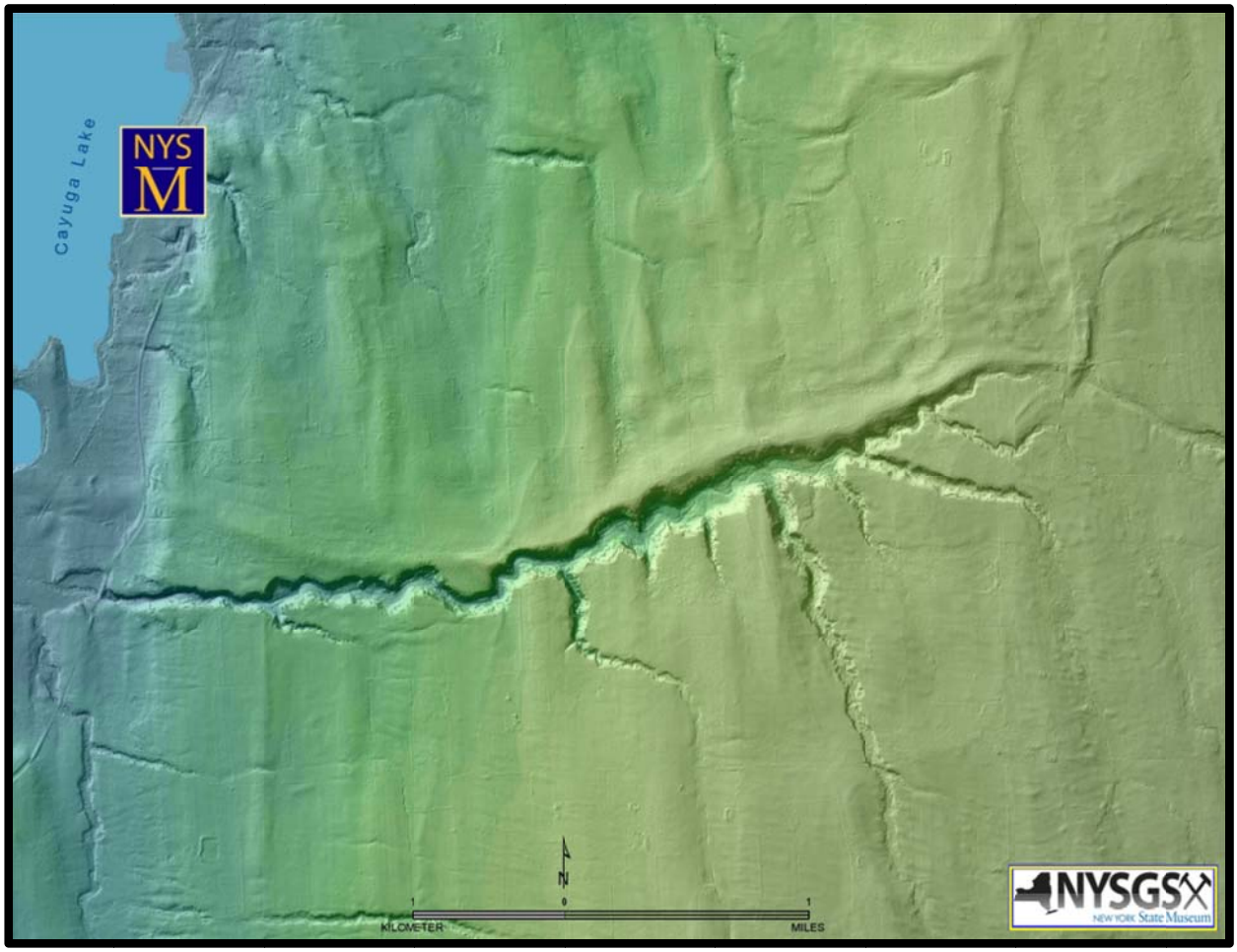
Glacial Geology of Cayuga County of the Eastern Finger Lakes: Lakes, Lore, and Landforms



77th Annual Reunion of the

North Eastern Friends of the Pleistocene Field Conference

June 7–8, 2014
Auburn, New York

Edited and Compiled by:
Andrew Kozlowski and Brandon Graham
The New York State Geological Survey
New York State Museum/New York State Education Department
Albany, NY 12230
http://www.nysm.nysed.gov/nysgs/

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77th Annual Reunion

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Trip Leaders:
Dr. Andrew Kozlowski – New York State Geological Survey
Dr. Brian Bird – New York State Geological Survey
Dr. Jonathan Lothrop – New York State Museum
Dr. Robert Feranec – New York State Museum
Dr. David Barclay – SUNY Cortland
Bill Kappel – USGS
Dr. Ed Evenson – Lehigh University
Nate Hopkins – Lehigh University
Brandon Graham – New York State Geological Survey
Andrew Clift – New York State Geological Survey

Contributions:
Dea Musa, Dr. Laura Sherrod– Kutztown University

Edited and Compiled by:
Andrew Kozlowski and Brandon Graham

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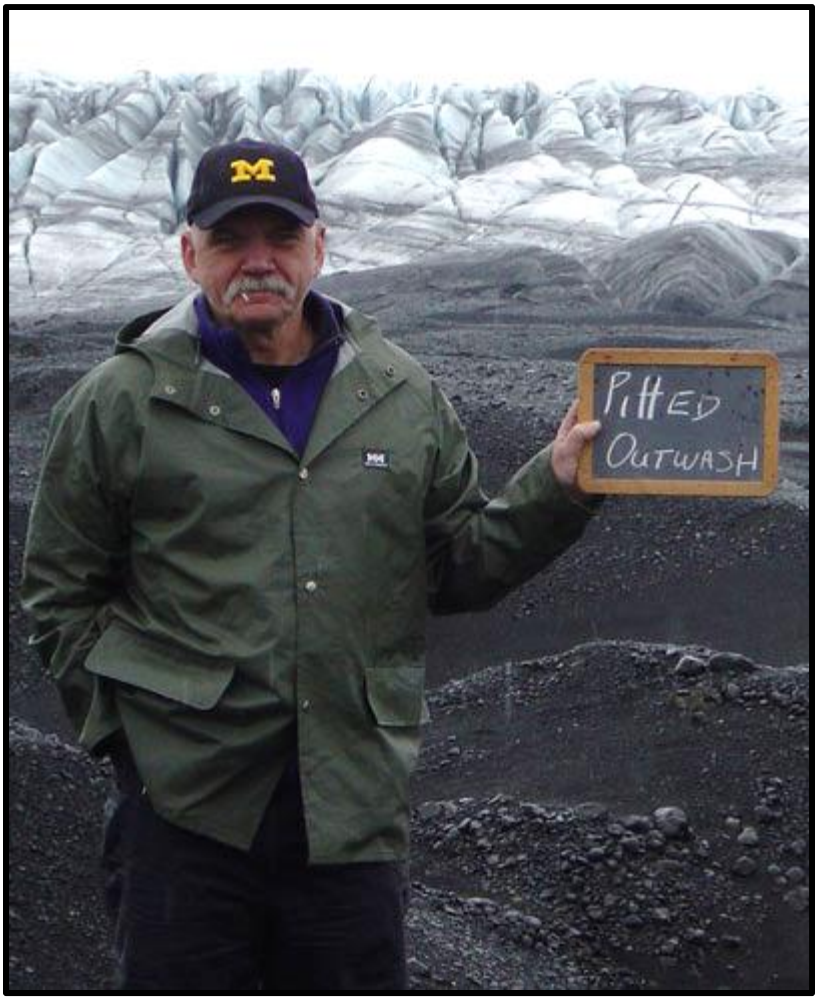


Dr. Edward B. Evenson has spent a lifetime exploring and researching glacial landscapes throughout the globe. Following Bachelor's and Master's degrees from the University of Wisconsin, Ed received his Ph.D. in Geology from the University of Michigan in 1972. After spending a few years working with the Exxon research laboratory, Ed came to Lehigh University in 1974. In the past forty years, his research has taken him to glaciated regions throughout the world, including the North American Rockies, the Andes of Argentina and Chile, Tierra Del Fuego, Alaska, Iceland, and Sweden. He has investigated subjects as wide-ranging as the subglacial hydrologic and sediment transport systems beneath modern glaciers, North American quaternary stratigraphy, supraglacial and periglacial debris systems, the chronology and extents of mountain glaciations, and the kinematics of till deformation and the origin of subglacial bedforms. Throughout this varied research career, he has maintained an active presence in the Finger Lakes region of New York conducting research on the area's glacial history and landforms and introducing countless undergraduate and graduate students to the region on annual field trips.

2014 NEFOP Site Addresses
Day 1
7:15 arrive at Holiday Inn Auburn
7:30 load bus
7:45-8:00 depart Holiday Inn

Stop 1 502 Locke Rd
Groton, NY 13073

Locke Delta Gravel Pit.

Stop 2 1996 Landon Rd
King Ferry, NY 13081

Poplar Ridge Pock Marks (Ice Walled Lake Plain?) Will simply drive past this point and NOT unload. This is a visual, show-n-tell of the local geology. It is requested that we go via state route 90 to the West entrance of Dills Road/Kings Ferry Rd.

Stop 3 Intersection of Dill Rd and Redmond Rd.
Union Springs, NY 13160

Great Gully part 1. Will unload at this point for main stop of day. This will be the lunch stop. 2 porta potties will be available as well as lunch and drink. We will then hike into Great Gully and tour the field sites. Once we return, we will reload the bus and travel to Stop 4 around the corner.

Stop 4 3994 Chase Rd
Union Springs, NY 13160

Great Gully part 2. Will unload at this point and be here for the second main stop of the day. We will hike into Great Gully again to tour the second half of the field sites. We will reload and continue to stop 5. The estimated time of reload will be 4:30-5:00

Stop5 4534 Cork St
Auburn, NY 13021

Dumond Wetlands. Will simply drive past this point and NOT unload. This is a visual, show-n-tell of the local geology.

Back to Holiday Inn Auburn

We estimate being back by 6:00 at the latest.

Day 2
7:15 arrive at Holiday Inn Auburn
7:30 load bus
7:45-8:00 depart Holiday Inn

Stop 1 W. Tyre Rd off of Highway 89
42.993722 N
076.820524 W

Flute Cross section. Arrive at the location and unload. It is a low travel road, but we will be along the road at a road cut. This should be a fairly short stop. Reload and proceed to the next stop.

Stop 2 11459 Lock Rd
Clyde, NY 14433

Lock 26. Arrive at the Erie Canal Lock and unload. We will talk about the glacial geology, the Paleo-Indian history of the region, as well as the groundwater and salinity. Reload the bus.

Stop 3 10550 Slayton Rd
Port Byron, NY 13140

Cato Drumlin. Last stop of the trip. Unload the bus and talk about the drumlin topography and recent study in the area. This being the last stop may take a longer time to finish due to discussions that normally ensue at the end of the trip. Our anticipated goal is to reload the bus by 12:00 to return to the Holiday Inn by 1:00 at the latest.

