird (1976) that oversteps the possible Crosby correlative in the Canandaigua Valley. The Starkey Black Bed of de Witt and Colton (1978), or more appropriately the base of this unit, is another possibly correlative bed, but more work is needed to understand the nature of this interval.
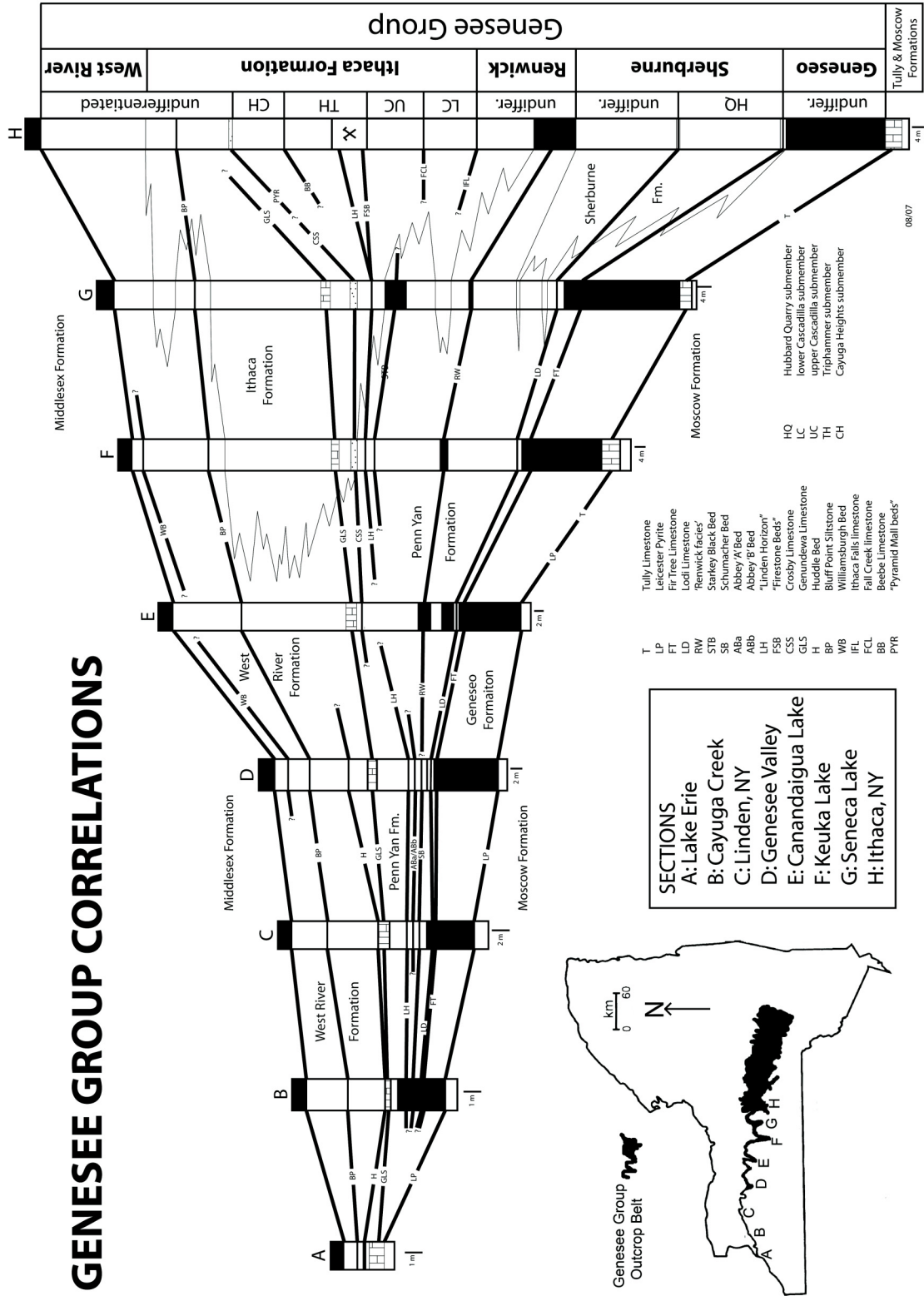
FIGURE 4—Correlations within the Genesee Group between Lake Erie and the Cayuga Lake Valley. This is modified from Kirchgasser (1985) with the data presented in this paper (see text for appropriate references).

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The Genundewa Limestone and associated North Evans lag bed is a unique and dominant feature of the Genesee Group succession in westernmost New York (see summary of issues in Baird et al. 2006). Space does not allow extensive description of this ultra-condensed interval of styliolinid carbonate and associated discontinuities, but it must be noted that it is enormously time-rich and in western sections may potentially be equivalent to a thick portion of the upper Ithaca succession. The North Evans bone/conodont bed encompasses up to seven conodonts zones mixed as a composite lag where this unit oversteps the entire lower Genesee Group succession near Buffalo. Both the North Evans disconformity closes eastward to continuity within the Genundewa succession and the Genundewa itself thickens into a splay of discrete styliolinid limestone beds in the same direction (Baird et al. 2006). In existing literature, the Genundewa is unknown east of a locality (Schuman Cemetery) near Gorham, NY (deWitt and Colton, 1978). However, ongoing unpublished fieldwork by the present authors shows at least some portion of the Genundewa (both with Styliolinid limestone and "North Evans" aspect) is traceable through the Keuka Valley to the west side of the Seneca Lake Valley. These discoveries show the potential of linking this important, time-rich, unit into the Ithaca area. However, just how such a bed will look at the delta-front is uncertain; proximity to the delta-front has had a substantial effect on the development and variability of the condensed intervals presented above. It may be that the Genundewa correlative is present in the upper-most "Pyramid Mall beds", especially if these represent more than one cycle. Finally, the current easternmost known limits of post-Genundewa divisions (conodont-glauconite-rich "Huddle Bed" of the basal West River Formation, goniatite-bearing Bluff Point siltstone bed of the medial West River Formation (see Kirchgasser, 1985), and pyrite clast/conodont-rich lag horizon of the Williamsburgh Bed of the topmost West River Formation) are also being extended toward the Ithaca area as part of ongoing fieldwork.

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REFERENCES CITED

Aboussalam, S.Z., Becker, R.T., and Schultz, H.P., 2001, The global Taghanic biocrisis in the Upper Givetian (Middle Devonian). – 15th International Senckenberg Conference; Mid-Paleozoic Bio- and Geodynamics, The North Gondwana-Laurassia Interactions, Abstracts: 1; Frankfurt am Main.

Aboussalam, S.Z., 2003, Das "Taghanic-Event" im hoheren Mitteldevon von West-Europa und Marokko. Munstersche Forschungen zur Geologie und Palaontologie: Heft 97. 330 p.

Baird, G.C., 1976, Coral encrusted concretions: a key to recognition of a "shale on shale" erosion surface. Lethaia: vol. 9, p. 293-302.

Baird, G.C., and C.E. Brett, 1986, Submarine erosion on the dysaerobic seafloor: Middle Devonian corrosional disconformities in the Cayuga Valley region, in Field Trip Guide for the 58th Annual Meeting of the New York State Geological Association, 58th Annual Meeting: Cornell University, Ithaca, New York, p. 23-80.

Baird, G.C., Brett, C.E. and Kirchgasser, W.T., 1988, Genesis and geochronology of black shale-roofed discontinuities in the Devonian Genesee Formation, western New York State, In McMillan, N.J., Embry, A.F. and Glass, D.J., eds., Devonian of the world; Canadian Society of Petroleum Geologists Memoir 14, p. 357-375.

Baird, G.C., and Brett, C.E., 2003, Shelf and off-shelf deposits of the Tully Formation in New York and Pennsylvania: Faunal incursions, eustasy and tectonics, Cour., Forsch.-Inst. Senckenberg: vol. 242, p. 141-156.

Baird, G.C., Kirchgasser, W.T., Over, D.J., and Brett, C.E., 2006, An Early Late Devonian Bone Bed-Pelagic Limestone Succession: The North Evans-Genundewa Story, in, Jacobi, R.D., ed., Field Trip Guide for the

74th Annual Meeting of the New York State Geological Association: University at Buffalo, Buffalo, New York, p. 354-395.

Caster, K.E., 1933, Upper Devonian Rocks at Ithaca, New York, Geological Society of America Bulletin: v. 44, no. 1, p. 201-202.

Caster, K.E., 1933a, The sedimentary succession and divisions of the Devonian in the Finger Lakes region, Guidebook 4 of the 16th International Geological Congress, p. 40-49.

Chadwick, G.H., 1933, Great Catskill delta and revision of late Devonic succession., Pan-American Geologist: v. 60, p.91-107.

Chadwick, G.H., 1935, Chemung is Portage, Geological Society of America Bulletin: v. 46, p. 343-354.

de Witt Jr., W., and Colton, G.W., 1959, Revised correlations of lower Upper Devonian rocks in western and central New York, American Association of Petroleum Geologists Bulletin: v. 43, p. 2810-2828.

de Witt Jr., W., and Colton, G.W., 1978, Physical Stratigraphy of the Genesee Formation (Devonian) in western and central New York., U.S. Geological Survey Professional Paper 1032-A, 22 p.

Ettensohn, F.R. (1985) The Catskill Delta Complex and the Acadian Orogeny: a model, In: Woodrow, D.L., and Sevon, W.D. (Eds.), The Catskill Delta, Geological Society of America, Special Paper, vol. 201:39-49.

Ettensohn, F.R., 1998, Compressional tectonic controls on epicontinental black shale deposition: Devonian-Mississippian examples from North America, in Schieber, J., Zimmerle, W., and Sethi, P.S., eds., Shales ad Mudstones: vol. 1 (Basin Studies, Sedimentology, and Paleontology), p. 109-128.

Ettensohn, F.R., Miller, M.L., Dillman, S.B., Elam, T.D., Geller, K.L., Swager, D.R., Markwitz, G., Woock, R.D., and Barron, L.S., 1988, Characterization and implications of the Devonian-Mississippian black-shale sequence, eastern and central Kentucky, U.S.A.: pycnoclines, transgression, regression, and tectonism, in McMillan, N.J., Embry, A.F., and Glass, D.L., eds., Devonian of the World, Canadian Soc. Petrol. Geol. Mem. 14, vol. II., p. 323-346.

Fox, I.W., 1932, Geology of part of the Finger Lakes region, New York, American Association of Petroleum Geologists Bulletin: vol., 16, no., 7, p. 675-690.

Grasso, T.X., Harrington, J.H., and Kirchgasser, W.T., 1986, Stratigraphy and Paleontology – around Cayuga Lake once again, in Field Trip Guide for the 58th Annual Meeting of the New York State Geological Association: Cornell University, Ithaca, New York, p. 23-80.

Harrington, J.H., 1970, Benthic Communities of the Genesee Group (Upper Devonian), in Heaslip, W.G., ed., Field trip Guidebook for the 42nd Annual Meeting of the New York State Geological Association: SUNY Cortland, Cortland, New York, p. A-1 to A-18.

House, M.R., and Kirchgasser, W.T., 1993, Devonian goniatite Biostratigraphy and timing of facies movements in the Frasnian of eastern North America in Hailwood, E.A. and Kidd, R.B., eds., High Resolution Stratigraphy, Geological Society Special Publication: No. 70, p. 267-292.

Huddle, J.W., 1981, Conodonts from the Genesee Formation in Western New York, U.S. Geological Survey Professional Paper 1032-B, 63 p.

Johnson, J.G., 1970, The Taghanic Onlap and the end of the North American Devonian Provinciality, Geological Society of America Bulletin: vol. 81, p. 2077-2106.

Johnson J.G., Klapper, G., and Sandberg, C.A., 1985, Devonian eustatic fluctuations in Euramerica, Geological Society of America Bulletin: vol. 96, p. 567-587.

Kindle, E.M., 1896, The relation of the Ithaca Group to the faunas of the Portage and Chemung, Bulletins of American Paleontology: vol. 2, 56 p.

Kindle, E.M., 1906, Notes on the Range and distribution of Reticularia laevis, Journal of Geology: p. 188-193.

Kirchgasser, W.T., 1985, Ammonoid Horizons in the Upper Devonian Genesee Formation of New York: Legacy of the Genesee, Portage, and Chemung, Geological Society of America Special Paper 201, p. 225-235.

Kirchgasser, W.T., 2000, Correlation of stage boundaries in the Appalachian Devonian eastern United States, Cour, Forsch.-Inst. Senckenberg: vol. 225, p. 271-284.

Kirchgasser, W.T., and House, M.R., 1981, Upper Devonian Goniatite Biostratigraphy, in Oliver Jr., W.A. and Klapper, G., eds., Devonian Biostratigraphy of New York, Part 1, Text: International Union of Geological Sciences, Subcommission on Devonian Stratigraphy, July 1981, Washington, D.C., p. 39-56.

Kirchgasser, W.T., Over, D.J., and Woodrow, D.L., 1994, Frasnian (Upper Devonian) Strata of the Genesee River Valley, Western New York State, in the Field Trip Guide for the 66th Annual Meeting of the New York State Geological Association, University of Rochester, Rochester, New York, p. 325-358.

Klapper, G., and Johnson, J.G., 1990, Revisions of Middle Devonian Conodont Zones, Journal of Paleontology: vol. 64, no. 6, p. 934-935, 941.

Rickard, L.V., 1964, Correlation of the Devonian rocks in New York State, New York State Museum and Science Service, Map and Chart Series 4.

Rickard, L.V., 1975, Correlation of the Silurian and Devonian rocks in New York State, New York State Museum and Science Service, Map and Chart Series 24, 16 p.

Rickard, L.V., 1981, The Devonian System of New York., in Oliver Jr., W.A. and Klapper, G., eds., Devonian Biostratigraphy of New York, Part 1, Text: International Union of Geological Sciences, Subcommission on Devonian Stratigraphy, July 1981, Washington, D.C., p. 5-22.

Smith, B., 1935, Geology and Mineral resources of the Skaneateles Quadrangle, New York State Museum Bulletin, 120 p.

Sutton, R.G., 1959, Use of Flute Casts in Stratigraphic Correlation, American Association of Petroleum Geologists Bulletin: vol. 43, no. 1, p. 230-237.

Sutton, R.G., 1963, Correlation of the Upper Devonian strata in south-central New York, in Shepps, V.C., ed., Middle and Upper Devonian stratigraphy of Pennsylvania and adjacent states – a symposium: Pennsylvania Geological Survey, 4th Series, Bulletin G-39, p. 87-10.

Sutton, R.G., Humes, E.C., Nugent, R.C., and Woodrow, D.L., 1962, New Stratigraphic nomenclature for the Upper Devonian of south-central New York, American Association of Petroleum Geologists Bulletin: vol. 46, no. 3, p. 390-393.

Torrey, P.D., Fralich, G.H., Young, W.H., Brewer, C., and Phillipi, P.M., 1932, The geology of New York and northern Pennsylvania, American Petroleum Institute, Div. Production, Paper 826-4A, 19 p.

Vanuxem, L., 1840, Fourth Annual Report of the geological survey of the third district, New York State Survey, Annual Report:, vol. 4, p. 355-383.

Ver Straeten, C.A., and Brett, C.E., 1995, Lower and Middle Devonian Foreland Basin Fill in the Catskill Front: Stratigraphic Synthesis, Sequence Stratigraphy, and the Acadian Orogeny, in Field Trip Guide for the 67th Annual Meeting of the New York State Geological Association.

Williams, H.S., 1884, On the Fossil faunas of the Upper Devonian, U.S. Geological Survey Bulletin: vol. 3, p. 55-86 (5-36).

Williams 1906, The Devonian Section of Ithaca, New York, Part I, Journal of Geology: vol. 14, no. 7, p. 579-598.

Williams, H.S., 1913, Recurrent Tropidoleptus Zones of the Upper Devonian in New York. United States Geological Survey, Professional Paper 79, 103 pp.

Williams, H.S., Tarr, R.S., and Kindle, E.M., 1909, Geologic Atlas of the United States, Watkins Glen-Catatonk Folio, Folio 169, United States Geological Survey.

Williams, G.Q., 1951, The Stratigraphy of the Type Ithaca Formation, unpublished PhD thesis, Cornell University.

ROAD LOG FOR TRIP B-4

RE-EXAMINATION OF THE TYPE ITHACA FORMATION: CORRELATIONS WITH SECTIONS IN WESTERN NEW YORK
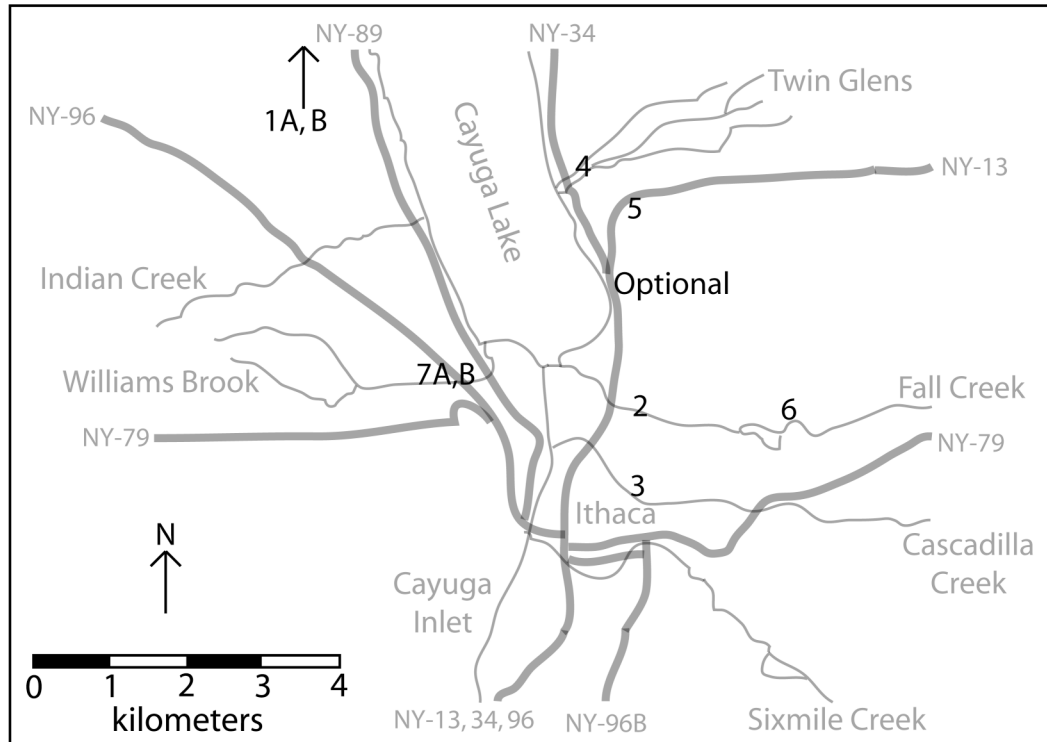
FIGURE 5—Location of stops to be made on this trip.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Begin roadlog at the entrance to SUNY Cortland, intersection of Tompkins St. (NY-13) and Pashley Dr. Turn right on Tompkins St. (NY-13 southbound).
1.3	1.3	Turn left following NY-13 southbound
1.8	0.5	Enter South Cortland
3.4	1.6	Cross Gracie/Webb Road.
4.5	1.1	Tompkins-Cortland County Line.
		0.25 mile long overgrown, but fossiliferous outcrop. The presence of *O. mesastrialis* and 'Pugnoides' places this within the Triphammer member, possibly the University Quarry submember (see Harrington 1970).
7.5	3.0	Pass entrance to Tompkins-Cortland Community College.

8.0	0.5	Enter the Village of Dryden.
8.4	0.4	Turn right, remaining on NY-13 southbound.
11.2	2.8	Pass Ringwood Road.
15.2	4.0	Cross Hanshaw Road.
16.3	1.1	Enter the Town of Lansing
17.1	0.8	Outcrop of West River equivalent strata at the intersection of Warren Rd.
17.7	0.6	Exit for Triphammer Road. Strata of the Ithaca Formation.
18.4	0.7	Exit for Cayuga Heights Road. Strata of the Ithaca Formation, this will be STOP 5.
19.7	1.3	Pass exit for NY-34.
20.5	0.8	Cross Dey/Willow Streets.
21.3	0.8	Turn Right on Buffalo St. (NY-89 northbound).
21.4	0.1	Turn right on NY-89 northbound (Taughannock Blvd.).
21.55	0.15	Bridge over Cayuga Lake Inlet.
21.6	0.05	Outcrops of Renwick Formation.
22.0	0.4	Cross Williams Brook (STOPS 7A and7B will be upper portion of this creek).
22.7	0.7	Enter Town of Ulysses.
25.4	2.7	Pass Cayuga Nature Center.
28.0	2.6	Enter Taughannock Falls State Park.
28.2	0.2	Park vehicles. Proceed along Taughannock Creek to waterfalls over the Tully Limestone

STOP 1A. TAUGHANNOCK STATE PARK (TULLY LIMESTONE)

Both STOPS 1A and B are used as an overview to put the Ithaca Formation in perspective with the underlying Middle Devonian strata. At the first stop, the Tully Limestone is accessible above the Moscow Formation of the Hamilton Group. As summarized in Baird and Brett (2003), the Tully represents both significant sedimentological and faunal changes in the Appalachian Basin. Sedimentologically, upper Tully divisions, commencing with the coral-rich Bellona Bed comprise a TST-succession that represents the onset of

the Taghanic onlap, a eustatic sea-level rise that locally within the Appalachian Basin resulted in the deposition of the black Geneseo Shale Formation and its correlatives (Harrell, Burkett, etc, see Baird and Brett 2003). Faunally, this onlap is significant because it is concurrent with late stages of the global Taghanic Bioevent, resulting in the suspected demise of Middle Devonian faunas worldwide. As will be demonstrated later in this trip, the demise of the Middle Devonian 'Hamilton-fauna' in the Appalachian Basin may not have taken place at this boundary; 'Hamilton'-characteristic fossils can be found throughout the Frasnian of New York (see Williams 1913 and references therein).

Return to vehicles and continue along NY-89 northbound.

| 28.5 | 0.3 | Turn right on Taughannock Park Rd. |
| 29.1 | 0.6 | Turn left into the falls overlook parking lot. |

STOP 1B. TAUGHANNOCK FALLS OVERVIEW (LOWER GENESEE GROUP)

The overlook at Taughannock Falls provides a stunning view of the lower Genesee Group; including the Geneseo Formation, Sherburne Formation, and the lowest Ithaca Formation (Figure 6). Within the walls of this hanging valley both the Fir Tree and Lodi Limestone Beds can be observed. These beds, or associated discontinuities, can be traced throughout most of New York (Baird et al. 1988). Also visible within the gorge wall are the siltstone lenses characteristic of the 'Renwick facies', which will be observed in detail at later stops. This is purportedly the highest waterfall east of the Mississippi River.

Return to vehicles and continue west along Taughannock Park Road
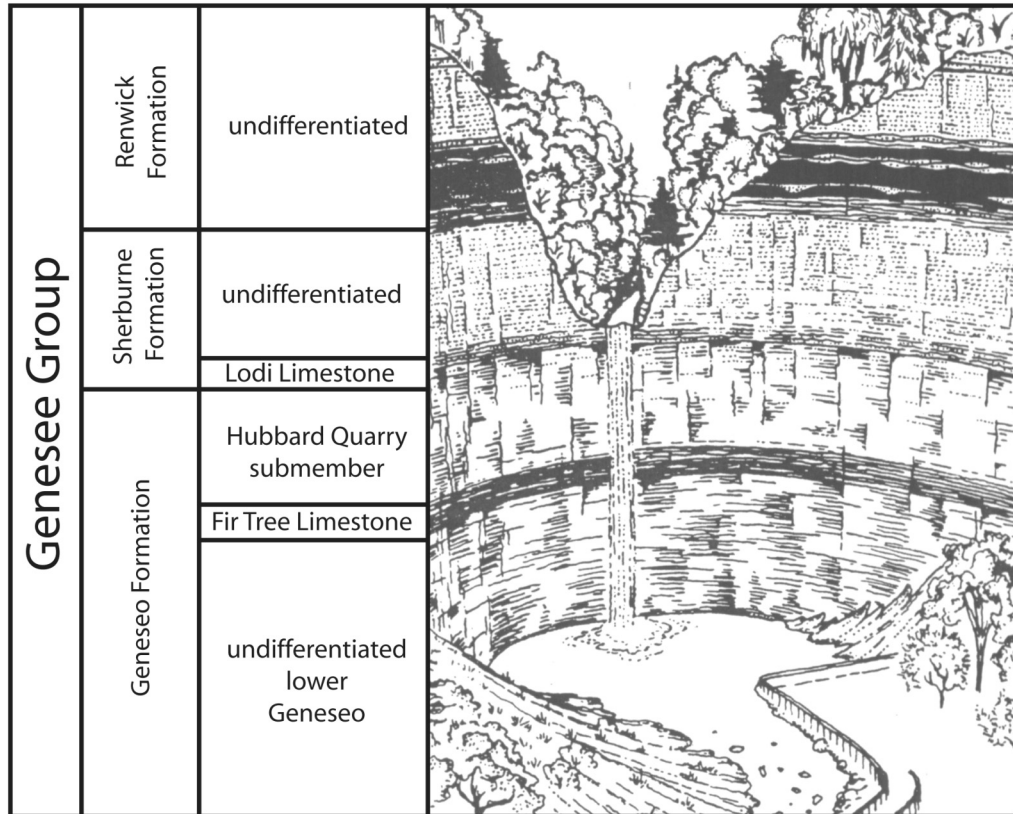
FIGURE 6— Diagrammatic representation of the view seen at the Taughannock Falls overlook. Modified from Baird and Brett (1986).

29.7	0.6	Turn left on Falls/Jacksonville Rd.
29.8	0.1	Cross Taughannock Creek.
30.3	0.5	Turn left on NY-96 southbound.
33.6	3.3	Pass Ithaca Antique Mall.
35.4	1.8	Pass Dubois Rd.
36.0	0.6	Pass entrance to The Paleontological Research Institution.
36.6	0.6	Enter the Town of Ithaca.
36.7	0.1	Cross Williams Brook and parking (STOPS 7A and 7B).
36.9	0.2	Enter the City of Ithaca. Williams Glen Road on left.
37.6	0.7	Pass outcrops of Renwick Formation.
38.1	0.5	Bridge over Cayuga Lake Inlet.
38.2	0.1	Turn left on NY-13/34 (northbound).
38.9	0.7	Bridge over Cascadilla Creek.
39.0	0.1	Right on Dey Street.
39.02	0.02	Left on Lincoln Street.
39.4	0.38	Proceed straight at the end of Lincoln Street, crossing Lake Street.
39.5	0.1	Park vehicles at Ithaca Falls Park.

STOP 2. ITHACA FALLS PARK, FALL CREEK (WARRENELLA ZONE SUBMEMBER)

This stop allows access to the Renwick Formation (Figs. 1 and 2). This section begins in the *Warrenella* Zone, near the base of the Renwick Formation. Above this and in the face of Ithaca Falls are the siltstone channel lenses characteristic of the 'Renwick facies'. Excellent collecting of the *Warrenella* Zone fauna is possible at this locality. The upper Renwick Formation is inaccessible at this locality, but can be accessed at the nearby gully running up through the old Ithaca Gun Factory, also known as Gun Hill. Due to reportedly high lead levels from test firing of rifles, this small gully and the Ithaca Gun Factory itself are currently posted.

Return to vehicles. Proceed left on Lake St.

39.6	0.1	Pass entrance of the Ithaca Gun Factory.
39.8	0.2	Right on University Avenue.
40.2	0.4	Right on Linn St. Park on the immediate left in the parking lot for the First Church of Christ Scientist.

STOP 3. CASCADILLA CREEK ('RENWICK FACIES')

This stop begins stratigraphically above the *Warrenella* Zone, within the siltstone lenses of the 'Renwick facies' (Figures 1 and 2). This is the "*Lingula* Shales" interval of H.S. Williams (1906). This outcrop offers collecting of the *Lingula* fauna, including *Plumalina plumaria* and articulated crinoids (*Eutaxocrinus ithacensis*).

Return to vehicles. Proceed with a right on Linn St.

40.3	0.1	Turn left on University Avenue.
40.7	0.4	Turn Left on Lake Street.
41.0	0.3	Bridge over Fall Creek (Ithaca Falls Park).
41.8	0.8	Turn right on NY-13 northbound, and park along the entrance ramp.

OPTIONAL STOP. NY-13 (RENWICK FORMATION)

   This optional stop shows a continuous section of the intervals seen at STOPS 2 and 3. The base of the section is in the *Warrenella* Zone (Figures 1 and 2). Above this, the siltstone channel interval ('Renwick facies') is accessible for collection of the *Lingula* fauna as well as *Plumalaria plumalina*. Although this outcrop does not reach the Ithaca Falls limestone, the upper portion of this cut is interesting in that it affords the opportunity for close examination of large ball and pillow structures. This is approximately the arbitrary lithological boundary given to the Renwick and overlying Ithaca in the past literature.

   Return to vehicles and continue along NY-13 northbound.

42.7	0.9	Enter Town of Lansing.
42.8	0.1	Exit at Cayuga Heights. The well-exposed outcrop along the cloverleaf of strata above the Renwick Formation will be examined later as part of STOP 5.
42.9	0.1	Turn right on Cayuga Heights Rd.
43.0	0.1	Turn right, returning onto NY-13, thistime southbound.
44.1	1.1	Exit at NY-34/Stewart Park.
44.3	0.2	Turn right following signs for NY-34 northbound.
44.5	0.2	Turn left on NY-34 northbound.
45.4	0.9	Park on side of NY-34 near (but 10 feet from) orange hydrant. Proceed on foot to trail along Twin Glens. This is private property, and permission must be granted for access.

STOP 4. TWIN GLENS (ITHACA FALLS LIMESTONE SUBMEMBER)

   The focus of this stop is the Ithaca Falls limestone submember as well as the recurrent 'Hamilton fauna' in the surrounding shales (Figures 1 and 2). At this locality the Ithaca Falls limestone is well developed, showing a sharp erosional base, erosionally lateral-truncation of turbidites, and an angularly truncated upper surface. This bed is excellent for collecting calcite replaced *Paleoneilo* sp. and Nuculiid bivalves. Above this limestone is a thin bed of crinoid debris that has yielded articulated cladid crinoids (at Fall Creek). Surrounding this bed is a dysaerobic fauna with some characteristic fossils of the 'Hamilton fauna'. This includes *Paleoneilo* sp., Nuculiid bivalves, *?Mucrospirifer mucronatas (?'Spirifer' posterus, ?Tylothyris mesacostales), Glyptotomaria capillaria*, and abundant pelmatozoan material.

   Return to vehicles and continue along NY-34 northbound.

45.5	0.1	Enter town of Lansing… once again.
46.5	1.0	Hard right onto Cayuga Heights Road.
47.7	1.2	Cross upper end of Twin Glens.
47.9	0.2	Turn left onto entrance ramp for NY-34 northbound. Park along side of entrance ramp.

STOP 5. NY-13 (UPPER CASCADILLA THROUGH CAYUGA HEIGHTS MEMBERS)

Excellent exposures are seen at the cloverleaf cut for the NY-13 interchange at Cayuga Heights Road. The lowest part of this exposure begins above the Ithaca Falls limestone. This section is continuously exposed upward to the "Pyramid Mall bed", therefore including the complete cycles represented by the upper Cascadilla, Triphammer, and Cayuga Heights members (Figures 1 and 2). At this section, the lithology of the University Quarry submember more closely resembles the "Firestone Beds" at Fall Creek, than to the Williams Brook Limestone (STOP 7A), and the Beebe Limestone is a calcisiltite rather than a well developed coquina as seen in the Fall Creek Gorge and at Sixmile Creek (see Figure 3). Most impressive at this section is the cross-sections of FSST siltstone channel complexes visible in the wall face. There is noticeably a large amount of covered section, on NY-13 between the optional stop and STOP 5, in fact larger than should be, given the correlations presented in this trip. Faunally, however, our correlations of the NY-13 section with the other sections studied do agree, and we presently interpret this large covered interval as possibly representing some unseen thrust-faulting resulting in stratigraphic duplication.

48.0	0.1	Return to vehicles and make a cautious u-turn along the entrance/exit ramp. Turn left onto Cayuga Heights Road, entering the Village of Cayuga Heights.
49.5	1.5	Ithaca City limit.
49.55	0.05	Turn left on Thurston Avenue.
49.8	0.25	To the right at the end of Highland Ave. is a suspension bridge and access into Fall Creek Gorge at the Cornell University Power Plant waterfalls and the 'type' "Firestone Beds". Proceed on Thurston.
50.1	0.3	Turn left onto Cradit Farm Road before the Thurston Avenue bridge.
50.4	0.3	Park in lot opposite observatory. Proceed along foot path to Beebe Lake.

STOP 6. BEEBE LAKE (BEEBE LIMESTONE SUBMEMBER)

This stop will show the Beebe Limestone in place near Beebe Lake (Figures 1 and 2). The lithology developed here is very dissimilar to the coquinitic limestone blocks seen downstream as isolated float below the Beebe Lake dam (see text and Fig. 3), but this section at least serves as an accessible type locality for this bed, which has been wanting ever since Caster (1933, 1933a) first named it. This stop also highlights local faulting within the Ithaca Formation. A down-dropped (?) fault block is seen just upstream of the Beebe Limestone.

Return to vehicles and turn right onto Cradit Farm Road toward Thurston Avenue.

50.7	0.3	Turn right onto Thurston Avenue.
51.2	0.5	Turn right onto Stewart Avenue.
52.8	1.6	Turn right onto NY-13 southbound.
54.1	1.3	Pass exit for NY-34.
55.8	1.7	Turn right onto NY-96 northbound; stay in left lane.
56.2	0.4	Pass outcrops of the Renwick Formation.
56.9	0.7	Pass Williams Glen Road.
57.1	0.2	Park on left side of road in pull-off at NY-96 and Hopkins Place. Proceed to end of pull-off and access Williams Brook upstream of NY-96.

STOP 7A. WILLIAMS BROOK (BEEBE LIMESTONE SUBMEMBER)

At Williams Brook, the Beebe Limestone is similar to that observed at STOP 6, a fossiliferous calcisiltite, but here more splayed open. The identification of this bed at this level agrees with Williams et al. (1909) in that the University Quarry interval is equivalent to the Williams Brook Limestone; and furthermore shows that both are distinctly older than the Beebe (Figures 1 and 2). Below the Beebe interval at this section, pyrite-coated concretions, similar to those reworked into the Beebe Limestone at Sixmile Creek, can be observed. If time permits, the "Pyramid Mall beds" are accessible just upstream of here.

Proceed across street and onto Private Property to access the Williams Brook Limestone; permission must be given by landowner.

STOP 7B. WILLIAMS BROOK (WILLIAMS BROOK LIMESTONE)

The Williams Brook Limestone (Caster 1933, 1933a) is one of the most impressive deposits to be seen on this trip. The interval of the coquina itself is approximately 3 meters thick, capping a small waterfall. Channelized bedforms, cross-bedding, truncated inter-bedded siltstones, and reworked concretions and rip-up clasts can be seen within the coquina. The local representation of the "Firestone Beds" at this locality, as well as the channelized bedforms observed in cross-section, suggest deposition within a submarine channel (Figures 1, 2, and 3). The beds above the coquina (Marathon Sandstone of Caster (1933, 1933a), '*Spirifer mesastrialis* Zone', and the '*mesastrialis* sands' of Williams et al. (1909) and Williams (1913)) contain abundant fossils at some horizons, including *Gomphoceras tumidum*.

Return to SUNY Cortland via NY-96 southbound to NY-13 northbound.

END OF FIELD TRIP

Trip A-5

APPLIED GEOMORPHOLOGY FOR LAND USE AND PLANNING IN CORTLAND, NEW YORK

JAMES E. BUGH

Department of Geology, State University of New York at Cortland, Cortland NY 13045

INTRODUCTION

The "science of landforms", geomorphology, is critical in the use of land. Through the last forty years, environmental geomorphology has come to the forefront of geology/physical geology because of the rapid growth of suburbia and the recognition of the impact of land use on stream flow. Central New York experienced rapid growth in the early 1970s because of many employment possibilities. In Cortland, we had Smith Corona Marchant (which was the world's largest manufacturer of portable typewriters at the time), Mack Trucks, Wilson Sporting Goods and Chris Craft (which manufactured pleasure boats). The Cortland-Homer valley was easy to develop because of its relatively flat landscape and the strength and thickness of the materials making up the landscape. After the 1970s, the county had a net loss in population when these companies left the area.

A GEOMORPHOLOGIST'S EXPERIENCE IN LAND USE AND PLANNING

Cortland County was extensively glaciated in the Pleistocene such that erosion created deep valleys which received thick glacial outwash forming flat or gently sloping valley train topography. In places this outwash has a thickness reaching 235 feet. Much of this sediment is very permeable and serves as the sole-source aquifer for Cortland and Homer. Soils developed on the outwash, when coupled with the shape of the land and their permeability, are the best agricultural soils in New York State. Cortland County has been a leading agricultural producer since the mid-nineteenth century. No farm located on the valley floor has had a failing year because of lack of water due to the proximity to the water table. However, farms on the hill-slopes average failing years once in every three years.

The factors that allow for excellent agricultural productivity also make for ideal urban expansion. The soils have the strength to support urban structures and the permeability does not allow for standing water after rain or snowmelt. Excavations are easy in the outwash so that foundations, basements, and underground utilities are not prohibitively expensive.