Back to Holiday Inn Auburn

Name	Organization
Participants	
Walter Aikman	Cayuga County College
Andy Bajc	Ontario Geological Survey
Michael Beardsley	NYS Archaeological Association
Steven Bill	Keene State College
Duane Braun	Bloomsburg Emeritus
Ruth Braun	
Julie Brigham-Grette	U Mass Amherst
Abigail Burt	Ontario Geological Survey
Lawrence Cathles	Cornell University
Rick Chormann	NH Geological Survey
G. Gordon Connally	
R Laurence Davis	U of New Haven
David De Simone	De Simone Geoscience Investigations
Robert Dineen	
Eugene Domack	U Florida
Priscilla Duskin	NYS DOT
John Esch	Michigan DEQ, Oil and Gas
Dru Germanowski	Lafayette College
Carol Griggs	Cornell University
Eric Hanson	Hanson Van Vleet, LLC
Gilbert Hanson	Stonybrook University
William Hecht	Cornell University
Lindi Higgins	AECOM
Carol Hildreth	CTH Enterprizes
Jason Hofgartner	Cornell University
Bryan Isacks	Cornell University
Samuel Kachuck	Cornell University
Dan Karig	Cornell University
Alan Kehew	Western Michigan University
Carl Koteff	USGS Emeritus
Becky Kranitz	Suny Plattsburgh
Lenka Lisa	Institute of Geo ASCR
Andrea Marich	Ontario Geological Survey
Shawn Moynihan	
Riley Mulligan	Ontario Geological Survey
John Rayburn	SUNY New Paltz

Name	Organization
Participants	
Jack Ridge	Tufts University
Hazen Russell	Geological Survey of Canada
David Sharpe	Geological Survey of Canada
Greg Sohrweide	NYS Archaeological Association
George Springston	Norwich University
Byron Stone	USGS
Janet Stone	USGS
Greg Taylor	
Woody Thompson	Maine Geological Survey
Daniel Tinkham	EGGI
Mike Wilson	SUNY Fredonia
Stephen Wright	University of Vermont
Richard Young	SUNY Geneseo

NYSM Trip Leaders and Staff	
Andrew Kozlowski	NYSGS
Brian Bird	NYSGS
Bob Feranec	NYSM
John Lothrop	NYSM
Meredith Younge	NYSM
Andrew Clift	NYSGS
Brandon Graham	NYSGS

Trip Leaders	
David Barclay	Suny Cortland
Ed Evenson	Lehigh University
Bill Kappel	USGS
Nate Hopkins	Lehigh University

NYSGS Interns	
Ashley Cirone	NYSGS
Bradley Sporleder	NYSGS
Lucas Oliveira	NYSGS
John Wiant	NYSGS Intern '12

INTRODUCTION

Greetings and welcome, the Geological Survey of the New York State Museum is delighted to be hosting the 2014 Northeast Friends of the Pleistocene (NEFOP) gathering, our 77th such meeting since the inception of this eminent non-organization. The Friends last visited Central New York (CNY) in 2005 in Onondaga County to the east with hosts Bill Kappel and Don Pair, prior to that a classic meeting in 1972 (the 36th) focused on sites within the Finger Lakes led by Art Bloom and John McAndrews . We are ecstatic to be back in the classic glacial terrains of the eastern Finger Lakes.

The Finger Lakes region of CNY has long been referenced as a classic landscape extensively modified by continental glaciation. The region drew early attention from major researchers in the emerging field of glacial geology starting more than 150 years ago (Vanuxem, 1842; Agassiz 1864). With the acceptance and ascendance of glacial theory, great debates between Diluvialists and Glacialists, and subsequently of equal intensity amongst "gougers" and the "sandpaperers", developed over the origins and processes associated with formation of the Finger Lakes landscape (Muller, 1965). From 1890-1920 a plethora of papers on the Finger Lakes and glacial geology associated with the region emerged. Notable among these are the 21 published works of R.S.,Tarr on glacial erosion (1893; 1894c; 1906), hanging valleys, deltas and waterfalls (1904b; 1905a and 1905c), moraines and glacial deposits (1904a; 1905e). By far the most prolific publisher was H.L. Fairchild with more than 100 papers, many specifically on Finger Lakes glacial phenomena such as: glacial lakes, water levels and meltwater drainage (1895; 1896; 1897; 1899; 1900; 1909; 1934), Drumlins (1900; 1907; 1911; 1929) and moraines (1900; 1923; 1932). Lesser known, Carney, published notable works on preglacial drainage (1904), glacial erosion (1907), pre-Wisconsin drift (1907b), meltwater routing and glacial deposits in the Moravia quadrangle (1909). Other notable contributions were made by D.F. Lincoln (1892; 1894; 1895a; 1895b), A.P. Brigham (1893), J.W. Spencer (1890; 1894; 1912), F.B. Taylor (1897; 1907; 1925), C.R. Dryer (1904), O.D. von Engeln (1918, 1926, 1931) and E.T. Apfel (1946).

The latter half of the 20th Century found E. H. Muller and his students at Syracuse University as the dominant workers in the region culminating in works such as (Muller & Prest, 1985; Muller & Cadwell, 1986; Muller & Calkin, 1993). Postulated reconstructions and dynamics of the Ontario Ice Dome lobe has been proposed by Ridky and Bindschadlier (1990) and multiple papers on the subsurface geology, stratigraphy, sequence of deglaciation and climatological events within of the Finger Lakes have been investigated by H.T. Mullins and his students (Mullins & Hinchey, 1989; Mullins et al., 1996;, Wellner et al., 1996; Wellner & Dwyer, 1996; Petruccione et al., 1996). The streamlined landscape of the northern edge of the Finger Lakes extending into the Ontario Basin has not gone unnoticed by more recent workers (Muller, 1974; White, 1985, Briner, 2007; Gentoso et al., 2012; Livingston et al,. 2013) and remains as much of interest today as it has in the past. Perhaps most elusive has been a better constraint on the origins and chronology of the Finger Lakes. Since the work of Maury (1908), speculation on the interglacial and glacial chronology has remained of interest to

researchers. Bloom's (1967) re-discovery of Fernbank and subsequent follow up studies by Karrow and others (2004, 2009) provide tantalizing clues to an older and richer glacial history, but a history that is not necessarily straightforward (Karig & Miller, 2013).

Our work and progress over the last six years has been possible largely in conjunction with the support of the United States Geological Survey's National Cooperative Geological Mapping Program, STATEMAP and the Great Lakes Geological Mapping Coalition elements. The results presented here started with work on mapping nine quadrangle area associated with the Montezuma Wetland Complex north of the Cayuga Lake Basin. That project expanded and we are presently working on 1:24,000 scale geologic mapping the Surficial Geology of Cayuga County. Our work has focused on a three – dimensional approach that incorporates subsurface characterization in addition to classic mapping of sediment-landform relationships. Where possible we have incorporated exploratory drilling, near-surface geophysics and a strong focus on improving the resolution of proposed models of glacial chronology. A fundamental tool in our arsenal for mapping has been the incorporation of high resolution LiDAR terrain models to evaluate the glacial landscape in unprecedented detail.

Beyond the pragmatic justifications that additional spatial information of geologic resources (aggregates, groundwater, etc) and geologic hazards will surely result from detailed three-dimensional geologic mapping, reevaluating surficial geology in this region has the potential to yield substantially more. Part of the lure of working on the glacial landforms and deposits in the central Finger Lakes area is that much of the previous work describes landforms, lake levels and clearly pertains to meltwater routing in the region. Yet despite these well documented proglacial lake systems and spectacular glacial landforms the weakest link of regional correlation in the Great Lakes Region is the age control of ice marginal positions and glacial events in the CNY (Fullerton 1980; note 35, Fullerton 1986, Ridge 2003, Ridge 2004, Karig and Miller, 2013). With the advantage of LiDAR based elevation models, our ability to recognize newly identified ice marginal positions, spatial patterns and landform suites have greatly improved; thereby creating enhanced opportunities to test previous hypotheses suggested for the glacial geology in the northern Finger Lakes area, and perhaps develop a few new ones along the way.

Before we proceed any further I would like to take a moment and thank numerous folks for their dedication and hard work that have resulted in the data and topics discussed here on this trip. The staff in the mapping program I supervise include Brian Bird, Andrew Clift and Brandon Graham. They are a dedicated and talented group of geologists and all together an outstanding group to work with over the last few years. Our office support staff, Donna Jornov, Christine Ryan and Dr. John Hart Director of Research & Collections, have spent countless hours navigating bureaucracy on our behalf to make this a successful workshop. My colleagues Robert Feranec and Jonathon Lothrop have graciously shared their expertise and provided contributions and I thank them. Dr. Hong Wang of the Illinois Geological Survey completed radiocarbon analyses as did Woods Hole NOSAMS. Shannon Mahan of the USGS Geochronology center in Denver provided exceptional guidance and consultation for the

collection and processing of Optically Stimulated Luminescence (OSL) samples and is an awesome person to collaborate with. Brian Slater from the NYSM\NYSGS Oil & Gas section came out and collected ubber resolution Giga pan photos of our sites in order to archive the stratigraphic sections and deserves many thanks. Mr. John Wiant worked the first field season and logged many hours in the field and laboratory, and deserves recognition. Bill Kappel USGS, Laura Sherrod & Dea Musa from Kutztown University, Ed Evenson and Nate Hopkins from Lehigh University all provided sections or lead discussions and serve to make this a better field guide than we could on our own. Lastly, and especially we thank the landowners and those whom gave us permission to work and map, John & Cheryl Roemmelt, James Young, Todd Dumond, the Nature Conservancy, Vitale & Robinson Companies, NYDEC Howland Island Northern Montezuma Wildlife Management area, NY Canal Corporation, US Fish & Wildlife Montezuma Refuge, (Particularly Linda Ziemba, Bill Stewart and Tom Jasikoff), the USGS National Cooperative Mapping Program and our oldest and dear friend Ezra Brooks without whom life and mapping would be less tolerable. My thanks to all who are in attendance.

Sincerely,
Andrew Kozlowski
New York State Museum/New York State Geological Survey

Please note: Almost all stops and points of access we are making during this field meeting are on Private Land. Written permission is required for all parcels and many are closed during hunting season. Some sites will remain as active research projects for years and we need to maintain access and good relations. Please DO NOT return with or without students without securing permission from the land owners. Lastly, we provide information in this guide in keeping with the spirit of the Friends of the Pleistocene; our hope and intent is to share our current state of knowledge and invoke good conversation. We have several manuscripts in process and in review, and ask that use or reproduction of information, figures, maps, data and/or tables not be reproduced without written permission of the authors. The articles are on the way; please be respectful and cite those. If you feel the need to utilize data presented here in an urgent fashion, please contact us, when and where possible we will try to oblige all requests.

Geologic Setting

Andrew Clift

Bedrock

Cayuga County is situated in the Finger Lakes region of central New York and has a total area of 2,237.7 km^2 (864 mi^2), of which about 20% is covered by water. It is partitioned through the middle by an east-west trending boundary between two major physiographic provinces in New York: the Erie-Ontario Lowlands in the north, and the Allegheny Plateau to the south (Figure 1). The overall topography dips gently to the north, with a total change in relief of approximately 500 meters (~1,640 feet) over 80 kilometers (~52 miles). There are six major watersheds in the county: Lake Ontario, Oswego River, Seneca River, Cayuga Lake, Owasco Lake and Skaneateles Lake. Current and historic mining has been prevalent throughout, and typical commodities are sand and gravel, gypsum-anhydrite, iron and sodium.

The bedrock in Cayuga County consists of Paleozoic sedimentary rock packages dating from the Late Ordovician to Late Devonian (Figure 2). During these periods, central New York was an inland basin and inundated by a broad sea. As sea levels fluctuated over time due to eustatic and orogenic processes, various depositional environments existed- some of which harbored a diversity of Paleozoic biota- and resulted in the distinct bedrock stratigraphy present today (Table 1). The dominant lithologies throughout Cayuga County are marine carbonates such as limestone and dolomite, as well as shale and sandstone. Fossils are abundant, particularly in the Devonian strata, which host some of the best preserved Devonian invertebrate fossil specimens and records in the world. The oldest rocks are exposed in the northern-most portion of the county, and become progressively younger to the south. The Queenston Formation is the oldest rock formation present (Late Ordovician), located at the extreme north of the county along the south shore of Lake Ontario. Silurian rock packages, beginning with the Medina Group (Early Silurian), sit unconformably above the Queenston Formation. The youngest Silurian bedrock present is the Akron Dolostone (Late Silurian), which sits unconformably below the Helderberg Group (Early Devonian). The youngest bedrock is the West River Shale (Late Devonian) and is found along the hilltops of southern Cayuga County.