Cortland County has an extensive mining program for its mineral resource: sand and gravel. Most of the mining today is confined to the southeast and southwest of Cortland because of the recognition that the best agricultural, urban, and mining lands are all the same tracts of land and are mutually exclusive. Mining had occurred north of Homer, but planning boards have not allowed for further mining while encouraging agriculture to continue. Many housing developments have been proposed on the farmland, but these too have been rejected.

The geomorphologist/geologist can become involved in sand and gravel mining in New York State by preparing a Geological Source Report. To more fully evaluate and ensure the quality of all material produced from granular deposits, a Geological Source Report must be submitted from each sand and gravel operation furnishing material for New York State Department of Transportation contracts. Also included in the Source Report are geologic cross-sections and petrographic analyses. Since April 1st, 1975, mines in New York State have been required to submit a Mined Land Reclamation plan to the New York State Department of Conservation. The Mined Land Reclamation Law seeks to foster and encourage the development of the state's

mining industry and mineral resources, prevent pollution associated with mining activity, and assure the reclamation of mined lands in a way the makes such lands suitable for future productive use.

The geomorphologist is an ideal person to write or, at least, to make a significant contribution to an environmental impact statement. The environmental impact statement is designed to help applicants and agencies determine, in an orderly manner, whether a project or an action may be significant. The question of whether or not an action may be significant is not always an easy one to answer, but the geomorphologist certainly can contribute to "impact on land" and "impact on water." For example, this writer was a contributor to a Generic Environmental Impact Statement on Sand and Gravel Mining in the Cortland-Tully Valley. This Generic Environmental Impact Statement was prepared to provide a single comprehensive document that addressed the impacts associated with a typical mining operation. The format allowed consideration of long-term, cumulative impacts, especially the loss of prime agricultural lands.

Many years ago several farm owners to the southwest of Cortland sold their lands for development. The Town of Cortlandville then zoned this land as "industrial." Much controversy has arisen from this action because the land overlies the aquifer for the Town of Cortlandville and the City of Cortland's aquifer recharge zone. Large industrial development both reduced the recharge and introduced contaminants to the aquifer. Development in Cortlandville has had a negative impact on the environment, but most often on the water supply. Problems such as trichloroethylene (TCE) and e. coli contamination developed, and solutions to these problems had to be found after the fact.

In 1987, for example, water tests showed fecal coli form bacteria in the private wells of 18 homes in a newly developed neighborhood. A week later the pollution had reached a total of 37 homes (Cortland Standard, August 25, 1987). Pollution testing expanded into the Town of Cortlandville and 27 more wells, close to the private wells, were found to be polluted. These homes were located on lots that had been deemed large enough for individual homes with septic systems. Engineers from the county Health Department had determined the lot size, and each homeowner or developer had been required to have a percolation test performed before the septic system was installed. However, the percolation was too fast because of the high permeability of the soil and lot size on that particular landscape, lateral moraine, should have been larger. As geomorphologists, we should be seeking to prevent problems from occurring by directing development in a sound environment way.

In 1992, Wendy's restaurant wanted to build on a vacant section of the then new Wal-Mart store property on Route 13, using just over 0.7 acres of land. The town's aquifer area is located along this already heavily developed area along Route 13, and the Cortland County groundwater management coordinator stated: "Eventually, you're going to hit the straw that breaks the camel's back" (Cortland Standard, June 18, 1992). A month later The Cortland County Planning Board recommended that Wendy's franchise not be built on this property. However, by 2007, the philosophy of the planning board had changed. Wal-Mart intends to build a super center on 33.7 acres (contingent on their getting all necessary permits) in close proximity to the 0.7 acres that Wendy's wished to develop and was turned down for. Despite the protests of a number of individuals and committees, the project will likely go forward. In this case, however, impact statements and snow collection and storm water management plans were scrutinized for three years before permits were issued. Storm water cannot be entered into Otter Creek before on site treatment. Also, Wal-Mart's tire and lube department will not be allowed on this property. Another major aspect before the permits were granted was the allowance for a landscaped green space between the Wal-Mart property and the adjacent assisted living facility. The planning board is currently in negotiation for outparcels of the project to be used as additional green space, rather than to be developed. It is possible to reduce the impact of development on the environment with sound planning.

ROAD LOG FOR TRIP A-5

APPLIED GEOMORPHOLOGY FOR LAND USE AND PLANNING IN CORTLAND, NEW YORK

Most of the stops on this field trips are located on private property. The leader of this trip has received permission to be on these sites for observations and measurements. The following individuals or companies should be contacted before on-site visits are made at a time other than this NYSGA field trip:

Steven G. Cleason, P.E.	Ralph Roe, Superintendent	Richard E. Schutz,
APD Engineering	Cortlandville Sand & Gravel	Operations Manager
3445 Winton Place - Suite 208	765 Route 13	Suit-Kote
Rochester, NY 14623	Cortland, NY 13045	P.O. Box 5160
(Wal-Mart Supercenter)		1911 Lorings Crossing
		Cortland, NY 13045

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Leave SUNY Cortland parking lot and proceed to Tompkins Street (Route 13). Turn right and cross railroad tracks.
1.2	1.2	STOP 1

STOP 1. THE USGS WELL, CT11.

The U.S. Geological Survey drilled a number of wells into the Cortland aquifer in the mid-1970s. Well CT11 is located nearest the City of Cortland wells. The U.S. Geological Survey has a number of reasons for drilling wells. The Survey may contract with communities to determine water supply potential or to develop a computer model of the aquifer based on the configuration and composition of the aquifer.

Before we begin our measurements, examine the well log and the landscape. What is the origin of this landscape?

TABLE 1—Log of Well CT11.

Depth (ft)	Description
10	coarse sand and pebbles
15	clay, sand and pebbles
20	medium to coarse sand and pebbles
25	coarse sand and pebbles
30	fine to coarse sand and pebbles
35	some clay, fine to coarse sand and pebbles
40	fine to coarse sand and pebbles
45	fine to coarse sand and pebbles
50	fine to coarse sand and pebbles
55	coarse sand and pebbles
60	medium to coarse sand and pebbles
65	silt and fine sand
70	fine to coarse sand and pebbles
75	fine to coarse sand and pebbles
80	fine to coarse sand pebbles
85	coarse sand and pebbles
95	medium to coarse sand and pebbles
100	clay, coarse sand and pebbles
105	clay and coarse sand

Measured data:

Surface altitude: _____   Depth to water: _____   Water table altitude: _____

Return to vehicles and continue southwest on Tompkins Street (Route 13). Development has progressed along Route 13 as strip development, which allows for easy access but is more costly with respect to utilities.

2.5	1.3	Route 13 bends at the intersection with Route 281.
2.7	0.2	Continue on Route 13 to the southwest, passing the existing Wal-Mart on your left.
3.4	0.7	Turn left onto Bennie Road.
3.7	0.3	Proceed past Walden Place Assisted Living to STOP 2.

STOP 2. SITE KNOWN AS THE POLO GROUNDS, SOON TO BE A WAL-MART SUPERCENTER

In March of 2003 preliminary plans for a proposed Wal-Mart Supercenter were submitted to the Town of Cortlandville to be located on this site. However, the site was not zoned for "big box" development and the project was put on hold pending changes in zoning. Wal-Mart requested that its project be termed a "Planned Unit Development" (PUD) and the request was granted after considerable controversy. A PUD can over-ride existing zoning, allowing Wal-Mart to submit a Draft Environmental Impact Statement (DEIS) for the project before a change in zoning occurred. The Final Environmental Impact Statement was submitted in November of 2006. The final approval by the Town was given in August of 2007. Throughout the four-year process, a local environmental organization fought the project, even bring a lawsuit against the Town. The problem with this site is that it is the most important tract of land over the aquifer.

Examine the test-pit data and the landscape. What is the origin of the landscape? The developers contend that the site is not a major recharge area because Otter Creek transports all of the runoff from the hills. We need to evaluate that contention.

TABLE 2—Test pit data.

Depth (ft)	Description
0-5	Gray-brown silt, little gravel, little sand, gravel pocket at four feet
5-6	Brown sandy silt and gravel
6-8	Gray silty sand and gravel

Measured data:

Surface altitude: _____   Depth to water: _____   Water table altitude: _____

Return to vehicles and drive back towards Route 13.

3.9	0.3	Turn left onto Route 13.
4.1	0.2	STOP 3 at 765 Route 13.

STOP 3 SOUTH CORTLAND SAND AND GRAVEL

To more fully evaluate and ensure the quality of all material produced from granular deposits, a Geological Source Report shall be submitted for each sand and gravel operation furnishing material to New York State Department of Transportation contracts. In the Source Report there must be a discussion of mode of deposition, a geologic cross-section, and a petrographic analysis.

What is the origin of this landscape?

We will analyze approximately 100 representative samples of gravel.

Measured data:

    Clastics: _____%   Carbonates: _____%   Crystallines: _____%

Return to vehicles and drive back towards Cortland on Route 13.

| 5.1 | 1.0 | Turn left at Lime Hollow Road. |
| 5.3 | 0.2 | Proceed to Stupke Road for STOP 4. |

STOP 4. USGS TEST WELL #381

In 1986 trichloroethylene (TCE) was discovered in water of private wells near this site. Well #381 was established to monitor water chemistry and provide data to determine the source of the TCE.

TABLE 3—Log of Well #381

Depth (ft)	Description
0-106	sand and gravel
106-110	till
110-175	varved silt and clay
175-216	sand and gravel
216-220	silt and clay
220-245	sand and gravel
245	bedrock

Measured data:

    Surface altitude: _____   Depth to water: _____   Water table altitude: _____

With three (3) measurements of the water table altitude, we may determine the slope of the water table and the direction of flow of the water and any contaminants. How is this done?

Return to vehicles and continue northwest on Stupke Road.

6.0	0.7	Turn right at McLean Road.
6.6	0.6	Left on Fairview Road crossing Route 222 to Highland Road.
7.6	1.0	STOP 5.

STOP 5. HIGHLAND ROAD DEVELOPMENT

    This development was considered to be ideal for many years until wells became contaminated. The County is now attempting to require much larger building lots in these "rural" areas. Larger building lots will allow for the owners to have wells adequately spaced for septic systems and avoid future contamination. When this development was completed, the lot size was deemed adequate.

    Continue north on Highland Road.

8.2	0.6	Right on Hoy Road and a quick left on to Kinney Gulf Road.
8.3	0.1	Right on Sweeny Road.
8.7	0.4	Right on Blue Creek Road.
9.1	0.4	STOP 6.

STOP 6. BLUE CREEK ROAD SUBDIVISION

When the Blue Creek Road Subdivision was first proposed, it had 16 building lots. The subdivision was then resubmitted to the Cortlandville Planning Board with 14 building lots. The problem with the subdivision was in the slope of the land and storm water runoff. At the conclusion of a extensive discussion regarding storm water, a motion was made by a planning board member to approve the subdivision, as requested, incorporating the disclaimer: "The design and Town of Cortlandville acceptance is not to be construed as a guarantee, expressed or implied, that the storm water management facilities will function properly, and the Town of Cortlandville assumes no liability should the storm water management system fail to function properly." The motion was seconded and the vote recorded with three "ayes" and two "nays".

Continue east on Blue Creek Road.

9.6	0.5	Turn left on Cosmos Hill Road.
11.3	1.7	Proceed to Route 90.
12.9	1.6	Turn right on Route 90 and continue to Route 11.
14.4	1.5	Turn left on to Route 11 and on to Suit-Kote gravel pits on the north side of the Village of Homer for STOP 7.

STOP 7. SUIT-KOTE HOMER GRAVEL PIT

Many gravel pits exist in this vicinity. The dollar value of aggregate in Cortland County is greater than what California has realized from gold; however, mining has prevented future agricultural activities or residential development. This is a major problem in land-use planning. If we don't mine here, where do we?

At this site we want to compare and contrast the aggregate with our sample analyzed at Stop 3.

Measured data:

Clastics: _____%   Carbonates: _____%   Crystallines: _____%

Return south through the Village of Homer.

16.0	1.6	Turn left at Albany Street.
16.6	0.6	At the Y-intersection, turn left and continue on Lighthouse Hill Road.
16.7	0.1	Turn right on to Route 13 and very shortly turn left on to Loring Crossing Road.

STOP 8. SUIT-KOTE LORINGS CROSSING PLANT.

Here we want to examine the location of the plant with respect to the Tioughnioga River. What safe-guards are in place to protect the river in case of the failure of an asphalt storage tank?

17.2	0.2	Continue on Route 112, over the Tioughnioga River, to East River Road.
19.3	2.1	Turn right on to East River Road and southward to Route 11.
20.0	0.7	Turn left on to Route 11 and proceed to the Polkville Plant.

STOP 9. SUIT-KOTE POLKVILLE PLANT

Suit-Kote is the largest producer of asphalt in central New York. Although business is important for the welfare of any community, this type of business creates problems for the community. We will examine such problems as: noise, dust, odor, loss of land from excavation, and water pollution.

Source Reports submitted to the Department of Transportation must contain a petrographic analysis of the materials before crushing. In addition to examining potential problems, we will compare aggregates from the face of the deposit to processed materials ready for market.

Measured data:

Clastics: _____%   Carbonates: _____%   Crystallines: _____%

We have now examined several aggregate sites from which gravels are mined for road construction.

What are the qualities we want in a durable aggregate?

What could have been the provenance for each of these deposits?

Return to SUNY Cortland.

END OF FIELD TRIP

Trip B-1

SUBSIDENCE AND LANDSLIDES IN TULLY VALLEY, CENTRAL NEW YORK

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INTRODUCTION

On this field trip we will examine two aspects of recent geologic activity in the Tully Valley: Fractures apparently opening in response to subsidence and slow-moving earth slides located on tributaries to the Tully Valley. The following guide is divided into two sections that reflect these two different features of the Tully Valley region. Please note that these stops are on private property and PERMISSION MUST BE OBTAINED PRIOR TO ACCESSING THE PROPERTY.


OPEN FRACTURES IN THE TULLY VALLEY

*[This section prepared by G. Gleason and W. Hackett]*

The purpose of this portion of the field trip is to examine the spatial relation between sets of open fractures and local subsidence in the Tully Valley, Onondaga County, NY. The study areas are located on the east and west slopes of the Tully Valley in Onondaga County (Figure 1). On the Otisco Valley USGS 7.5 minute topographic quadrangle, this location is at the north end of what were two brine fields. On this field trip we will visit the study area on the east side of Tully Valley (STOP 1 of Figure 1).

Local Geology of the Fractures

Exposed bedrock in the study area is Middle Devonian Hamilton Group shale. The strata are horizontal to sub-horizontal. Older strata at depth include layered rock salt deposits of the Upper Silurian Salina Group (~1,200 to 1,400 ft below grade) that have been mined by solution methods in the last 150 years. The valley was carved out by glaciers during the last ice age (10,000 to 20,000 years ago), and as a result has the typical
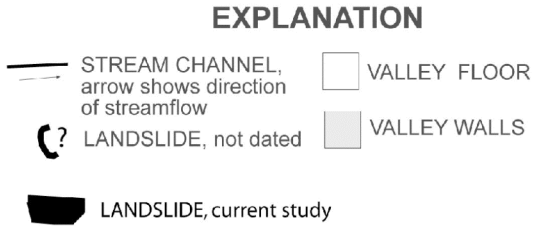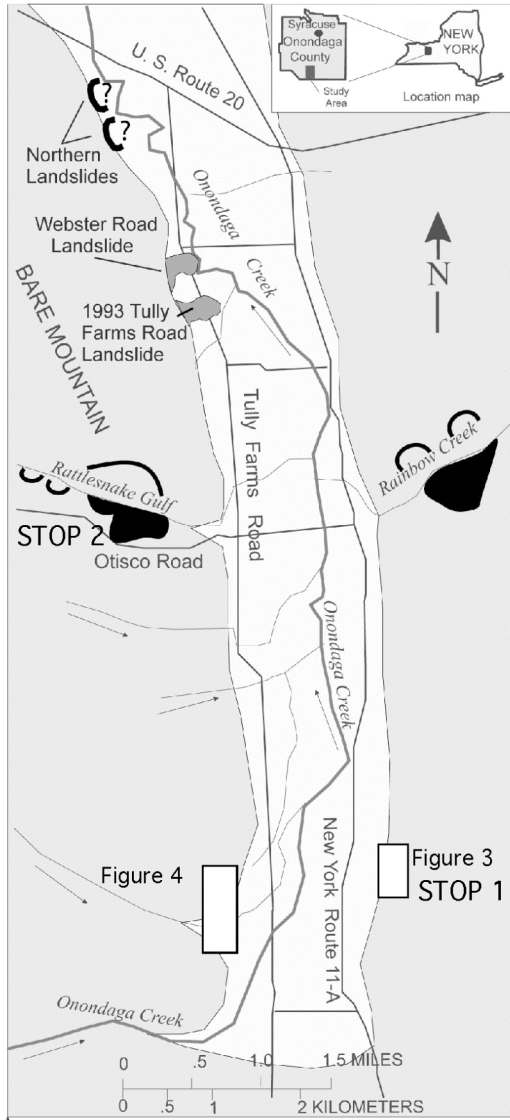
FIGURE 1—Locations of field trip stops. Mapped fractures are in boxes labeled Figure 3 & 4. Location of Rattlesnake Gulf and Rainbow Creek landslide areas in relation to the southern end of the Onondaga Creek Valley (Tully Valley), in central New York.
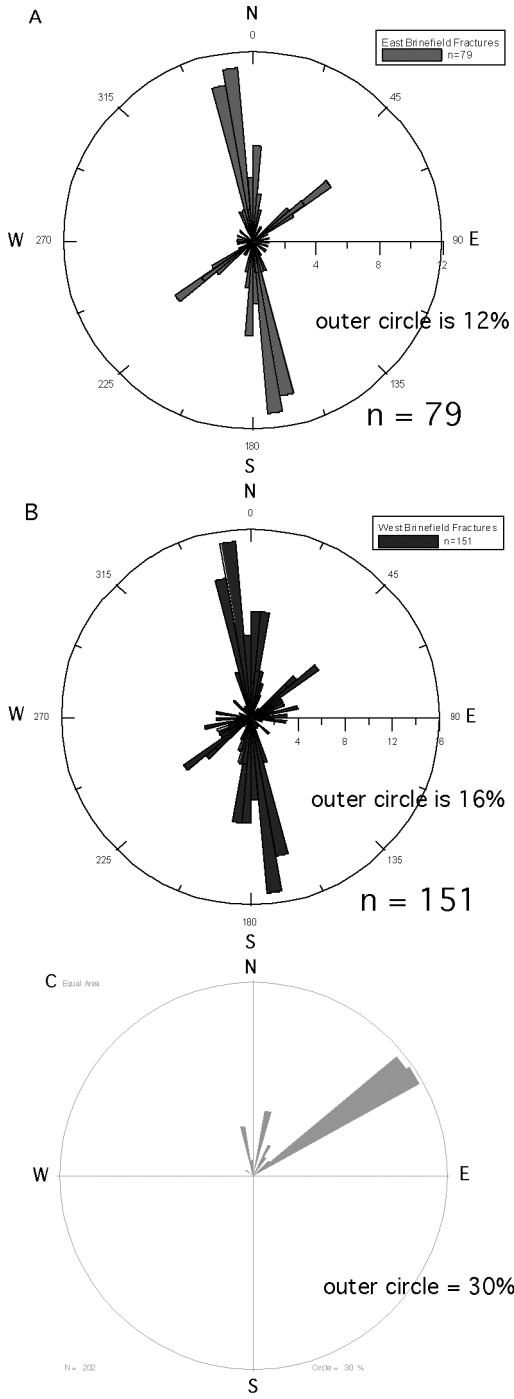
FIGURE 2—Rose diagrams of trends of fractures in Tully Valley. A) Number of fractures on East side. B) Number of fractures on West side. C) Length of fractures on East side. In all plots, data are in bins of 5 degree increments.

U-shape created by valley glaciers. Subsequently, the valley was partially filled with glaciolacustrine and fluvial deposits related to the glacial meltwater (Kappel & Miller 2003).

In Central New York, three joint sets have been identified and studied (e.g., Parker 1942; Engelder 1982; Engelder et al. 2001). Following Parker (1942) we will refer to these as set I striking roughly N10W; set II striking E-W, and set III striking roughly N65E. Sets I and II have been interpreted as cross-fold and strike joints developed during the Alleghanian orogeny as they have a geometric relation to other structures of that time (Parker 1942; Engelder and Geiser 1980). The origin of set III is more enigmatic. Crosscutting relations between the joint sets have not been well exposed. A correlation between the orientation of set III and the horizontal shortening direction of the current stress field lead Engelder (1982) to propose that the set was neotectonic. However, further work by Gross and Engelder (1991) on neotectonic joints in other localities demonstrated that the set III joints could not be young. More detailed work on the set III joints in the Finger Lakes region of NewYork revealed that these joints are offset by sets I and II, thus pre-dating the Alleghanian orogeny (Engelder et al. 2001).

Brine Mining History

Solution brine mining began in the Tully Valley in the late 1800's. Solution brine wells remove material by pumping water into the desired halite bed, allowing it to dissolve the halite, and then pumping brine back out through a different chamber in the well. The material was removed from halite beds in the Syracuse Shale, which are found at approximately 1,200 to 1,400 feet of depth. This method was further expanded through the technique of drilling wells in linear patterns allowing the dissolution cavities to connect (Yanosky and Kappel 1998). "Wild Brining", the practice of allowing unknown amounts of groundwater to infiltrate a well and dissolve halite as well as allowing brine to leave the well, was also practiced in the Tully Valley (Briggs and Sanford 2000). This technique did not allow for an understanding of cavity geometry or size and therefore potentially increased the possibility of collapse.

The brine wells were drilled on both sides of the valley floor at the south end of the valley. The fields were appropriately named the east and west brinefield. Mining began first in the east field around 1888. Production in both fields peaked around 1950-1960 before the east field was closed in 1960. The west field continued production until 1988 when the last wells were shut down. Ultimately, the brining period resulted in the removal of approximately 31,000 acre-feet of halite. With the removal of this quantity of material, widespread land surface subsidence occurred. Some areas on the valley floor were shown through repeated surveying to have subsided up to 40-50 feet over several decades. Subsidence observed on the valley floor was widespread, and open bedrock fractures appeared upslope on the valley walls of both the east and west fields.

Description of the Fractures

The fractures are either in exposed bedrock or observed as linear depressions ("coffin holes") in the covering soil. Fractures in the bedrock are typically open, and have gaps from 10 to 60 cm wide. Depths of these openings vary from 0.5 to 15 m. Many times the linear depressions are along a line with level "bridges" or covered segments between them. The width of these "coffin holes" probably reflects collapse of the soil into the bedrock fracture below. The linear depressions are typically 1 to 3 m wide, although their lengths vary considerably (2 to 10 m).

The trends of the on-line linear depressions were measured, and dip was assumed to be vertical. When the fractures in the bedrock were exposed, both trend and dip were measured on the fracture surface. On both sides of the valley, the dominant set of fractures trends just west of north-south (NNW; Figures 2a & b) corresponding to set I of Parker (1942). A second set trends 060˚, corresponding to set III of Parker (1942). A few E-W trending fractures were also recorded (set II of Parker, 1942). However at least on the east side, the 060˚ direction clearly dominates with 60% of the **total length** (+/- 5˚), and each other direction accounts for less that 10% of the total (Figure 2c).

In the east brine field, the open fractures are grouped up-slope (to the east) from three sinkholes, two of which are easily seen on the map (Figure 3). This spatial relationship to sinkholes is also observed in the west field, where fractures are more numerous above the three sinkholes in that field (Figure 4).
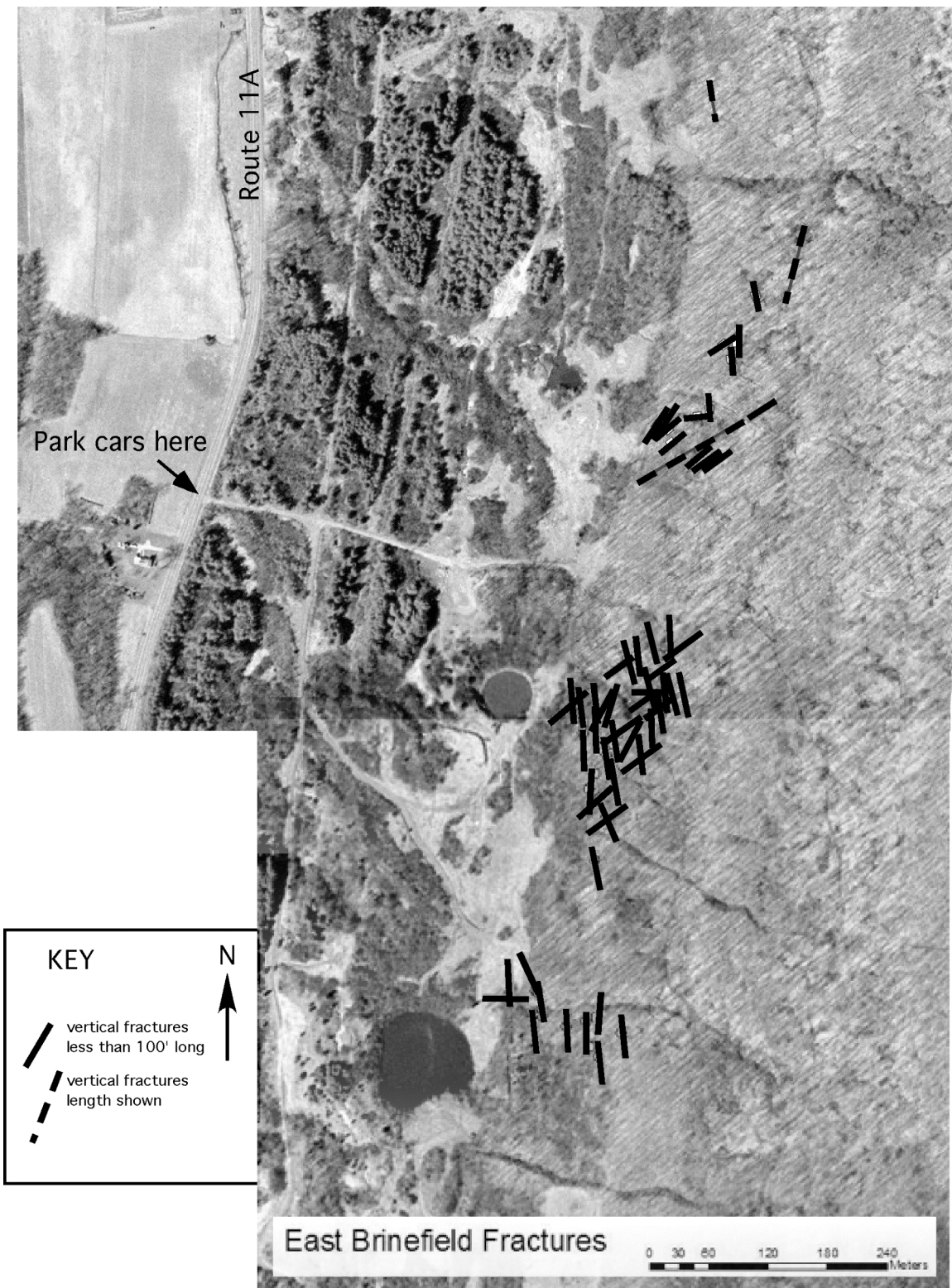
FIGURE 3—Aerial photo of STOP 1 on the East wall of Tully Valley. Open fractures cluster upslope from sink holes

FIGURE 4—Aerial photo of the West wall of Tully Valley showing distribution of fractures.

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PRESENT CONDITIONS AND HISTORIC DENDROGEOMORPHOLOGICAL ASSESSMENTS OF THE RAINBOW CREEK AND RATTLESNAKE GULF LANDSLIDES

[This section prepared by K. Tamulonis]

Introduction to the Landslides

The Tully Valley, New York is a 6-mile-long glacial trough located in the eastern Finger Lakes region of the Allegheny Plateau and has a landslide history dating back to 9,870 ^{14}C yr B.P. In 1993, the largest landslide in the state since the early 1900's occurred on the west wall of the valley. Presently, two slow-moving earth slides, Rainbow Creek and Rattlesnake Gulf, are located in tributary valleys to the Tully Valley (Figure 1)**.** Precipitation, ground-water level, and land-surface movement measurements and dendrogeomorphology indicate that movement on the Rainbow Creek and Rattlesnake Gulf landslides occurs on two scales: shallow displacement following precipitation events, and deep-seated, rotational movement, which was determined through analysis of tree ring data, a five-year moving precipitation average, and time at the Rattlesnake Gulf landslide.

Rainbow Creek Landslide

The main body of the Rainbow Creek landslide is located on the south side of Rainbow Creek, covers approximately 34 acres, and is at an elevation of 820 feet to 1180 feet above sea level. Rainbow Creek is the only major tributary to the Onondaga Creek on the eastern Tully Valley wall. The landslide surface slopes between 20°NW to 45°NW, and the landslide material is primarily laminated clay and silt and well-sorted fluvial sand and gravel. The unconsolidated sediment is underlain by stable Middle Devonian shale and siltstone of the Delphi Station and Lower Pompey members of the Skaneateles Formation of the Hamilton Group (personal communication, Gordon Baird, June, 2007). Throughout the summer and fall of 2006, landslide material along the northernmost scarp was composed of more than ten feet of laminated clay and silt overlain by a 40-foot-thick layer of interbedded, well-sorted silt, sand, and gravel, which coarsened upward. This material was rotated and oriented at N9°-75°E, 12°-34°SE. Following the 2006-2007 winter, portions of the fluvial silt, sand, and gravel package along the northern scarp were eroded, exposing approximately 25 feet of rotated, laminated silt and clay, underlain by eight feet of poorly sorted glacial till. This silt and clay is oriented at N59°-61°E, 9°-20°SE. Smaller slumps are located on both the north and south sides of this tributary stream valley for a distance of at least 0.5 miles east of the main landslide body, and several landslide 'scars' (where bedrock is exposed within the sediment covered stream banks) also surround the landslide.

Rattlesnake Gulf Landslide

The Rattlesnake Gulf landslide covers approximately 23 acres, and the landslide surface slopes 15° to 45° northwest at an elevation of 1140-1260 feet above sea level. Unlike the Rainbow Creek landslide, the Rattlesnake Gulf landslide does not have smaller active landslides upstream of the main body, though several scars are located upstream and on the north side of the stream valley several bedrock landslide scars are seen directly across from the active landslide. Landslide material is primarily composed of laminated silt and clay, although well-sorted fluvial sand is exposed at several locations within the landslide. The rotated silt and clay is oriented at N46°-88°W, 14°-55°SW, and the middle Devonian Delphi Station, Pompey, and Butternut members of the Skaneateles Formation underlie the unconsolidated, sliding material (personal communication, Gordon Baird, June, 2007).

Landslide Data Collection

Beginning in Summer 2006, displacement and ground-water level measurements have been recorded at the two landslides, and precipitation data recorded from the Tully Valley floor (Table 1). Data records from both landslides show that shallow displacement corresponds to precipitation events and heightened ground-water levels, and increased ground-water levels in the shallow, unconsolidated sediments lag precipitation events by up to two days. The Rainbow Creek landslide and the northwest portion of the Rattlesnake Gulf landslide experienced the most activity in July 2006, which was the wettest month for the 11-month data collection period (Figures 5 and 6). The center of the Rattlesnake Gulf landslide was most active in March 2007 (Figure 7). Displacement magnitude is greatest in the northwest portion of the Rattlesnake Gulf landslide due to proximity

to the over-steepened scarp and the unstable toe, and there is an eight-month lag time for shallow stress release to translate up slope from the northern scarp to the center of the landslide. The activity pattern at the Rainbow Creek landslide is similar to that of the northwest portion of the Rattlesnake Gulf landslide.
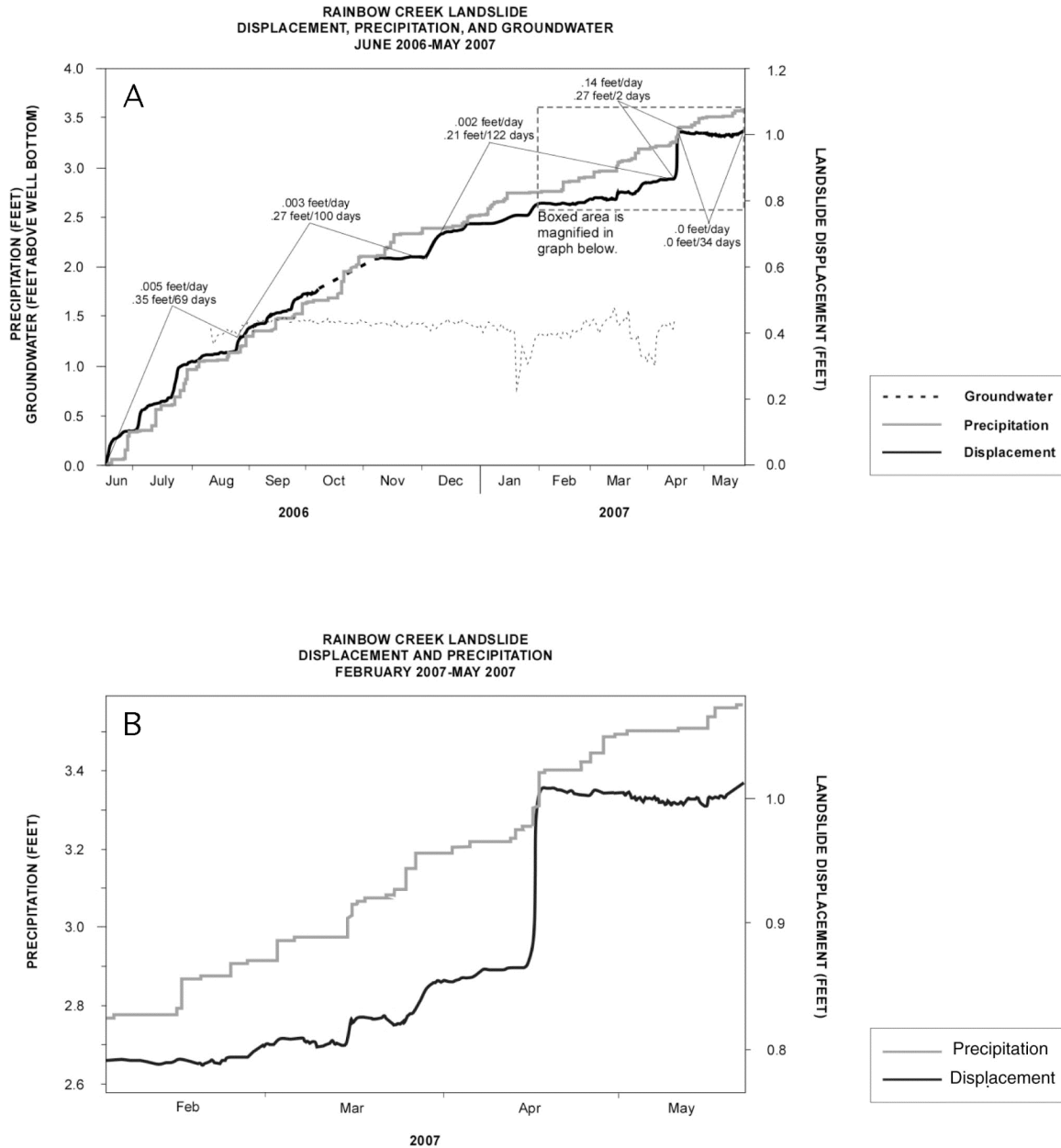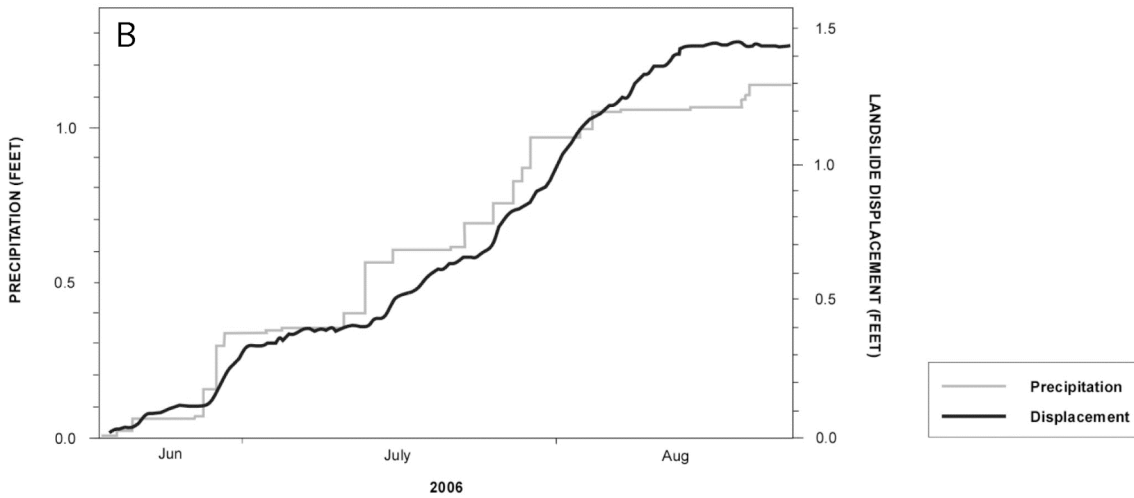

FIGURE 5—A) Graph of precipitation, ground-water level, and land-surface displacement in the Rainbow Creek Landslide area – June, 2006, through May, 2007. B) Detail of precipitation and land surface displacement from February, 2007 through May, 2007.
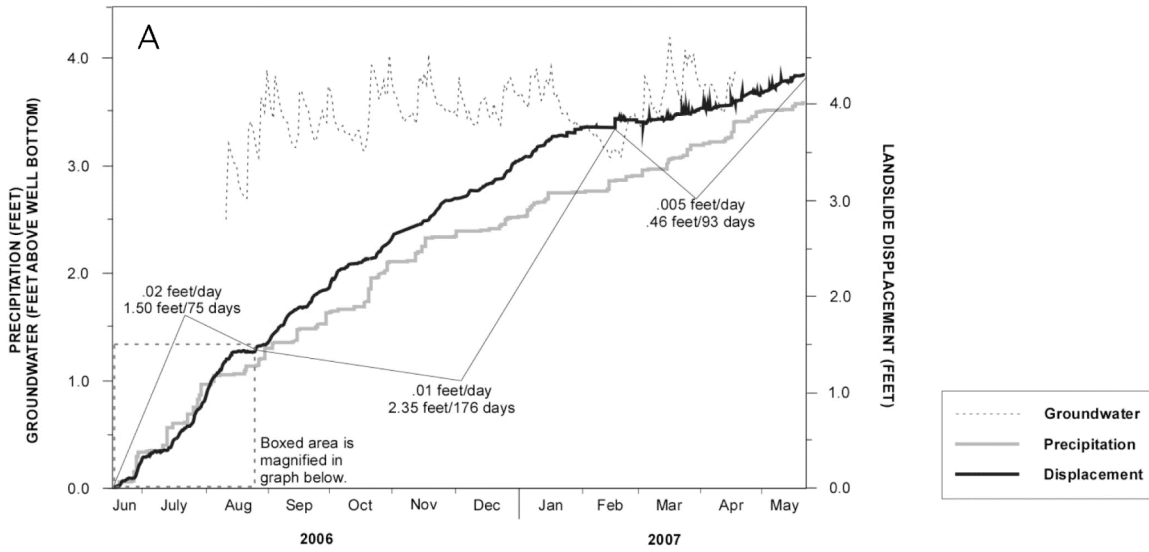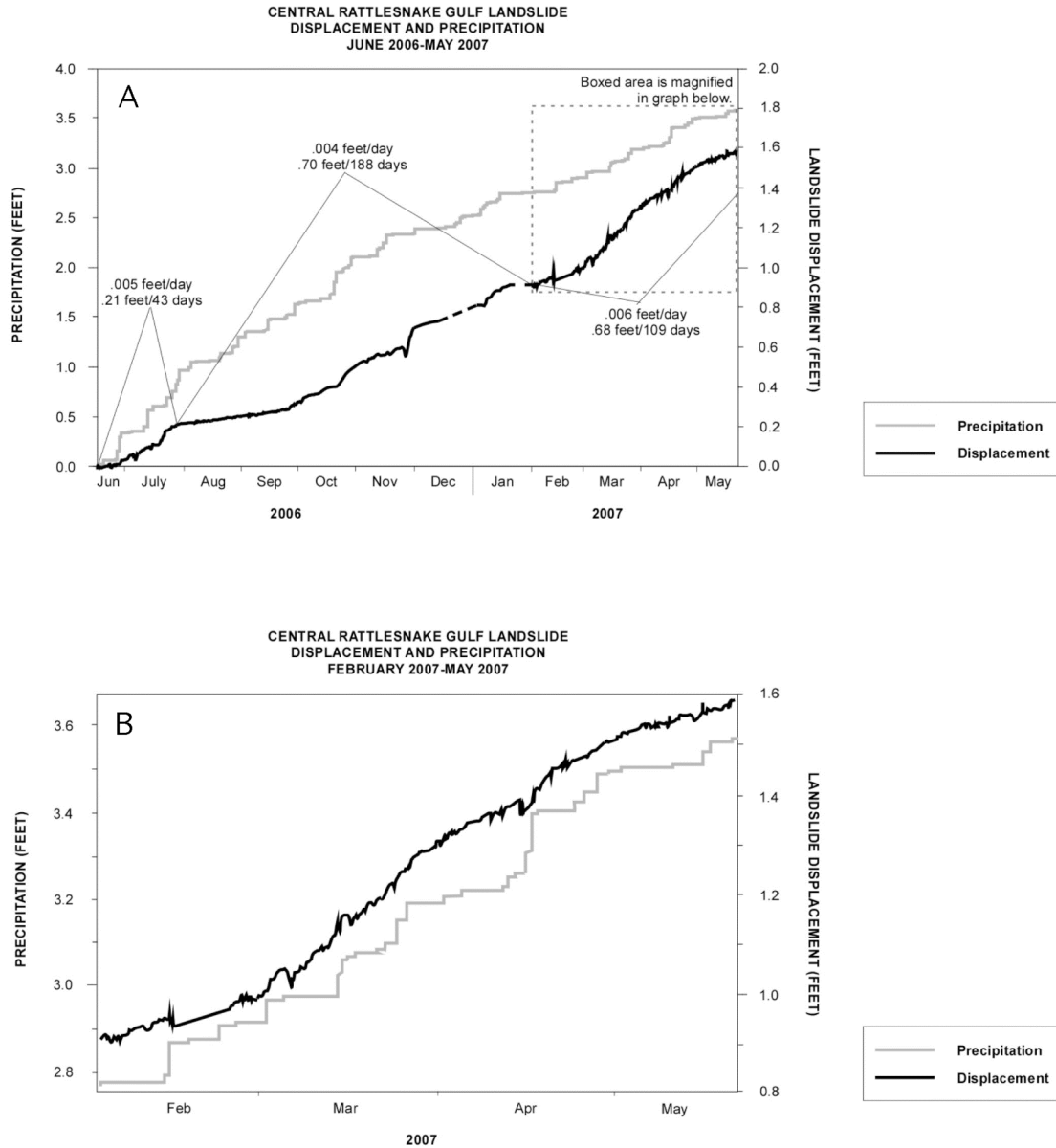
FIGURE 6—A) Graph of precipitation, ground-water level, and land-surface displacement in the northwest part of the Rattlesnake Gulf landslide, -- June, 2006 through May, 2007. B) Detail of precipitation and land surface displacement from June, 2006 through August, 2006.

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FIGURE 7—A) Graphs of precipitation, ground-water level, and land-surface displacement in the central part of the Rattlesnake Gulf landslide—June, 2006 through May, 2007. B) Detail of precipitation and land surface displacement from February, 2007 through May, 2007.

TABLE 1—Monthly precipitation and displacement for the Rainbow Creek and Rattlesnake Gulf Landslides Tully Valley, New York.

|	Precipitation (inches)	Rainbow Displacement (inches)	Northwest Rattlesnake Displacement (inches)	Central Rattlesnake Displacement (inches)
2006				
June*	3.95	1.20	3.6	0.48
July	7.61	2.52	8.4	2.04
August	4.01	1.32	6.0	0.36
September	4.03	1.20	7.2	0.96
October	5.68	1.08	6.0	2.16
November	2.81	0.24	4.8	2.28
December	2.12	1.20	4.8	1.32
2007				
January	2.83	0.72	3.6	1.32
February	1.77	0.12	1.2	1.08
March	3.42	0.72	1.2	3.60
April	3.73	1.68	2.4	2.40
May**	0.93	0.12	2.4	1.08
Total	42.89	12.12	51.6	10.08

* Precipitation and displacement record-June 19-30, 2006

** Precipitation and displacement record-May 1-21, 2007

Dendrogeomorphology

Dendrogeomorphology uses dendrochronology (tree ring study) to determine landform evolution because tree rings provide dates for significant historic geomorphic events. A total of 97 tree cores and cross sections were collected from and surrounding the two active landslides. Tree rings were measured and reaction wood was identified in order to determine years when tree ring growth changed from concentric to eccentric (Figure 8), and it was assumed this change in growth pattern occurred due to landslide movement. With this data, event indexes were generated to temporally analyze landslide activity. At the Rattlesnake Gulf landslide, a multiple Fourier function regression model correlates the event index, five-year moving-precipitation average, and time, implying there is a multi-year lag between precipitation and deep ground-water discharge into the landslide, which causes more deeply-seated movement. Model cyclicity also suggests that Rattlesnake Gulf landslide activity has an approximate 70-year cycle, with destabilizing years noted in 1927 and 2000. Drought periods followed by persistent (3 month or greater) above-average monthly precipitation, corresponding high stream discharge eroding the landslide toe, and harvesting of mature trees above and within this landslide may also be displacement triggers. At the Rainbow Creek landslide, there is no clear correlation between annual precipitation, three or five-year moving precipitation averages, time, annual average temperature, and the event index.

ACKNOWLEDGEMENTS
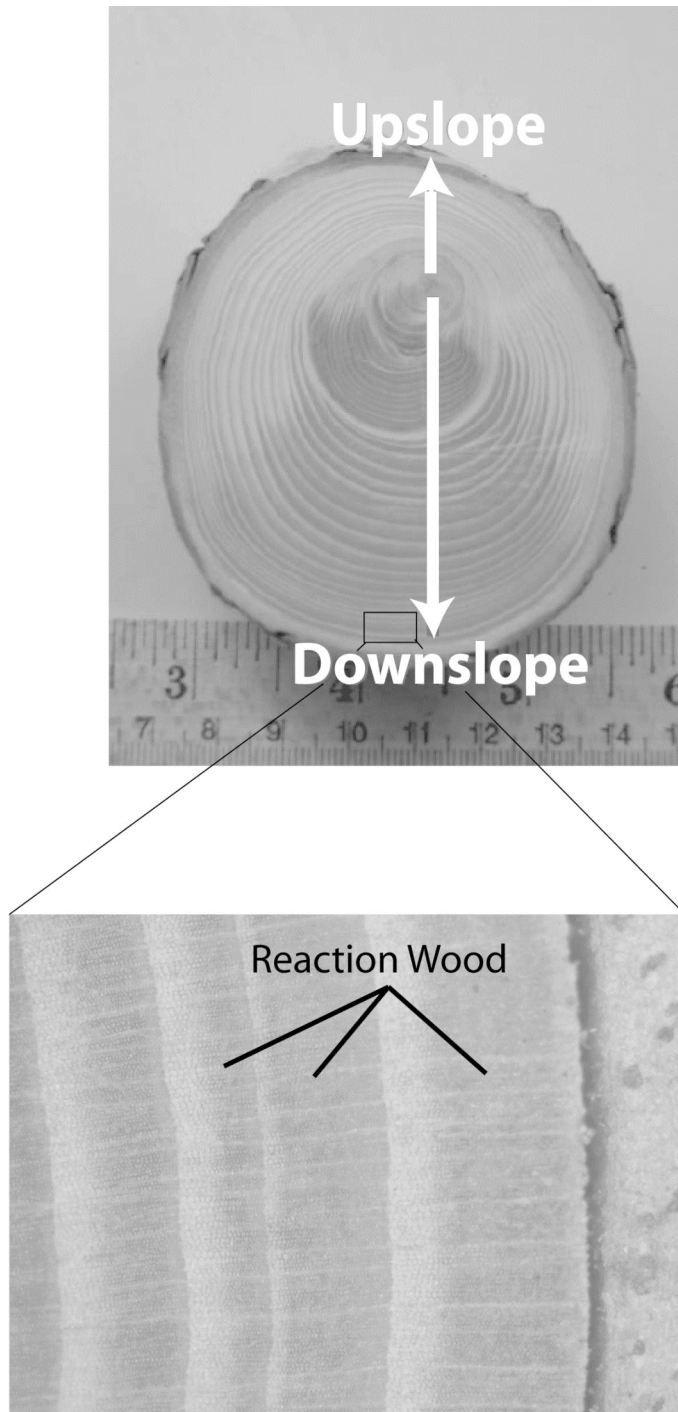
FIGURE 8—Example of a small trees cross-section showing the change from concentric tree-ring growth to eccentric tree-ring growth due to the tree becoming tipped over in a landslide and correcting its orientation back to a near-vertical position via reaction wood growth.

REFERENCES CITED

Briggs, P and Stanford, K., 2000. The Ups and Downs of Post-Closure Subsidence- Monitoring at the Tully Valley Brine Field, New York. Solution Mining Research Institute. Fall meeting

Engelder, T., 1982. Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America? Tectonics, 1, 161-177.

Engelder, T. and Geiser, P., 1980. On the use of regional joint sets as trajectories of paleostress fields during the development of the Appalachian Plateau, New York, J. Geophys. Res. 85, 6319-6341

Engelder, T., Haith, B.F. and Younes, A., 2001. Horizontal slip along Alleghanian joints of the Appalachian plateau: evidence showing that mild penetrative strain does little to change the pristine appearance of early joints, *Tectonophysic*s, 336, 31-41.

Gross, M. and Engelder, T., 1991. A case for neotectonics joints along the Niagara Escarpment. Tectonics, 10, 631-641.

Kappel, W., & Miller, T., 2003. Hydrogeology of the Tully Trough: Southern Onondaga County and Northern Cortland County, New York. Water-Resources Investigations Report 03-4112, USGS.

Parker, J.M., 1942. Regional systematic jointing in slightly deformed sedimentary rocks. Geol. Soc. Am. Bull., 53, 381-408.

Yanosky, T. and Kappel, W., 1998. Tree Rings Record 100 Years of Hydrologic Change Within a Wetland. U.S. Geological Survey Fact Sheet 057-97.

ROAD LOG FOR TRIP B-1

SUBSIDENCE AND LANDSLIDES IN TULLY VALLEY, CENTRAL NEW YORK

The two field sites described in the road log are on private property. As such, it is imperative that we respect the rights and wishes of these land owners. Please do not visit these sites on your own, as it may jeopardize future field-trip opportunities. Over the past 15 years it has become increasingly difficult to maintain our access agreements to these sites as individuals and even groups of people have entered these properties without obtaining land-owner permission. We strive to maintain good relations with the land owners and do not want the inappropriate actions of a few to ruin the educational opportunities for many others who wish to enter these areas. The USGS can, and does act as the 'point of contact' for the property owners and we would be happy to facilitate your future access to these sites. Please contact Bill Kappel (USGS, Ithaca, NY) for any questions as to access for you and your classes.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Start at parking lot north of Old Main building on SUNY Cortland campus. From parking lot turn left (E) on Gerhart Dr., at the yield sign turn left (N) onto Graham Ave.
		At bottom of hill (traffic light) turn right (E) on to Groton Ave, (also Rt. 222).
		At Main St. (traffic light) go straight through intersection, road becomes Clinton Ave. Stay on Clinton Ave. Bear left (N) at the Mobil gas station (traffic light), road is still named Clinton Ave. and go under I-81 overpass.

0.0	1.6	CORTLAND CAMPUS TO I-81N
15.9	14.3	Turn left (W) onto I-81 North entrance. Exit at Tully exit (exit 14). At stop sign at end of ramp, turn left (N) on to Rte 281 N.

I-81N TO TULLY INTERCHANGE

At traffic light turn left (W) on to Rt. 80 and go through the I-81 overpass and take an immediate right (N) on to Route 11A. Go down the back side of the Tully (Valley Heads) Moraine and continue north until you see a red house on your left. Just ahead on your right you will see a red pipe gate, pull over and park to the side of the locked gate. You need prior permission to enter this property.

18.4	2.5	NYS-RTE 80/RTE 11A TO STOP #1 - BRINEFIELD FRACTURES

Walk east on the dirt road to base of slope (~1200 ft), proceed north (left) on dirt road approximately 300 ft. Stop where fracture from northeast intersects road.

STOP 1A. LONG, NORTHEAST TRENDING FRACTURE NEAR EAST BRINEFIELD

This is one of the largest fractures in this area, both in exposed length and depth. This vertical fracture trends 060˚and is exposed for 540 feet along strike. The width of the fissure opening is about 0.3 m (1 ft) and the measured depth is 11.5 m (38 feet). The orientation of this fracture correlates to set III of Parker (1942). Possible fringe cracks intersect the southeast wall of the fissure, trend 075˚, and are vertical.

Walk in the 060˚ direction along strike to place where tree roots jut over the fissure, but the tree is no longer there.

A tree trunk was removed from this spot last spring for tree ring analysis. The time of root exposure, and thus the time of fracture opening, may be determined by recognition of a change in the characteristics of the tree rings of the roots. Analysis is still in progress as of this writing. The hypothesis is that the fracture opened in response to collapse at depth after the final surge in salt removal by solution mining about 1950-1960.

Walk down slope to road, continue North to small east-west trending gully.

STOP 1B. CROSS-CUTTING RELATIONS BETWEEN FRACTURE SETS

On the South side of gully the two sets of open fractures intersect. The longer vertical fracture trends 055˚, is open about 0.3 m (1 ft), and is straddled by a tree—not unlike what the tree at the previous stop looked like before removal for tree ring analysis. Again, oriented as set III of Parker (1942). Possible fringe cracks intersect the southeast wall of the fissure, trend 083˚, and are vertical. A small fracture oriented 352˚/86˚E (corresponding to set I of Parker, 1942) appears to offset the 055˚ fracture. If this is correct then the timing relationship is consistent with Engelder et al., (2001); that is, set III is older than set I and II hypothesized to be related to the Alleghanian orogeny.

Walk west to return to dirt road. Walk south 275 m (~900 ft) past intersection with dirt road to next group of open fractures.

STOP 1C. CLUSTER OF FRACTURES UPSLOPE FROM SINK HOLE

Note that no fractures occur along this road between the long fracture of STOP 1A, and this cluster. This cluster of open fractures appears to be spatially related to the sinkhole just to the west on the valley floor. Walking uphill along an east trending dirt road, one can see several fractures of the three orientations on both sides of road.

Return to dirt road by walking west, downhill. Walk north 75 m (250 ft) to intersection with dirt road we came in on. Turn west, walk back to US11A and cars.

23.9	5.5	BRINFIELD FRACTURES TO STOP #2 – RATTLESNAKE GULF LANDSLIDE

Turn around on Route 11A back (S) to Solvay Road (~1 mile) and take a right. Travel to the end of Solvay Rd and take a right on Tully Farms Road (N). Travel about 3 miles until the intersection of Otisco Road. At Otisco Rd, turn left (W) and travel uphill ~1 mile to a small red house on the right. You will need prior permission to enter this property as the driveway is small, and the home owner does not want surprise 'guests'!

At the west side of the house follow the road down the hill along the tree line to the NW. At the stream channel, turn uphill SW and cross the channel at grade. Follow the logging road and take the second spur to the right (N then NW). Follow that road until you see a wooden recorder box on your right (about 10 feet lower than the road). The slide is further to the north, downslope – again it is quite dangerous to walk around the area as many fractures are hidden in the understory and the slope is quite unstable and slippery, when wet.

STOP 2. RATTLESNAKE GULF LANDSLIDE

At this site we will observe the ongoing land-surface movement of the Rattlesnake Gulf Landslide area. Review of New York State aerial imagery (circa 1937 to present) indicates that this slide area has slowly coalesced into the large slide we will see today. Also, the use of dendrochronology to determine previous landslide movement allows the casual observation of 'J-ayed' trees to be refined to specific years of movement. The genesis of the slide continues today at two different scales – shallow displacement related to rainfall infiltration into the shallow weathered fine-grained soils, and more deeply seated ground-water induced movement. Also the effects of toe-cutting of the slope can be observed along Rattlesnake Gulf Creek which influences the movement of slide material as it makes its way from the upper scarp to, and into the creek, to be carried down to the floor of the Tully Valley and Onondaga Creek.

To return to SUNY Cortland campus, take Otisco Road downhill to T
Farms Road, turn right (S) and take Tully Farms Road to Solvay Road
left (E) and at the intersection with NYS-Route 11A, turn right (S) and
uphill to the Burger King and NYS-Route 80. Cross Route 80 and take
entrance ramp on to I-81 South – left hand entrance. Travel south on I
back to the Homer exit. Exit at Homer, and at the 'T' (NYS-Route 28
intersection, turn left on to Route 281 (S). Follow 281 until it intersect
Route 222 (Groton Road at 3rd traffic light) – get in left hand turning l
the light. Turn left on to 222 (Groton Road) and go to Graham Road, t
the hill to the SUNY Cortland campus.

45.1	21.2	LANDSLIDE TO SUNY CORTLAND campus

END OF FIELD TRIP

Trip B-2

MINERALIZATION AT THE KNOX UNCONFORMITY
AND THE WESTERN ADIRONDACK PALEOSURFACE

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INTRODUCTION

The Ordovician Knox unconformity (Landing, 2003) is well exposed in the bed of Roaring Brook, on the eastern margin of Tug Hill in the Black River Valley. Here basal arkosic sandstones and minor pebble conglomerates of the middle Ordovician Pamelia Formation rest on middle Proterozoic feldspar-quartz gneiss. Immediately below the nonconformable contact, the gneisses show strong evidence of spheroidal weathering (Orrell and Darling, 2000) which is interpreted to be of middle Ordovician age because the spheroidal weathering diminishes with depth below the unconformity and other exposures of feldspar-quartz gneisses in the area show no evidence of modern-day spheroidal weathering.

The Knox Unconformity in New York State represents a period of subaerial exposure when upper Cambrian and lower Ordovician sediments were eroded from the region (Isachsen et al., 1991). Subsequent submergence resulted in deposition of the Black River and Trenton Limestones which was followed by a deepening foreland basin and subsequent collision during the Taconic Orogeny. Paleomagnetic studies demonstrate that this part of the North American continent was located in tropical to subtropical latitudes south of the equator (Niocaill, et al., 1997), an interpretation consistent with the development of extensive chemical (spheroidal) weathering of felsic gneisses.

The preservation of Ordovician age chemical weathering is significant because other mineralogical evidence (in the form of cryptocrystalline quartz + iron hydroxide veins) preserved in gneisses at Roaring Brook may have formed from chemical weathering as well. Similar microcrystalline quartz + iron hydroxide "rusty" veins are preserved in elevated regions of the Adirondack Mountains further to the east. The specific age of these veins is unknown but are herein interpreted as having formed during Ordovician chemical weathering.

On this trip, we will examine spheroidal weathering and cryptocrystalline quartz veins hosted by middle Proterozoic gneiss at Roaring Brook. We will then drive to and climb Bald Mountain, near Old Forge, NY (in the west-central Adirondacks) and examine several "rusty" quartz veins on the way to and at the summit. We will then discuss the eastward projection of the Knox unconformity over the west-central Adirondacks and the possibility that rocks at the summit of Bald Mountain and other summits in the area were located only a short vertical distance (10's of meters?) below a middle Ordovician paleosurface, and that gradually decreasing summit elevations in the western and central Adirondacks may, indeed, represent a partly eroded relict erosional surface (Whitney et al., 2002, p.5; Miller, 1910, p. 39). Lastly, we will discuss this evidence in light of Isachsen's model (1975; 1981) of Tertiary to Holocene doming of the Adirondack region.

SPHEROIDAL WEATHERING AT ROARING BROOK

Figure 1 shows spheroidal weathering preserved in middle Proterozoic felsic gneiss in the stream bed of Roaring Brook. The spheroidal weathering is characterized by closely spaced (3-4 mm) bands of iron-hydroxide, the bands extending a few centimeters into the gneiss (from the joints). Microscopically, the bands are characterized by fine-grained iron-hydroxide, calcite, serpentine? and chlorite? Locally, the bands are filled with medium-grained calcite, suggesting an open fracture at some point.

At Roaring Brook, the lowermost strata of the Pamelia Formation (Middle Ordovician) rest directly on top of Proterozoic gneiss. Spheroidal weathering occurs directly below the (well exposed) nonconformable contact, but is observed only during low water levels (normally late summer). The spheroidal weathering directly below the nonconformity and ~30 meters downstream (location of Fig. 1) are the only places where it is observed. Both are located within one vertical meter of the nonconformity. Exposures of felsic gneiss farther downstream, which are a few meters below the projection of the unconformity, show little or no evidence of spheroidal weathering. The proximal relationship between the nonconformity and the spheroidal weathering is interpreted as evidence of Ordovician age chemical weathering.


FIGURE 1—Vertical view onto surface of spheroidal weathering preserved in middle Proterozoic felsic gneiss just below the Knox unconformity at Roaring Brook (Stop 1). Hammer for scale.
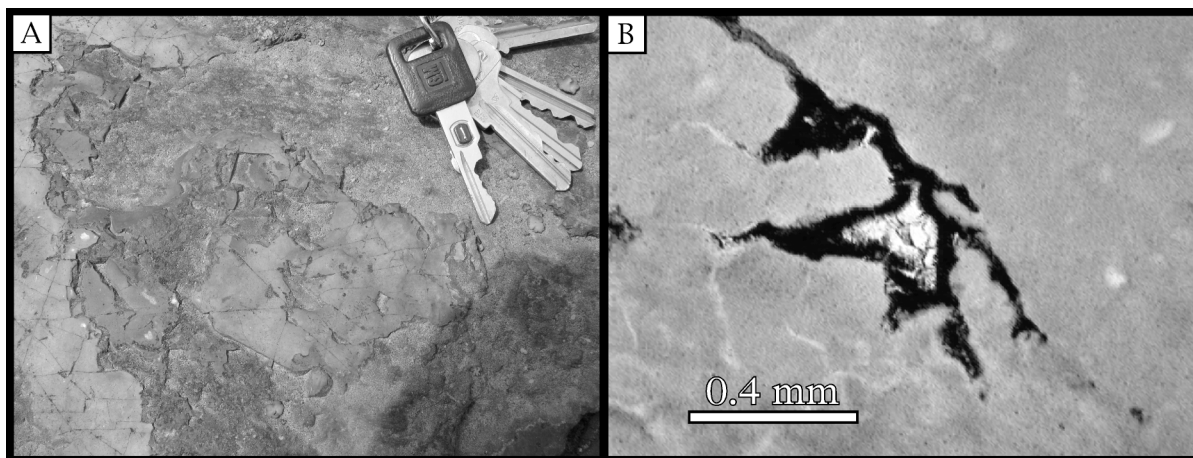
RUSTY QUARTZ VEINS AT ROARING BROOK

About 100 meters northwest of this location, where Canaan Road crosses over Roaring Brook, exposures of felsic gneiss (under the bridge on the northwest side) contain smoothly worn, sub-horizontal, thin (2 - 4 mm) veins of dark greenish-gray to rusty-brown cryptocrystalline quartz and iron hydroxide (Fig 2A). In thin section, quartz grains are barely resolvable, averaging perhaps 2 to 5 micrometers across. Some goethite? lined micro vugs are present which are then filled with slightly coarser-grained quartz (Fig 2B). The cryptocrystalline quartz and iron hydroxide veins are located approximately 3 to 4 meters vertically below the projection of the unconformity.

The origin of these veins is no doubt debatable, but I interpret them as having an origin related to Ordovician chemical weathering. Chemical weathering of feldspar (the dominant mineral in the hosting gneisses) normally yields silica according to the simplified reaction:

$$feldspar + acid + water \rightarrow clay + a\ dissolved\ cation + silica \qquad (1)$$
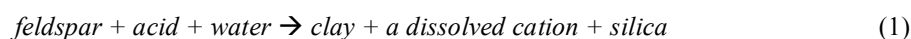
Product silica in reaction (1) can be in the form of silicic acid (H_4SiO_4) or precipitated quartz (chalcedonic or opaline). The exact mechanism of quartz precipitation is unknown. The sub-horizontal orientation (of the veins) is interpreted to reflect original pressure-release fracturing (sheeting) during middle Ordovician unroofing. Also, note the presence of three sets of steeply dipping joints preserved in the smoothly worn veins (Fig. 2A) indicating an age older than regional near-vertical jointing (i.e. pre-Alleghanian; Engelder et al., 2001).

FIGURE 2—A) smoothly worn, sub-horizontal, cryptocrystalline quartz vein located where Canaan Rd crosses Roaring Brook (Stop 1). Note faint joints cross-cutting vein. Keys for scale. B) Photomicrograph of cryptocrystalline quartz vein showing vug lined with goethite? and infilled with coarser-grained quartz.

RUSTY QUARTZ VEINS ON BALD MOUNTAIN

This popular mountain summit is characterized by an unusually large amount of exposed bedrock and therefore offers a greater opportunity to observe features that might not be seen on most western Adirondack summits. Note that Bald Mountain is located about 42 km east of Roaring Brook, but similar quartz + iron hydroxide veins are found here.

At a number of places along the ridge trail and at the summit, steeply dipping, thin (2–10 millimeter wide) rusty quartz veins cut across feldspar-quartz gneiss (Fig. 3). Foliation in the gneiss strikes approximately N60E,

but many of the rusty quartz veins strike northwest (Fig. 4). I know of no geological description of these veins except possibly a reference in David H. Beetle's (1948) book "Up Old Forge Way" where he (p. 95) describes, "…near the top, are thin rusty, veins of basalt." The rusty-brown color of the quartz + iron-hydroxide veins does, indeed, resemble typical rusty-brown weathering of basaltic rocks, but they are not basalt.

A total of 19 different rusty quartz veins were observed along the ridge northeast of and at the summit of Bald Mountain. *Yet, rusty quartz veins are rare, if not absent, in Adirondack gneisses at lower elevations*. I have searched for similar veins in exposed gneisses in western Adirondack rocks but have observed them only at or near mountain summits. For example, a few rusty quartz veins (3–4 mm wide) occur near the summits of Bare Mountain (~15 km southwest) at N 43.67501, W 75.07321 and Wakely Mountain (~32 km east) at N 43.73628, W 74.51366. A thin (1 mm) rusty quartz vein occurs at the summit of Stillwater Mountain (~17 km northwest) at N 43.86155, W 75.03381.


FIGURE 3—Vertical rusty quartz vein in felsic gneiss located ~10 meters south of Bald Mountain fire tower. Vein is approximately 8 mm wide. Hammer for scale.

The rusty quartz veins are characterized by microcrystalline quartz and iron-hydroxide. The iron-hydroxide is identified only by its brown, rusty color and flakey texture in thin section. Its specific mineralogy is unknown at the time of this writing but may likely be goethite. The iron-hydroxide does not occur as a staining of earlier formed quartz; it is intergrown with quartz and is interpreted to have formed syngenetically. The relative amount of quartz and iron-hydroxide are variable such that veins dominated by quartz are more resistant than surrounding gneisses (like that shown in Fig. 3) whereas iron-hydroxide rich veins are less resistant than surrounding gneisses.
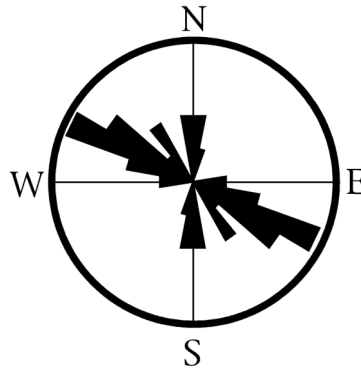
FIGURE 4—Rose diagram showing the strike of 19 rusty quartz veins measured on Bald Mountain. Dip is steep to vertical. Note prominent NW strike. Petal (or bin) size = 10°. Rose diagram created using R. Allmandinger's "Stereonet v 6.3 X" available at http://www.geo.cornell.edu/geology/faculty/RWA/programs.html

In thin section, the quartz and iron hydroxide grains are microcrystalline, averaging perhaps 15 to 30 micrometers across. In many of the veins, sharply angular fragments of coarser-grained (60 to 300 micrometers) quartz are present (Fig. 5A). Badly decomposed feldspar with a similar texture and grain size is locally present. The *coarser* quartz grains locally contain fluid inclusions of a similar composition and liquid-vapor phase ratios to that of typical fluid inclusions in quartz from Adirondack rocks (Lamb et al, 1991; Darling and Bassett, 2002; McLelland et al., 2002) and, therefore, are interpreted to have originated from the host gneisses. However, the coarser grains are *occasionally* rounded (Fig. 5B) which supports the idea that the veins may have originated, in part, as sandstone (or neptunian) dikes. Sand-filled dikes have been described in Proterozoic rocks of the Adirondack lowlands (Selleck, 2005).
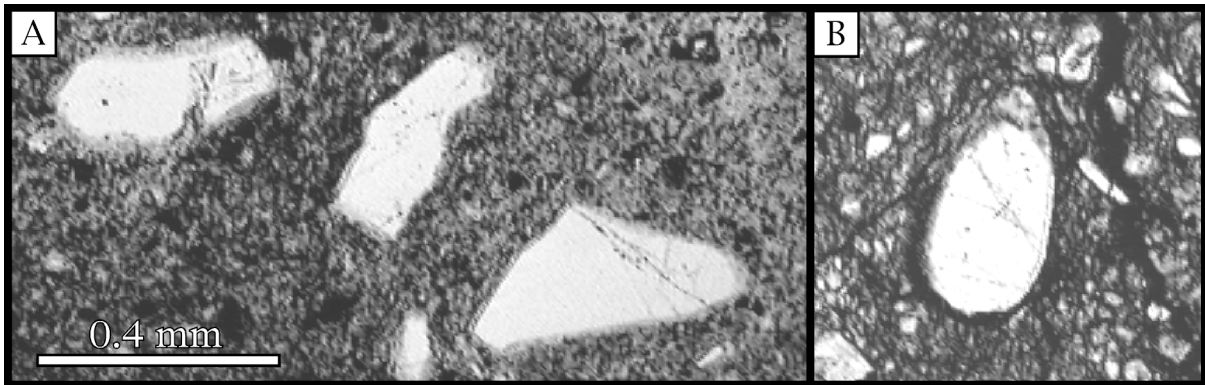
FIGURE 5—A) photomicrograph of rusty quartz vein showing microcrystalline matrix comprising quartz and iron hydroxide and angular fragments of coarser-grained quartz. B) similar photomicrograph showing a rounded grain of quartz (scale same as in A).

The mineralogy and texture of the rusty quartz veins coupled with the observation that they are most abundant at western Adirondack mountain summits suggest they have an origin related to chemical weathering similar to what is observed at Roaring Brook. This may be significant because it would represent evidence that the projected eastward extension of the Knox unconformity over the western and central Adirondacks (from the Black River Valley) is located at an elevation not far above (~10's of meters?) the present-day summit of Bald Mountain.

A MIDDLE ORDOVICIAN PALEOSURFACE?

If one looks southeastward from the summit of Bald Mountain, the summits of the western and central Adirondacks decline in elevation towards the west. A composite panoramic image of this western Adirondack vista (as viewed from Bald Mountain) is shown in Figure 6. A much better panoramic view (created by Mr. Carl Heilman of www.carlheilman.com) can be viewed at www.naturepanoramas.com/14001114.html.

A plane can be drawn through western Adirondack summits to show the westward decline in elevation, and Whitney et al. (2002, p. 5) suggested that this plane lies close below a Cambrian erosion surface in the Fulton Chain-of-Lakes area. They also noted that the plane projected close to the Precambrian / Paleozoic contact in the Black River Valley to the west. Miller (1910, p. 39) described the Precambrian surface of the southwestern Adirondacks as a peneplain and suggested that the region had been an extensive smooth surface near sea level prior to Paleozoic sedimentation.

FIGURE 6—View looking east (left), southeast (center) and south (right) from Bald Mountain summit showing the summits of the western Adirondacks declining in elevation to the west. Fourth Lake (left), Third Lake (center), First and Second Lakes (right). Little Great Range at distant horizon between Third and Fourth Lakes.

The plane connecting western Adirondack summits (Whitney et al., 2002, p. 5) is herein interpreted as a *middle Ordovician paleosurface* which I suggest extends from the "Little Great Range" comprising Squaw, Snowy, Lewey, Blue Ridge, and Pillsbury Mtns (west of the Indian Lake fault) westward to the Knox unconformity in the Black River Valley. In large-scale topographic maps of the Adirondacks, the paleosurface, as defined above, is striking, with more extensive erosion and higher summits in the Little Great Range and less eroded (but still abnormally high elevations) in the West Canada Lakes region. The fact that rusty quartz veins have been observed near the summit of Wakely Mountain supports this idea, however, I was unable to locate rusty quartz veins on the summits of Black Bear Mountain and Blue Mountain.

Using Surface III+ software (Kansas Geological Survey; www.kgs.ku.edu/Tis/surf3/surf3Home.html) the slope of the western Adirondack paleosurface can be approximated by regressing a simple "trend surface" through the highest elevations of the region (as defined above). As illustrated in Figure 7, first and second order regressions result in a planar and a concave downward curved paleosurface, respectively, but both result in a generally westward dipping paleosurface decreasing in elevation about 11 to 12 meters per kilometer. However, the subsurface slope of the unconformity west and south of the Black River Valley, which can be estimated from deep drill holes in central NY, dips southwestward at a slope of about 12-13 meters per kilometer in the region just east and south of Lake Ontario and 25-26 meters per kilometer immediately south in the Mohawk Valley (Rickard, 1973; Isachsen, 1975).

The elevation of the Precambrian surface, particularly in areas farther southwest reflects a Cambrian erosion surface as upper Cambrian and lower Ordovician sedimentary rocks rest unconformably on Proterozoic gneiss (Rickard, 1973). However, even if the thickness of the upper Cambrian and lower Ordovician sediments are taken into account, the slope of the Knox unconformity is at least *equal to* or *steeper* in the subsurface (farther to the southwest) than its inferred position over the western and central Adirondacks.