The depth to bedrock as measured from the surface elevation varies throughout Cayuga County. Borehole data provided by the New York State Department of Environmental Conservation (NYSDEC) was used to create a county-wide depth-to-bedrock map (Figure 3) by interpolating bedrock depth data in ArcMap (v. 10.1). This map shows that bedrock is generally deepest in areas overlain by drumlin clusters, and that the overall depth-to-bedrock increases northward. Similarly, a bedrock elevation map was interpolated from these data as well, and shows that the elevation of the bedrock generally compliments the surface elevation. This suggests that the bedrock is a dominant influence on the overall topography of this region, with the exception of local surficial features such as moraines and drumlins.

Table 1: Rock Packages in Cayuga County

Paleogeography	Age	Formation	Description	Environment	Fossils	Thickness
	Late Devonian	West River Shale	Shale, limestone	Deep basin	Low/moderate, bottom dwellers	Up to 45 m.
	Middle Devonian	Tully Fm.	Limestone	Deep basin	Bottom dwellers	Up to 300 m.
		Moscow Fm.	Shale, sandstone	Deep basin	Abundant	Up to 850 m.
		Ludlowville Fm.	Limestone, shale	Deep basin	Abundant	Up to 150 m.
		Skaneateles Fm.	Shale	Deep basin	Abundant	Up to 150 m.
		Marcellus Fm.	Shale	Deep basin	Low/moderate	Up to 270 m.
		Onondaga Fm.	Shale, limestone	Shallow sea	Abundant	Up to 150 m.
	Early Devonian	Oriskany Sandstone	Quartz sandstone	Shallow, near shore	Brachiopods	Up to 150 m.
		Helderberg Group	Limestone, dolostone	Shallow to moderate depth	Abundant	Up to 90 m.
	Late Silurian	Akron Dolostone	Dolostone, shale	Shallow shelf	Sparse	Up to 210 m.
		Camillus Shale	Shale, dolostone	Shallow shelf	Sparse	Up to 300 m.
		Vernon Formation	Silty shale, dolostone	Coastal plain, shallow shelf	Rare	Up to 4 m.
		Lockport Group	Limestone, dolostone	Shallow shelf	Corals, mollusks	Up to 60 m.
	Early Silurian	Medina Group	Sandstone, siltstone, shale and mudstone	Near-shore marine, delta, muddy offshore, braided streams, shore zone	Ostracodes, Arthrophycus, brachiopods, claims, snails, nautiloids, trilobites and crinoids	Up to 50 m.
	Late Ordovician	Queenston Formation	Red shale, red siltstone, red sandstone	Nonmarine to shallow marine; part of large delta	Few possible spores of oldest land plants	Up to 300 m.

North Arrow (approximate)
New York State outline
Cayuga County

* As has been documented throughout central New York
Paleogeographic maps modified from: Colorado Plateau Geosystems, Inc., 2014
Table modified from: Isachsen et al., 2000
Bedrock unit references from: Fisher et al., 1970

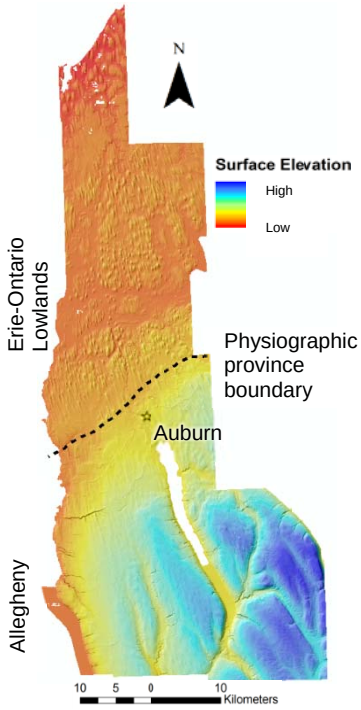


Figure 1: Physiographic map of Cayuga County showing the boundary between the Erie-Ontario Lowlands and the Allegheny Plateau provinces.

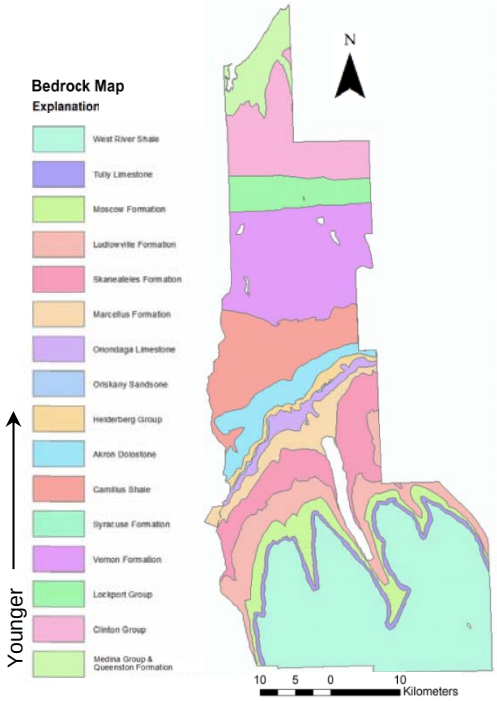


Figure 2: Bedrock geology of Cayuga County.

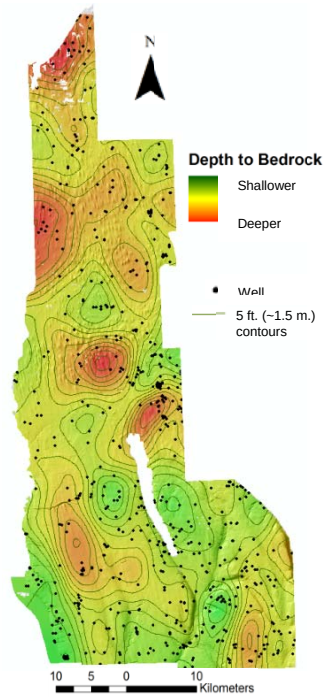


Figure 1: Depth to bedrock map of Cayuga County.

An Overview of Pleistocene Events in the Finger Lakes

Andrew Kozlowski

The following overview is a discussion of proposed evidentiary chronologic events in the eastern Finger Lakes and very closely follows the framework of information laid out by Muller & Calkin (1993). As we move through this outline we may also refer to chronologic events utilizing diachronic revisions for the Late Quaternary time-stratigraphic classification proposed by Karrow et al. (2000). This work is in harmony with that of Johnson et al. (1997) for major portions of the glaciated Midwest that adjoin the Great Lakes and although the classification of Karrow et al. (2000) (Figure 4) is centered in Ontario, the Ontario Lobe of the Laurentide Ice Sheet is equally responsible for the prodigious landscape observed in north central New York. Despite obvious gaps in glacial chronology in Central New York, as we move forward with our mapping program and participate in the Great Lakes Geologic Mapping Coalition inclusive of eight other states in the Great Lakes Region and our colleagues from Canada it makes sense for us to integrate and to utilize the most updated and applicable time-stratigraphic classification system. Our story revolves around the Cayuga Lake basin which is 60 km in length and has a maximum depth of erosion of 358 meters in the southern half of the basin, it extends to a depth of 242 meters below sea level. It represents the lowest of the Finger Lakes or the "axial depression" as described by Fairchild (1934) all ice advances and proglacial lakes likely impacted or were infectious to this basin and thus it's a record of such events should be present.

Pre-Wisconsin events in the Finger Lakes

Description by Maury (1908) of freshwater invertebrate fauna has many similarities to the Interglacial beds of the Don Valley near Toronto (Kerr-Lawson et al. 1992). McAndrews identified a pine – spruce interstadial pollen assemblage above a dominantly deciduous interglacial association in a 10 m thick sequence between glacial units. Variations exist between the pollen spectra only in that of the abrupt change from interglacial to interstadial assemblages. Radiocarbon ages greater than 54,000 and 52,000 have been reported on wood in the lower interglacial zone, an age greater than 50,000BP was from the top unit of the upper interstadial zone. The reexamination by Karrow et al. (1990, 2009) including a single thermoluminescence date, amino-acid analyses in corroboration with Maury's findings to believe the beds are correlative with the Sangamonian interglacial age. Further, the Sixmile Creek and Great Gully have yielded ages > 35,000 years old. Muller & Calkin (1993) interpret these as minimum ages because the valleys into which they are inset are wider and deeper than their postglacial analogs and are assumed to have been cut during the Sangamonian Interglacial (von Engeln 1929,1961: Muller 1957). Yet, a lack of sufficient age control has yielded other interpretations that imply perhaps the Finger Lakes were created more recently or at least all deposits contained within are attributed to the Late Wisconsin events (Mullins et al., 1996).

Middle Wisconsin – Finger Lakes –

In Millport, NY wood fragments retrieved from a depth of 46 meters from well cuttings passing through till and sand yielded a radiocarbon age of > 40,270 BP implying the age of the sediment bearing unit is

at least Middle Wisconsin in age. Exposures at Sixmile Creek near Ithaca described by Schmidt (1947) consist of varve sets separated by sand and gravel units there he described one varve set between two till units contains detrital wood with ages of 41,900 and > 35,000. Each varve set documents an interval when lake level was raised either because the rock threshold at the north end of the Cayuga trough stood higher than at present, or because the ice sheet blocked the north end of Cayuga trough.

More recently Karig & Miller (2013) documented additional Mid-Wisconsin ages in Sixmile Creek for lacustrine deposits and interpreted overlying tills to represent Mid-Wisconsin ice advance to the southern end of the Finger Lakes. This suggestion is in harmony with interpretations of Young & Burr (2006) for a Mid-Wisconsin ice advance into the region.

Late Wisconsin – Finger Lakes –

As stated by Fullerton (1986) numerous dates have been reported across NY but as of yet no singular synopsis about ice sheet advance has been postulated for CNY the exception being events in the Western Mohawk Valley (Muller et al., 1986; Ridge et al., 1991). That being said it is well documented that the during the Nissouri Phase the central Finger Lakes region was overridden by southward advancing ice as it made its way to north central Pennsylvania to the LGM around 24,000-20,000 ^{14}C B.P. (Ridge 2003). Little is known regarding early deglaciation in the south-central New York and few radiocarbon dates exist to help substantiate that chronology. After the Nissouri Advance, ice retreated northward into the Ontario Basin to some undefined northern latitude, but far enough north to allow free eastward drainage into the Mohawk River Valley associated with the Erie Phase (Erie Interstade) (Ridge, 1997). This period of free drainage was followed by a substantial ice advance back to the southern end of the Finger Lakes to the position of the Inner and Outer Valley Heads Moraines, and is likely equivalent to the Port Bruce phase, Salisbury or perhaps Hinckley-St. Johnsville advances of Ridge (2003 and reference therein). Generally the age of the Valley Heads Moraine has been suggested to be around 14,400 ^{14}C B.P. Retreat and advance of ice within the Finger Lakes has a well-known association of high elevation proglacial Lakes (Fairchild, 1909; Bird this volume). Chronologic data to constrain deglaciation in CNY after those proposed for the Valley Heads has remained elusive (Karig & Miller, 2013) (Figure 4-5).

This guidebook would be incomplete without reference to the efforts Mullins (1996 and references therein). Mullins et al., (1996) utilizing seismic reflection data in conjunction with drill core in the Finger Lakes identified\interpreted six consistent first order depositional sequences deposited during a brief interval during the Late Wisconsin (14.4-13.9 ka). On the basis of stratigraphic relationships, they attribute basal unit (sequence I) as being equivalent with coarse-grained water laid sediments comparable to those mapped as Valley Head deposits or Kame Moraines (Muller & Cadwell, 1986) at the southern end of the Lakes. They are specific to clarify that the presence of the Late Wisconsin infill does not preclude the possibility that erosion within the Finger Lakes has been the product of multiple glacial cycles, but only that any pre-late Wisconsin sediments that may have existed within the basin must have been removed prior to the deposition of Late Wisconsin sequences.

Seismic Sequence II and III are inferred to represent finer-grained outwash and subglacial meltwater deposits in deep proglacial lakes. Sequence IV is the subsequent glaciolacustrine unit deposited in high-standing proglacial lakes resulting from ice dam blockage of the northern end of the Lake Basins. The next sequential unit (Sequence V) is associated with a rapid (instantaneous) drop in base level as the Ontario Lobe ice retreated and initiated (or re-initiated) northward drainage. Radiocarbon dates collected by Wellner and Dwyer (1996) as a parallel study to Mullins et al. (1996) have been used to suggest that the corresponding drainage reversal and sediment input occurred by 13.9 ka and at this time glens, waterfalls and deltas were developing in response to this base level drop. The final and upper most sequence (sequence VI) represents organic-rich post glacial sediment (Figure 6).

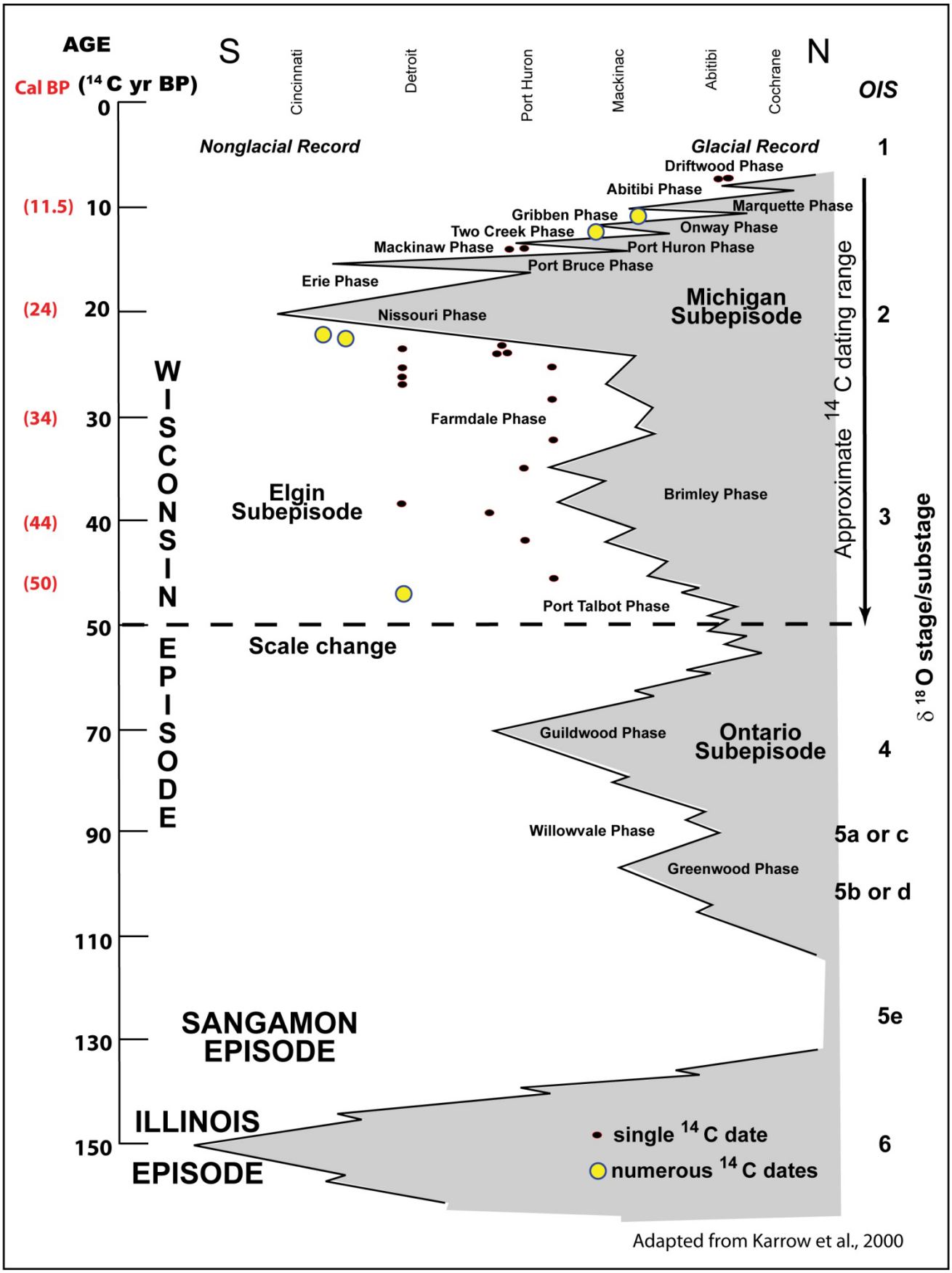


Figure 4. Time – distance graph developed for Eastern Great Lakes region

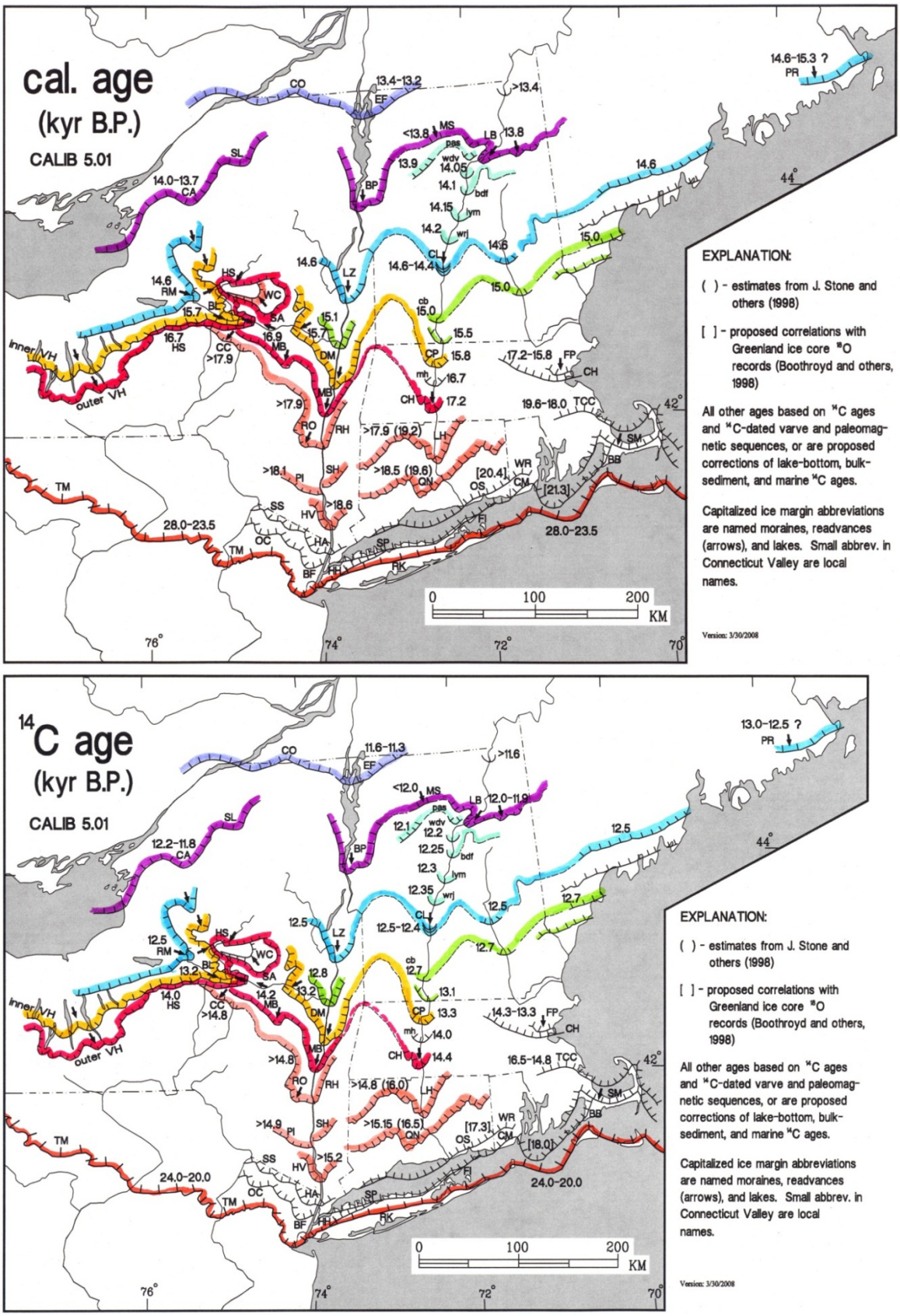
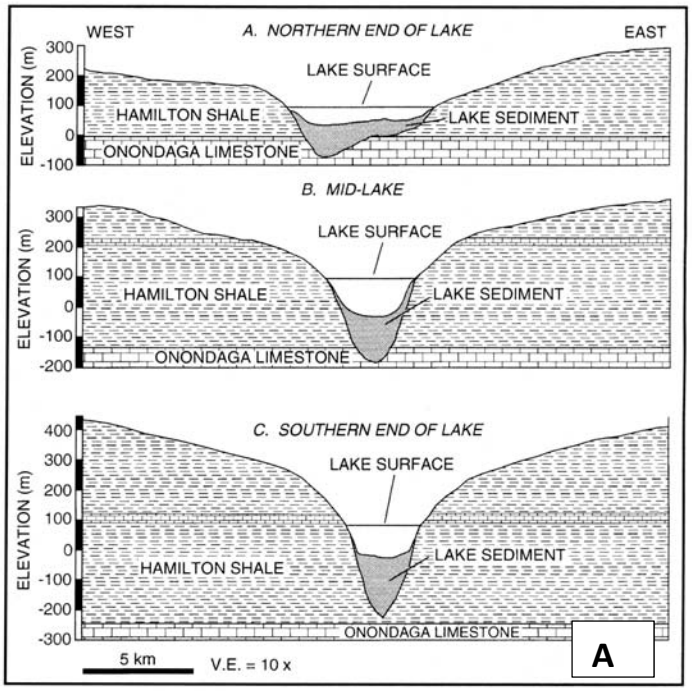
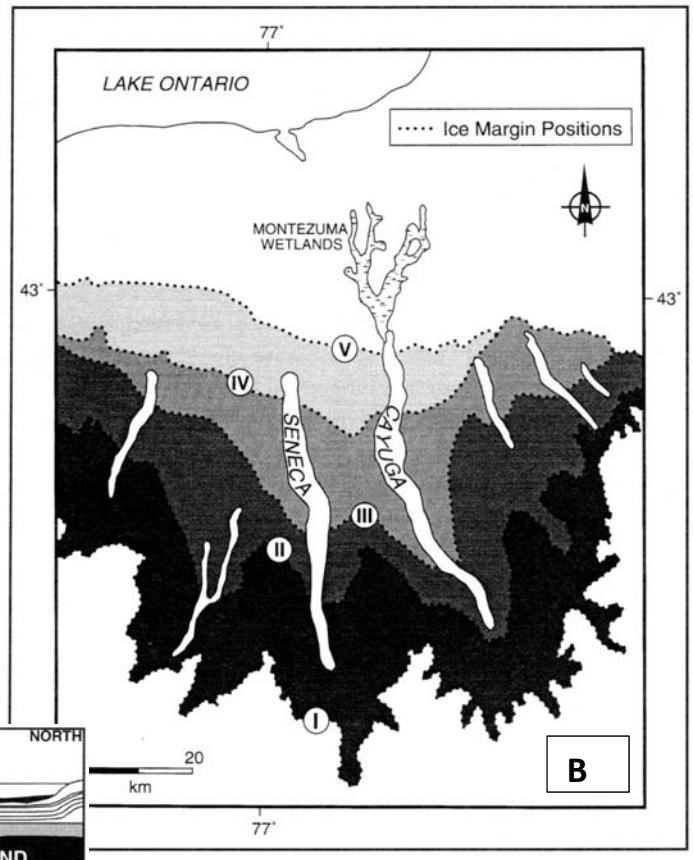


Figure 5. Calibrated (upper) and uncalibrated ^{14}C (lower) Ice Margin positions in the Northeast (from Ridge, 2004)