The difference in slope of the surface of the Precambrian rocks has traditionally been attributed to the deep erosion of the Adirondacks and the inference that a steeply sloping paleosurface has long been eroded away (Miller, 1910). If, however, the *middle Ordovician paleosurface* has not been deeply denuded and is preserved in the present-day western and central Adirondack summits, as interpreted here, then the difference

(or similarity) in slope would seem to argue against the hypothesis that the Adirondack uplift resulted from largely vertical crustal movements in the Tertiary to Holocene (Isachsen, 1975, 1981). Such uplift should have steepened the middle Ordovician paleosurface in the Adirondacks as compared to areas outside the Adirondack dome assuming an originally flat unconformable surface (Isachsen et al., 1991, p. 44).
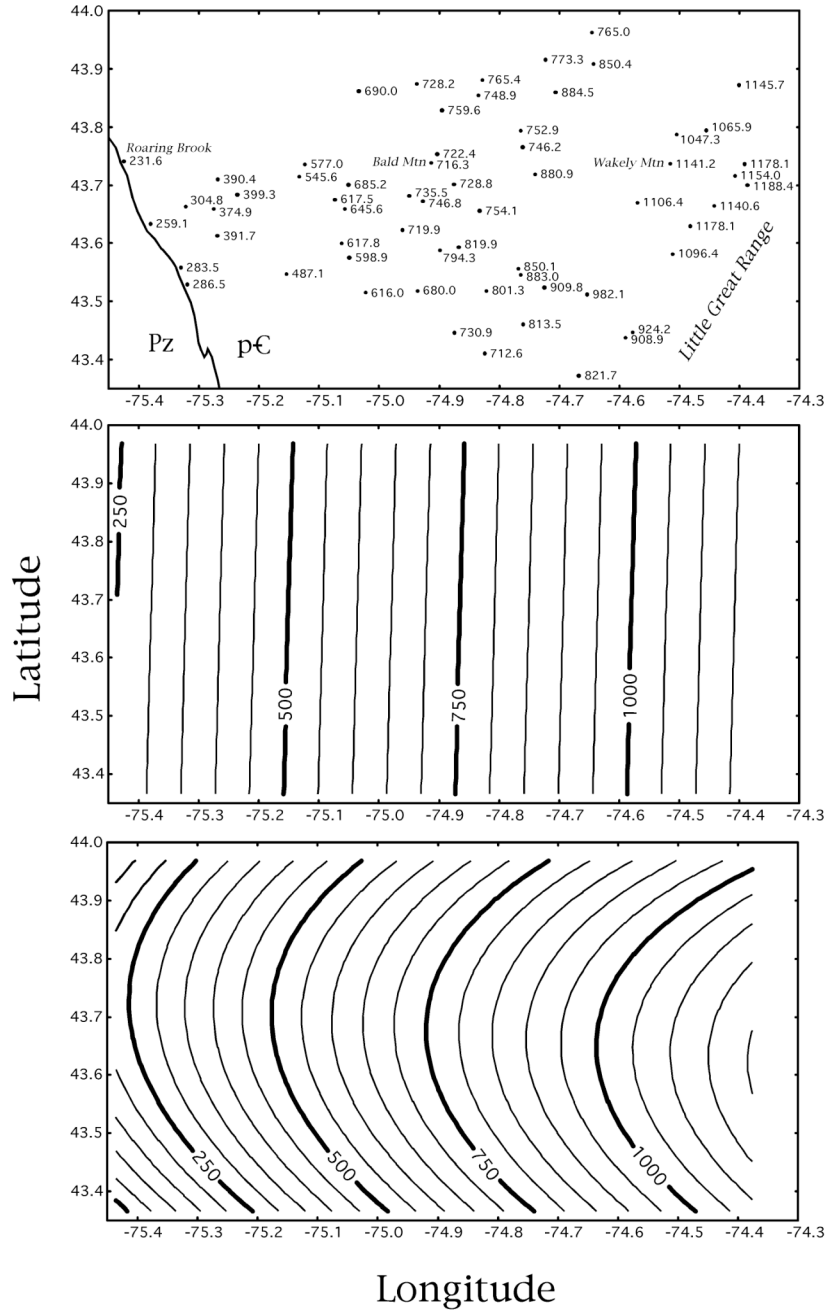
FIGURE 7—The western Adirondack paleosurface defined by bedrock-cored summit elevations on Proterozoic metamorphic rocks (top). Note locations of Roaring Brook, Bald Mtn, and Wakely Mtn. First order regression of data (middle) resulting in west-dipping plane. Second order regression of data (bottom) resulting in concave downward, west-dipping, curved plane. Scale: at 43.75° latitude, one degree longitude = 80.4 km.

ACKNOWLEDGEMENTS

REFERENCES CITED

Beetle, D.H., 1948, Up Old Forge Way, Louis Heindl & Son, Rochester, NY, 183 p.

Darling, R.S., and Bassett, W.A., 2002, Analysis of natural H_2O+CO_2+NaCl fluid inclusions in the Hydrothermal Diamond Anvil Cell: American Mineralogist, v. 87, p. 69-78.

Engelder, T., Haith, B.F., Younes, A., 2001, Horizontal slip along Alleghanian joints of the Appalachian Plateau: evidence showing that mild penetrative strain does little to change the pristine appearance of early joints: Tectonophysics v. 336, p. 31-41.

Isachsen, Y.W., 1975, Possible evidence for contemporary doming of the Adirondack Mountains, NewYork and suggested implications for regional tectonics and seismicity: Tectonophysics, v. 29, p. 169-181.

Isachsen, Y.W., 1981, Contemporary doming of the Adirondack Mountains: further evidence from releveling: Tectonophysics, v. 71, p. 95-96.

Isachsen, Y.W., Landing, E., Lauber, J.M., Rickard, L.V., and Rogers, W.B., 1991, Geology of New York: A Simplified Account, New York State Museum, Educational Leaflet No. 28, 284 p.

Lamb, W.M., Brown, P.E., and Valley, J.W., 1991, Fluid inclusions in Adirondack granulites: implications for the retrograde P-T path: Contributions to Mineralogy and Petrology, v. 107, p. 472-483.

Landing, E., 2003, Sauk "Sequence" and the Knox Unconformities: Practical Base of the Laurentian Upper Ordovician: Geological Society of America, Abstracts with Programs, Paper No. 216-10.

McLelland, J., Morrison, J., Selleck, B., Cunningham, B., Olsen, C., and Schmidt, K., 2002, Hydrothermal alteration of late- to post-tectonic Lyon Mountain Granitic Gneiss, Adirondack Mountains, New York: Origin of quartz-sillimanite segregations, quartz-albite lithologies and associated Kiruna-type low-Ti Fe-oxide deposits: Journal of Metamorphic Petrology, v. 20, p. 175-190.

Miller, W.J., 1910, Geology of the Port Leyden Quadrangle, Lewis County, N.Y., New York State Museum Bulletin 135, 61 p.

Niocaill, C.N., van der Pluijm, B.A., and Van der Voo, R., 1997, Ordovician paleogeography and the evolution of the Iapetus ocean: Geology, v. 25; no. 2; p. 159–162.

Orrell, S.E., and Darling, R.S., 2000, Survey of the Geology of the Northwestern Adirondack Mountains and Black River Valley, in Woodrow, D.L., and McKinney, D.B., (eds.) Field Trip Guide, 72nd Annual Meeting of the New York State Geological Association, p. 3-11.

Rickard, L.V., 1973, Stratigraphy and Structure of the subsurface Cambrian and Ordovician carbonates on New York: New York State Museum and Science Service Map and Chart Series 18, 26 p.

Selleck, B., 2005, Exploring the root zone of an ancient fault-driven hydrothermal system in the Adirondack lowlands, New York: in Valentino, D., (ed.) Field Trip Guide, 77nd Annual Meeting of the New York State Geological Association, p. 11-28.

Whitney, P.R., Fakundiny, R. H., Isachsen, Y.W., 2002, Bedrock Geology of the Fulton Chain-of-Lakes Area: New York State Museum, Map and Chart 44. 124 p.

ROAD LOG FOR TRIP B-2

MINERALIZATION AT THE KNOX UNCONFORMITY AND
THE WESTERN ADIRONDACK PALEOSURFACE

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Start from the corner of Graham Ave and Gerhart Dr. on the SUNY Cortland campus. Head downhill on Graham Ave.
0.1	0.1	Turn right onto Groton Ave.
0.5	0.4	Continue straight through Main St. staying on Clinton Ave.
1.2	0.7	Turn left onto Pomeroy St. at major intersection.
1.4	0.2	Pass under Interstate 81.
1.5	0.1	Turn left onto the Interstate 81 on-ramp heading north.
30.4	28.9	Exit right onto Interstate 481 (it heads only east from here).
40.2	9.8	Take Exit 6 for the NYS Thruway. Get ticket at toll booth and head toward Albany on Interstate 90 East.
65.2	25.0	Get off the Thruway at Exit 33 (Verona). Take Rt. 365 East toward Rome. As you approach Rome, continue passed the Rt. 49 exit that heads into Rome, and stay on Rt. 365 East.
76.8	11.6	Take exit for Barneveld / E. Dominick St.
77.4	0.6	Take right onto E. Dominick St (heading east toward Barneveld).
79.4	2.0	Turn left (this is still Rt. 365 East). Pass through the towns of Floyd and Holland Patent.
91.1	11.7	Turn left in downtown Barneveld (this is still Rt. 365 East).
92.0	0.9	Turn right onto Rt. 12 / 28 North on-ramp. Head North on Rt. 12 / 28.
101.9	9.9	Rts. 12 & 28 split. Continue straight (North) on Rt. 12.
116.3	14.4	Traffic light in Port Leyden (continue North on Rt. 12)
128.4	12.1	Take right onto Canaan Rd just before crossing bridge over Roaring Brook.
128.5	0.1	Pull off Canaan Rd on left side next to Roaring Brook.

STOP 1. THE KNOX UNCONFORMITY AT ROARING BROOK.

Get out of the vehicles and walk down the stream bank to the reddish-colored exposures along the bed of Roaring Brook (watch out for poison ivy!). These are feldspar-quartz gneisses that strike N40E. Note the development of spheroidal weathering between joint sets in the gneiss (Figure 1).

Walk ~30 meters upstream. The lowermost strata of the Pamelia Formation are exposed in the stream bed. More spheroidal weathering occurs directly below the (well exposed) nonconformable contact, but is observed only during low water levels (late summer, normally).

Walk back to Canaan Road and walk northwestward (~100 meters) until you cross a small bridge (over Roaring Brook). Walk down to the exposures of gneiss under the bridge (on the northwest side). Note the presence of smoothly worn, sub-horizontal thin (2 - 4 mm) veins of dark greenish-gray to rusty-brown cryptocrystalline quartz and iron hydroxide.

128.5	0.0	Turn around, and head back to Rt. 12.
128.7	0.2	Turn left onto Rt. 12 south.
140.9	12.2	In the village of Port Leyden, turn left at the traffic light onto E. Main St.
141.4	0.5	Turn right onto River Rd at "T" intersection.
141.6	0.2	Turn left onto Moose River Rd.
155.3	13.7	Turn left onto Rt. 28 North (at McKeever). Pass through Thendara and Old Forge.
170.6	15.3	Turn left onto Rondaxe Rd.
170.7	0.1	Turn left into Rondaxe Trailhead Parking area.

STOP 2. RUSTY QUARTZ VEINS ON BALD (RONDAXE) MOUNTAIN.

Walk southwestward on the trail to Bald Mountain. As you approach the summit along the ridge trail, look for thin (2–10 millimeter wide) rusty-brown veins cutting across feldspar-quartz gneiss. One rusty-brown vein (striking N49W) is offset on an unmineralized joint (striking N20E) about 50 meters northeast of the tower. Veins can be found both NE and SW of the Bald Mountain fire tower.

Return to vehicles, head back to Route 28 south and back to Cortland.

END OF FIELD TRIP

Trip B-3

COMPARATIVE SEDIMENTOLOGY OF THE HELDERBERG GROUP
IN CENTRAL NEW YORK

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INTRODUCTION

"What am I supposed to be looking at here?" The students have just assembled at the outcrop and, invariably, this question floats up from somewhere. This question drives some instructors wild, but most just chuckle to themselves. The answer to that question is, of course, the point of the field trip stop. If the stop is in sedimentary rocks (as on this trip) the immediate question: "what am I supposed to be looking at here?" breaks down into four questions: (1) what kind of rock is this, i.e. what was deposited?; (2) how did it get deposited?; (3) when did it get deposited?, and; (4) what happened after the sediments were deposited? A better way to phrase question two would be: what was the environment of deposition of these rocks? The answers to these questions are the core questions of *sedimentology*, the study of sediment, specifically the nature and origin of unconsolidated sediments and consolidated sedimentary rocks. The purpose of this trip is to show you how to systematically go about answering question two: the series of steps necessary to diagnose the *paleo-environment* of deposition.

PRELIMINARIES 1: STRATIGRAPHY (WHEN DID IT GET DEPOSITED?)

The question: "how old are these rocks?"; is the purview of a branch of sedimentology known as *stratigraphy*. One of the great scientific achievements of the nineteenth and early twentieth centuries was mapping and establishment of the relative age of all of the different rocks that cover the Earth, resulting in the *geologic column* or *stratigraphic column*. The stratigraphic column was assembled from information on: (1) the fossil content of sediments and sedimentary rocks, and; (2) application of a few common sense rules most important of which are that sediments and sedimentary rocks were deposited layer upon layer on gently inclined surfaces, and that discordances in the geologic record represent time gaps. In the early to mid twentieth century, development of radioactive isotopic dating techniques for rocks allowed the relative time scale based on fossils to be absolutely fixed in time. Moreover, rocks that did not contain fossils could now be dated within the limits of accuracy of the various radiometric dating techniques. This combination of relative and absolute age dating resulted in the *geological time scale*. There is no outcrop of rock anywhere on planet Earth whose position in the geologic column and geologic time scale is not known.

In particular, the rocks that are the subject of this field trip: the Helderberg Group, are Upper Silurian to Lower Devonian in age, or roughly 415 million years old (Fig. 1). The Helderberg Group of New York has been studied since the 1800s and is discussed in many historical geology and sedimentology text books. The classic work on the stratigraphy of the Helderberg was by Rickard (1962, 1973). Laporte (1967) provided an excellent early paleo-environmental interpretation that tackled many of the issues discussed here 40 years later.

In eastern New York, the Helderberg Group is up to 120 m thick and comprises five separate formations. Around Syracuse, the Helderberg comprises only one formation: the Manlius Limestone and is generally 10 m thick or less. The boundary between the underlying Rondout Formation and the Manlius Limestone occurs at each field trip site. East of Syracuse, at Chittenango Falls State Park (Stop 5 of the trip) a few meters of the Coeymans Limestone overlies the Manlius Limestone. The Manlius Limestone around Syracuse is reckoned to be younger than the Manlius Formation in the Hudson Valley and is divided into a number of members. However, the Manlius Limestone has only a few types of fossils and the true ages and stratigraphic relationships of the members of the Manlius Limestone and formations of the Helderberg Group is an area of active research in New York. Another interesting aspect of the Manlius Limestone around Syracuse is that there is a significant *unconformity* separating the Helderberg formations and the Oriskany Sandstone and another unconformity separating the Oriskany Sandstone from the Middle Devonian Onondaga Limestone. In all, a few millions of years of erosion and/or nondeposition are **not** recorded in these rocks. The Oriskany Sandstone itself is discontinuous in this part of the world and is difficult to find in some of the outcrops on this fieldtrip. The Middle Devonian Onondaga Limestone occurs at the top of the section at all of the stops in this trip.
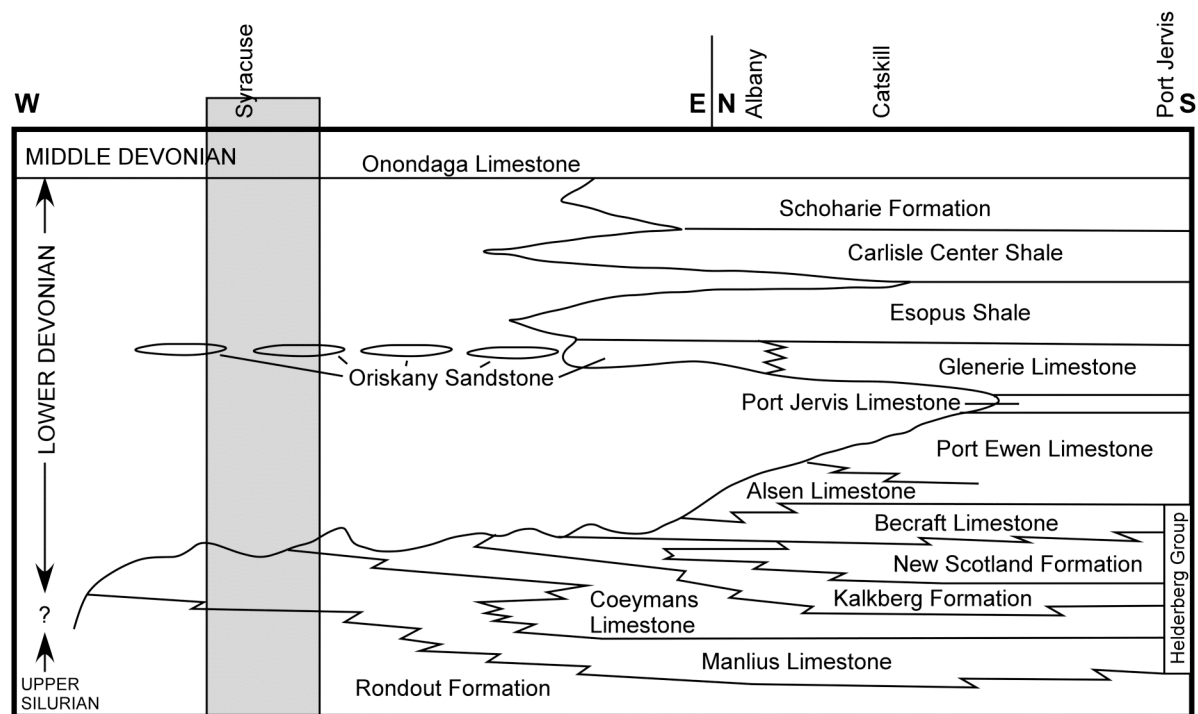
FIGURE 1—Stratigraphy of the Lower Devonian rocks in New York State showing Rickard's (1975) interpretation of the time-transgressive nature of the Helderberg Group rocks. The vertical axis here is time, so thicknesses of the formations are not determinant. The boundary between the Silurian and the Devonian is reckoned to be in the Manlius Limestone in eastern New York and in the Rondout Formation in western New York. The shaded area indicates the stratigraphy of the rocks in this field trip area.

It is our contention that the depositional paleoenvironments of the Manius Limestone can be diagnosed without worrying about the detailed stratigraphy at the various outcrops. In fact, the stratigraphy confuses the issue in some respects. The starting point of any study of sedimentary rocks is an analysis of the depositional environments represented.

PRELIMINARIES 2: PETROLOGY (WHAT WAS DEPOSITED?)

The first question posed in the introduction was: "what kind of rock is this?" The answer of for this field trip is fairly straightforward: carbonate rocks. Carbonate rocks are chemical and biochemical sediments composed of calcium (Ca^{2+}), magnesium (Mg^{2+}) and carbonate (CO_3^{2-}). In particular there are two minerals to worry about: calcite and dolomite. Calcite is most important in the Manlius Limestone and usually weathers a bluish gray to gray. Most sedimentary calcites older than Cenozoic have $< \sim 4\%$ magnesium in them and are technically known as low magnesium calcites ($Ca_{0.96}Mg_{0.04}CO_3$). Dolomite is a calcium, magnesium double salt ($CaCO_3 \quad MgCO_3$) and in this part of the world commonly weathers tan to brownish due to a few percent iron in the lattice.

Two important features of modern chemogenic and biogenic carbonate sediments are: (1) they originate as either biochemical precipitates or as physico-chemical precipitates in the depositional environment where they accumulate, and; (2) the majority of the grains are subjected to physical transport and deposition, although some of the grains may remain at the precipitation site. In particular, tropical, shallow-water environments with low terrigenous-sediment supply host a veritable 'carbonate factory' of biogenic and non-biogenic carbonate sediment production, both today and in the past. Biochemical production of calcium carbonate is controlled mainly by water temperature, salinity, depth, water clarity and residence time.

There are several types of carbonate sedimentary grains that can be recognized in modern and ancient deposits: mud, skeletal grains, pellets, ooids, grapestones, and intraclasts. Much of the calcium-carbonate mud and sand on modern shallow marine tropical environments is aragonite (another $CaCO_3$ mineral) and is produced by *green algae* such as *Halimeda* and *Penicillus* that need warm, shallow, clear water. Macroscopic *skeletal grains* are also produced by hermatypic corals that typically occur in tropical shallow-water environments, and these have symbiotic photosynthetic unicellular microorganisms known as *Zooxanthellae* in their tissues. These are dinoflagellate protozoans and, like green algae, require warm, sunlit (i.e. shallow and clear) and siliciclastic-mud-free water to thrive. Modern tropical shelves also contain a wide range of other bottom-dwelling organisms, mainly molluscs, red algae, and foraminifera that also produce silt- to gravel-sized skeletal grains composed of aragonite and high magnesium calcite. High magnesium calcite has up to a few tens of percent magnesium in the lattice and is the most common calcite produced by modern organisms. The warm conditions of tropical, shallow seas also favor non-biogenic precipitation of calcium carbonate both as sedimentary particles and as early diagenetic cements. Calcium-carbonate precipitation as mud takes place in *whitings*, clouds of suspended aragonite needles ~ 4 microns long that occur in surface water on shallow-marine carbonate shelves (and in lakes). Whitings occur where deep ocean water rises onto a shallow platform, warms, begins to evaporate, and degasses CO_2. Whitings are perhaps triggered by blooms of unicellular photosynthetic organisms that also take CO_2 out of the water. *Ooids* are concentrically-laminated, sand-sized grains considered to be chemical precipitates created as sea-water warms, evaporates, and degasses as it washes onto a shallow platform. Ooids typically accumulate as tidal shoals or beaches at the shelf margin. Carbonate cementation on and just below the sea floor is also important in shallow, tropical-marine carbonate deposition. Carbonate mud on modern tropical shallow shelves is commonly aggregated into *pellets* by both deposit-feeding and filter feeding organisms. Some pellets become cemented on the sea floor and aggregated into lumps known as *grapestones*. Erosion of cohesive mud and cemented sediments results in various types of *intraclasts*.

The most important carbonate rocks are: (1) grainstones; (2) packstones; (3) wackestones, and; (4) mudstones. *Grainstones* comprise well-sorted sand- and gravel-sized components surrounded by calcite cements. As cements must be introduced after deposition, grainstones originally comprised mud-free accumulations of carbonate particles. In *packstones*, the sand- and gravel-sized components are in grain-to-grain contact but carbonate mud fills the space between the grains. In *wackestones*, the sand- and gravel-sized components 'float' in mud, whereas *mudstones* are composed of uniform crystals that cannot be seen in a hand lens. In other words, if you don't see grains it is most likely a mudstone. The rocks that contain sand- and gravel-sized debris can be further classified by adding one or two modifiers to the rock name based on the dominant sand- or gravel-sized components: e.g. *ooid grainstone*, *skeletal packstone*, or *intraclast wackestone*. It is commonly assumed that a mud-supported limestone was deposited in the absence of turbulent water currents, and *vice versa* for a mud-free limestone. However, the texture and fabric of biogenic and chemogenic grains are difficult to relate to transport mechanisms, because of the possibility of *in-situ* accumulation with

little transport, and because of original rounded shapes. Furthermore, carbonate mud may be produced diagenetically (by a process known as 'micritization') and bioturbation commonly mixes separate sand and mud layers into packstones and wackestones. Classification is commonly made difficult by diagenetic modification of original textures and fabrics, which is especially true where limestones (rocks dominately composed of low magnesium calcite) have been altered into dolomite.

The most common rock types in the Manlius Limestone are mudstones and wackestones. Carbonate muds are renowned for preserving delicate sedimentary features unlike fissile siliciclastic mudstones. The exact reasons for preservation of such exquisite detail in carbonate muds is a matter of some debate and need not concern us here. The carbonate mudstones in the Manlius Limestone also contain various amounts of terrigenous mud (mostly comprising clay minerals). In some cases Manlius Limestone mudstones can be quite shale-like with fissility and cleavage development. In addition, the organisms that produced the skeletal grains in the Manlius Limestone were from quite different groups than those found on modern carbonate shelves. Skeletal grains in the Manlius Limestone included ostracod, brachiopod, and gastropod fragments. Stromatoporoids (a calcium-carbonate secreting sponge) can be quite common in some layers where corals also occur. The Coeymans Limestone (exposed at Stop 5 at Chittenango Falls State Park) is a skeletal grainstone to packstone. Common skeletal debris in the Coeymans Limestone includes crinoid and brachiopod fragments.

COMPARATIVE SEDIMENTOLGY:
DIAGNOSING ANCIENT ENVIRONMENTS

The only rational way of interpreting the origin of ancient sedimentary deposits is to compare them with modern sedimentary deposits: an approach referred to as *comparative sedimentology* by Robert Ginsburg (1974). An *environment* is a part of the Earth's surface where erosion and deposition are proceeding that has a distinctive association of physical, chemical and biological landforms and processes. Examples of environments are rivers and adjacent floodplains, glaciated regions, deserts, lakes, beaches, tidal flats, rocky coasts, continental shelves, coral reefs, and ocean basins. Environments, in turn, can be broken up into a series of smaller, more specific subenvironments. For example, a river can be divided into channel thalwegs (the deepest scoured areas), point bars, channel cross-over areas, partly to completely abandoned channels (sloughs), and levees. On the adjacent floodplains there are oxbow lakes, crevasse-splays, marshes or forests, and lakes. Tidal flat environments commonly can be subdivided into subtidal, intertidal and supratidal flats that are traversed by meandering tidal channels. In each case, the distinctive set of physical, chemical and biological processes operative in each subenvironment leaves behind a deposit with a distinctive set of physical, chemical and biologic sedimentary structures, sedimentary grain types, sedimentary textures (grain size, shape, sorting), body fossils, trace fossils and paleocurrent indicators. Understanding ancient environments from study of sedimentary rocks, using modern counterparts as comparative guides, is the essence of comparative sedimentology.

The most important step in interpreting the depositional paleoenvironment of ancient sedimentary rocks is accomplished by dividing the outcrop up into elemental rock units that are commonly referred to as *facies* or *lithofacies*. Facies are usually between a few tens of centimeters up to a few meters thick and in carbonate rocks are characterized first by their assemblages of physical, chemical and biogenic sedimentary structures and early diagenetic features and secondarily by the rock type, fossil content, grain size and shape, and paleocurrent indicators. Demicco and Hardie (1994) provide an introduction to sedimentary structures and early diagenetic features of shallow marine carbonate deposits. *Each facies ideally should represent the depositional record of an ancient subenvironment*. It is the lateral and vertical arrangement of facies at an outcrop that unambiguously point to the depositional environment of the rocks.

In order to interpret the depositional environment of a formation it is necessary to describe the three-dimensional distribution of the facies (commonly referred to as a *facies association*). In this part of the world, large, three-dimensional outcrops are rare. Instead, the three-dimensional details of facies distribution must be pieced together from widely-spaced, one- and two-dimensional outcrops that (we hope) can be correlated. This is a tricky business (see below). With this in mind, the description of an outcrop follows a logical sequence.

Step 1: Outcrop Reconnaissance (a.k.a. 'Scratch & Sniff')

The first thing you should do is a brief reconnaissance overview of the outcrop to gain an impression of the main rock types, sedimentary structures, and body and trace fossils. In deformed rocks, it will be necessary to establish which way is stratigraphic 'up' but in this part of the world the rocks are in place and become younger higher and higher in the section. For some purposes (and on many field trips) this step is all that is done, and, by the time you have walked through the outcrop, you will have a 'shopping list' of sedimentary features that will allow you to make a pretty good guess as to the broad depositional environment. For example, the 'shopping list' of common sedimentary features in the outcrops of the Manlius Limestone around Syracuse include: (1) laminated mudstones; (2) desiccation cracks; (3) 'ribbon rocks' - alternating thin beds of small-scale cross stratified grainstones alternating with dolomitic mudstones and shales; (4) wave ripple marks; (5) current ripple marks; (6) meter-thick layers rife with skeletons of the stromatoporoid sponge *Syringostroma barretti* nearly to the exclusion of other fossils; (7) microbial tufas; (8) stromatolites; (9) burrow-mottled skeletal-peloidal wackestones in thin-bedded sets of strata, and; (10) a low diversity of fossils including one or two species of brachiopods and ostracods. All these features have long been recognized as the hallmarks of carbonate tidal flat deposits. A useful descriptor for such rocks is 'peritidal': literally 'around the tides'.

In the reconnaissance walkthrough you will commonly notice that certain sedimentary features occur associated with each other, i.e. comprise facies. For example, one of the most famous facies of the Manlius Limestone comprises thick cosets of planar laminated mudstones, stromatolitic-laminated mudstones, millimeter thick lenses of intraclastic conglomerates, and desiccation cracks. The laminae commonly are very fine (less than one millimeter) alternations of limestone (calcite) and dolostone (dolomite). Another notable facies of the Manlius Limestone is the meter-thick stromatoporoid sponge wackestone beds. The emphasis of this field trip will be the recognition and interpretation of the facies of the Manlius Limestone.

Step 2: Measuring Sections and Constructing Photomosaics ('Getting Serious')

The only real way to interpret sedimentary rocks is to construct measured sections (also called stratigraphic logs) of the vertical succession of facies, and, where appropriate, construct two-dimensional scale drawings of the outcrops from photomosaics. Measuring the section forces you to come to grips with the problem of defining facies by putting your nose directly on the rocks. In addition, correlation between outcrops is impossible without a to-scale, detailed representation of the facies at the various outcrops (commonly referred to as *graphic logs*). Sedimentological logs normally include measurement of the upward variation through the sedimentary sequence of stratal thickness, texture, color, composition, fossils, sedimentary structures and paleocurrent directions. Measurement of true stratal thickness in dipping strata is accomplished using a level on a ranging pole. Location of positions in sedimentological logs can now be measured using differential GPS. Figure 2 illustrates how logged sedimentary features can be represented graphically. The legend for Figure 2 is a typical legend for sedimentological logs, but there are many other legend designs to suit specific needs. *We would like to encourage anyone interested in trying to measure a section to have a go at it at Stop 2 in Split Rock Quarry*.

It is not uncommon to encounter rocks where the sedimentary features cannot be resolved at the outcrop. The rocks may be covered by mosses or lichens, or poorly weathered. In these cases, oriented samples of the rocks are taken back to the laboratory where they can be sawn, polished and etched in acids to bring out the sedimentary details. We have a number of slabs of some Manlius Limestone facies from poorly weathered portions of the outcrops. This iterative process of measuring, sampling, and laboratory study ultimately results in a complete, detailed graphical log of the outcrops. We find that graphical logs are best where drawn by hand, that the symbols for the different rock types look like the outcrop rocks, and that annotation of important features be made directly on the logs.

With large, continuous exposures of unconsolidated sediments and sedimentary rocks, a series of two-dimensional sections can be produced using photomosaics in combination with detailed logs of the sedimentary features. Figure 3 provides an example from Stop 4 of this field trip (Clockville). When constructing photomosaics, it is necessary to minimize the distortion in the photos related to varying distance of the outcrop from the camera. As a rule, the line of sight of the camera should be normal to the outcrop face, there should be 50% overlap of adjacent photos, and two ranging poles should be included in each frame for scale and to facilitate aligning the photos. In order to assemble photomosaics, it is necessary to establish a datum that should be surveyed during photography, possibly using a level or GPS and a laser range-finder for proper positioning

of outcrop images. Digital photographs or scanned photos can be assembled into photomosaics using computer software. Photomosaics are analyzed in the laboratory and in the field, and it is common to construct overlays for marking surfaces and sedimentary facies.
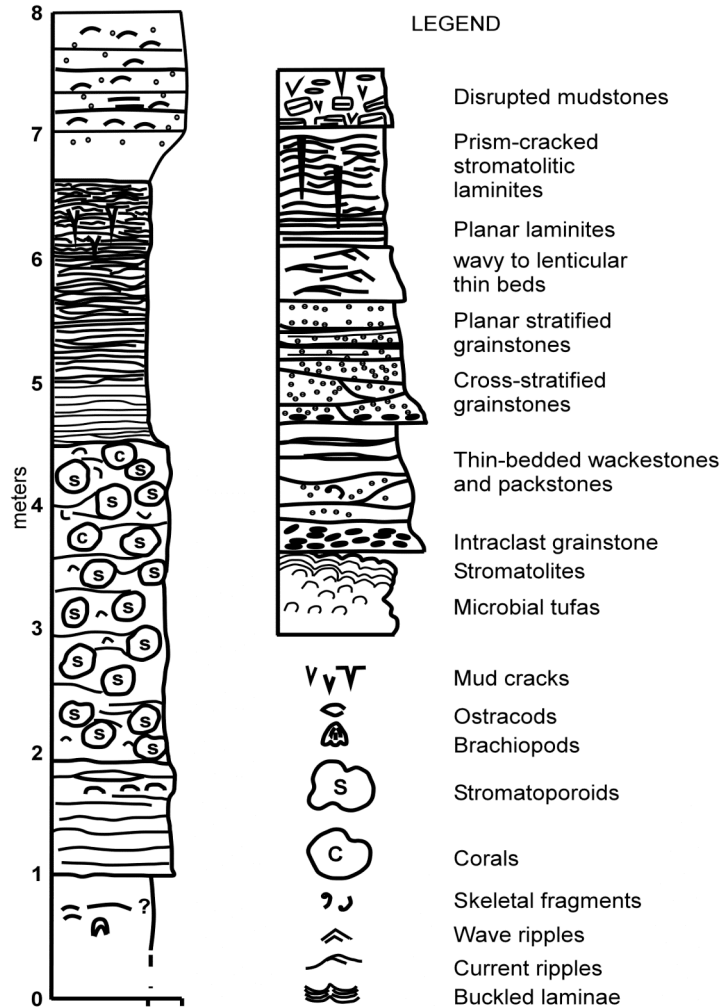
FIGURE 2—The right column is a graphic log of measured stratigraphic section at Nedrow, New York (Stop 1). Legend for the measured sections in this guidebook shown on right

Sedimentary rocks tend to occur in repetitive sequences of facies (this point is further discussed below). In the Hudson Valley, the Manlius Limestone is famous for its repetitive sequences. Thus, after the initial stages, it is not necessary to determine all of the sedimentary features that occur in facies. Indeed, once the facies have been established for a formation, it is easier to note the differences in facies within an outcrop or from outcrop to outcrop. Unfortunately, in this part of the world, the outcrops of the Manlius Formation are fairly thin and generally do not show repetitive sequences of facies (although there are some!). For this field trip then, we will be looking for different facies to recur from outcrop to outcrop instead of repetitively in the same outcrop.

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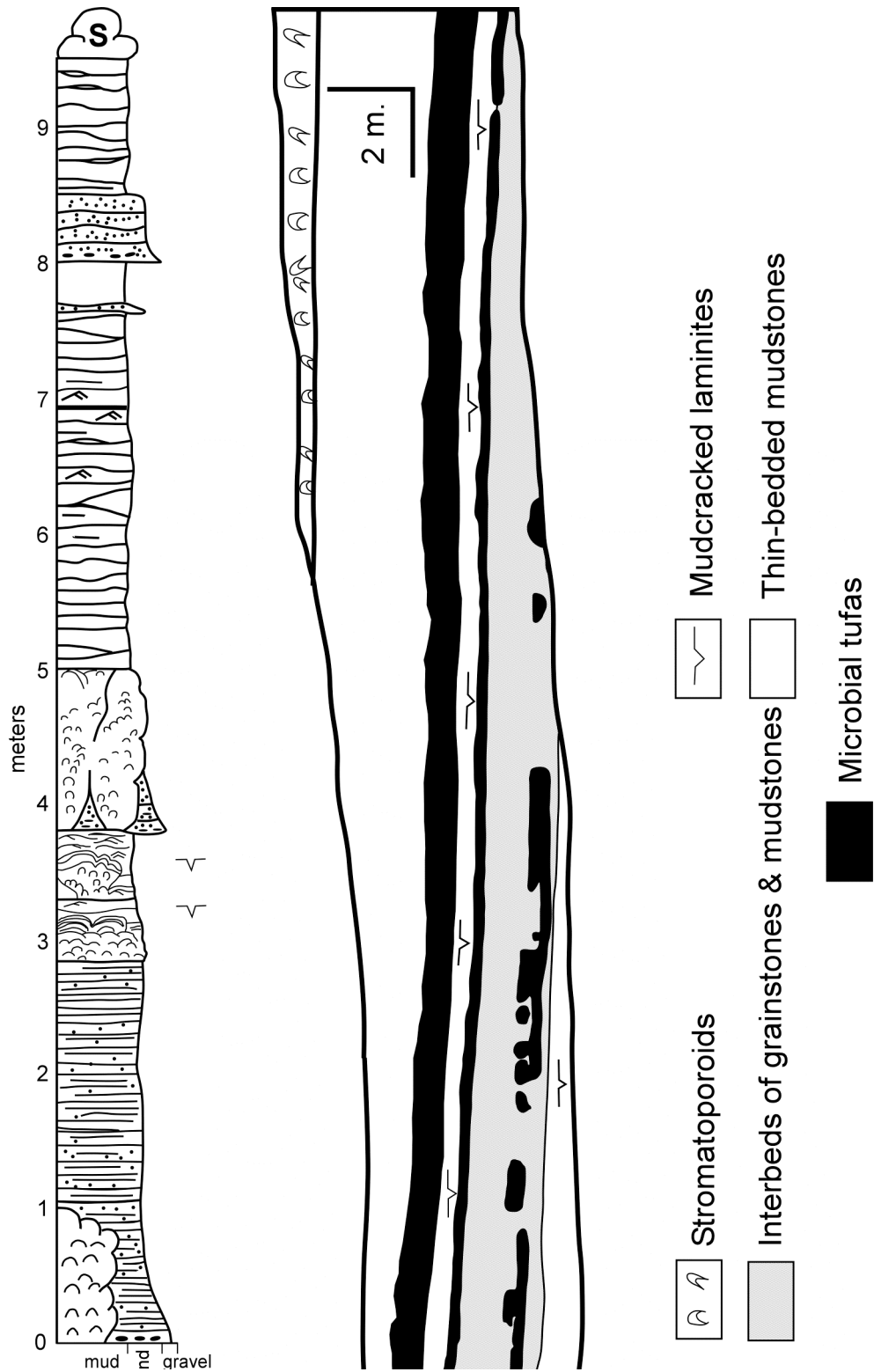
FIGURE 3—Measured section (left) and scaled outcrop diagram (prepared from photomosaics) of the Manlius Limestone at Clockville, New York (field trip Stop 4). Horizontal and vertical scales are the same. Modified from Browne (1986) and Browne and Demicco (1987).