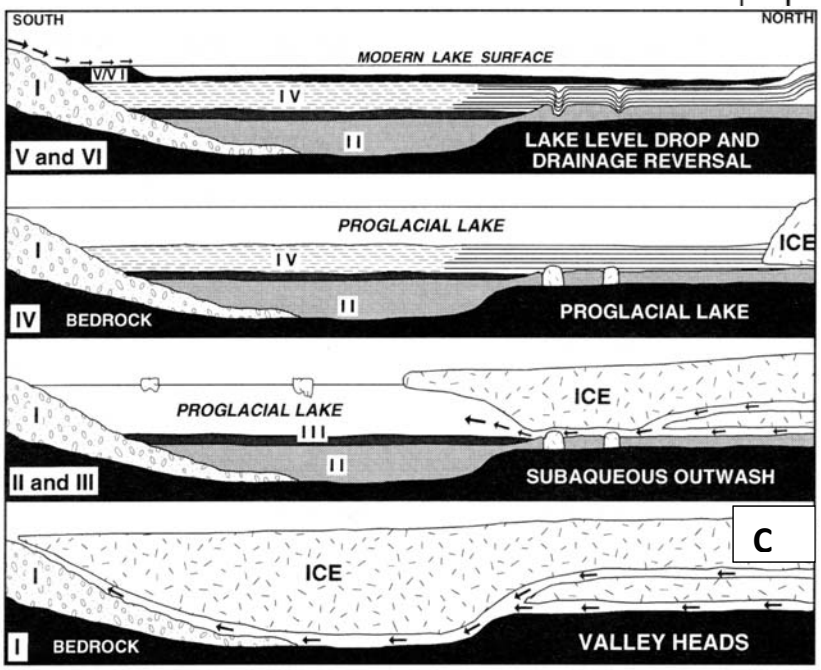


Figure 6. Summary Illustrations adapted from Mullins et al.,(1996) displaying schematic transverse profiles across Cayuga Lake (A) and aerial (B) and longitudinal reconstructions of sediment infill history of the Finger Lakes.

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Proglacial Lake Reconstruction in Cayuga County

The following is largely a recap of Bird, B. C., and Kozlowski, A. L., 2014, Using LiDAR to Reconstruct Glacial Lakes in Cayuga County, New York: Geological Society of America Abstracts with Programs. Vol. 46, No. 2, p.71

Brian Bird

Introduction

Just south of Cayuga County, the Valley Heads Moraine (*outer VH* in Figure 11: Deglaciation of Northeastern U.S. by Ridge, 2004) delineates an ice marginal position at the southern end of the Finger Lakes about 14.0 (16.9 cal) ka BP (Ridge, 2004). As ice retreated northward the isostatically depressed landscape and ice dammed outlets created a series of proglacial lakes. From the highest lake, glacial Lake Ithaca, to the lowest in the study area, glacial Lake Iroquois, nine lake levels were identified in Cayuga County. Close inspection of 2.0 meter LiDAR digital elevation model for Cayuga County reveals numerous geomorphic features related to former lake levels (deltas, strandlines, and wave eroded topography) along with clearly defined ice marginal positions (Figure 7: Lake Shore Features in Cayuga County). While there are many discernable ice margins, the abundance of features defining the lake level elevation is much less, especially for the older, higher elevation lake levels in the southern portion of the county. Complicating the reconstruction, outlets for each lake level are outside the county boundaries and for those, 10 meter digital elevation model data were used (Figure 8: Outlets of Glacial Lakes). Beyond these shortcomings, the high resolution LiDAR elevation data coupled with field verification of the geomorphic features allows for the reconstruction of lake levels far beyond what is possible with previous USGS 1:24,000 topographic maps. Fairchild (1934) described a sequence of lake levels in the Cayuga Trough which included a series of rising lake levels. Muller and Prest (1985) refined this sequence of lake levels using data from western New York and Canada. While there is evidence of a readvance of glacial ice in Cayuga County (Fairchild, 1932; Kozlowski et al., 2014 this volume), the progression of lake levels which include a rising and subsequent falling lake level sequence is difficult to reconstruct. Without abundant age control for the lake levels the timing of most of these lakes is quite speculative.

Methodology

Using ArcMap 10.1, an initial scan was made to identify and note the elevation of strandline features, deltas, fans, and channels associated with lake levels on the 2.0 m resolution LiDAR elevation model (Figure 9: Locations of Shore Features). Next, the ice marginal positions and shore feature elevations were used to establish a relative chronology of lake stages across the county. Together the ice margins and shore elevations can establish which features are associated with the various lakes. Once established, the elevation of shore features was plotted on the map. Lines of equal elevation were drawn across the lake, often a three point problem was used to calculate where the line should be drawn. In this investigation the isobases trended east-west across the county. An elevation grid surface with a cell size of 2.0 m was created from these contours. This sloping surface was then subtracted from the LiDAR elevation model to reconstruct the isostatically depressed land surface for

each age (time) equivalent lake stage. At this point the topography was "filled" to the elevation of the lake level.

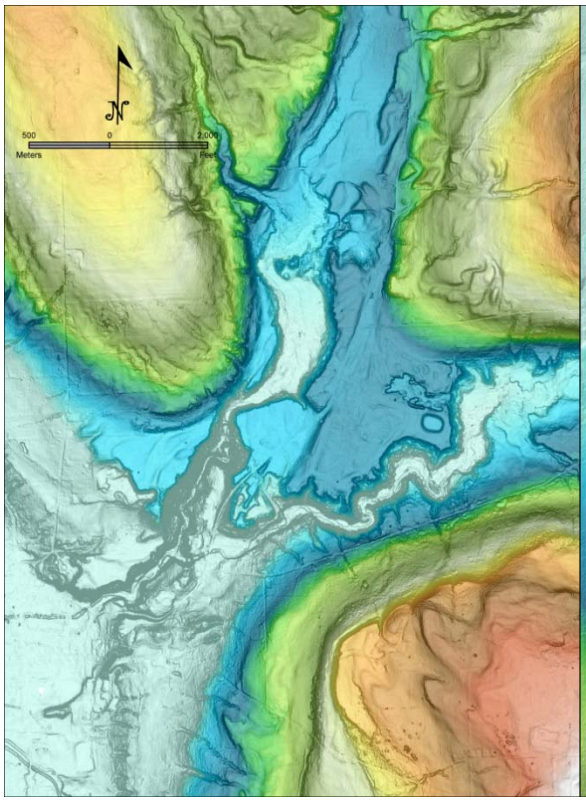
Results

Nine different lake levels utilizing four outlets were created using the elevation of shore features in Cayuga County. Figure 10 shows the sequence of lake levels decreasing in elevation as ice retreated northward. The highest lake level, with an outlet at 307 meters above mean sea level (mamsl) was Lake Ithaca while the lowest level recognized was Lake Iroquois draining through the Mohawk Valley at 130 mamsl. The drainage lowered progressively from White Church (307 mamsl) to Horseheads (280 mamsl) to Batavia (204 mamsl) to Gulf Channel (190 mamsl) and ending at the Rome Outlets (130 mamsl) (Table 2: Proglacial Lakes and Outlet Elevations). Corresponding to each lake a strandline was created. The slope of the strandline was calculated from a linear trend line connecting the shore feature elevation data. The shallowest slope was 0.11 m/km while the steepest was 0.9 m/km (Figure 11: Strandlines for Proglacial Lakes in Cayuga County).

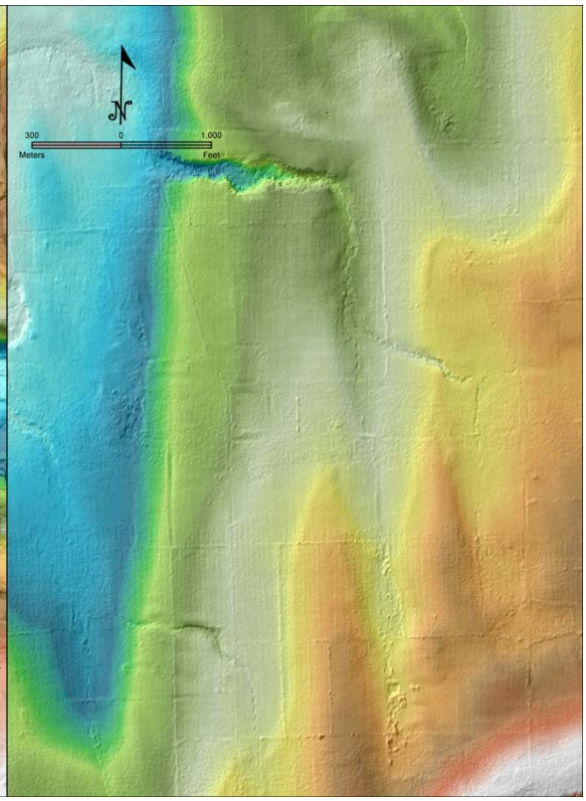
Discussion

As ice retreated from the Valley Heads Moraine the first lake present in Cayuga County would have been Lake Ithaca accompanied by smaller water bodies (Figure 10a). Confined in the Cayuga trough, Lake Ithaca would have drained southward through the White Church Outlet (Fairchild, 1934) at a modern elevation of 307 (mamsl). The most prominent feature for Lake Ithaca is the Locke Delta in the southern portion of the county. This ice contact delta occupies a large portion of the glacial trough connecting the towns of Locke and Groton, NY and is composed of stratified and cross bedded sands and gravel in excess of 30 m. The flat top of this delta would represent Lake Ithaca level (336 mamsl modern). The retreat of the ice margin from this point occurred relatively rapidly across Cayuga County as there are few deltas and essentially no well-defined beach features or deposits found here. Following Lake Ithaca, Lake Newberry was created when the outlet switched from White Church to Horseheads (280 mamsl) as Lake Ithaca merged with Lake Watkins (Fairchild, 1934). At this point there is evidence (Figure 12) of flow into the north end of Owasco Lake from the east. Lake Newberry existed until which time ice retreated to near Batavia, NY exposing lower elevation (204 mamsl) which shifted drainage to the west, ultimately across Michigan creating Lake Hall (Muller and Prest, 1985; Fairchild, 1934). The reconstruction for Margin 5 (Figure 10e) is likely at the point just before drainage shifts from south to west. At the time of this shift, the slope (0.11 m/km) of the water plane constructed for this stage is the shallowest of all investigated in Cayuga County and represents the deepest phase of Lake Hall just before the drainage through Batavia opened and may indicate fluctuation at the ice margin. As Batavia opened, the much lower outlet allowed for a nearly 62 meter drop in lake level from Lake Hall to Lake Warren. It is near this sequence that Fairchild (1934) describes a falling and subsequent rising of Lake Vanuxem. The determination of this event must be made outside Cayuga County as little evidence of a distinct, separate lake phase exists here. Lake Warren is represented in Cayuga County by a well-defined strandline and delta near Union Springs, NY. The

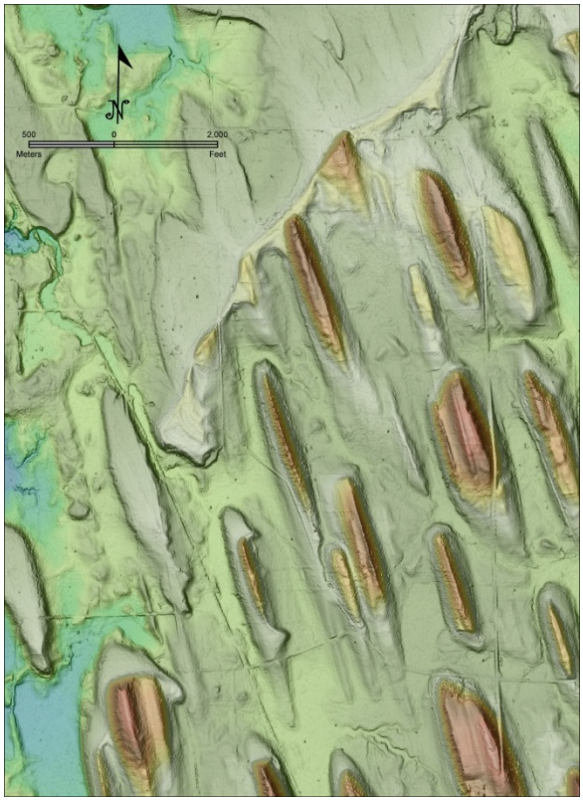
strandline traverses Great Gully at an elevation of 209 mamsl and is accompanied by a delta at the gully. Traceable for nearly 10 km the strandline helps define a water plane with a slope of 0.58 m/km. Preceding the next drop to Lake Dana level, meltwater was able to flow eastward through the Syracuse Channels, modern elevation ranging from 241 mamsl to 168 mamsl (Hand and Muller, 1972; Hand, 1978). Hand (1978) outlines a readvance model to account for Rock Cut Channel being lower in elevation than Nottingham to the north. This readvance may be the cause of the rising and falling of Lake Vanuxem discussed by Fairchild (1934). Continued northward retreat of ice opens a channel just east of the county called the Gulf Channel (190 mamsl) near Skaneateles, NY (Fairchild, 1934; Muller and Prest, 1985). This lake level would correspond to Lake Dana. Muller and Prest (1985) note the complexity of the relationship these lakes had to one another. Lake Dana in Cayuga County is well defined by another delta and strandline at Great Gully about 50m meters lower than Lake Warren and sloping at 0.84 m/km. Next lower in elevation, the extension of the Fairport channels cut across Cayuga County about 57 meters below the level of Lake Dana. With a swath about 7 km in total width, the drainage braids across in channels about 2 km wide. With the ice margin proximal to the flow and quite possibly the edge of the channel in some locations the water plane is the steepest in the county at 0.90 m/km. This event was the precursor to the final part of the sequence, Lake Iroquois. With ice now well away from the escarpment near Syracuse, water filled the area while draining eastward through Rome, NY down the Mohawk Valley toward Albany. Numerous drumlins in the north central portion of the county show wave erosion features (Figure 13: Wave Cut Drumlins). Strandlines connecting these areas throughout the northern and central portion of the county delineate the lake level and define a water plane with a slope of 0.78 m/km. Many of the drumlins in the northern portion of the county formed an “archipelago” in Lake Iroquois.



a. Deltas at various elevations



b. Strandline of glacial Lake Warren



c. Wave eroded drumlins of glacial Lake Iroquois

Figure 7: Lake Shore Features in Cayuga County

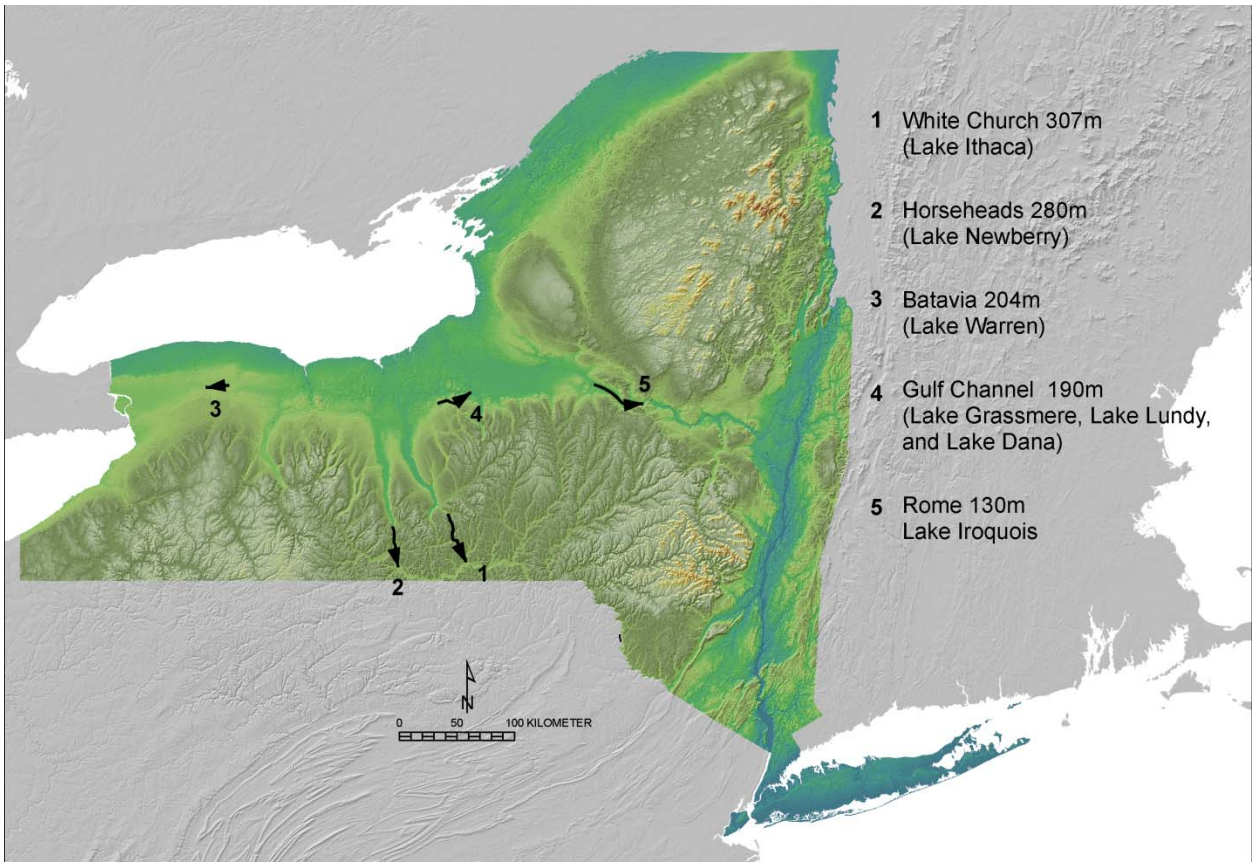


Figure 8: Outlets of Glacial Lakes (Fairchild, 1934; Muller and Prest, 1985)

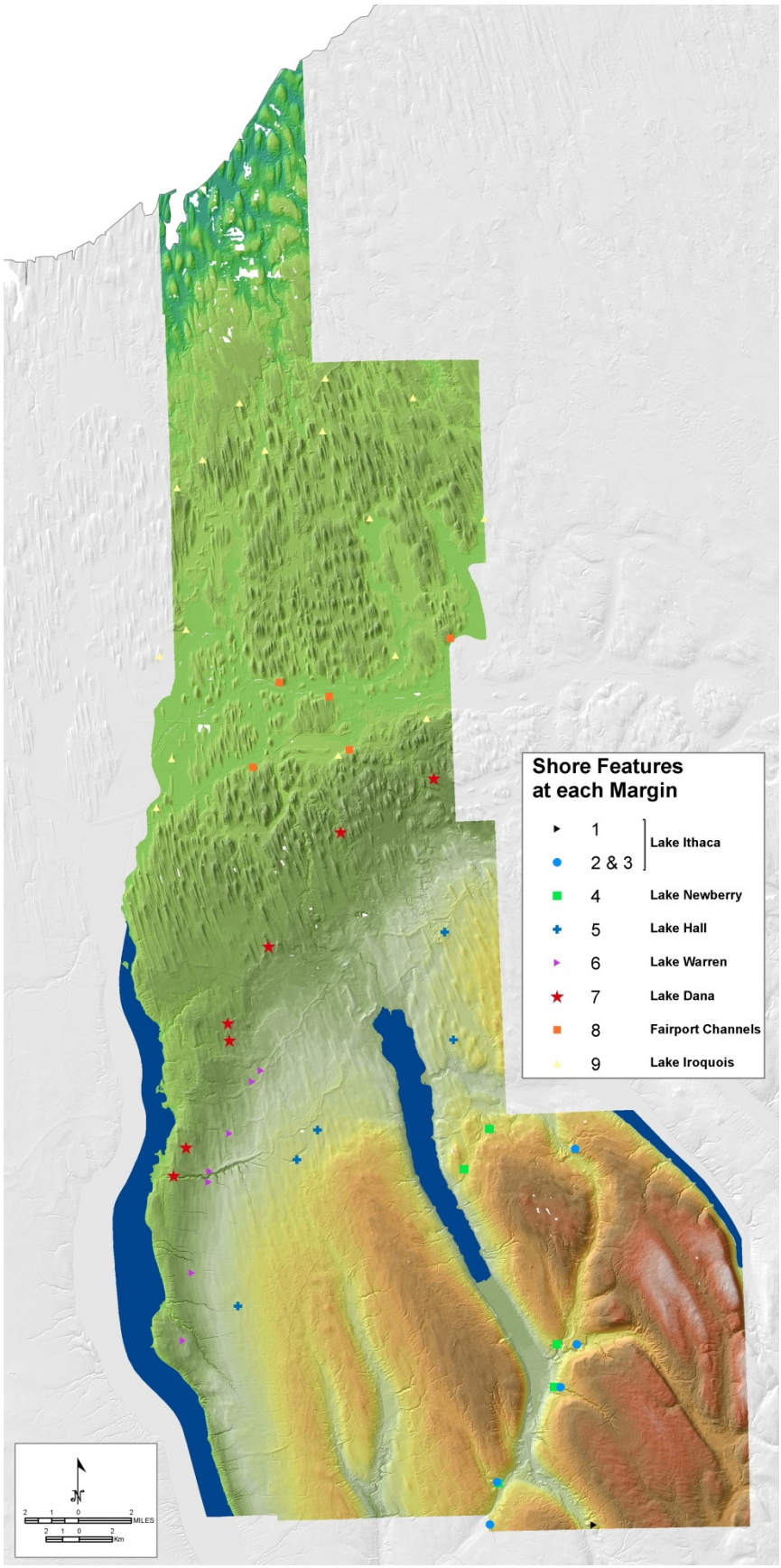
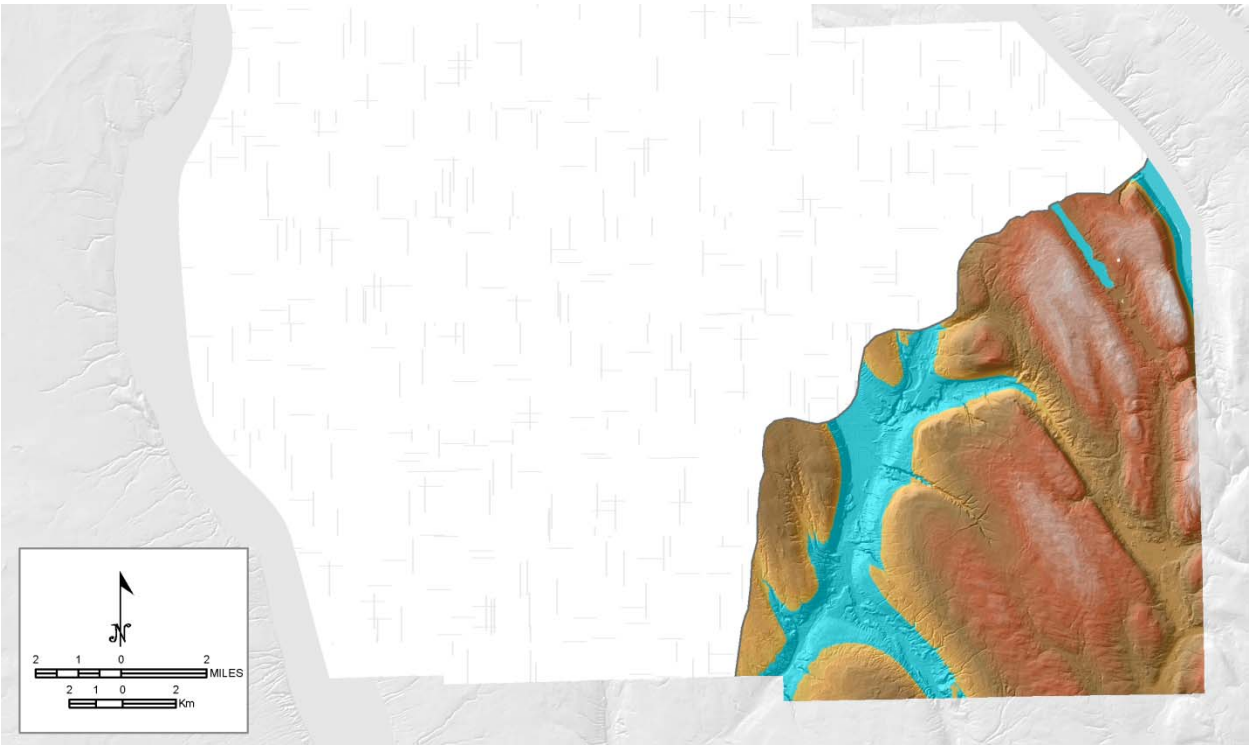
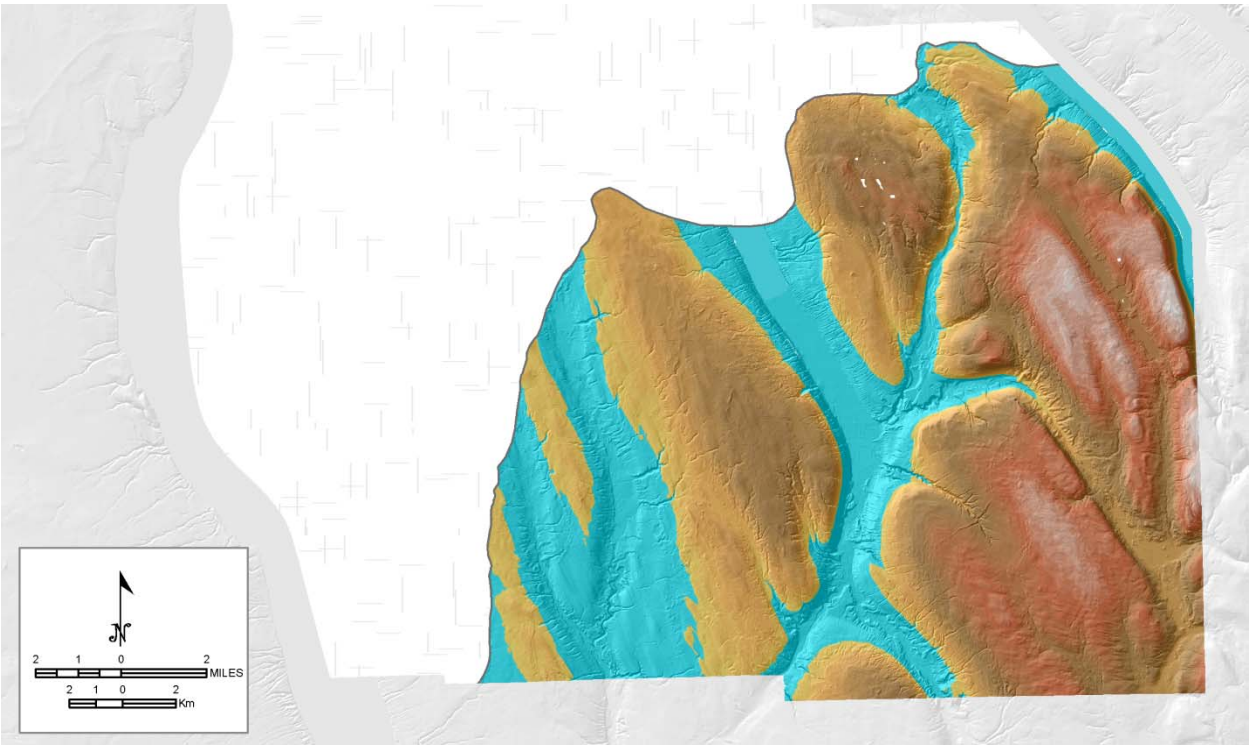