The link between the vertical and lateral arrangement of facies in an outcrop and ancient subenvironments that comprised the ancient environment is Walther's Law of Facies (Fig. 4). Walther's Law of Facies states that only those deposits that were laterally adjacent can be superimposed conformably one upon the other. Said another way, if there are no erosional breaks between strata, a vertical succession of facies represents an original horizontal distribution of subenvironments. Figure 2 shows the distribution of facies in the Manlius Formation at Stop One along U.S. Route 11 just south of Nedrow, New York. Here the Manlius can be divided up into three facies: (1) approximately 2 m of thin interbeds of packstones and grainstones with rare stromatoporoids; (2) 2.5 m of stromatoporoid – coral wackestone, and; (3) 2 m of planar-laminated to wavy-laminated mudstones. The laminated mudstones have rare desiccation mud-cracks. Note that: (1) there are no clear erosional breaks; (2) fossil-bearing beds are overlain by laminated mudstones with very rare fossils, and; (3) desiccation features are restricted to the top of the section. These observations suggest that the section records increasing exposure upwards and less habitable subenvironments. Figure 5 is a core taken through a tidal flat island in Florida Bay. Note that, aside from the stromatoporoids, this core has most of the features of this outcrop and we use it to guide us in our interpretation of this outcrop. We interpret this outcrop to represent subtidal pond (thin-bedded facies) and intertidal to supratidal mud flat covered by a cyanobacterial mat (laminated mudstones). The exact depositional subenvironment of the stromatoporoid wackestones is controversial and they may represent either: (1) patch reefs, or; (2) migrating tidal channels. However, the fact that it occurs within this shallowing-upwards succession radically restricts the subenvironment of deposition of this facies.
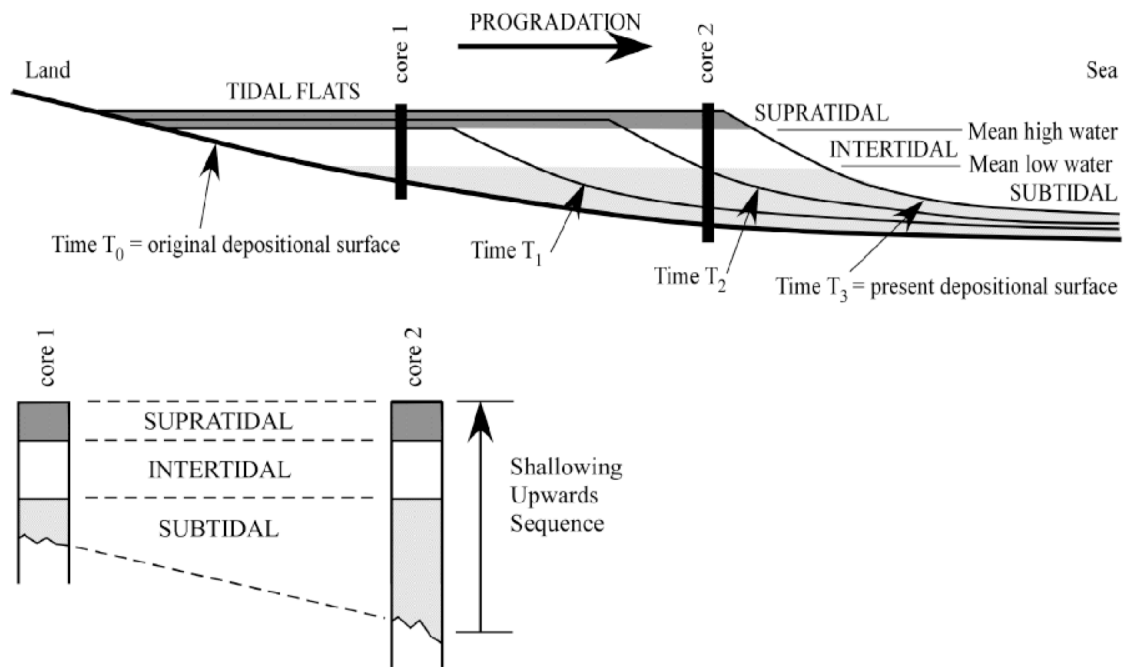
FIGURE 4—Walther's Law of Facies and the generation of shallowing-upwards sequences by tidal flat progradation (from Demicco and Hardie, 1994). We will follow this same approach at each outcrop. We have a number of samples from modern carbonates (including the core shown in Figure 5) that have been embedded in epoxy that we will use at the various outcrops to guide our subenvironmental interpretations.

Step 4: Correlations ('The Big Picture')

Once individual outcrops have been logged and their vertical and lateral associations of facies interpreted, it is time to move up to the next step; correlation of the graphic logs among the outcrops. It cannot be stressed enough that, unless you can actually walk layers out between outcrops, any correlation of facies between outcrops is *an interpretation*. Indeed, for most formations, the outcrop (or well) information that is used to diagnose the depositional environments and correlate facies among outcrops is only a tiny fraction of the formation! In the absence of: (1) reliable biostratigraphic data; (2) continuous exposure to walk out facies relationships, or; (3) or discreet 'event' beds, such as ash fall layers, *all* correlations (including those in Figure 1) are interpretations.

Compare our measured section from Stop 1 (Nedrow) to your measured section from Stop 2 (Split Rock Quarry) a distance of some 10 km. It is rarely possible to correlate facies from outcrop to outcrop in the Manlius Limestone either locally or across the state, a fact recognized by Leo Laporte (1967) some 40 years ago. However, the lack of correlation among outcrops itself may actually provide a clue as to the bigger depositional picture. Clearly, during deposition of the Manlius Limestone in central New York, we did not have

areas where large tidal flat coastlines prograded laterally for hundreds of kilometers leaving behind sheet-like tidal flat deposits correlative from outcrop to outcrop. Modern areas where this style of deposition occurs include tidal flats of southwestern Andros Island in the Bahamas and the famous 'sabkhas' of the Arabian Gulf that comprise the United Arab Emirates. Instead, we picture the Manlius Limestone around Syracuse as analogous to the modern Florida Bay (Fig. 6). Florida Bay comprises a restricted area of linear mudbanks, ornamented with tidal flat islands that separate the shallow subtidal areas into semi-circular areas known as 'lakes'. Mud banks in the eastern parts of Florida Bay (Fig. 6A) are linear features a few hundred meters across, 1 -2 m high, and up to kilometers long, that run between tidal-flat islands. The tidal-flat islands are up to a few square kilometers in area and mostly covered in mangroves, although many have open intertidal to supratidal ponds covered with cyanobacterial mats (Fig. 6C). In the western portions of Florida Bay, the mud banks are larger and more irregular in shape, and the islands are scattered on them. Florida Bay mud banks were once thought to be a product of baffling and subsequent deposition of transported sediment by the sea grasses on the banks, with additional sediment produced in place by a host of small, calcareous organisms (known as epibionts) that lived on the grass. However, it appears likely that the mud mounds in Florida Bay are tombolos and spits that are influenced by wave refraction around the tidal-flat islands.

Cores taken from the banks reveal bioturbated mud with sand- and gravel-sized skeletal fragments and the rhizomes of sea grasses (Fig. 6B). Cores through the tidal-flat islands (Fig. 5) show that they are mostly underlain by thin-bedded pond deposits and cyanobacterially-laminated muddy tidal-flat deposits, indicating that they have always been tidal flats that have migrated laterally perhaps a few hundred meters at most (Enos and Perkins, 1979). Mud banks in Florida Bay comprise tombolos that generally run between mangrove-covered tidal flat islands. These mud banks are probably related to longshore transport of mud controlled by wave refraction around the islands.

The deposits of an area like Florida Bay would comprise isolated tidal flat islands that develop by in place vertical growth (aggradation) as well as some lateral progradation. The deposits of such an area would produce a complicated three-dimensional mosaic of facies that could not be correlated over areas greater than 10 kilometers or so. This is exactly the situation of the facies of the Manlius Limestone in central New York. The restricted fauna of the Manlius Limestone suggests that the waters of the basin were inimical to invertebrates; i.e. they were either too salty, or too brackish. Florida Bay is brackish, and is backed up by the Everglades, a vast area of fresh-water carbonate deposition in periphyton marshes. It would not be at all surprising if some of the more exotic facies of the Manlius Limestone (such as the microbial tufas we shall visit at Stop 4 in Clockville) were, in fact, fresh water deposits.
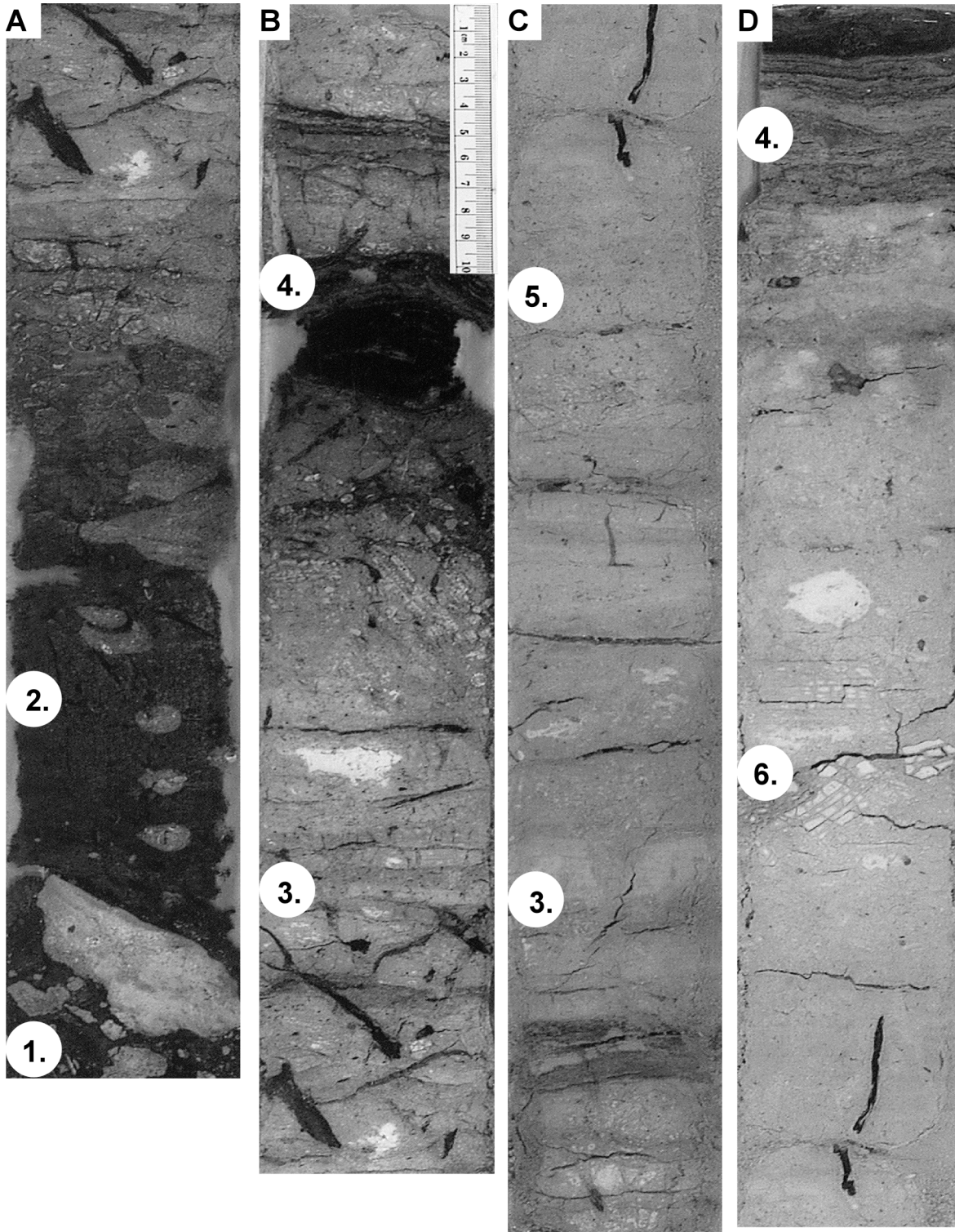
FIGURE 5—Epoxy embedded core 1.7 m thick through tidal flat island from Florida Bay. Base of core in A top of core in D. Circled numbers denote: (1) pieces of Pleistocene rock; (2) freashwater peat; (3) thin-bedded intertidal pond sediments; (4) two layers of cyanobacterial laminated muds of supratidal marsh; (5) burrowed mud of intertidal pond with indistinct thin beds; (6) high magnesium calcite (so-called protodolomite) cemented crust broken up during coring.
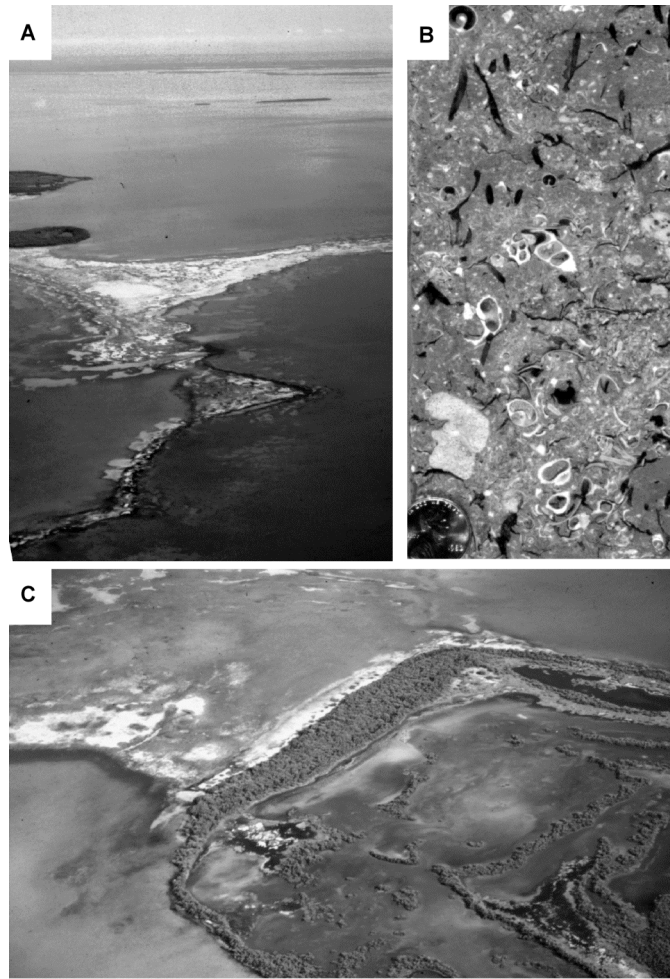
FIGURE 6—(A) Linear subidal mud banks of Florida Bay connecting tidal-flat islands. The freshwater marl prairie of the Everglades is in the far distance. (B) Epoxy embedded core of coarse skeletal wackestone from mud bank seen in A. (C) Tidal-flat island in Florida Bay showing open intertidal pond covered by cyanobacterial mats. Adjacent mud bank runs off to the upper right.

CONCLUDING REMARKS

It is our contention that understanding of the origin and significance of primary sedimentary structures and early diagenetic features is vital to unraveling the origin and significance of carbonate deposits in the geologic record. Without such an understanding at the individual sedimentary structure scale, we cannot hope to accurately reconstruct the large scale accumulation history of carbonate deposits or to decipher the roles of sea-level changes, sedimentation rates, subsidence rates, and tectonics in determining the facies stratigraphy, cyclostratigraphy, and sequence stratigraphy of these buildups. The message seems crystal clear, if we do not get the little things right we may not be able get the big things right (Demicco and Hardie, 1994, p. 242).
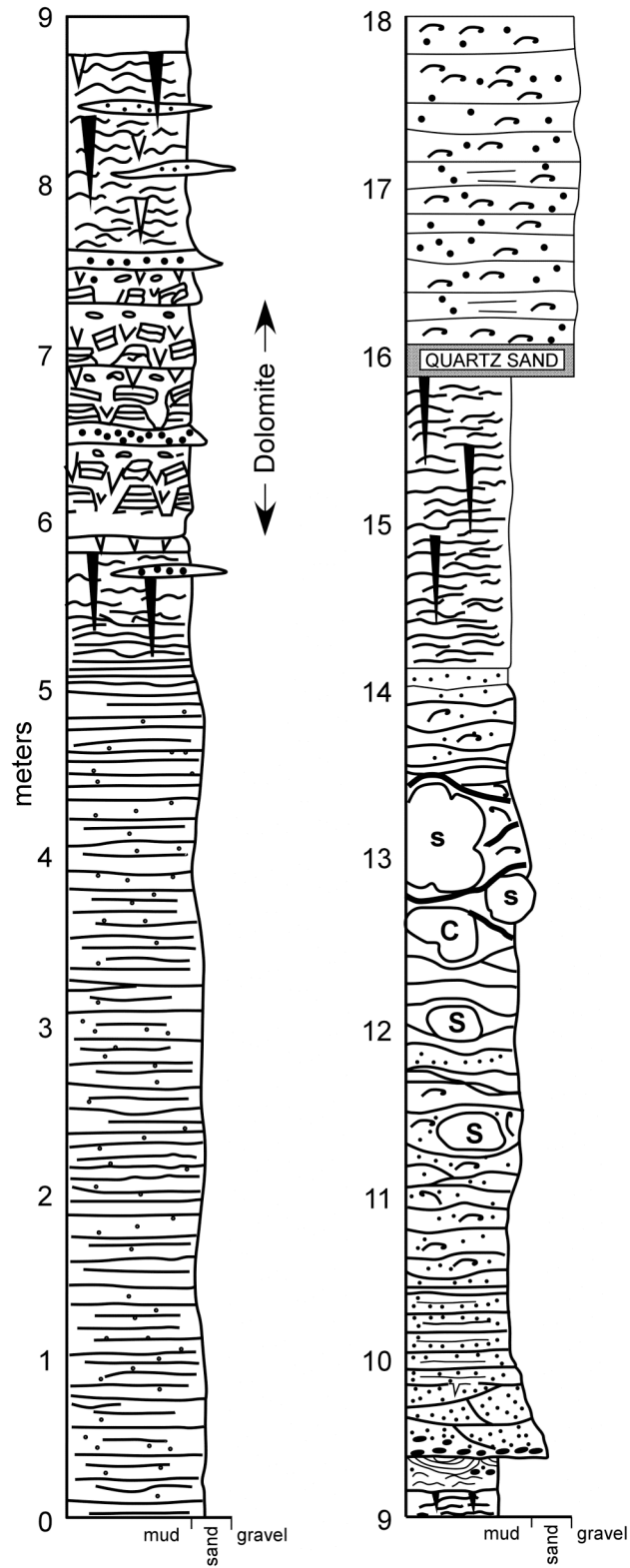
ACKNOWLEDGEMENTS

FIGURE 7—Measured section from Clark Reservation, New York (field trip stop 3).

REFERENCES CITED

Browne, K.M., 1986, Thrombolites of the Lower Devonian Manlius Formation of central New York: Unpublished M..A. thesis, The State University of New York at Binghamton, 61 p.

Browne, K.M., and Demicco, R.V., 1987, Thrombolites of the Lower Devonian Manlius Formation of central New York: Carbonates and Evaporites, v. 2, p. 149- 155.

Demicco, R.V., and Hardie, L.A., 1994, Sedimentary Structures and Early Diagenetic Features of Shallow Marine Carbonate Deposits: SEPM (Society for Sedimentary Geology) Atlas Series 1, 265 p.

Enos, P., and Perkins, R.D., 1979, Evolution of Florida Bay from island stratigraphy: Geological Society of America Bulletin, v. 90, p. 59-83.

Ginsburg, R.N., 1974, Introduction to comparative sedimentology of carbonates: American Association of Petroleum Geologists Bulletin, v. 58, p. 781-786.

Laporte, L.F., 1967, Carbonate deposition near mean sea-level and resultant facies mosaic: Manlius Limestone (Lower Devonian) of New York State: American Association of Petroleum Geologists Bulletin, v. 51, p. 3-101.

Rickard, L.V., 1962, Late Cayugan (Upper Silurian) and Helderbergian (Lower Devonian) stratigraphy in New York: New York State Museum and Science Service Bulletin Number 386, 157 p.

Rickard, L.V., 1975, Correlation of the Silurian and Devonian Rocks in New York State: New York State Museum and Science Service Map and Chart Series Number 24, 16 p.

ROAD LOG FOR TRIP B-3

COMPARATIVE SEDIMENTOLOGY OF THE HELDEBERG GROUP OF CENTRAL NEW YORK

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Enter I-81 northbound from NY St. Rte. 11 in Cortland
24.4	24.4	Exit I-81 at Exit 16 to U.S. Rte 11 at Nedrow, NY
24.5	0.1	Turn left at end of exit ramp onto Rte. 11 toward Nedrow, NY
25.0	0.5	Pull off Rte 11 on right – outcrop along east side of Rte. 11

STOP 1. NEDROW (Graphic Log Figure 2)

25.0	0.0	Proceed north on Rte. 11 through Nedrow into Syracuse, NY
27.7	2.7	Turn left (west) onto West Seneca Turnpike (NY St. Rte 173)
29.5	1.8	Junction NY 173 and NY 175 - bear right on 173 now called Onondaga Road
32.4	2.9	Turn left (south) on Onondaga Blvd.
33.1	0.7	Park at dead end – at top of access road take path to left into old quarry

STOP 2. SPLIT ROCK QUARRY (Do it yourself?)

33.1	0.0	Return north on Onondaga Blvd.
33.8	0.7	Turn right (east) on NY 173 and retrace route back into Syracuse
38.5	4.7	Intersection of NY 173 and US 11 – continue straight on NY 173 now called East Seneca Turnpike
39.0	0.5	Cross under I-81
39.2	0.2	At light, turn right (south) on 173 (East Seneca Tpk.) toward Jamesville, NY
41.0	1.8	Turn left (north) into Clark Reservation State Park – go to entry kiosk.
41.2	0.2	Park in lot, walk right to trial down to 'glacial' lake

STOP 3. CLARK RESERVATION STATE PARK (Graphic Log Figure 7)

41.2	0.0	Return out entry road
41.4	0.2	Turn left (east) on NY 173 (East Seneca Tpk.) and proceed through Jamesville
42.8	1.4	Intersection NY 173 and NY 91 – stay on NY 173 east
44.7	1.9	Junction NY 92 and NY 173 in Manlius – continue east on NY 173 (now Brinkerhoff Hill Rd.)
53.8	9.1	End of NY 173– proceed straight (east) on NY 5 & US 11 through Chittenango
54.6	0.8	Bear right NY 5 (east) & US 13 (north) (E. Genessee St.)
60.2	5.6	Turn right (south) on Oxbow Road in Canastota
63.0	2.8	Pull out into parking area going up the hill

STOP 4. CLOCKVILLE (Interpretive photomosaic and graphic log Figure 3)

63.0	0.0	Return north on Oxbow Road to Canastota
65.8	2.8	Turn left (west) onto NY 5 (west) & US 13 (south) (E Genessee St.)
71.4	5.6	Turn left (south) on NY 5 (west) & US 13 (south) into Chittenango, NY
72.1	0.7	Turn left (south) on US 13
77.3	5.2	Turn right into Chittenango Falls State Park – go to entry kiosk
77.4	0.1	Pull into parking lot and stop

STOP 5 – CHITTENANGO FALLS STATE PARK

Return out access road and turn right (south) – follow US 13 ~ 6.5 miles back to intersection of I-81 and US 13 (starting point of trip).

END OF FIELD TRIP

Trip B-4

PALEOENVIRONMENTAL GRADIENTS ALONG A RAMP TO BASIN TRANSITION IN THE MIDDLE DEVONIAN LUDLOWVILLE FORMATION OF CENTRAL NEW YORK STATE

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INTRODUCTION

The richly fossiliferous Middle Devonian Hamilton Group strata of central New York State form a natural laboratory for the study of paleoecology and the relationship of organisms to paleoenvironmental changes during cycles of varying magnitude. Indeed, these rocks have formed the basis for numerous detailed studies about paleoecology (including; Brower, 1987; Bower and Nye, 1991; Brett et al., 1990). These studies have focused largely on vertical gradients of faunal replacement and make a tacit Waltherian assumption that the vertical relationships seen in medium and small-scale cycles within the Hamilton Group reflect lateral changes; however, it is rarely possible to actually test this assumption. Much of this recent work has been framed in the context of sequence stratigraphy and the presence of small-scale sequences and cycles in central New York State this forms part of the basis for our somewhat revised model of siliciclastic-carbonate depositional sequences in foreland basins (see Brett and Baird, 1996).

The Middle Devonian Ludlowville Formation in the eastern Finger Lakes region provides a particularly excellent opportunity for examination of these issues (Brett et al., 1986). The rocks are relatively simple structurally, and are well-exposed in numerous creeks and shoreline bluffs along Owasco, Skaneateles, and Otisco lakes (Owasco, Skaneateles, and Spafford 7.5'quadrangles). These lakes, with relatively northwest-southeast orientation oblique to facies strike, provide natural cross-sections across rapid facies changes particularly within units of the lower Ludlowville Formation. Those units, comprising the Chenango - Centerfield succession and the overlying Otisco Member, are the subject of the present study. In particular, we focus upon the lowest 2-3 cycles of the Ludlowville Formation, consider their sequence stratigraphic context and mode of origin, and describe an unusually steepened gradient within one of these units, the Stag Horn Point submember of the Otisco Member. Together, these units provide insights into both vertical and lateral changes in benthic marine fossil assemblages that are a response to relative water depth as well as variations in sedimentation. We begin with an overview of the physical and sequence stratigraphy of the Chenango to lower Otisco interval and then focus on the details of the Stag Horn submember in the region of Skaneateles and Otisco lakes. The result of this study is a fairly detailed description of lateral as well as vertical gradients of faunal change during the deposition of this interval.

GEOLOGIC SETTING

The rocks under study in this report belong to the Middle Devonian Hamilton group and are of Middle Givetian age (*timorensis* to *rhenanus-varcus* conodont zones). This interval is estimated to be about 383 million years old. The total interval may represent a bit over half a million years based on the inference that it includes the base of one large third-order sequence, the Ludlowville Formation sequence, as well as portions of two smaller fourth- and fifth-order cycles. Hamilton group sediments were deposited in an active foreland basin as debris eroded from the rising Acadian highlands during the pronounced second tectophase (collisional pulse) of the Acadian Orogeny during the latest Eifelian (Fig. 1; Ettensohn 1998; Kaufmann 2006). The foreland was presumably created by thrust loading in the orogenic belt to the east-southeast, and the basin underwent various episodes of movement during this time (Fig. 1). These are viewed as far-field tectonic responses to Acadian thrust loading and probably included reactivation of older basement faults, which was particularly notable during deposition of the upper Hamilton and Tully intervals in New York (Heckel 1973). During deposition of the Ludlowville Formation, progradation of clastics into the foreland basin appears to have been particularly active, resulting in successively more westwardly migrating packages of coarser silts and silty mudstones through the course of deposition of the unit. This is a fact of some consequence to the present interpretations. It appears that major regressive or falling stage intervals triggered progradational pulses, which brought silt to the approximate meridian of present day Otisco and Skaneateles lakes during deposition of the lower Ludlowville sediments, and to the vicinity of Cayuga Lake by the end of Ludlowville deposition. These siliciclastic silts formed relatively shallow, somewhat wave influenced platforms and pass westwardly into the subsiding center of the Appalachian foreland basin, which may have been located in the Cayuga to Seneca valley region during this time (Fig. 1).

In some cases, rapid progradation and/or minor synsedimentary tectonic adjustment on faults may have resulted in over-steepening of prograded clastic wedges, which in turn fostered submarine erosion. Such a case is suggested by the details of the so-called "Stag Horn Platform", a 10 m-thick, upward-coarsening mudstone to siltstone package that is capped with concretionary, heavily bioturbated silts.

The supply of siliciclastics into the Acadian foreland may have been interrupted periodically by intervals of relative sediment starvation. A plausible explanation for such intervals is rapid rises in sea level, and subsequent flooding of bays and estuaries producing coastal clastic traps and resulting in strong curtailment of siliciclastic sedimentation in offshore settings. Conversely, periods of sea level drop may have triggered more rapid progradation of sediments and led in some cases to unstable conditions.
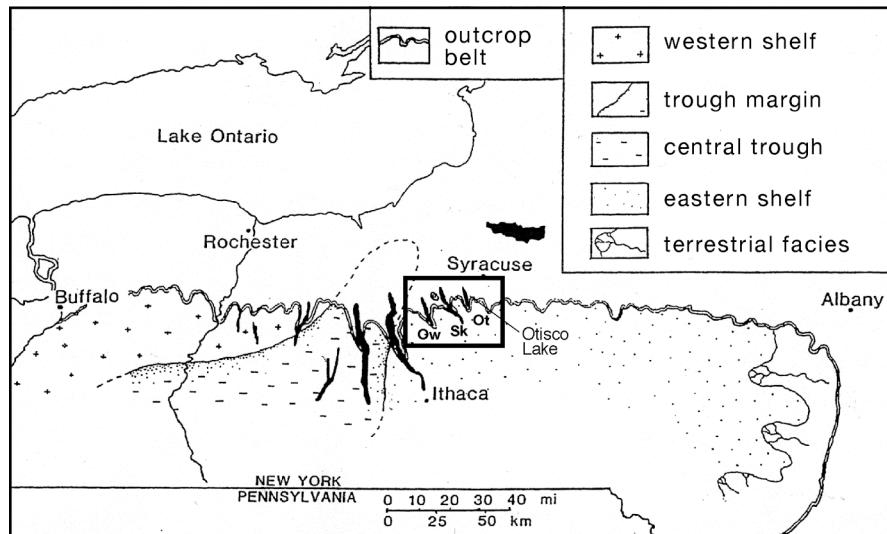
FIGURE 1—Generalized map of New York State showing paleogeography for middle Givetian time, during Ludlowville Formation deposition. Note position of deltaic shoreline and main axis of foreland basin. Study area indicated with box; eastern Finger Lakes initialed: Ow: Owasco Lake; Sk: Skaneateles Lake; Ot: Otisco Lake. Dashed line indicates projected position of Appalachian basin north of the outcrop belt.

Benthic marine communities responded in various ways to changing water depth and sedimentation patterns. Deeper areas of the Acadian foreland basin were occupied much of the time by dysoxic to anoxic water conditions, resulting in deposition of dark, poorly burrowed to laminated shales with monospecific, low diversity assemblages, dominated by the brachiopod *Eumetobolotoechia* and styliolines. Intermediate depth ramp environments, represented by medium to dark gray claystone and mudstone facies, were populated by moderate diversity assemblages of medium to small brachiopods, bivalves, trilobites, small rugose corals and other invertebrates. Shallow shelf areas were populated, during times of low sedimentation, by diverse assemblages of rugose and tabulate corals, bryozoans, brachiopods, and crinoids. Conversely, under higher sedimentation regimes the comparable depth zones favored lower diversity assemblages of brachiopods, such as *Mucrospirifer* and *Tropidoleptus*, and large bivalves. Generally, these biofacies relationships have been inferred from vertical successions of biotic replacement. In the case of the Stag Horn Point submember, however, a faunal gradient is more directly observable because of abrupt lateral facies changes along a rather well-constrained time parallel surface.

PHYSICAL STRATIGRAPHY

The units considered herein are presently assigned to the uppermost portion of the Skaneateles Formation and lower Ludlowville Formation of the Middle Devonian (Givetian) Hamilton Group (Fig. 2).

Butternut and Chenango Members, Skaneateles Formation

The upper Skaneateles Formation is comprised of the Butternut Member, dark gray to nearly black, platy silty shale that carries a sparse to abundant low diversity fauna dominated by the rhynchonellid brachiopod *Eumetabolotoechia multicostum*. The upper portion of the Butternut Member becomes increasingly silty and carries thin beds, typically around 1-2cm, of tabular, laminated siltstone. In some localities east of the study area, thicker siltstone bundles form two minor cycles within the Butternut Member.

The dark, silty upper Butternut Member shales are abruptly and perhaps disconformably overlain by a distinctive shell-rich bed, which was defined as the base of the Centerfield Member by Gray (1986) and subsequently redefined as the base of the Chenango Member of the Skaneateles Formation by Bartholomew et al. (2006). This bed, the Peppermill Gulf Bed (Figs. 2, 3), is somewhat calcareous, blocky mudstone typically about 30-50 cm thick and carries an exceedingly rich fauna of brachiopods, bryozoans, crinoid debris, small- to medium-sized corals, bivalves, and trilobites. This bed shows a marked contrast to the sparsely fossiliferous upper portion of the Butternut Member. The Peppermill Gulf Bed, in turn, is overlain by 4-5m of medium gray slightly silty mudstone that locally carries a moderate fauna of ambocoeliid and chonetid brachiopods, auloporid coral thickets and other faunal elements typical of deeper, slightly dysoxic facies. However, *Eumetabolotoechia* is rare in these beds. Two richer horizons occur near the top of this interval that appear to correspond to the Salt Creek and Browns Creek beds of western New York State (see Savarese et al., 1986). These horizons are calcareous siltstones or silty mudstones that carry a rich fauna typified by atrypid brachiopods and the distinctive pearly brachiopod *Pholidostrophia nacrea*, which also characterizes this interval throughout western New York State. In some areas, abundant proetid trilobites also occur at the level of the Browns Creek horizon as they do in the calcareous facies of the same level in the west. These beds occur at a relatively marked change from mudstone into siltstone that tends to coarsen upward through the remaining 9-10m of the Chenango member. Most of these beds are sparsely fossiliferous and contain a fauna rich in the brachiopod *Tropidoleptus*, along with locally abundant bivalves and the large spiriferid brachiopod *Spinocyrtia*, which occurs in in-situ clusters in some horizons. To the west, this interval also yields rugose and tabulate corals and appears to thin markedly and merge westward into an interval referred to as the Triphammer Falls beds (Saverese et al., 1986) in the Genesee Valley; this is a coral biostrome, generally less than a meter in thickness. The hard upper beds of the Chenango Member, composed of siltstone and sandstone, typically form the caps of waterfalls and a bench in the stream bed.
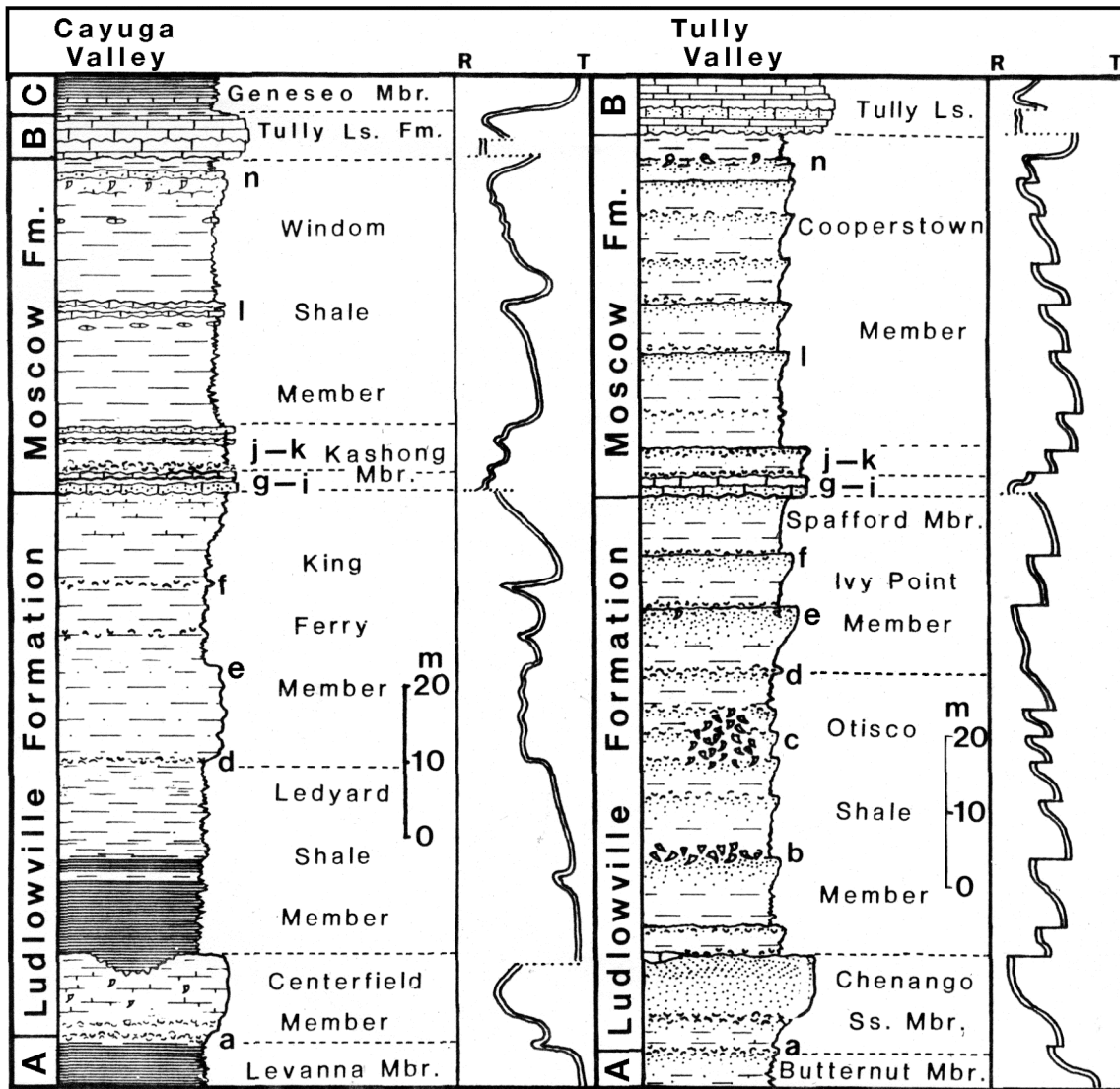
FIGURE 2—Lithostratigraphy and relative sea-level curve for the upper Hamilton Group in the Cayuga Lake and eastern Finger Lakes area. Lettered bed levels are as follows: A) Skaneateles Formation; B) Tully Formation; C) Genesee Formation; a) Peppermill Gulf bed; b) Stag Horn Point coral biostrome; c) Joshua coral biostrome; d) Mt. Vernon (Elmwood Point) bed; e) Ensenore Ravine shell/coral bed; f) Bloomer Creek shell bed; g-i) Portland Point Member= g) Tichenor Ls., h) Deep Run Shale, i) Menteth Ls.; j) RC shell bed; k) Barnes Gully bed; l) Bay View shell bed; n) South Lansing shell bed.