Figure 9: Locations of Shore Features

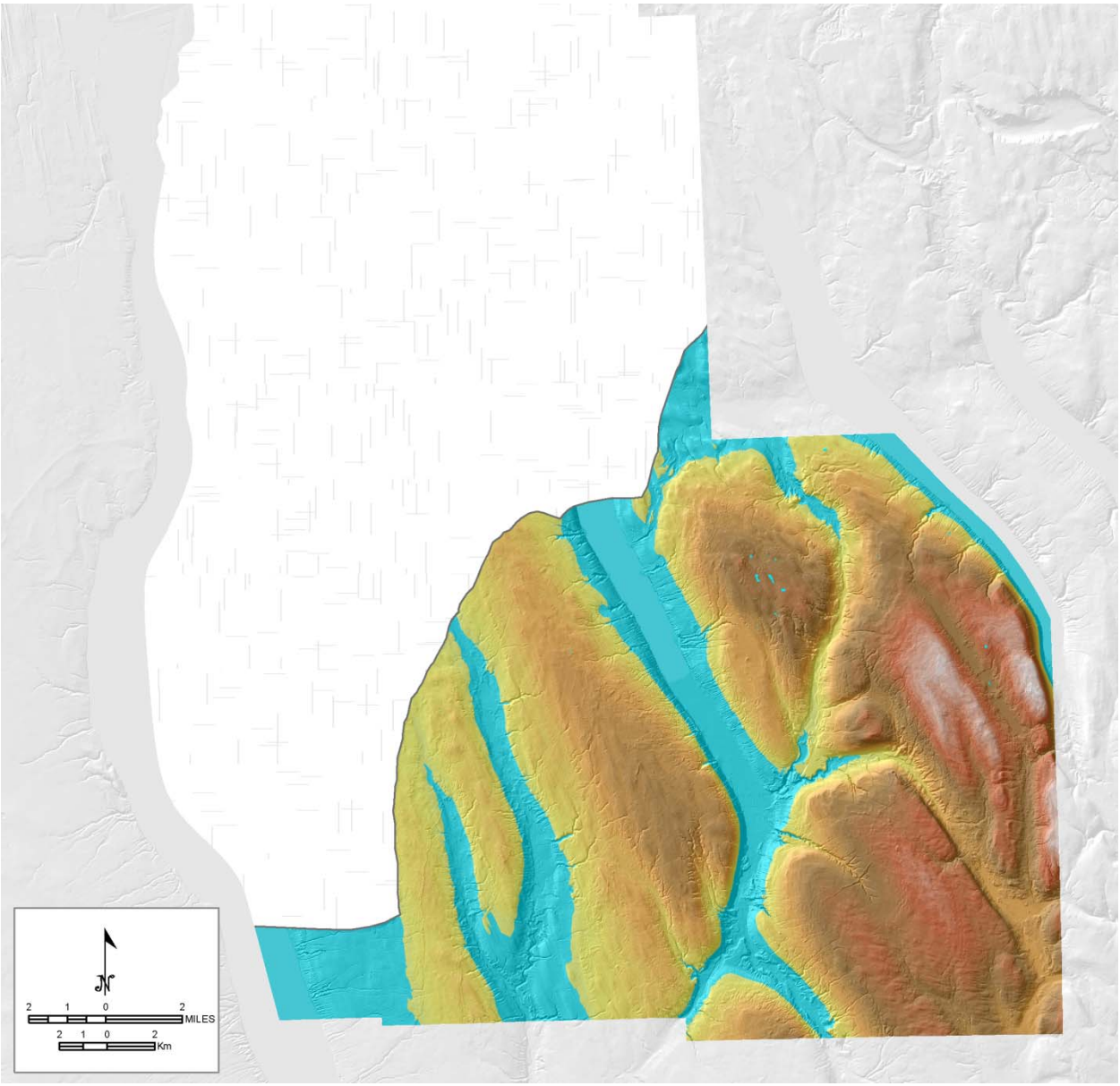
a. Ice Marginal Position 1, Locke Pit Delta at Southern Edge