Centerfield Member

The Chenango Member is capped by an interval typically about 0.5 m-thick which is heavily bioturbated and shows exquisite spreite structures of the trace fossil *Zoophycos* (Figs. 2,3) Amidst these burrows are occasional large tabulate corals (i.e.: *Favosites*) up to 1.5 m in diameter. Locally this interval merges into or is overlain by skeletal grainstones containing abundant corals and crinoid debris up to several centimeters in thickness referred to in the Hamilton area as the Stone Mill Limestone Member. The Stone Mill is typically quite thin and discontinuous and this bed (or correlative disconformity surface) marks the base of the Centerfield Member of the Ludlowville Formation. This surface is sharply overlain by soft, highly fossiliferous, slightly silty mudstones presently assigned to the upper, or Halls Landing, submember of the Centerfield Member, named for exposures along the east side of Skaneateles Lake south of Hall Creek. The Halls Landing

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beds, up to about 3 m-thick, carry a very diverse fauna of brachiopods including large *Pseudoatrypa*, *Athryis*, *Mediospirifer*, *Megakozlowskiella*, *Fimbrispirifer*, *Strophodonta demissa*, *Megastrophia*, *Tropidoleptus*, and many others. These brachiopods are admixed with a rather diverse bivalve fauna characterized by *Actinopteria*, *Cornellites*, *Modiomorpha* spp. *Cypricardinia*, and *Cypricardella*. In some levels corals are relatively common and are particularly characterized by large specimens of the domical tabulate *Pleurodictyum*. Some beds carry scattered large rugosans, (*Heliophyllum*, *Cystiphylloides*, and *Heterophrentis*) as well as tabulates, including large favositids. The crinoid grapnel *Ancyrocrinus* is also abundant and large in the Halls Landing beds. The Halls Landing beds, in turn. are capped by a thin, highly fossiliferous coquina horizon carrying locally large rounded phosphatic pebbles, termed the Moonshine Falls Phosphate Bed for exposures along Paines Creek at Moonshine Falls on the east side of Cayuga Lake. This important bed marks the upper boundary of the Centerfield Member and the base of the Ledyard/Otisco members.

Ledyard Member

In the vicinity of Aurora on the east side of Cayuga Lake (Ledyard Township, Cayuga Co.), is the type section of the Ledyard Shale Member (Fig. 3A;Cooper, 1930). In this area, as typified by the section at Moonshine Falls, the Halls Landing Beds of the Centerfield Member are abruptly (disconformably) overlain by hard, strongly jointed dark platy shales that carry a low diversity fauna including small nuculid bivalves, the brachiopod *Eumetabolotoechia*, and gastropods including *Paleozygopleura*. A few higher beds are more diverse. For example, one horizon, approximately 14 m above the Moonshine Falls Bed, locally yields abundant ambocoeliid brachiopods in a fossil debris ("hash layer"); this horizon may correlate eastward to the Stag Horn Point submember of the Otisco Member (see below). Similar more fossiliferous medium gray mudstone horizons occur at at least two other levels within the Ledyard Member, perhaps corresponding to the Glen Cove and Joshua submembers of the Oticso Member. Ledyard deposits at the Cayuga Valley range from 15 m in thickness near Romulus to the northwest to about 40 meters at Aurora to the southeast; these dark shales pass westward into predominantly gray mudstones in the Genesee Valley-Erie County region. In this area, the Ledyard typically is marked by a horizon of abundant pyrite (Alden Pyrite Bed) within 10 m of its base and a dark gray to black interval near the middle subdividing the Ledyard into the Alden and Elma submembers of McCollum (1982).

Otisco Member

The Ledyard Member passes laterally within a relatively short distance (~20 km) into highly fossiliferous, medium gray, silty mudstones, assigned to the Otisco Member (Figs. 2, 4). The Otisco Member makes its westernmost appearance at Portland Point near the south end of Cayuga Lake where the Fir Tree anticline brings the upper beds of this interval into view; these beds are rich in the brachiopods *Athyris*, *Mucrospirifer*, chonetids and others. These beds have already undergone the transition to Otisco facies although they are only about 35 km due southwest of the type Ledyard Shale.

Exposures in the northern portions of the Owasco and Spafford quadrangles show a dark, more fossiliferous facies characterized by an increased abundance of *Eumetabolotoechia*. However, to the south-southeast, particularly in outcrops along stream gullies and bluff exposures along Owasco, Skaneateles and Otisco Lakes, the Otisco Member becomes increasingly fossiliferous. It was divided into component submembers by Smith (1935).

Owasco Lake exposures show transitional facies between typical Ledyard and Otisco shale. In this area, the lower 30 m and the middle shales are dark, platy, and carry an abundance of the brachiopod *Eumetabolotoechia multicostum*. However, other horizons, particularly 30 m or more above the base, have already passed into gray mudstones with abundant brachiopods, typified by *Athyris*, and bivalves, especially *Modiomorpha subalata* and *Cypricardella bellistriata*. Certain horizons yield an abundance of the small rugose coral, *Stereolasma rectum*.

The lower (Stag Horn Point) submember comprises about 8-10 m of gray mudstone with scattered small ellipsoidal concretions (Fig. 2, 4). Immediately overlying the Halls Landing Beds, the interval is fossiliferous and contains a great abundance of the small brachiopod *Arcuaminetes scitulus*. Occasional horizons bear large brachiopods, such as atrypids and *Athyris*, and even small rugosans. These shales pass upward into the darkest portion of the interval, which still yields moderately abundant *Eumetabolotoechia*; this interval, in turn, shows a coarsening-upward pattern into silty mudstones and siltstones that typically cap a small bench-forming often referred to as "the Stag Horn platform". The siltstone, like the upper portion of the underlying Chenango
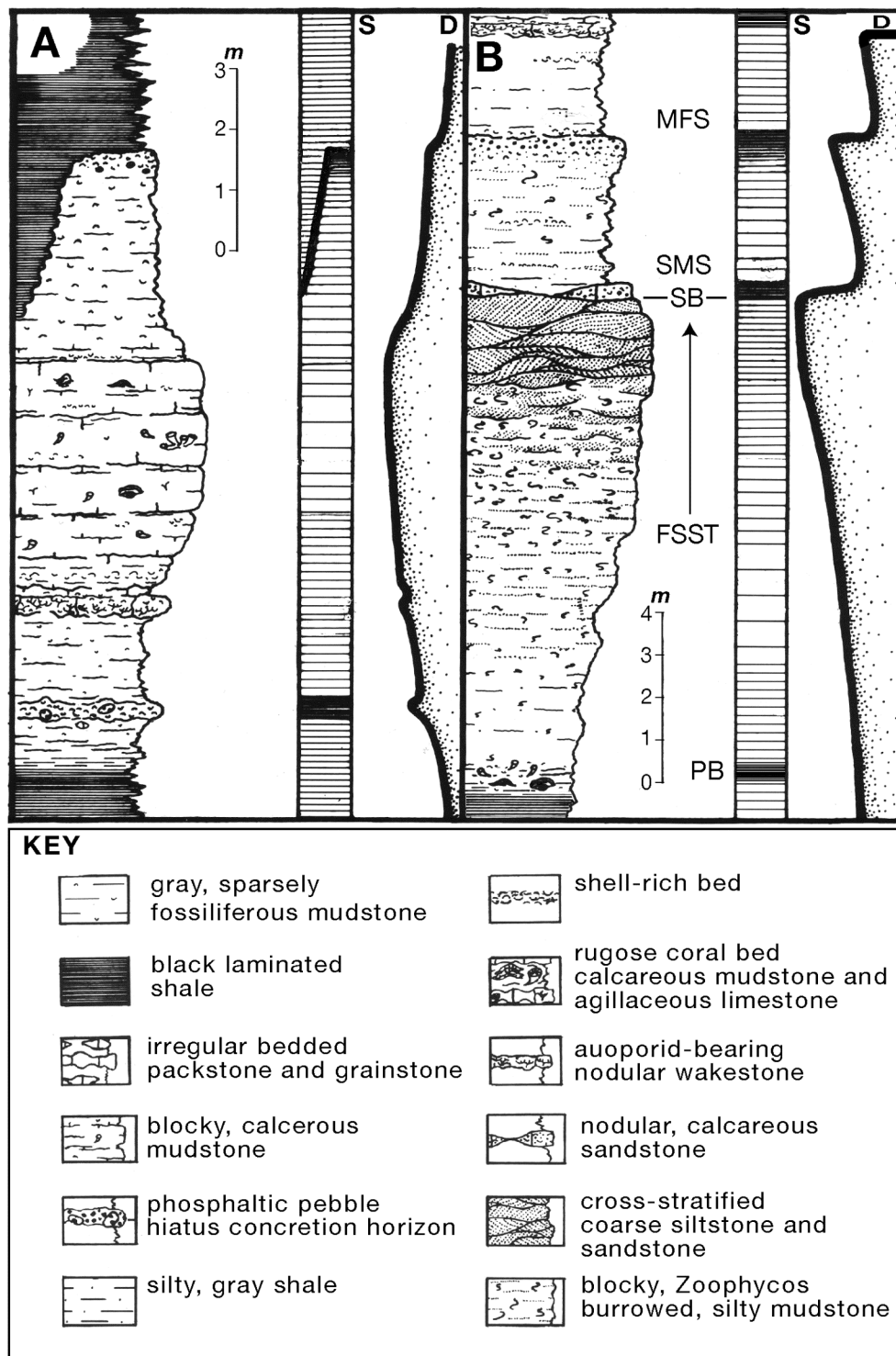
FIGURE 3—Details of the Chenango-Centerfield interval of the upper Skaneateles-Ludlowville Formation boundary in: A) Carbonate rich facies of Seneca-Cayuga Lake area; B) coarser siliciclastics of eastern Finger Lakes area. Bar scales next to columns show spacing of hypothetical equal time units; note condensation shown by tight spacing. PB: precursor bed; FSST: falling stage; SB: sequence boundary; SMS: surface of maximum starvation; MFS: maximum flooding surface. S= shallow; D= deep on relative sea-level curves.

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Member, is heavily bioturbated with *Zoophycos* which show excellent spreiten structures. The upper part of the siltstone carries occasional imbedded rugose corals and small phosphatic nodules, and, indeed, a phosphatic nodule and corroded coral layer typically mantles the Stag Horn siltstone bench. The bench in turn is overlain south of Stag Horn Point and in outcrops to the east-northeast to the Oran quadrangle by a biostrome up to 2m thick of densely packed corals dominated by the rugosans *Heliophyllum*, *Cystiphylloides*, and *Siphonophrentis*. This biostrome, the Stag Horn Point Coral Bed (*sensu* Smith, 1935; Oliver, 1951), is well exposed along the eastern shore of Skaneateles Lake from Stag Horn Point south for about a kilometer to Willow Point. To the north of Stag Horn Point, this interval abruptly changes to an interval of individual thin coral beds, separated by mudstone intervals with occasional tabular siltstones. Corals here occur in wedge and pod-shaped accumulations up to a meter thick but thinning laterally over a few 10s of meters into single 5-10 cm coral horizons. The basal horizon typically carries abundant phosphatic pebbles. This succession is described in more detail below. Creek exposures of the lower Otisco Shale to the north-northwest of the Stag Horn Point area show a subtle coarsening-upward pattern. Dark shales, rich in *Eumetabolotoechia* grade upward into *Zoophycos*-churned silty mudstones, which in turn appear to be capped with shell-rich horizons containing a particular abundance of *Athyris* brachiopods and the small rugose coral *Stereolasma*; rare phosphate nodules have been observed in the shales immediately overlying the bioturbated silty mudstones.
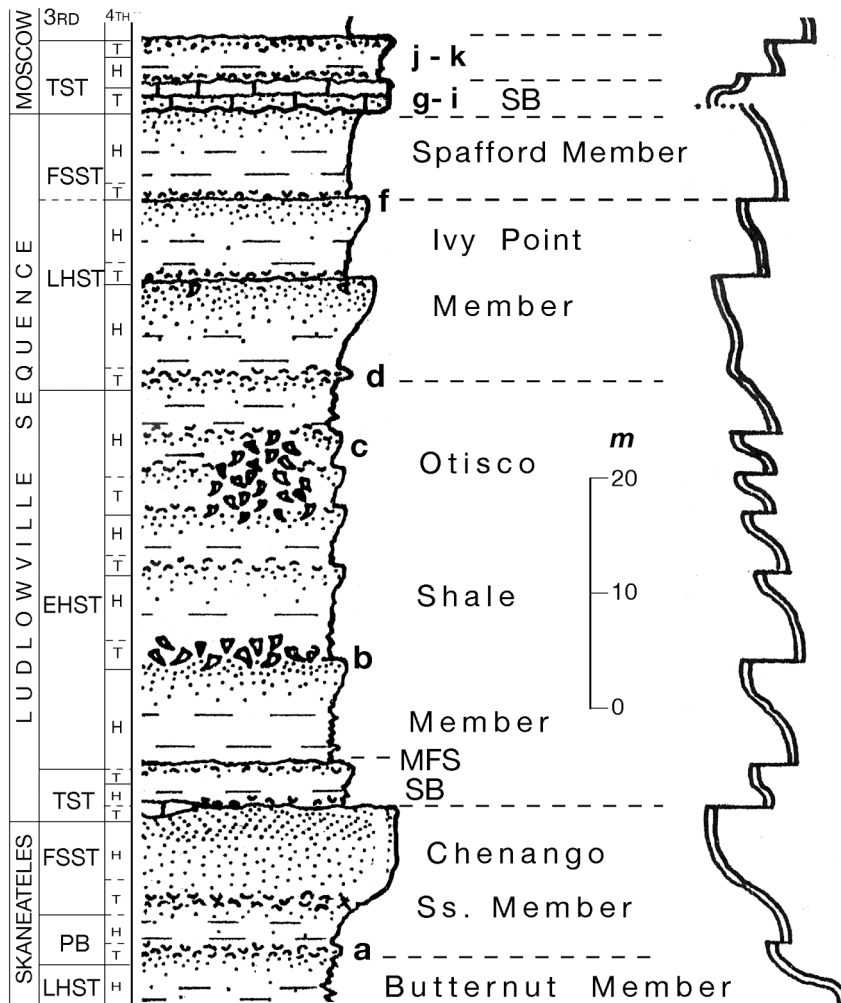
FIGURE 4—Physical and sequence stratigraphic interpretation of Butternut–Chenango-Centerfield –Otisco interval in the study area. Systems tracts for 3rd order (3rd) sequences/formations: TST: transgressive; EHST: early highstand; LHST: late highstand; PB: precursor bed; FSST: falling stage (regressive). Systems tracts for 4th order (4th) sequences; T: transgressive lag; H: highstand-regressive. SB: sequence boundary; MFS: maximum flooding surface.

Approximately 4 meters above the Stag Horn equivalent shell beds in many localities around Skaneateles and Owasco lakes, a second bench-forming silty mudstone interval is overlain by another set of *Athyris* brachiopod- and *Stereolasma* coral-rich horizons. This bed is referred to herein as the Glen Cove shell bed. As with the underlying Stag Horn Point interval, shell horizons may be scattered through approximately one meter of section overlying the silty bench-forming beds. About 3-4 meters higher is yet another subtle bench, that is never associated with a major siltstone (Fig.4). This, is in turn overlain by a variable interval up to 10m thick at Ten Mile Point Ravine at Skaneateles Lake but as little as a few centimeter in thickness in other sections, that carries abundant large rugose corals, as well as *Stereolasma*, *Athyris*, *Pseudoatrypa*, strophomenids, and other diverse brachiopods. This coral-rich division, up to 10m-thick at Ten Mile Point Ravine at Skaneateles Lake but as little as a few centimeter in thickenss in other sections, is referred to as the Joshua submember for exposures along Rte. 80 at Lords Hill near Joshua, in the South Onondaga quadrangle where corals are found in a road-side ditch through approximately 25m of section (Grasso 1970). The uppermost Otisco interval is composed of fossiliferous silty mudstones that carry an abundant shelly fauna including particularly toward the top abundant *Tropidoleptus*, *Athyris*, and atrypid-rich shell beds. The Otisco Member, as does the Ledyard Member, terminates at a silty, blocky mudstone horizon referred to as the Elmwood Point bed. This shell bed has been shown to correlate westward with Grabau's (1898-1899) *Strophalosia* (or *Truncalosia*) bed, referred to as the Mount Vernon bed by Kloc (1983). This forms the base of the Wanakah Member in the west and the Ivy Point Member in the east.

SEQUENCE STRATIGRAPHIC INTERPRETATION

Sequence stratigraphy constitutes a predictive framework for interpreting the sea-level and sedimentational history of the the Devonian strata (see Coe, 2004 and Cateneanu, 2002 for recent summaries). The entire Ludlowville Formation, from the upper (post Chenango) portion of the Centerfield Member, upward to the base of the Tichenor Limestone is considered to record a single third-order (~1-2 myr) depositional sequence (Fig. 3B). The underlying Butternut Member is considered to represent the terminal stages of the highstand portion of the Skaneateles third-order sequence, and is the highstand of an internal fourth-order sequence that begins with a stacked set of shell-rich beds termed the Marietta Beds by Baird et al. (2000). The abruptly upward-coarsening succession of the Chenango Member is regarded as a third-order falling stage (regressive) systems tract.

As noted above, in central New York the base of the Chenango is marked by up to 50 cm of highly fossiliferous mudstone (Figs. 2-4), the Peppermill Gulf bed of Gray (1991). Near Cayuga Lake, the Peppermill Gulf Bed locally carries reworked concretions encrusted with auloporid corals and bryozoans that have been exhumed due to erosion of the upper Butternut Member (Gray, 1991). The Peppermill Gulf Bed passes westward into a thin (1-2cm) pavement of ambocoeliid and other brachiopods that lie sharply on black upper Levanna Shale (Butternut equivalent) shales.

The Peppermill Gulf Bed is a "precursor bed" (*sensu* Brett and Baird, 1996; Fig. 4). By this term, Brett and Baird implied a bed showing: a) faunal/ sedimentological evidence of abrupt shallowing, with b) a sharp contact on underlying beds, a facies discontinuity between that bed and underlying strata, and c) significant condensation within the bed. The precursor bed concept suggests that this type of horizon represents the initiation of the falling stage, i.e. first surface of forced regression. As sea level drops abruptly during this phase, the sea floor comes into increased storm wave action, which serves to erode older muds. Interestingly, the initial portion of the forced regression is marked by condensation, whereas later portions are associated with increased progradation of siliciclastic sediments. Brett and Baird (1996) attributed this condensation to the disequilibrium conditions that followed relatively abrupt lowering of sea level. The sharp basal erosion surface may involve storm erosional bypass along a west-facing paleoslope without replacement of sediment during the initial period of drop prior to progradation, possibly coupled with a certain degree of subaerial accommodation associated with re-grading of stream channels during this time. At another scale of process the precursor bed may represent the transgressive lag deposit of a smaller, higher-order sequence and it is clearly overlain by finer grained sediments that signal a return to deeper conditions, though not as deep as those of the underlying succession (Fig. 4). These beds, reflecting the highstand of a fourth-order sequence, show an abrupt upward change into a shallowing- and coarsening-upward succession that marks the later portion of the falling stage or regression. In essence, the falling stage of a fourth-order cycle is superimposed on the falling stage of a third-order cycle to yield a very abrupt and rapid progradation.

Correlation of the Chenango Member into western New York sections shows that it corresponds to a small portion of the "lower Centerfield" cycle (Savarese et al., 1986; Fig. 3). Moreover it is clear that in the western portion of the state, this interval was relatively turbidity-free, allowing for the growth of corals, whereas in central New York State, the large quantity of prograded silty sediments largely overwhelmed bottom communities giving rise to lower diversity assemblages with *Zoophycos*-producing deposit feeders, bivalves, and quasi-infaunal brachiopods as the dominant elements. The sparse, but well-preserved fossils within the Chenango Member suggests rapid sedimentation and an absence of time averaging, as preservation of patchy, in-situ organisms (such as clusters of *Spinocyrtia* in life position), suggests rapid burial of a sea floor mosaic. The shift to more calcareous siltstones with scattered corals in the upper meter or so of the upper Chenango Member suggests turn over of the depositional cycle from falling stage to either lowstand or early transgression (Figs. 3, 4).

In areas where the Stone Mill bed is developed, a condensed trangressive limestone, comparable to that of the Mottville and Tichenor members of the underlying and overlying formations is present. This shell-rich bed records the accumulation of skeletal material during a time of rising sea level (Figs. 3, 4), but still shallow water condition when seas were relatively free of siliciclastics as a result of the sequestering action created by the drowning of river mouths. Hence, the uppermost Chenango and Stone Mill are interpreted as early TST or lowstand deposits. The Halls Landing beds record later phases of transgression to near maximum flooding. During this time in still relatively shallow waters, seas remained rather free of siliciclastics, although an influx of muds had commenced. It is notable that this interval remains more uniform in thickness and facies across a major swath of western and central New York than does the lower Centerfield or Chenango, a consequence of the marked asymmetry in sedimentation rates of transgressive vs.regressive phasaes, in areas proximal to siliciclastic sources as opposed to western areas where siliciclastics failed to have as strong an impact, even during periods of overall regression.

In one sense, the entire upper Ludlowville Formation may be considered the highstand and falling stage of a third-order sequence (Fig. 4), but of course, this is a highly over simplified way of looking at the pattern. In fact, there are a number of fourth-order sequences that feature prominently within the upper Ludlowville Formation. In the Ledyard and Otisco members at least two and perhaps as many as three such sequences are present. The first begins with the upper Centerfield and Halls Landing beds and has a major flooding surface at the phosphatic Moonshine Falls bed. However, actual maximum flooding may occur somewhat higher, above a package of chonetid-brachiopod bearing shales. In the middle portion of the Stag Horn Point submember, dark platy shales with abundant *Eumetabolotoechia* probably signify the deepest and most dysoxic portion of the cycle.These beds grade upward abruptly into silty shales and mudstones and siltstones with *Zoophycos* churning, much like the underlying Chenango Member, and signify the late highstand or falling stage of the Stag Horn Point fourth-order sequence (Fig. 4). Obviously, shallowing triggered progradation of a local siliciclastic wedge well into the foreland basin, but it appears that during the time of the deposition of the "Stag Horn platform", there may have been some active tectonism in the basin which caused local steepening of the prograded wedge and initiated erosion. In a broad sense, the Stag Horn coral biostrome and its lateral equivalent shell beds record sediment starvation associated with initial transgression of the next overlying small-scale depositional cycle. The following succession shows a change upward into sparsely fossiliferous shales and these, in turn, into silty mudstones and siltstones that cap the Glen Cove submember.

THE STAG HORN POINT SUBMEMBER: A CASE STUDY IN SEA FLOOR DEPOSITIONAL PROCESSES AND FAUNAL GRADIENTS

As already noted, the Stag Horn Point cycle represents a probable fourth-order cycle of approximately 100-400 ka duration. It commences with Moonshine Falls, shelly beds formed during sea level rise, deepens upward into mudstones and shallows into siltstones. Where completely developed, bench-forming siltstone, approximately 3m-thick, occurs near the top of the cycle, immediately below the Stag Horn Point coral biostrome (Figs. 6, 7). Large, pyrite-cored concretions are abundant, particularly within its upper portion (0.5-1m below the upper contact). This interval is overlain sharply by the Stag Horn Point coral biostrome in the southern and eastern portion of the study region (Smith 1912, 1935; Oliver, 1951; Brett et al., 1986). As noted above, the biostrome and coeval transported debris are interpreted as having formed during a time of relative sediment starvation associated with sea level rise (Fig. 4).

FIGURE 5—Map of study area in the Owasco-Skaneateles-Otisco Lake region. Inset shows a portion of the Spafford 7.5' Quadrangle with key localities along east side of Skaneateles Lake; arrows denote unnamed coves between Jenney and Stag Horn Points, discussed in the text. Also note Glen Cove on west side of lake.

A remarkable series of outcrops occurs along the shore of Skaneateles Lake northwest from Stag Horn Point (Smith, 1912, 1935; Oliver, 1951; Brett et al., 1986; Figs. 5, 6). At Stag Horn Point, the full development of the siltstone platform with its sharp, concretionary top and overlying coral biostrome is in place. At the contact between the biostrome and the platform, the siltstone is heavily bioturbated by *Zoophycos* and carries corroded corals and phosphatic nodules, an indication of sediment-starvation and time-averaging. The phosphatic nodules include steinkerns of enrolled trilobites, trilobite cephala, conulariids, and other fossil objects, but most are amorphous, black fluorapatite or collophane. The occurrence of corals attached to phosphatic nodules is significant as it indicates that the period of phosphatic nodule growth preceded the development of the coral biostrome. Indeed, the phosphatic nodules formed interstitially within the upper portions of the underlying silts during a period of low sediment input and at a redox boundary; they must have been reworked as many of them are pebbles that have been smoothed, corroded, bored, and encrusted. Pebbles may have provided the initial substrates of the founding corals of the biostrome. However, coral skeletons soon

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provided a hard substrate for subsequent generations; a classic case for taphonomic feedback (Kidwell and Jablonski 1983). The upper portion of the coral bed appears to be rather abruptly overlain by sparsely fossiliferous gray mudstone. In places, the coral thicket shows horizons of concentrated corals that may represent periods of storm reworking.

Within the ~800m stretch between Stag Horn Point and Jenney Point a remarkable series of outcrops expose what appears to be the edge of the platform with down-draped, onlapping beds that represent locally transported debris shed from the platform edge (Figs. 5, 6). Exposures are poor immediately north of Stag Horn Point but minor outcrops on a bluff between cottages in this stretch along Chase Point show a fairly typically developed platform of siltstone overlain by coral biostrome. However, approaching the north edge of Chase Point, corals overlying the platform occupy a thinner interval. At the north edge of the small cove immediately north of Chase Point, the last portion of the normal platform is exposed with concretionary silts underlying the typical siltstone platform, which is nearly denuded of corals and immediately overlain by dark, middle Otisco Shale Fig. 6). Northward from this area, a gently sloping truncated platform edge occurs toward the middle of the cove with a slope of approximately five degrees representing a truncation surface that cuts down approximately 2-2.5m below the Stag Horn bench elevation. It is particularly notable that the beds in this area appear to dip in a northward direction counter to the regional dip (Fig.6). These beds include a series of packages of coral rich rubble (Fig. 6) of which the lowest, exposed typically below lake level of Skaneateles Lake, carries a corroded coral fragment and phosphatic pebble lag at its base and passes upward into about 2 cm of hummocky-laminated siltstone.
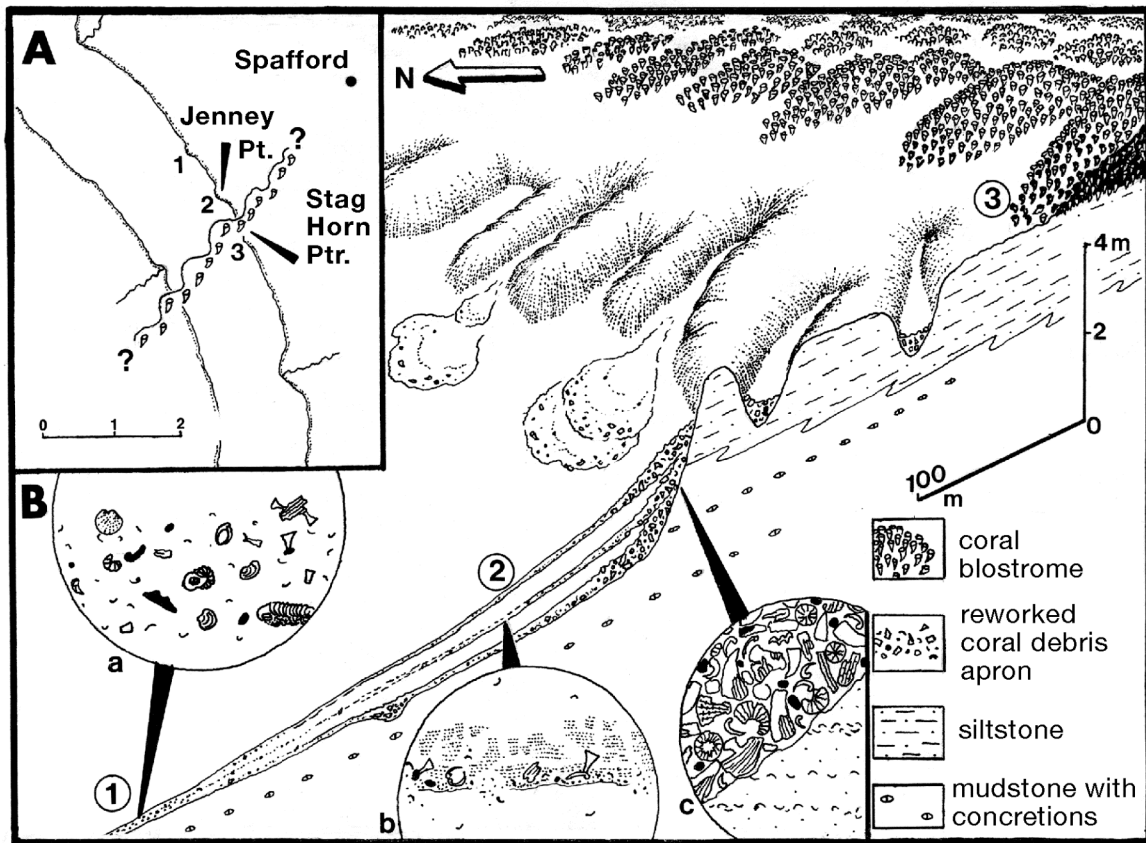
FIGURE 6—Schematic reconstructed cross section of truncated edge of Stag Horn Point siltstone platform as exposed along the east shore of Skaneateles Lake from Stag Horn Point northwestward to north of Jenney Point. Inset circles show details of sedimentary fabric along gradient: a) distal section showing bioturbated sediment with small corals, brachiopods and trilobites; note black phosphatic nodules; b) sharply based bioturbated siltstone bed; note small corals attached to phosphatic nodules; c) sharp contact between lower Otisco mudstone and base of coral debris apron. Modified from Brett et al. (1986).

A series of higher beds of siltstone and coral rubble alternate with more sparsely fossiliferous medium gray mudstone with scattered, transported, rugose corals (Fig. 6), as well as indigenous shelly fauna including athryid and chonetid brachiopods, *Greenops* and *Eldredgeops* trilobites, and others. The intervening beds include four siltstone layers that can be traced laterally along the shore line and appear to thicken toward the platform. Immediately adjacent to the base of the truncated platform is a thick wedge of packed corals. These appear to have been transported off the platform and have accumulated at its margin as a debris apron. The siltstone debris layers appear to finger out from this wedge and become separated by the aforementioned intervening gray mudstones. At the north end of this cove, a reverse pattern is observed where siltstone and coral debris layers rise upward onto the truncated edge of the platform again. It is unlikely that this reappearance of the edge of the platform is a result of the curvature of the cove since this area would, if anything, project outward and should be farther away from the platform edge. Conversely, it is probable that a spur of the platform edge extended out on the opposite side of a large channel-like feature that occupies much of the wall of the cove (Fig. 6). Extending into the next cove that lies to the southeast of Jenney Point, again, the reverse succession is seen. At the south end of this cove, the siltstone platform remains partially truncated; about 30m further to the south, the pyrite-cored concretions, which normally lie below the platform are directly overlain by coral rubble. In places, the coral rubble thickens to about one meter as the truncation surface runs downward from the pyritic concretions to nearly the level of Skaneateles Lake. Once again, a series of graded rubble beds and siltstones emanate from this region and extend out along the cliff in this cove. At the far north end, immediately adjacent to the southern cottage on Jenney Point, the edge of the platform again appears to be in sight and thickened rubble beds thicken out from it and slope gently to the south. Once again there is a suggestion of a large channel that is separating two spurs of an eroded platform edge.

The Staghorn Point interval is exposed in a large cliff, north Jenney Point but with no siltstone platform present (Fig. 6). Rather, a succession of gray mudstones with thin siltstones showing planar to hummocky lamination and coral fragments is present. Near the base of the succession, a horizon yielding abundant coral fragments and phosphatic nodules appears to be correlative with the base of the Stag Horn biostrome further up on the platform; a distinctive hummocky bedded siltstone band occurs slightly about this. From this region northward, the siltstone platform is no longer present and instead, the Stag Horn interval is represented by a 1.5m-thick succession of shell beds. As far north as Ten Mile Point Creek (~2.5 km north), the horizon of the Stag Horn interval still contains abundant phosphatic nodules as well as corroded and non-corroded corals, athyrid brachiopods and other fossils (Figs. 6, 7).

On the west side of Skaneateles Lake, the Stag Horn platform is nowhere visible, having dipped below the surface of Skaneateles Lake southward of the first exposures of the lower Otisco Member in Glen Cove ravine. At this section, the uppermost beds of the Halls Landing submember are exposed in low banks, followed by chonetid-rich shales. Approximately 10 m above this succession there occurs a thin, highly fossiliferous shell bed containing abundant brachiopods including *Athyris*, *Mediospirifer*, as well as small and large coral fragments. This clearly represents the Stag Horn biostrome position, although larger corals are only represented by scattered, corroded "biscuits" and are evidently allochthonous. Another resistant bench capped by abundant stereolasmatid corals occurs approximately 4m above the Stag Horn Point horizon occurs.

Northward to Carpenter Point at Bear Swamp Creek, the Stag Horn-equivalent bed maintains much of its character, with scattered corroded corals, *Athyris*, small stereolasmatid corals, as well as varied bivalves. At Fall Brook, ~ 1.5 km farther northwest, the bed is barely discernable as a thin series of *Athyris*-rich shell horizons with very rare stereolamatid corals. However, there are thin siltstones, possibly emanating from the eroding silty platform to the southeast, and at least one of these showed a small coral at its base as well as an articulated trilobite. Layers rich in columns of *Eutaxocrinus* also provide a link between this succession of the Stag Horn submember and those seen in outcrops along Owasco Lake (Fig. 7).

On Owasco Lake, the interval of the Stag Horn Point submember is present on several creeks, most notably Seward Point ravine on the east side, and Casowasco ravine on the west side; the latter outcrop has been slumped over in recent years and no longer exposes this interval. At both of these sections phosphatic nodules are rare, but shell pavements rich in *Athyris* and *Eumetabolotoechia* brachiopods are associated with abundant *Eutaxocrinus*, some of them preserved as clusters of articulated fossils. Both Casowasco ravine and Seward
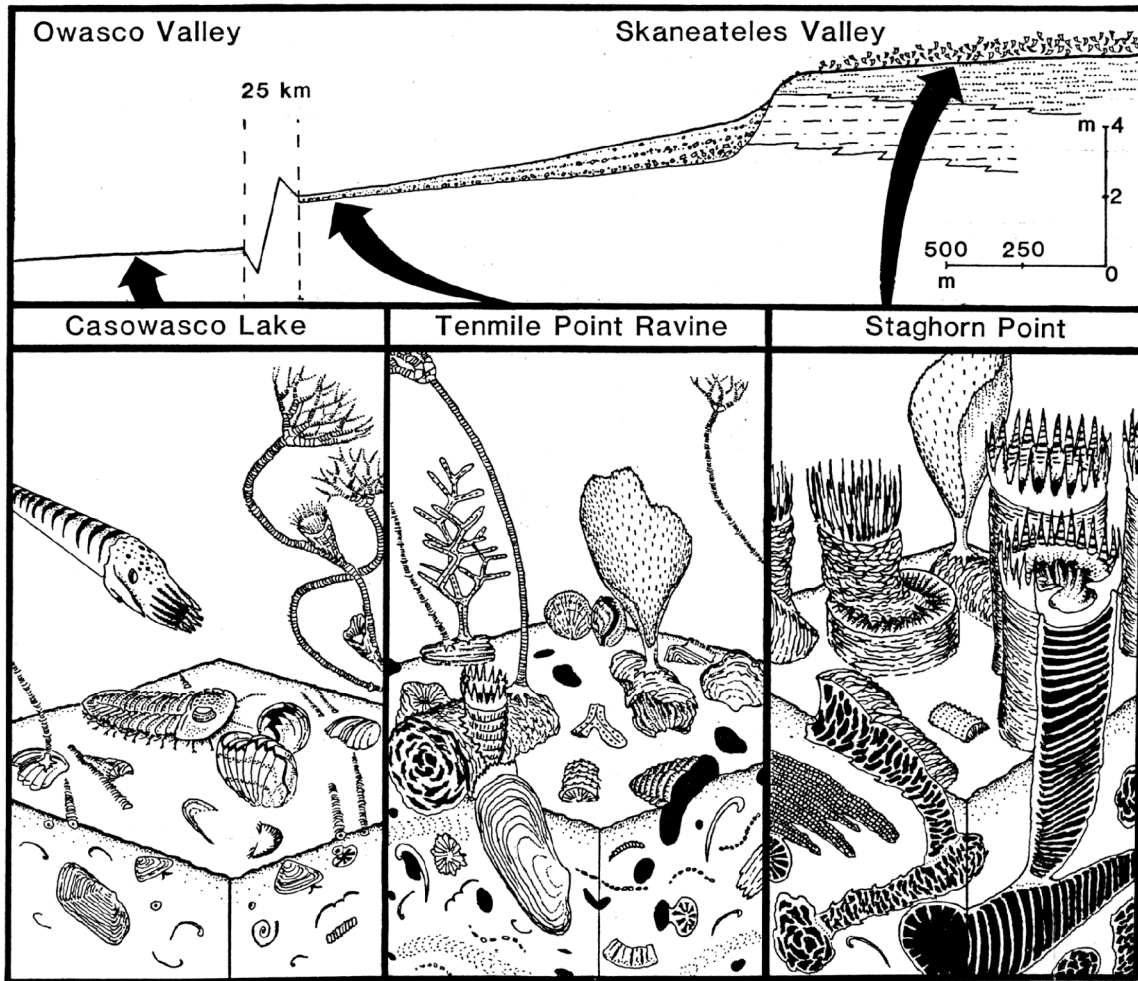
FIGURE 7—Schematic west-east transect of paleoenvironments during Stag Horn coral bed deposition in the Owasco-Skaneateles Lake region, showing reconstructions of seafloor communities at different positions of gradient from bank top coral biostrome (right side) to deeper ramp. **Casowasco Lake**) deeper water assemblage characterized by leiorhynchid brachiopod *Eumetabolotoechia*, phacopid trilobite, burrowing *Cypricardella* and small flexible crinoid, *Eutaxocrinus;* note attachment of small *Stereolasma* corals to crinoid stems; orthoconic nautiloids are common. **Ten Mile Point Ravine**) intermediate depth communities, including brachiopods such as *Rhipidomella*, burrowing bivalve, *Modiomorpha* bivalves, encrusting *Taeniopora* and fenestrate bryozoans and crinoids attached to reworked corals. **Staghorn Point**) Stag Horn Point coral biostrome with large rugose corals *Cystiphylloides* and *Siphonophrentis*. From Brett et al. (1990).

Point ravine have yielded an intriguing occurrence of stereolasmatid corals. The latter are not found commonly attached to benthic shells, but rather occur encrusting along the sides of columns of *Eutaxocrinus* (Fig. 7), which being preserved in an articulated condition evidently were buried very shortly after death. Hence it is clear that these corals attached to the sides of the upright, living crinoid columns. This was possibly a strategy to elevate the corals above a relatively dysoxic, unfavorable sea bottom and allowed them to live in a marginal environment at the outer limit of the Stag Horn coral biostrome. No reworked coral fragments have been found in these sections.

Exposures along Otisco Lake, east of Skaneateles Lake (Fig. 5), provided further information on the orientation of the Stag Horn platform. Oliver (1951) surveyed a series of gullies primarily along the west side of Otisco Lake and in these he was able to map out both the Stag Horn Point and Joshua coral beds. With respect to the Stag Horn submember, he found a siltstone platform and dense overlying coral biostrome in all but the

northern-most sections. In the latter, the platform was present, but corals were relatively rare. However, in outcrops to the north in twin-glen tributaries of Willowdale Ravine, the Stag Horn Point coral bed interval is represented only by thin shell horizons with *Athyris* and stereolasmatid corals that overlie a more sparsely fossiliferous silty mudstone. The main body of the Stag Horn cycle, in this area, is occupied by dark gray shales with abundant *Eumetabolotoechia* and *Pustulatia* (*Vitulina*). In a small stream along Coon Hill Road, the Stag Horn level is nearly unrecognizable, although a slightly more fossiliferous zone with some *Athyris* brachiopods occurs between dark, *Eumetabolotoechia*-bearing shales, and seems to mark the position of this interval. Again, in this area, a rapid north-northwest transition from the richly fossiliferous coral platform facies to deeper basin-margin sediments is evident. Together, these outcrops help to constrain the orientation of the Stag Horn silty platform margin and thereby the limit of the coral biostrome to the northwest. Based upon all available information, the line of the front edge of the siltstone escarpment must run north-northeast to south-southwest and cut diagonally across from the north end of Otisco Lake to the east side Skaneateles Lake approximately at the position of Chase Point. From there, it dips out of sight beneath Skaneateles Lake and plunges into the subsurface (see Fig. 6).

The transitional strata northward of the platform provide an excellent transect of faunas from high density, coral-rich beds, to coral rubble beds populated by in-situ stereolasmatid corals, *Athyris*, and *Eutaxocrinus* (Fig. 7). The latter assemblage persists onward into the basin with the addition of *Eumetabolotoechia* in the vicinity of Owasco Lake exposures and the diminution in abundance of *Athyris* and other brachiopods. Corals were able to inhabit these environments only by attaching themselves to the elevated columns of *Eutaxocrinus* (Fig. 7).

SEQUENCE OF EVENTS IN THE STAG HORN POINT SUBMEMBER DEVELOPMENT

Based upon the available evidence enumerated in the previous paragraphs, we visualize the following series of events in the formation of the Stag Horn siltstone platform, coral biostrome, coral rubble beds, and their lateral equivalents (Fig. 8).

Stage I—Initially, a minor fall in sea level initiated progradation of silty muds and silts into the basin at least to the position of the Stag Horn platform line shown on Figure 6. It is possible that very rapid progradation of silts resulted in an unstable, northwest-facing, basinward slope that was somehow subjected to later, transgression-related erosion (Fig. 8A).

Stage II—Reworking and winnowing associated with the lowstand of this cycle may have resulted in the presence of the heavily bioturbated siltstone that caps the platform. During this time and also during the ensuing interval of sediment starvation, carbonate concretions may have begun to nucleate in the sediment about a meter below the top of the platform around pyrite nodules, forming pipe-like to ellipsoidal concretions.

Stage III—Sediment-starvation, probably associated with a sea-level rise of a few meters, resulted both in sediment- stabilization on the top of the platform and reduction of sediment influx owing to transgression-related reduction of delta progradation and the stabilization of the sea floor. During this interval of time, phosphatic nodules apparently developed within the uppermost portion of the silty sediment. In some cases, this phosphate nucleated both within and around the skeletons of various animals, such as trilobites, etc. This interval was followed by a period of minor erosion and reworking, probably the result of further and more extensive sediment-starvation coupled with minor current activity. This resulted in the reworking of phosphate nodules as pebbles on the sea floor. These pebbles were exposed and subsequently bored into by worms that formed *Trypanites* borings, and encrusted by bryozoans and corals. This period of sediment-starvation appears to have been widespread, with the occurrence of phosphatic nodules in silty mudstone northward of the edge of the platform (Fig. 7).

Stage IV—During the next phase, corals grew abundantly and prolifically upon the upper portion of the platform (Fig. 8B). It is likely that at this time, deeper portions of the sea bottom were beginning to be colonized by athyrid brachiopods, stereolasmatid corals, bivalves, and various other organisms. Coral growth on the platform may have been initiated by the presence of pebbly phosphatic substrate that provided areas for settlement of pioneering coral larvae, but as noted, their skeletons would have provided much larger substrates for subsequent generations during the formation of the biostrome. Intermittently during the course of development of the Stag Horn bioherm, the coral thickets were swept by strong, storm-generated currents. These events reworked corals, piling them together, and, in some cases, buried corals in an upright-standing

orientation. At other times mud infiltrated into baffles of upright-standing, *in situ* corals. A few brachiopods and bivalves dwelled within this mudstone but the predominant organisms living on the Stag Horn bench were solitary rugosans.
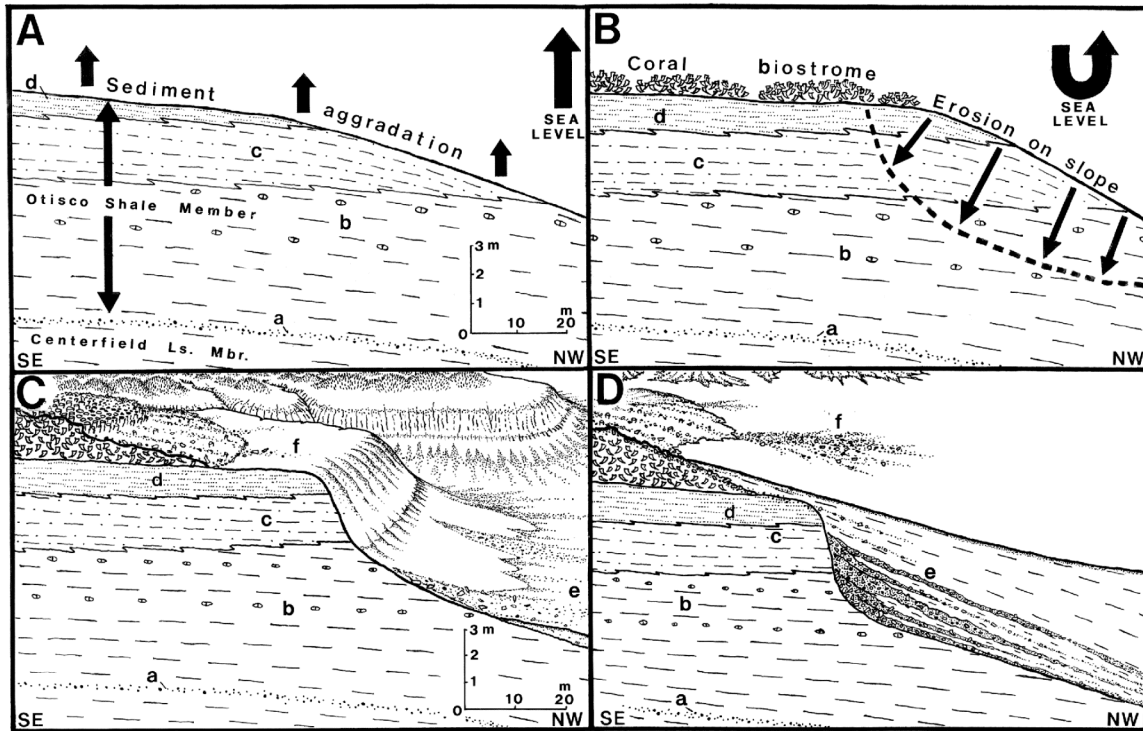
FIGURE 8—Reconstructed series of events in the development and burial of Stag Horn Point siltstone platform and coral biostrome. A) progradation of siltstone tongue into central New York area during regression (falling stage). B) colonization of upper surface of siltstone bench by rugose coral biostrome and storm erosional scouring and steepening of downslope end of submarine siltstone platform. C) development of spur and groove topography of siltstone platform edge, reworking of coral debris from biostrome into surrounding area. D) burial of coral biostrome and debris apron by highstand muds.

    Stage V—It is possible that, at approximately this time, a minor "down-to-the-northwest" faulting event may have taken place, producing a northwest-facing submarine scarp (Fig. 8B). Following such an event, the scarp would have retreated due to differential erosion and associated sediment-starvation and bypass on the up-thrown side. Such erosion, controlled by deep-storm action and tide-generated currents, may explain the formation of the aforementioned channel-like features along the platform edge. It seems likely that this only would have occurred if the platform already had a steep gradient, but once started, this erosional furrowing could propagate by headward erosion cutting into the front edge of the platform to form a series of spur and groove like features (Figs. 6, 8C). Channels several tens of meters across developed between ridge/pinnacle-like elevated remnants along the platform edge.

    During the interval of luxuriant coral growth on the platform, large storms intermittently battered this surface, knocking corals over and burying some and also producing a flow of rubble off the truncated platform margin; this produced accumulations of abundant reworked coral skeletons particularly in the channel-like areas between platform spurs. More distally, coral rubble layers graded laterally (downslope) into graded coral and siltstone beds up to 3-4 cm thick. The presence of hummocky cross-stratification and escape burrows indicates rapid deposition as single graded units during bottom impinging storm events. The occurrence of reworked silt in the higher beds of coral rubble suggests that siltstone flooring the escarpment lip was being actively dissected (Fig. 8C). These tabular siltstones extend hundreds of meters, if not over a kilometer out into the basin as suggested by the occurrence of similar silts in the section at Fall Brook on the west side of Skaneateles Lake.

All of this growth and reworking of corals occurred within the context of relative sediment-starvation probably produced by a rise of base level by a few meters which flooded coastal areas of the Catskill Delta complex and produced estuaries and bays that sequestered sediment temporarily. Biostrome growth may have continued for several millennia before being exterminated and buried by muddy sediments. Intermittent mud influx was ongoing as evidenced by the sparsely fossiliferous mudstone layers that lie between coral and siltstone event beds that lie adjacent to the Stag Horn platform (Fig. 8C, D). Distal ends of the same events may have been responsible for the mass mortality and burial of the thickets of *Eutaxocrinus* farther to the northwest.

Stage VI. Ultimately, the growth of the Stag Horn Point coral bed was terminated (Fig. 8D). This change was probably brought about by a combination of increased water depth, decreasing turbulence and oxygen levels, and increasing turbidity as rates of base level rise slowed, allowing increasing amounts of muddy sediment to prograde offshore. This change was marked by a return to moderate diversity assemblages of small brachiopods and bivalves.

CONCLUSIONS

The middle portion of the Hamilton Group in the eastern Finger Lakes area provides an excellent case study in the interactions of relative sea level fluctuation, related changing patterns of sedimentation, and concurrent biotic responses. Decameter and smaller scale sedimentary cycles are superimposed on a larger scale pattern of sea-level rise and fall, represented by the upper Skaneateles Formation and the Centerfield and Ledyard/Otisco members of the Ludlowville Formation.

Sea-level highstands in this area of the foreland basin were characterized by low diversity assemblages of brachiopods and small mollusks, adapted to dysoxic mud substrates. Falls in relative sea level not only led to shallowing of the seafloor but also increased progradation of muds and silts into the basin. Faunas changed upward to moderate diversity assemblages of larger semi-infaunal brachiopods and bivalves, as well as the deposit feeding organisms responsible for the complex spreiten trace *Zoophycos*.

Conversely, rises in relative sea level were associated with abrupt reduction in sediment supply. Concretions and phosphatic nodules may have developed in older sediments during an interval of sediment starvation representing rapid sea level rise. Highly diverse assemblages including, but not limited to, abundant rugose and some tabulate corals, occupied the shallowest water settings during relatively clean siliciclastic free intervals. In the case of the lower Otisco Stag Horn cycle, a thriving coral biostrome was developed on a shallow prograded siltstone platform during this phase. This example also shows that during times of relatively low sediment input, normal submarine processes such as storm-generated waves and currents may be effective in eroding previously deposited sediments and transporting them further basinward, particularly along the sloping edge of a prograded platform.

The rapid lateral facies changes and constrained stratigraphy of the lower Ludlowville Formation in central New York provide excellent opportunities to examine lateral-, depth- and sedimentation-related gradients of environmental and biofacies change. This is an area wherein exceptional preservation and exposure can be used to test general hypotheses about the connection between physical environmental factors and biotic change on the scale of millennia to hundreds of thousands of years.

ACKNOWLEDGMENTS

REFERENCES CITED

Baird, G.C., Brett, C.E. and Ver Straeten, C.A. 2000. Facies and fossils of the lower Hamilton Group (Middle Devonian) in the Livingston County-Onondaga County region. New York State Geological Association 72nd Annual Meeting Field Trip Guidebook, p. 155-175.

Bartholomew, A. J., Brett, C.E., and Baird, G.C. and DeSantis, M. 2006. Sequence stratigraphy of the Middle Devonian at the border of the Michigan: correlations with New York and implications for sea-level change and paleogeography. Northeastern Geology, v. 28, 2-33.

Bonuso, N., Newton, C.R., Brower, J.C., and Ivany, L.C., 2002a, Does coordinated stasis yield taxonomic and ecologic stability?: Middle Devonian Hamilton Group of central New York: Geology, v. 30, p. 1055-1058.

Bonuso, N., Newton, C.R., Brower, J.C., and Ivany, L.C., 2002b, Statistical testing of community patterns: uppermost Hamilton Group, Middle Devonian (New York: USA): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 185, p. 1-24.

Brett, C. E. and Baird, G.C., 1985, Carbonate-shale cycles in the Middle Devonian of New York: an evaluation of models for the origin of limestones in terrigenous shelf sequences, Geology, v.. 13: 324-327.

Brett, C. E. and Baird, G. C., 1995, Coordinated stasis and evolutionary ecology of Silurian to Middle Devonian marine biotas in the Appalachian basin: *in* Erwin, D. and Anstey, R., eds., New Approaches to Speciation in the Fossil Record. Columbia University Press, New York, 285-315.

Brett, C.E., Ivany, L.C., and Schopf, K.M., 1996, Coordinated stasis: An overview: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 127, p. 1-20.

Brett, C.E., Miller, K.B., and Baird, G.C., 1990, A temporal hierarchy of paleoecological process in a Middle Devonian epeiric sea: *in* Miller, W., III, ed., Paleocommunity Temporal Dynamics: The Long-Term Development of Multispecies Assemblages: Paleontological Society Special Publication, v. 5, p. 178-209.

Brett, C.E., Speyer, S.E. and Baird, G.C., 1986. Storm-generated sedimentary units: tempestite proximality and event stratification in the Middle Devonian Hamilton Group of New York. **In** Brett, C.E., ed., Dynamic Stratigraphy and Depositional Environments of the Middle Devonian Hamilton Group in New York State Part 1. N.Y. State Mus. Bull. 457, p.129-156.

Brower, J.C., 1987, Multivariate analysis in community paleoecology: Compass v. 64, 89-102.

Brower, J.C., and Nye, O.B., Jr., 1991, Quantitative analysis of paleocommunities in the lower part of the Hamilton Group near Cazenovia, New York. **In** Landing, E. and Brett, C.E., eds., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part II. New York State Museum Bulletin 469, p. 37-74.

Brett, C.E., Bartholomew, A.J. and Baird, G.C.. 2007. Biofacies Recurrence in the Middle Devonian of New York State: an example with implications for evolutionary paleoecology. Palaios v. 22, p.306-324.

Brett, C.E., Miller, K.B. and Baird, G.C., 1986, Sedimentary cycles and lateral facies gradients across a Middle Devonian" shelf-to-basin ramp, Ludlowville Formation, Cayuga Valley: *N.Y. State Geological Association 58th Annual Meeting Field Trip Guidebook*, Ithaca, N.Y. p. 80-128.

Cateneanu, O., 2002, Sequence stratigraphy of clastic systems: concepts, merits, and pitfalls: *Journal of African Earth Sciences*, v. 35, p.1-43

Coe, A. L., ed., 2003, The Sedimentary Record of Sea-Level Change, Cambridge University Press, 287 p.

Cooper, G.A., 1930, Stratigraphy of the Hamilton Group of New York, Parts 1 and 2, American Journal of Science, 5th Series, vol. 19: 116-134.

Eldredge, N. and Gould, S.J. 1972, Punctuated equilibrium, an alternative to phyletic gradualism, 540-547, **In** Schopf, T.J.M. (ed), *Models in Palaeobiology*, Freedman, Cooper, San Francisco.

Ettensohn, F.R. 1998, Compressional tectonic controls on epicontinental black shale deposition: Devonian-Mississippian examples from North America, 109-128, **In** Schieber, J., Zimmerle, W., and Sethi, P.S. (eds),

Shales and Mudstones, vol 1 (Basin Studies, Sedimentology, and Paleontology), E. Schweitzerbart'sche Verlagsbuchandlung (Nägale und Obermiller), Stuttgart.

Ettensohn, F. R., 1998, Compressional tectonic controls on epicontinental black shale deposition: Devonian-Mississippian examples from North America. Schieber, J., Zimmerle, W. and Sethi, P. S., eds., Shales and Mudstones, vol., 1 (Basin Studies, Sedimentology, and Paleontology): 109-128: E. Schweizerbart' sche Verlagsbuchhandlung (Nagele u. Obermiler); Stuttgart, Germany.

Gray, L.M. 1991. The paleoecology, origin, and significance of a shell-rich bed in the lowermost part of the Ludlowville Formation (Middle Devonian, Central New York), **In** Landing, E., and Brett, C. E., eds., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part II, New York State Museum Bulletin, v. 469, pp. 93-106

Kaufmann, B. 2006, Calibrating the Devonian time scale: A synthesis of U-Pb ID-TIMS ages and conodont stratigraphy, Earth Science Reviews, vol. 76: 175-190.

Kidwell, S.M. and Jablonski, D. 1983. Taphonomic feedback: ecological consequences of shell accumulation: *in* Tevesz, P.J., and McCall, P.J., eds., Biotic interactions in recent and fossil benthic communities: Plenum Press, New York, p. 195-248.

Kloc, G.J. 1983. Stratigraphic distribution of ammonoids from the Middle Devonian Ludlowville Formation. Unpubl. MS Thesis, SUNY at Buffalo, 73 p.

Grabau, A.W., 1898-1899. Geology and paleontology of Eighteen Mile Creek and the lake shore sections of Erie County, NewYork. Buffalo Society of Natural History Bulletin 6, 403 p.

Grasso, T. X., 1970, Paleontology, Stratigraphy, and Paleoecology of the Ludlowville and Moscow Formations (upper Hamilton Group) in central New York, D1-D17, **In** Heaslip, W.G. (ed), *Field trip Guidebook*, New York State Geological Association, 42nd Annual Meeting, Cortland.

Heckel, P.H., 1973, Nature, origin, and significance of the Tully Limestone, Geological Society of America, Special Paper, vol. 138, 244p., Boulder.

McCollum 1980. Paleoecology of the Ledyard Shale Member (Middle Devonian: Givetian) of New York. Unpubl. PhD Dissertation, SUNY at Binghamton, 143 p.

Miller, K.B., 1991, High-resolution correlation within a storm-dominated muddy epeiric sea: Taphofacies of the Middle Devonian Wanakah Member, western New York.: In Landing, E. and Brett, C.E., eds., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part II. New York State Museum Bulletin 469, p.129-152.

Oliver, W. A., Jr., 1951, Middle Devonian coral beds of central New York. American Journal of Science, v. 249, p. 705-728.

Savarese, M., Gray, L.M., and Brett, C.E., 1986, Faunal and lithologic cyclicity in the Centerfield Member (Middle Devonian: Hamilton Group) of western New York: A reinterpretation of depositional history: In Brett, C.E., ed., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part I. New York State Museum Bulletin, v. 458, p. 446-454.

Smith, B., 1912, Observations on the structure of some coral beds in the Hamilton Shale. Proceedings of the Philadelphia Academy of Natural Sciences, 1912,

Smith, B., 1935, Geology and mineral resources of the Skaneateles Quadrangle. New York State Museum Bulletin vol. 300, 120 p.
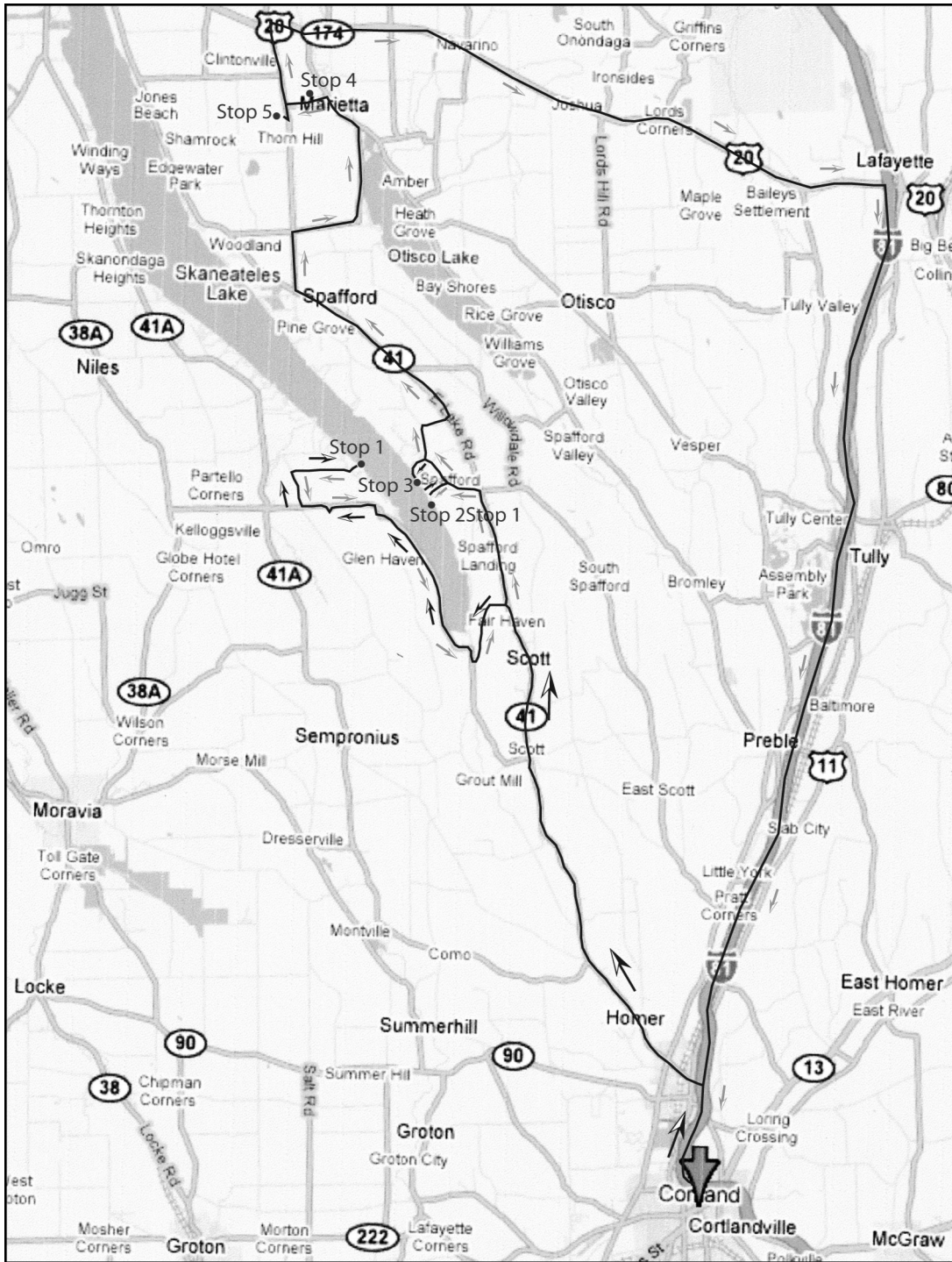
FIGURE 9--Map of study area showing field trip route; numbers correspond to stops on field trip, indicated in road log.

ROAD LOG FOR FIELD TRIP B-4

RECONSTRUCTING A MIDDLE DEVONIAN SUBMARINE ESCARPMENT: CORALS, PALEOSLOPES, AND DISCONTINUITIES WITHIN THE OTISCO SHALE, EASTERN FINGER LAKES REGION

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.00	Jct. Rte. 13 and I-81N
0.4-0.8	0.4-0.8	Outcrops of Ithaca Fm.
1.15	0.7	Homer Exit off I-81N
2.2	1.05	Jct. I-81 and Rte. 11 & 41 – take a left at end of exit loop
2.6	0.4	Jct. Rte. 90, stay straight
2.9	0.6	Jct. Rte. 41 and 11, bear left at fork on Rte. 41
3.5	0.6	Jct. Rte. 281, stay on Rte. 41
4.8	1.3	View of bedded kame gravels to left
5.0	0.2	Jct. Rte. 41A
7.3	2.3	View of bedded kame(?) gravels to left
11.85	4.55	Jct. Ripley Hill Rd., stay on Rte. 41
13.3	1.45	Onondaga Co. line. Turn LEFT onto Vincent Hill Rd.
13.7	0.5	High vista of Skaneateles Lake ahead
14.4	0.7	Old quarry to left in Tully Limestone
14.55	0.15	Contact of Fall Brook and Fisher Gully submembers of the Windom Member of the Moscow Fm. in ditch
14.8	0.25	T-Jct. at Fairhaven Rd. Turn RIGHT (north)
15.0	0.2	Glen Haven Road. Turn LEFT.Town of Scott
15.4	0.4	Cayuga Co. line
15.7	0.3	View of Skaneateles Lake
15.85	0.15	Glenhaven
16.2	0.35	Outcrop of Ivy Point Member siltstone
16.65	0.45	Cross gully. View of Spafford Landing cliffs to east across lake
17.2	0.55	Outcrop of Ivy Point Member siltstone
17.55	0.35	Contact between the Owasco Member of the Ludlowville Fm. and the Tichenor Member of the Moscow Fm.
17.7	0.25	Roadside shale pit in lower Windom Shale rich in *Ambocoelia* roadside shale pit in middle Windom Shale; Bay View bed rich in brachiopods and small corals is exposed in floor adjacent to road; higher banks are in the Bear Swamp submember of the Windom.
18.0	0.3	Three Mile Point ravine; exposures of middle and upper Windom to west of road
19.2	1.2	Upper end of Glen Cove ravine

19.6	0.4	Turn RIGHT onto Carver Road.
20.3	0.7	Glen Cove Road. Turn RIGHT and proceed slowly to bottom of hill and marina at Skaneateles Lake shore
21.7	1.4	Sevy's Boatyard; park vehicles.

OPTIONAL STOP 1. GLEN COVE CREEK

If water levels in the creek are not too high we may proceed on foot to the mouth of Glen Cove Creek behind the marina. Beds exposed near the creek mouth are in the lower Otisco slightly above Moonshine Falls phosphate bed; these shales are rich in the small brachiopod *Arcuaminetes scitulus* and carry small concretions and pyritic nodules. Upstream a few hundred feet the Stag Horn Coral horizon is exposed near the base of a small falls. The bed here yields primarily *Athyris* and other brachiopods and small corals. Occasional corroded fragments of larger rugose corals may be obtained.

Hiking upstream and climbing the relatively gentle small waterfall provides access the Glen Cove shell bed (type section). Here the bed is rich in *Athyris*, the small rugosan *Stereolasma*, the bivalve *Cypricardella* and *Modiomorpha* among others. These exposures provide a view of relatively fossiliferous down ramp facies of Otisco shelly horizons.

NOTE: if weather permits we will take a boat trip from Sevy's Marina across Skaneateles Lake to the vicinity of Stag Horn Point and Jenney Point; see descriptions under driving directions for STOPS 1 and 2. If so, we will be on the boat for approximately an hour.

23.3	1.6	Return to Carver Rd., Turn LEFT
24.0	0.7	Turn RIGHT onto Glen Haven Rd.
24.6	0.6	Cross upper end of Glen Cove ravine
25.75	1.15	Cross Three Mile Point ravine
26.15	0.4	Large roadside quarry in lower Windom Shale rich in *Ambocoelia*
26.25	0.2	Contact between the Owasco Member of the Ludlowville Fm. and the Tichenor Member of the Moscow Formation
28.8	2.55	Cortland County Line
29.15	0.35	Fair Haven Rd., turn RIGHT
29.35	0.2	Vincent Hill Rd., turn LEFT
29.6	0.25	Contact of Fall Brook and Fisher Gully submembers of the Windom in ditch
29.75	0.15	Old quarry to right in Tully Limestone
30.9	1.15	Jct. Rte. 41, turn LEFT; Onondaga Co. line
33.2	2.3	Jct. Bacon Hill-Cold Brook Rd., turn LEFT on Bacon Hill Rd.
33.5	0.3	Turn RIGHT on Morris Run Rd. (Road to Stag Horn Point)
34.05	0.55	Holzworth Rd., stay RIGHT
34.5	0.45	Park in pull-off to left at bend; remainder of road is too steep for most non-four wheel drive vehicles; so proceed down to shore of Skaneateles Lake (~0.25 mi.), passing by exposures of upper Ivy Point Member, Spafford, and Portland Point members in Chase Point Ravine; note exposures of Otisco Shale on steep road down to shore; past the mouth of Barber Creek proceed past the first cottage (permission for access from homeowners is required); bear left

(southeast) and onto bench formed by Stag Horn siltstone. (Alternatively, this section may be accessed by boat).

STOP 2. STAG HORN POINT

Here the upper beds of the siltstone are exposed and show excellent spreiten of *Zoophycos*; occasional phosphatic nodules and corroded corals are present in the top of the silt. The bench is abruptly overlain by about 1.5 m of the Stag Horn Point coral biostrome; densely packed solitary rugosans, including *Heliophyllum*, *Cystiphylloides*, and *Siphonophrentis*, the latter up to about 50 cm long. Corals range from corroded flattened fragments to complete and in situ uncorroded skeletons; they are packed in a mudstone matrix; upper part of the bank shows normal middle Otisco Shale. The coral biostrome continues to the south for nearly a kilometer before dipping below lake level.

34.9	0.4	Jct. Bacon Hill Rd., turn LEFT
35.4	0.5	Booth Rd., turn LEFT (Road to Jenney Point)
35.5	0.1	Fork, stay LEFT; proceed cautiously (steep road)
36.1	0.6	Jenney Point, find parking (permission for access from homeowners is required); proceed, if possible, southwestward along shoreline. (Alternatively, this section may be accessed by boat).

STOP 3. COVES SE OF JENNEY POINT: EDGE OF STAG HORN PLATFORM

The first cove southeast of Jenney Point displays the lower Otisco Shale at about the level of the Staghorn biostrome. Depending upon lake level it may be possible to walk along the shore to the next point and into the next cove to examine features of the transition of the platform edge into ramp sediments. Corals are present here in lenticular, wedge-shaped masses and layers that abut the truncated edge of the siltstone platform. Adjacent to the first cottage a large mass of rugose coral corals dips to the south about 8-10° along an apparent erosional surface. At the south end of this cove, the siltstone platform remains partially truncated; about 30m further to the north, near the middle of the cove, the pyrite- cored concretions, which normally lie below the platform are directly overlain by coral rubble. In places, the coral rubble thickens to about one meter as the truncation surface runs downward from the pyritic concretions to nearly the level of Skaneateles Lake.

The next cove to the south between an unnamed point and Chase Point, north of Stag Horn Point exposes a similarly intriguing series of beds including a series of thin graded beds of siltstone, locally with corals in their bases that appear to be storm event beds of silt eroded from the truncated platform margin (see text and Figures 6-8 for details). Together, these two cove sections provide a view of an eroded platform edge that was shedding coral debris and silt layers into the surrounding ramp area.

Return to vehicles and reverse directions back to Bacon Hill Road

36.85	Jct. Bacon Hill Rd., turn LEFT
38.2	Jct. Rte. 41, turn LEFT
40.95	Borodino mud mound (micritic bioherm) in Tully Formation
41.7	Upper Windom Shale outcrops
42.2	Village of Borodino, turn RIGHT onto Rte. 174 North
43.3	Turn RIGHT, follow Rte. 174 North
44.3	Chenango Siltstone outcrop

44.6	0.3	Sharp bend, view of Otisco Lake
46.1	1.5	Road follows along Otisco Lake shore
46.5	0.4	North end of Otisco Lake
46.6	0.1	Turn LEFT following Rte. 174 North; Village of Marietta
47.35	0.75	Coon Hill Rd., turn LEFT
47.7–47.95	0.35-06	Disembark near base of roadcut and walk upward to top along the road; vehicles will park at upper end of exposure

STOP 4. ROAD CUT ALONG COON HILL ROAD, MARIETTA, NY

This long cut exposes the complete transition from the upper Skaneateles Formation to the base of the Ludlowville Formation. Lower portions of the roadcut expose black, silty Butternut Shale. These shales yield a low diversity assemblage of *Eumetabolotoechia*. Its basal contact is exposed in a stream on the south side of the road. The upper 10 meters of the unit contains thin 1 to 2 cm tabular, laminated siltstone beds. The Butternut Shale is abruptly overlain by the Peppermill Gulf Bed, a 30 cm interval of medium gray calcareous and highly fossiliferous mudstone at the base of the Chenango Member.

From this point the Chenango Member, ~10 meters thick, shows an upward-coarsening succession from shale and silty mudstone to *Zoophycos*-churned siltstone. Fossils are generally scattered but the brachiopod *Tropidoleptus* is common at many levels and an *in situ* cluster of the large brachiopod *Spinocyrtia* was found about midway through the succession.

Near the upper end of the roadcut the upper Chenango weathers as a distinct bench. Large favositid and rugose corals present in the upper calcareous sands and overlying thin, silty limestones. These beds signify the basal transgressive systems tract of the Ludlowville Formation. The Halls Landing beds (upper Centerfield Member), mudstones with thin shell rich horizons form the uppermost unit of the cut.

Uppermost exposures in the adjacent stream bank opposite the upper part of Coon Hill Road show the lower Otisco Shale. The Stag Horn siltstone bench is absent at this section and its position is only suggested by beds of silty shale rich in *Athyris* brachiopods.

Return to vehicles and proceed west on Coon Hill Road to junction with Rose Hill Road

48.2	0.25	Bear LEFT at Y onto Rose Hill Rd.
48.4	0.2	Turn RIGHT onto hidden dirt drive into quarry in village of Rose Hill; drive in to parking area at east side of shale pit
48.5	0.1	

OPTIONAL STOP 5. SHALE PIT AT ROSE HILL

This shale pit exposes richly fossiliferous beds of the middle-upper portion of the Otisco Member. Shales contain diverse brachiopods, bivalves and other fossils and a horizon with scattered rugose corals toward the upper part of the pit appears to record the Joshua Coral Biostrome level, although it is poorly developed compared with exposures to the east on Otisco Lake. This section provides an excellent opportunity to collect fossils from the Otisco beds.

48.6	0.1	Turn LEFT on Rose Hill Rd.
50.4	1.8	Jct. U.S. Rte. 20, turn RIGHT; Village of Clintonville; red clay banks
55.9	5.4	Hogsback Rd. and Peppermill Gulf, stay on Rte. 20; type locality of

		Peppermill Gulf Bed at base of Chenango Siltstone
59.0	3.1	Lords Corners, Jct. Rte. 80, stay on Rte. 20
59.1	0.1	Case Hill Rd., stay on Rte. 20
60.6	0.7	Tully Valley, Village of Cardiff, site of famous Cardiff Giant hoax
63.1	2.5	McDonald's on left, turn RIGHT onto ramp to I-81 South, large cut of Butternut Member of the Skaneateles Fm.
69.3	6.2	Windom Shale outcrops
69.7	0.4	Exit for Tully, stay on I-81 South, note large exposure of uppermost Windom Shale behind Best Western Inn to east of highway
71.1	1.4	Cortland Co. line
81.4 – 82.4	10.3-11.3	Roadcuts in Ithaca Formation
84.4	2.0	Exit #11 to Rte. 13 at Cortland

END OF FIELD TRIP

Workshop B-5

CREATING A VIRTUAL FIELDWORK EXPERIENCE AT A FOSSIL-RICH QUARRY

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INTRODUCTION

What is a Virtual Fieldwork Experience?

Fieldtrips have long been an essential part of a hands-on Earth science course, and some teachers have long lamented increasing restrictions on getting students into the field. These trips have ranged from active problem-solving expeditions to less active show-and-tell. Technology in the 1990's began making creation of "virtual" fieldtrips, and sharing them electronically, relatively straightforward for anyone with basic computer skills and some time. Although hundreds of virtual field trips are now available over the web (some for the Northeast, for example, are at the PRI website at www.priweb.org/ed/earthtrips/northeast/northeast.htm), many of these are not much different than a slide show of someone's trip (though these too have their place). The term "virtual field*work*" is intended to have a slightly different connotation: the emphasis is on doing rather than seeing (Duggan-Haas and Ross, 2007).

People who do fieldwork are usually exploring a new place, collecting data for research, or both. "Virtual fieldwork experiences" provide opportunities for your students to experience problem solving in nature when you can't actually get them there. It provides an opportunity to connect your lessons with features your students might encounter in their own lives, and to apply what they've been learning to the way the world really looks outside the classroom walls. Ideally virtual fieldwork has some of the same active experiences as those of a geologist investigating a place for the first time. This suggests opportunities to explore and discover, to ask questions, and to look for observations relevant to answering a question about a locality. While "show and tell" at a locality can be useful in some contexts, "virtual fieldwork" is less about showing features to students and more about student activity.

Consider what we might hope students can do at a real field site several years after they have left our classrooms. Try to imagine a scenario by which they employ knowledge and understanding from your classroom in a way that's useful to themselves and others. The simplest situation perhaps is that they can see an Earth science phenomenon, in their own backyard, in their neighborhood, in a park, at a lake, at a construction site and know what meaningful questions they might ask and what observations they can make to answer them. The most basic of questions is "Why does this place look the way it does?" This question incorporates both the questions "What is the history of Earth processes that happened here?" and "What is going on here now: today and over the course of the year?" One of us (RMR) gives interpretive walks at local parks and elsewhere around Central NY. Many of the places look quite different one from the next, but the series of questions asked during the walks are almost the same, and that is the point: understanding of a set of concepts, and knowing what questions to ask in each place, are transferable across broad regions.

But how does one bring the field experience to the classroom on a regular basis, to help students with such real-world applications? This is, of course, a challenge, because transporting the complexity of even a small outcrop, and the myriad observations that might be made there, from the field to the classroom seems overwhelming. But with the availability of digital cameras, software such as PowerPoint, and greater availability of computers and projectors in classrooms, it has never been easier to present large numbers of photos, together with on-line maps, data, and any specimens brought back from the field. Virtual fieldwork experiences take some effort to do well, and never can replace the real thing, but the opportunity for creating something useful that reflects real environments has never been greater.

The beauty of creating a virtual fieldtrip thoroughly for one particular locality is that this locality can be revisited by students numerous times over the course of the year. The Earth is a system, after all, and at many places one may be able to study a wide range of phenomena from sedimentary rocks to meteorology to weathering to stream flow. Visiting a site numerous times helps students to discover how different Earth phenomena interact with each other, "multi-purposes" the effort one puts into creating a virtual fieldwork experience, and helps one to concentrate on getting to know just one site very well. In principle, colleagues creating virtual fieldwork experiences can share materials, or even be guest experts in each other's classrooms.

Research has been done into what characteristics of virtual experiences (which are used in a wide variety of educational contexts) makes them most highly effective as educational tools, but relatively little of that research has been done on Earth science virtual trips, or how they tie into teaching inquiry- and systems-thinking in Earth science. There is substantial room not only to create experiences that will complement what you are already doing in your classroom, but to create innovative approaches that might benefit your colleagues in NY State and beyond.

Application of a virtual fieldwork experience to Rose Hill Quarry, a fossil-rich Devonian shale

We will work together to consider how to bring the Rose Hill Quarry back to your classroom. Before determining what kinds of pictures and objects you want to choose to represent the site, you will need to think about the lessons to which you can apply study of Rose Hill Quarry. The most obvious units would be sedimentary rocks, fossils, and geologic history, but several other units could also be approached. Let's assume that you are introducing sedimentary rocks. You may be concerned with introducing understanding what sedimentary rocks are, how they form, and how to know them when you see them. Keeping in mind your long-term goals, you will want students to know how to ask meaningful questions about them some years down the road, even after their memories of vocabulary and facts have begun to wane.

Duggan-Haas (2005) has created an excellent vignette (in the style of the National Science Education Standards [NAS 1996]) of how a virtual field trip experience might intersect with curriculum and teacher approach. His example just happens to be a small Devonian quarry, so it is quite relevant in some respects to Rose Hill Quarry. Although there are many ways to lead an open-ended inquiry or guided inquiry investigation of the site, some of the more general virtual field trip questions that would be associated with the question "Why does this place look the way it does?" include:

- What kind(s) of rock(s) are found in the area? How do you know? What environment did these rocks probably form in?
- Describe the arrangement and variety of rocks shown in the photographs.

- Tell a story of how these rocks may have formed referring back to the photographs and what you have determined about the rock sample(s).

- What has happened to this area to make it look the way it does today? (That is, what has happened to the area since the rocks formed?) Why do you think so (what is the evidence for your claim)?

- If you could go to the site, what else would you want to do to answer the above questions?

GENERAL GEOLOGIC AND PALEONTOLOGIC BACKGROUND

As part of the workshop, we will be visiting a fossiliferous outcrop in which part of the Middle Devonian Hamilton Group is well exposed. Although much of the information provided below applies to this particular succession of sedimentary rocks and its contained fossils, the general sorts background information for other localities is easily obtainable from available literature. More specific data will be collected in the field that will lead to paleoenvironmental and paleoecological conclusions in the laboratory portion of the workshop.

Paleogeographic and Geologic Setting

During the Middle Devonian, some 385 million years ago, much of what is now central and western New York was situated in the Appalachian Foreland Basin that extended roughly northeast—southwest from southern Quebec to the southwest, around present day Alabama, where it opened into the Iapetus Ocean (Fig. 1). Paleogeographic reconstructions place central New York in a low, sub-tropical latitude (30º—35º S) during Middle Devonian time (e.g., Blakey, 2005). The Appalachian basin was formed as a response to lithospheric flexure associated with tectonic loading of the Acadian Orogeny (Ettensohn, 1985, 1987) and was a site of significant accumulation of sediments shed from the eroding Acadian Mountains situated to the east. Entering through large river systems and their associated deltas, the siliciclastic sediments (mostly mud, silt and sand) were deposited on broad shelf areas later to become the shale siltstone, and sandstones that are exposed today throughout much of the region.

FIGURE 1—Paleogeographic map of Appalachian Basin Foreland Basin of eastern Laurentia (modified from Blakey, 2005).

The Hamilton fossil record represents one of the best-preserved, most ecological diverse well-studied marine faunas of the Paleozoic. The fossils of the Middle Devonian Hamilton Group of New York has been of considerable interest since the pioneering works of James Hall, who published between 1847 and 1894 a

beautiful set of 13 monographs largely devoted to the Hamilton fauna. Within the past 50 years, Hamilton fauna has contributed greatly to our understanding within a wide range of disciplines, including stratigraphy and sedimentology, evolutionary theory, paleoecology, and the quality of the fossil record itself. Although published studies on this fauna are too numerous to list here, some of the more accessible and overarching articles include Brett (1986), Landing and Brett (1991), and, more recently, Brett et al. (2007).

The Hamilton Group shales of central NY contain fossils typical of middle Paleozoic continental seas. The Hamilton faunas in New York include many thousands of bottom-dwelling species of brachiopods, bryozoans, trilobites, corals, crinoids, and molluscs including bivalves and gastropods. Distribution of this marine fauna within the marine rocks of central and western New York is controlled by many paleoenvironmental factors, but factors that were especially important include: (1), water depth, (2), dissolved oxygen in marine water, (3), the clarity/turbidity of bottom waters, and (4), the sedimentary substrate upon (or within) which the animals lived. For example, corals and crinoids seem to favor shallow-water settings removed from significant muddy/silty input, and certain brachiopods and bivalves are more common in near-shore turbid waters and muddy and silty substrates.

Description of Rose Hill Quarry

The Rose Hill Quarry is a relatively small, privately owned, shale quarry within the hamlet of Rose Hill, Onondaga County, New York State. The quarry sits approximately mid-way between Skaneateles and Otisco Lakes, and approximately 25 km southwest of the city of Syracuse (Fig. 2).
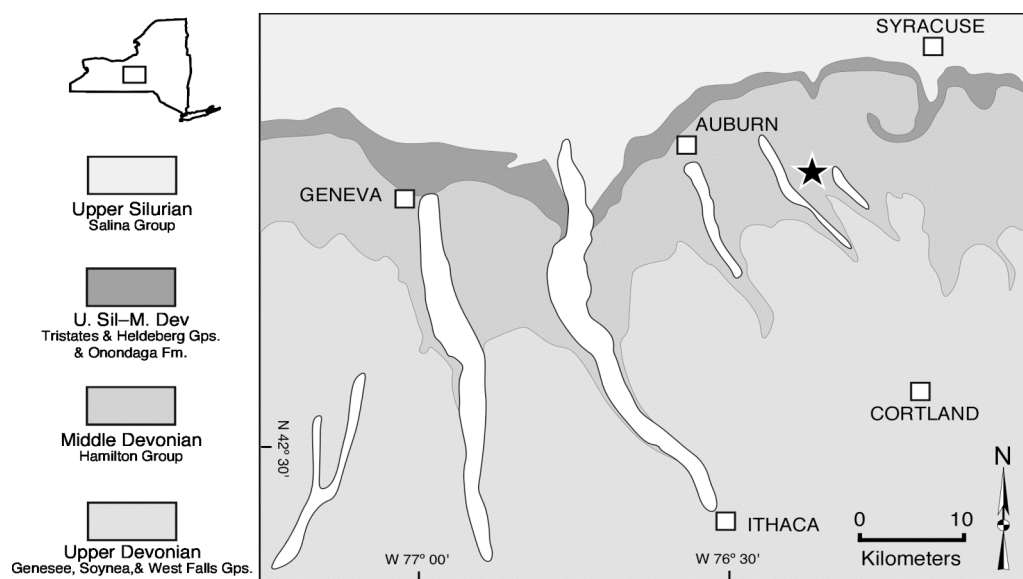

FIGURE 2—Generalized bedrock geology of the Finger Lakes Region, New York (modified from Rogers et al., 1990). Rose Hill locality indicated by star.

The quarry exposes approximately 10 meters of the upper part of the Ludlowville Formation and contains the upper part of the Otisco and lower part of the Ivy Point members (Figs. 3, 4). At Rose Hill, the lower 6.7 m of the outcrop belongs to the Otisco Member (first named by Smith, 1935) and is comprised of fissile dark gray mud shale that contains a diverse, abundant and well-preserved marine fauna. The fauna from these beds is largely contained in multiple shell beds and is dominated by spiriferid and strophomenid brachiopods with fewer numbers of bivalve mollusks, trilobites, bryozans, crinoids, and anthozoan corals. Elsewhere the Otisco Member is well known for its Staghorn coral bed in the lower part (see Trip B-2 this volume); the Rose Hill Quarry contains a few thin-beds of corals that likely correlate to the Joshua Coral Bed of the upper Otisco Member (see Oliver, 1951). Within the region, the Joshua Coral Bed can achieve thickness of up to 15 m (e.g., at Lords Hill; see Oliver, 1951), yet at the Rose Hill Quarry it is comprised of only 1-3 very thin (5 cm-thick) rugosan and tabulate coral-bearing horizons. Overlying the Otisco Member at Rose Hill is approximately 3.2 m

of rusty-brown silt-shale that is attributed to the Ivy Point Member (also first named by Smith, 1935). Preservation of fossils in the Ivy Point Member is largely moldic and, with the exception of corals, which are absent in this member, contains many of the taxa found in the underlying Otisco Member.

Sedimentological and paleontological evidence you will see at Rose Hill Quarry suggests that the site was situated on what was the middle to outer edge of the marine shelf that was moderately well-oxygenated and weakly influenced by wave and storm activity.
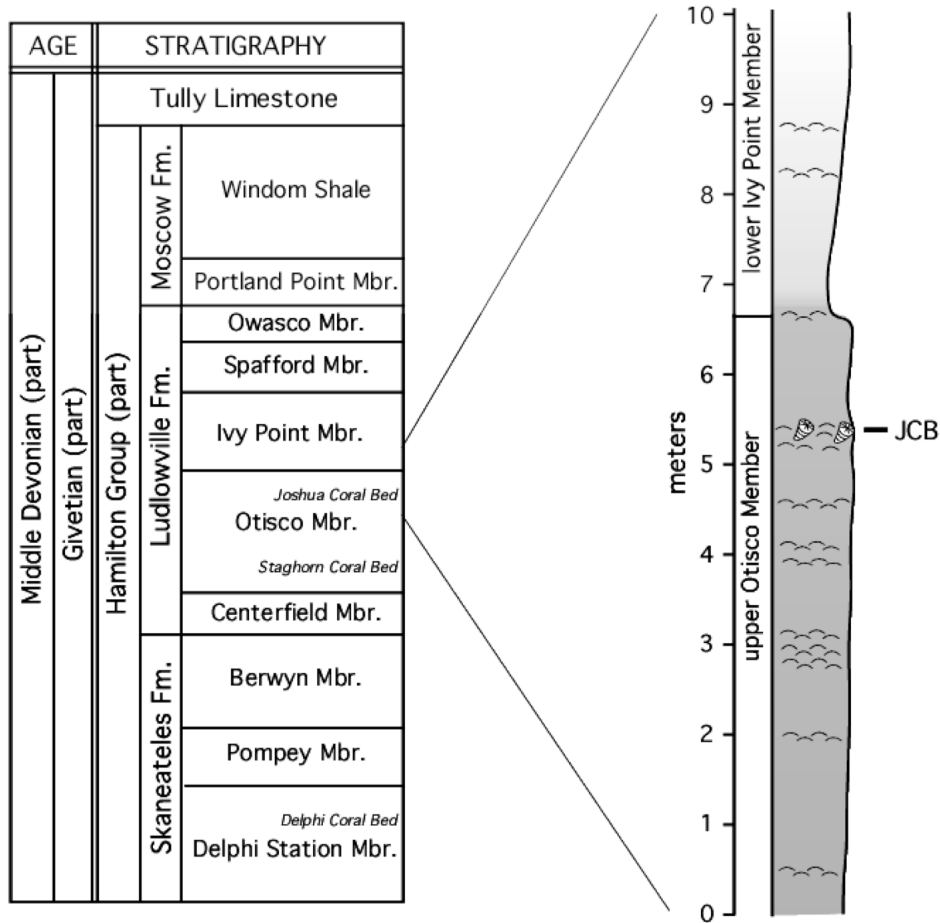
FIGURE 3—Generalized stratigraphy of part of the Hamilton Group (left) and stratigraphic section of the Rose Hill Quarry showing exposed parts of the Otisco and Ivy Point Members of the Ludlowville Formation (right). JCB = Joshua Coral bed.

A

Ivy Point Mbr.

Otisco Mbr.

B
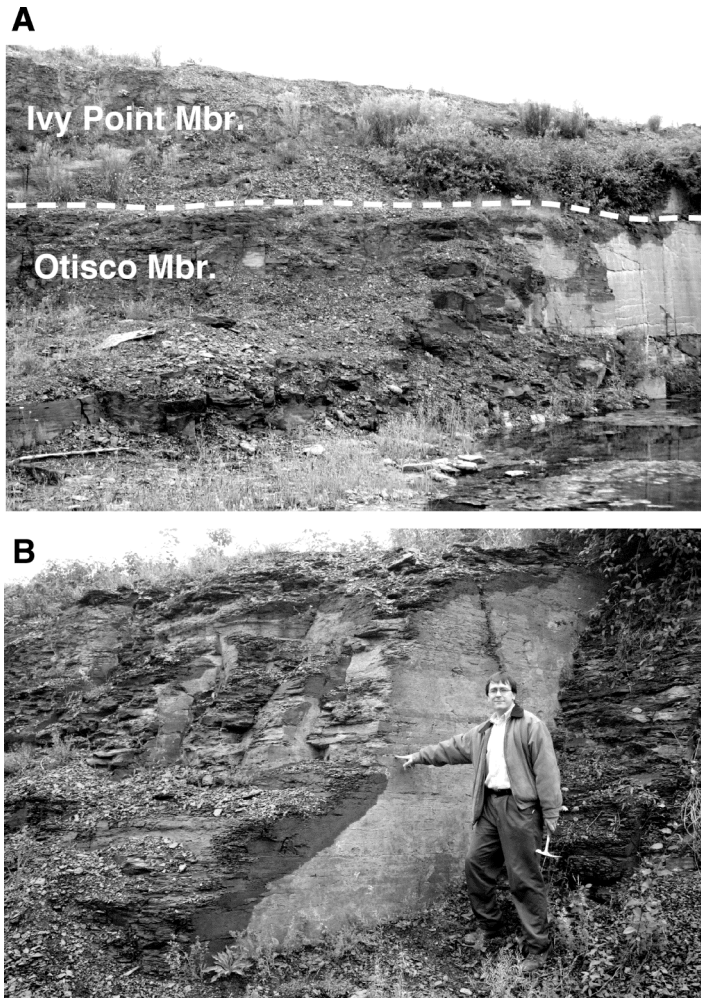
FIGURE 4—Outcrop photographs of the Rose Hill Quarry. A) approximate boundary between the Otisco Member and the Ivy Point Member. B) Showing position of the Joshua Coral bed. In the Otisco Member.

Basics in the field for creating your virtual field trip

You may have lots of field experience already, in which case you won't need reminders on what to bring and what to do at the outcrop. (If you want a fairly thorough guide, Compton [1985] is the classic.) In creating a virtual fieldtrip, however, you will need to be very systematic about the observations you make, the specimens you collect, and how all of your data is recorded, as you are doing it in proxy for your students. You should, in fact, include photographs and specimens not only of the best examples, but of a broad enough set to give a general feel for what's at the site.

Getting started

Though it may seem too intuitively obvious to note, the first step is to wander the site and get to know it yourself. (That is, resist the urge to start collecting fossils from one fossil-rich area, to the exclusion of other areas that may have fewer fossils but other kinds of geological information.) Start out by looking across the visible outcrop for signs of structures and for broad spatial changes in rock types -- both laterally and vertically. Now see if you can spot a way to move from lower in the rock section to higher and "get your nose to the rocks." Observe how rock types, fossils, and small sedimentary structures change vertically, which of course reflects how environments changed through time.

Photographs.—It will be helpful to keep careful track of the location and orientation of the pictures you take. This will help you remember how the pictures go together, and likewise help your students visualize the site. You might want to hold off on taking pictures of your site until you've explored the site and have a plan for how to communicate the site to your students visually. Use or make a map to indicate the place and direction at which your pictures are taken; in most places, and at the small scale you may be working, you will have to draw your own map, but you may be able to get topographic or aerial maps for other locations.

Your pictures should have a scale that makes it obvious the size of the geologic features in the photos. Common scales are people (especially yourself, since you will be interesting to your students), a hand pointing, or a standard rock hammer. For close-up photos traditional objects are quarters, Swiss Army knives, or rulers with a series of centimeter-sized blocks.

Notes, drawings, and Polaroid photos.—Now matter how many digital pictures you take, you are likely going to need to make some drawings, at the very least to record information about your pictures and the specimens. It will be helpful to use either a 'Rite in the Rain' notebook or to bring a large transparent plastic bag big enough to hold your notebook in case it rains. It can be helpful to use a Polaroid camera so that you can write notes directly onto photos. Polaroid photos are somewhat expensive and the quality isn't high, so these pictures do not replace others that you are taking for your virtual fieldtrip.

Measuring the section.—In research it is generally important to measure precisely where observations are made in a vertical section, thus it is necessary to measure any particular section from the bottom as a means of documentation. For teaching purposes it is possible to use basic tools such as meter sticks and long tape measures to make sure that your samples are collected at appropriate intervals and that your outcrop descriptions are accurate. Figure 3 shows a measured section for the outcrops we are visiting.

Collecting specimens and bulk samples.—You will undoubtedly want to collect specimens, especially fossils, at the Rose Hill Quarry site. Some of these will be useful simply as examples of the kinds of fossils that can be found in Central New York. You can also collect directly from the outcrop, either for individual samples or for "bulk samples," which are samples of a certain volume of rock or sediment (irrespective of the fossils that are in it). Bulk samples are what we will use to study changes in the fauna through the section. Collecting directly from the outcrop is often more work and less productive than collecting from the loose shale, if you are simply trying to create a large collection of diverse Middle Devonian marine fossils. But of course if you are trying to look for changes through time within the rock section you will need to collect from specific points along the outcrop.

Keeping locality data with your specimens is key to having a scientifically useful collection (that is a model to your students) and being able to use your specimens with your virtual fieldtrip. Even if you collect loose specimens from the piles of broken shale at the base of the cliffs, while you will *not* know precisely from which layers the fossils originated, you *will* know that they came from the Otisco or Ivy Point member of the Ludlowville Formation at Rose Hill Quarry. Since specimens so easily become separated from the label that identifies their locality, it is useful to number your specimens in some way. Putting numbers on specimens with a permanent marker right in the field, and identifying the position of the samples in the field on a diagram, is an excellent way to make sure locality documentation takes place.

Materials to bring into the field

To make a virtual fieldtrip you will need essentially the same equipment you might bring into the field anyway, except that you might bring more material for documentation. Here is a brief checklist.

- A digital camera of any kind. Film cameras will do if you are willing to pay for or make digital photos from the negatives, but with a digital camera you'll feel free to take any number of photos and you can use them on your computer immediately after being in the field.

- Digital video cameras are terrific if you have one and know well how to edit the results on your computer.

- A note book and pencils. 'Rite in the Rain' brand notebooks (www.riteintherain.com) are a good investment for long-term use, as they can get wet and still be used.

- Topographic map of the area, including a copy that you can write on.

For collecting specimens:

- A rock hammer is helpful (chisel-head is better for sedimentary rocks). While you can collect loose material that is useful at many outcrops, you need a rock hammer to collect fresh samples directly from the outcrop, for example if you are collecting from a specific point. Chisels made for rock are also useful, and butter knives are handy for splitting shale.

- Goggles or other eyewear protection

For collecting and bring specimens home you'll need:

- Zip-lock bags and small specimen boxes; tissues and paper towels are good for fragile specimens.

- Soda flats (shallow cardboard boxes) or inexpensive small plastic totes.

- Sharpie (permanent) markers to mark bags and rocks with specimen numbers.

A few other items that are useful in the field:

- Hand lens (about 10X).

- Pocket knife.

- Brunton compass if you have access to one (for measuring strikes and dip angles).

- Tape measure and/or measuring stick.

Special Projects

To provide more of a research experience for the class or for specific student projects, you may choose to go beyond an inquiry-based discussion of the site to have students collect data toward investigating a problem. Most paleontological research today starts with a specific question, for which the answer provides information about a key process of interest, such as how and when evolution and extinction occur, how rapid sea level changes during certain time intervals and what environmental impacts it has, and so on. To give a data-collection project meaning, it is important to frame it within one of these larger questions. Ideally, students might choose the question they wish to investigate, suggest a hypothesis that they will be testing, and figure out what sort of data to collect to test their hypothesis. It is possible to collect data for exploratory purposes such as documenting changes in faunas through the section and laterally along the section without a specific hypothesis in mind, but for scientific meaning one must still assume that there is a larger context in which the data might be of interest. This approach is typical primarily of those describing a rock section for the first time; such an empirical activity would have more meaning for your students if you make it clear ahead of time that you intend for your class to provide detailed descriptions of the site, toward a particularly purpose, that will be of eventual interest to scientists.

The significance of Central New York Fossils

One of the reasons New York Middle Devonian Hamilton Group shales are so important in paleontology and geology is that the geological sequences are thick, well-preserved, and fossil-rich, and therefore record in a relatively detailed way the changes in environments and faunas that occurred over a span of several million years in the Paleozoic Era. Although there are fossil marine organisms preserved all over the world in every interval of time, in many places there is not as much sedimentary rock per unit time interval (not as much sediment deposited), there are significant missing intervals of time (unconformities), or fossils were not well-preserved. Although in Central NY the shales and fossils within them look somewhat similar throughout the

Hamilton Group, there are some subtle changes that are important clues to changes in water depth, sedimentation rate, and other key variables. There are cycles of sea level change of at least 10 or 20 meters; it isn't clear what caused them, but some believe it may have been Milankovitch cycles (the same cycles that cause glacial-interglacial cycles of the Quaternary). On the whole, the taxonomic composition of the faunas stays fairly stable throughout the time interval, and the species are probably tracking (moving along with) their preferred environments. Species themselves seem to be rather stable morphologically. For most places and times represented by Hamilton Group rocks, changes through time have not been documented in a great amount of detail, primarily because there aren't enough people to do all the work (this turns out to be true for much of our study of nature). Then at the end of the interval, after deposition of the Tully Limestone, the faunas change dramatically, many species go extinct, and the marine faunas are never again quite the same.

You and your students can collect data in detail in one sample from the Rose Hill Quarry or elsewhere, looking in detail at the composition of skeletonized fauna at one point in the Devonian. If you and your colleagues take several samples and work collaboratively, you can look at what sort of changes occur through time at the Rose Hill Quarry outcrops. This is what we will begin to do together in the lab and is the subject of the next section.

Collecting samples to document faunal change

While it is fun to simply collect as many different kinds of fossils at a site as possible, unless they are simply doing reconnaissance work, paleontologists normally sample the fossils in some regular way in order to understand better what the fossils indicate about evolution and ecology. In order to do this, samples are taken from a specific layer or thickness of shale (say, 2 cm), and roughly the same volume of rock might be removed for each sample. These samples represent some limited amount of time -- perhaps hundreds or even thousands of years, but still much less than the whole stratigraghic section. Samples are typically taken at regular intervals, for example, one sample per meter, or every ten centimeters. It would be convenient to take samples every unit interval of time, such as every 10,000 years, but we can't currently date these sedimentary rocks so precisely, so we use thickness of accumulated sediment as our best approximation. We will collect samples like these from the Rose Hill Quarry section and compare the faunas at different points in time, and at roughly the same time but a several meters apart to look at spatial variability. Note that there is no published data from this site, and so whatever we learn will be new. The same could be said about similar projects applied to many other stratigraphic sections that you could study with your students.

The most straightforward way to describe changes in faunas is to count specimens of different taxonomic groups that are reasonably recognizable. Examples include brachiopods, bivalves, gastropods, trilobites, and so on. As a gross generality, different taxonomic groups tend to prefer different ecological conditions (though there can of course be *lots* of variation with taxonomic groups too), so simply by counting how many of each group are in your samples you can get some feel for how the faunas are changing. In addition to the "body" fossils, you can also have your students count "trace fossils" -- evidence of movement such as tracks and burrows that are rather common. We often don't know what kind of organism left them, yet the abundance and shapes of these trace fossils do correlate with environmental characteristics.

Paleontologists will typically do such an analysis using individual species rather than higher taxonomic groups, and your students can too if they have time to try it. We have tried this approach with 4th to 9th grade students and assessed resulting data quality (e.g., Harnik and Ross, 2003). You can start with a list of species that are likely to be found based on faunal lists from similar places. It is not strictly necessary to know species names, however, unless you wish to share the data with others, as the species names are not important for the students to learn; in fact, they can intimidating. Alternatively, you can ask the students to determine themselves what the species are and to give each species a letter or number. This may seem complicated, but it is what people tend to do on their own informally (though with some error) when they go fossil collecting. In this case you might create a two-step process, firstly coming up with a set of species that everyone agrees upon, and then counting how many of each of these species is in each sample. The students will need to make interpretations along the way, e.g., determining whether groups of specimens are from one variable species or more than one species, and making estimates of species identification when individual fossils are not well preserved.

Once you have collected your data, either higher taxonomic groups or species-level data, it is time to try to make sense of it. Firstly, do the samples differ in species present and in their relative abundance? Secondly, if they vary, is it in a way that seems to make sense ecologically? Are there any changes in the texture or color of

the rocks that correlate with changes in the fossils? Is your data reliable? – that is, are there any biases, either in the way the sample was preserved or in the way that it was collected and studied, that could account for the changes you see?

Collecting samples to document changes in shape and size of species

There have been numerous studies done of changes in shape and size of individual fossil species in particular places around the world, yet such studies have been made for a very tiny proportion of the world's species. Some places are better than others for collecting such data; again, places that have more samples per time interval and better-preserved samples are more likely to preserve detail of change through time. Some species are also better for evolutionary study than others, in the sense that some are more likely to be represented by large enough numbers of specimens for us to feel confident that we have some feeling for the real range of variability within the species. It may be better to look along a bedding plane to be confident that you are investigating something like a "population" of the species, but often this is not practical, and one must aggregate specimens from across a couple of centimeters of shale that might represent organisms accumulated over hundreds of years. The kinds of changes we are likely to see in the Hamilton Group shales might be as likely to be "ecophenotypic" as evolutionary. This means that the changes in size and shape are due to ecological changes that affect the growth of animals such that species locally and temporarily become smaller or even differently shaped. An example in our own lives are crops that don't grow as large in poor soils or arid summers, or oddly-shaped trees growing in marginal environments. It is important to note that species are not necessarily in constant evident morphological change; this does not mean that evolution does not happen, but is important data that helps us understand when and where evolutionary change does occur.

To do this kind of study you will need to make a decision about which species, if any, are abundant enough to follow through time. Then, rather than just looking visually for differences among samples, measure specific characteristics (such as length and width, which show size) or ratios of characteristics (say, length divided by width), which describe shape. Note that some species that are commonly found broken, such as trilobite exoskeletons and "winged" spirifirid brachiopods, can still be measured if you choose to make your measurement on just a commonly preserved piece of the broken shell.

ACKNOWLEDGMENTS

REFERENCES CITED

Blakey, R., 2005, North American paleogeographic maps, Middle Devonian (385 Ma.) Paleogeography and geologic evolution of North America, updated December 4, 2006, http://jan.ucc.nau.edu/~rbc/namD385.jpg, accessed August, 2007.

Brett, C.E., (ed.) 1986, Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State (Part I), New York State Museum Bulletin, v. 457, 156 p.

Brett, C.E., Bartholomew, A.J., and Baird, G.C., 2007, Biofacies recurrence in the Middle Devonian of New York State: An example with implications for evolutionary paleoecology, Palaios, v. 22, pp. 306-324.

Compton, Robert R. 1985 Geology in the Field John Wiley and Sons Ettensohn, F.R., 1985, Controls on the development of Catskill delta complex basin facies, pp. 85-115, *in* Woodrow, D.L. and Sevon, W.D. (eds.), The Catskill Delta, Geological Society of America Special Paper, v. 201.

Duggan-Haas, D., 2005, The virtual fieldtrip version of the quarry query. http://people.colgate.edu/ddugganhaas/WhatIsThisPlace/VFTVignette.htm. (accessed September 2007)

Duggan-Haas, D. and Ross, R.M., 2007, Fieldwork, real and virtual, as a driver of inquiry in education: Why does this place look the way it does? (abst.) Geological Society of America Program with Abstracts

Ettensohn, F.R., 1987, Rates of relative plate movement during the Acadian Orogeny based on spatial distribution of black shales, Journal of Geology, v. 95, pp. 582-582.

Harnik, P.G. and Ross, R.M., 2003, Assessing data accuracy when involving students in authentic paleontological research. Journal of Geoscience Education, v. 51(1), pp. 76-84.

Heckel, P.H. 1973, Nature, Origin, and the Significance of the Tully Limestone, Geological Society of America Special Paper 138, 244 p.

Landing, E., and Brett, C.E. (eds.), 1991, Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State (Part I), New York State Museum Bulletin, v. 469, 177 p..

NAP (National Academy of Sciences), 1996, National Science Education Standards. National Academies Press: Washington, DC

Oliver, W.A., 1951, Middle Devonian coral beds of central New York: American Journal of Science, v. 249, pp. 705-728.

Rogers, W.B., Isachsen, Y.I., Mock, T.D., and Nyahay, R.E., 1990. New York State Geological Highway Map. New York State Department of Education, Educational Leaflet 33, 1:1,000,000.

Smith, B.S. 1935, Geology and Mineral Resources of the Skaneateles Quadrangle., New York State Museum Bulletin 300, 120 p.

ROAD LOG FOR WORKSHOP B-5
CREATING A VIRTUAL FIELDWORK EXPERIENCE AT A FOSSIL-RICH QUARRY

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Leave SUNY Cortland parking lot and turn left on to Graham Avenue
<0.1	<0.1	At light, turn left (west) on to Groton Avenue.
0.9	0.9	Turn right on to Rt. 281 North (West Avenue).
4.4	3.5	Turn left on to Rt. 41 North. At approximately 21.1 [cumulative] miles, is the well known outcrop of Tully Limestone consisting of bedded crinoidal grainstones and carbonate mud mound (see Heckel, 1973). Continue north on Rt. 41 to the Village of Borodino.
22.3	17.9	Turn right (northeast) on to Rose Hill Road (Rt. 474) towards Marcellus.
23.5	1.2	Continue straight on Rose Hill Rd. as Rt. 474 branches to the right
25.8	2.3	Turn left onto unmarked private gravel road across from yellow house.
26.0	0.2	STOP 1. **Please note, this is a private quarry and permission from the owner should be granted before entering.**

Return to vehicles and return to SUNY Cortland for the remainder of the workshop.

END OF FIELD PORTION OF WORKSHOP