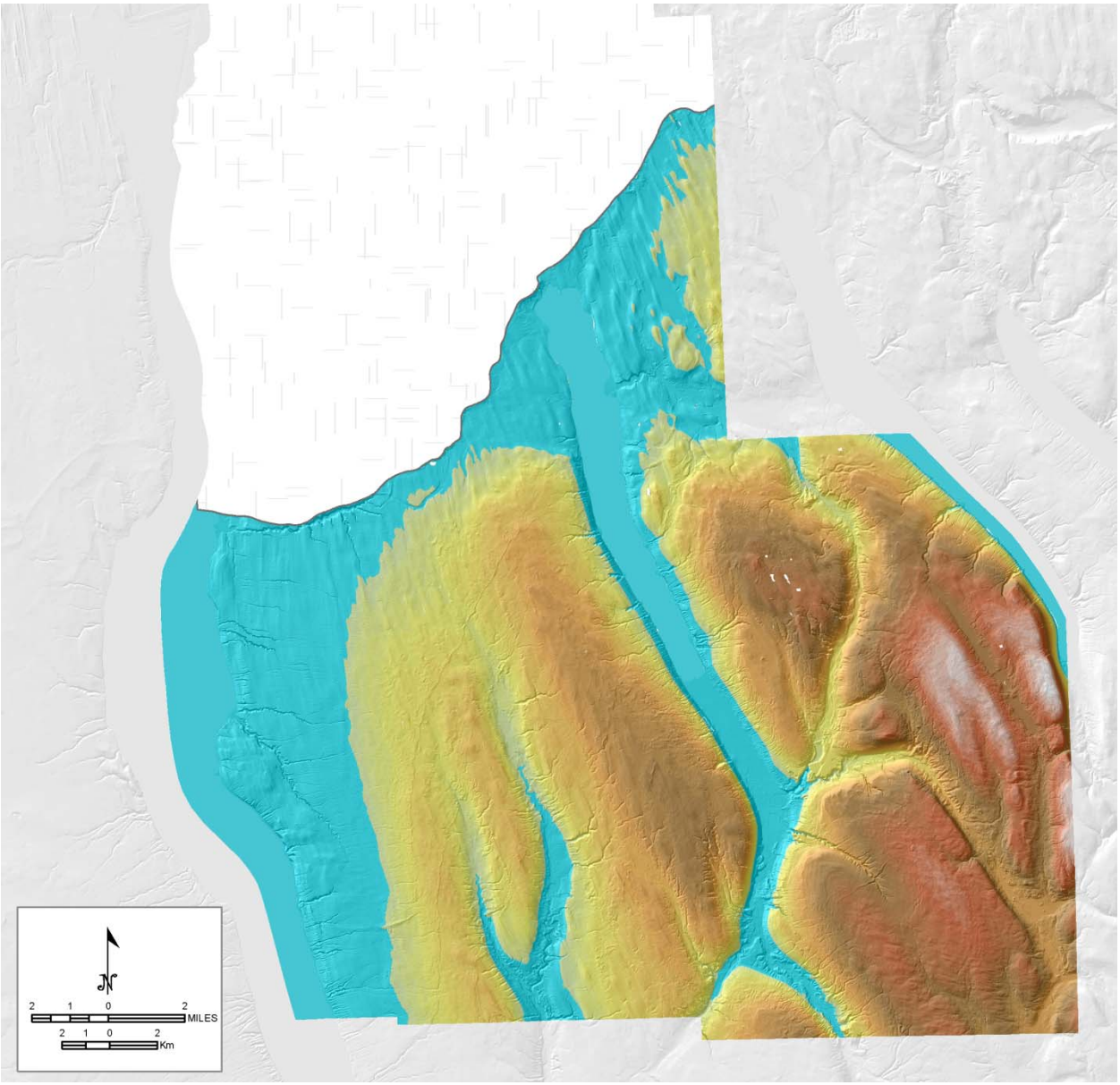
b. Ice Marginal Position 2, Lake Ithaca



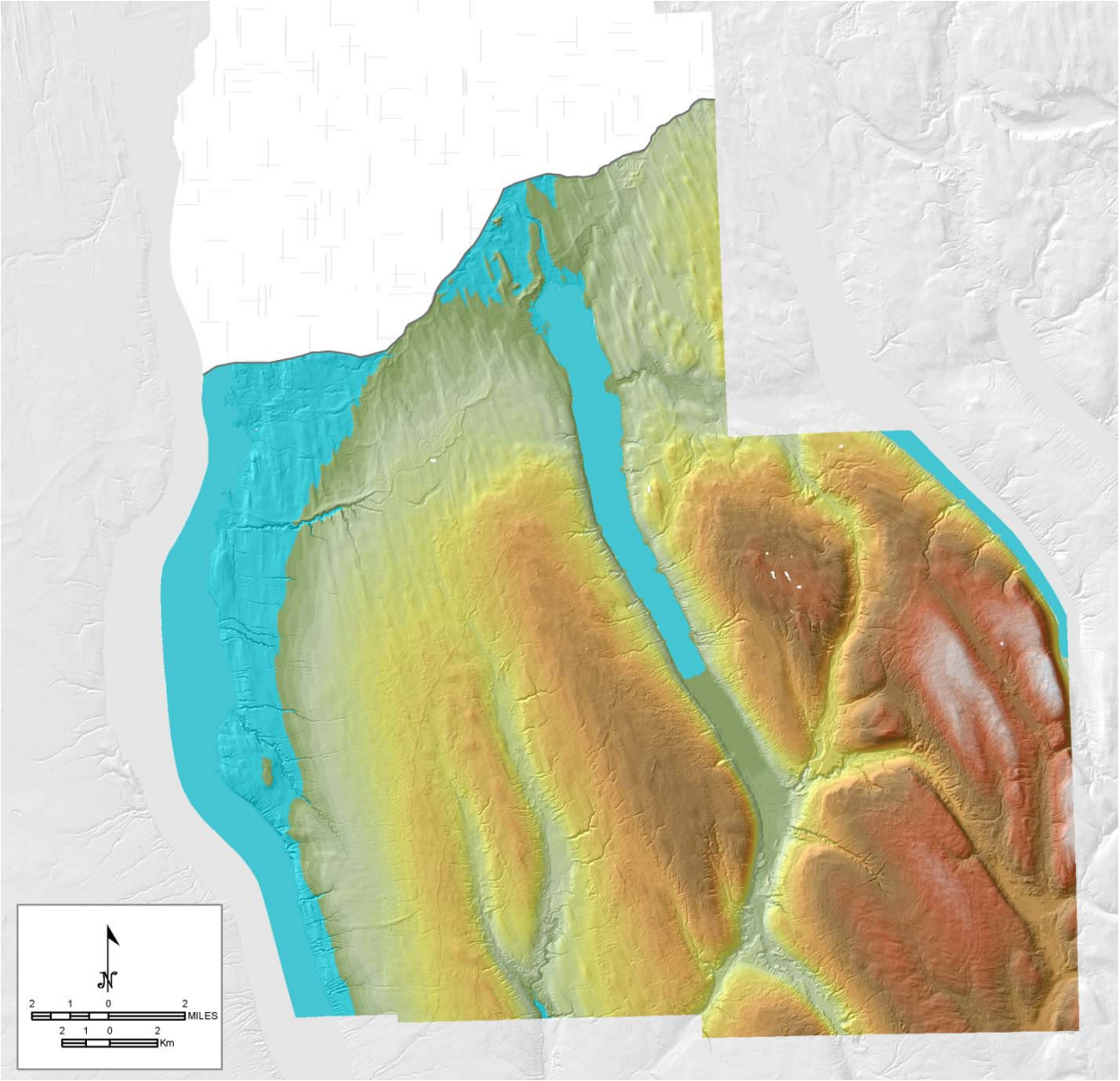
a. Ice Marginal Position 3, Lake Ithaca



d. Ice Margin Position 4, Lake Newberry



e. Ice Marginal Position 5, Lake Hall (Mapleton Moraine of Shumaker, 1957)



f. Ice Marginal Position 6, Lake Warren (Waterloo Moraine of Fairchild, 1934; Shumaker, 1957)

g. Ice Marginal Position 7, Lake Dana

h. Ice Marginal Position 8, Channels

i. Glacial Lake Iroquois

Figure 10: Sequence of Glacial Lakes in Cayuga County

Sequence	Outlet (Elevation (m))	Flow Direction	Lake Equivalent*
1			
2	White Church (307)	↓	Ithaca
3			
4	Horseheads (280)	↓	Newberry
5			Hall
6	Batavia (204)	←	Warren
7	Gulf Channel (190)	→	Grassmere, Lundy, and Dana
8			Channels
	Rome (130)	→	
9			Iroquois

* From Fairchild (1934) and Muller and Prest (1985)

Table 2: Proglacial Lakes and Outlet Elevations

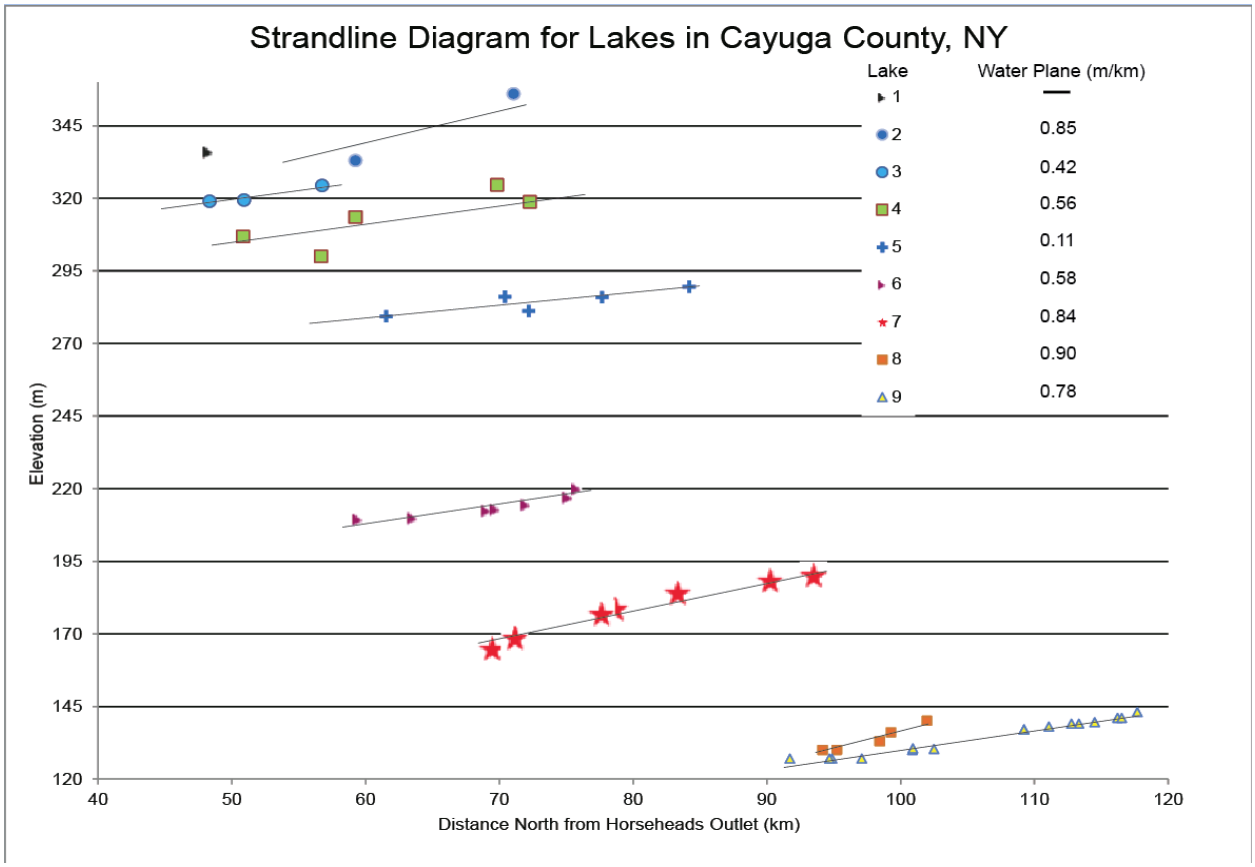


Figure 11: Strandlines for Proglacial Lakes in Cayuga County

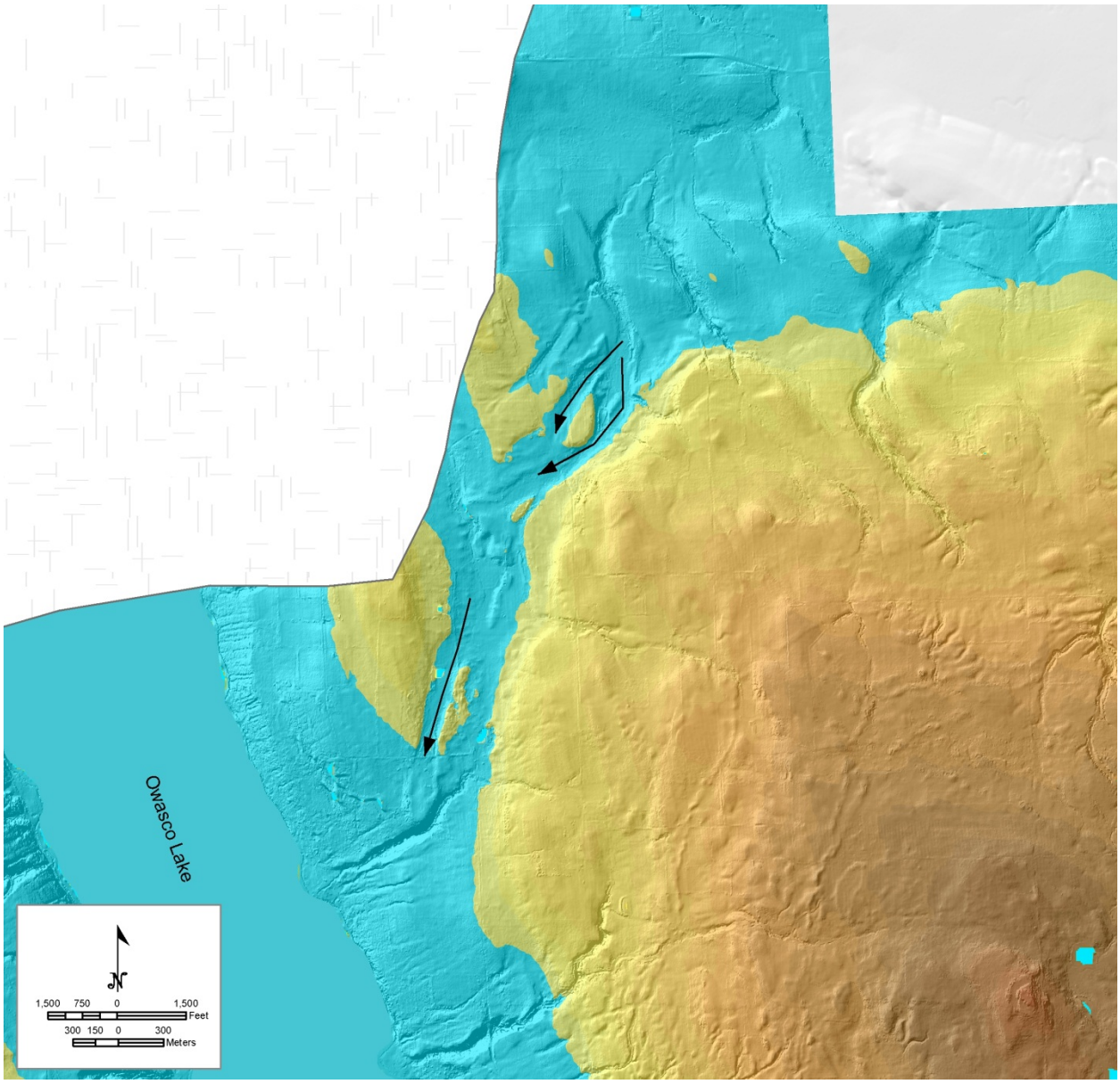


Figure 12: Westward flow into Owasco Lake basin at the time of glacial Lake Newberry

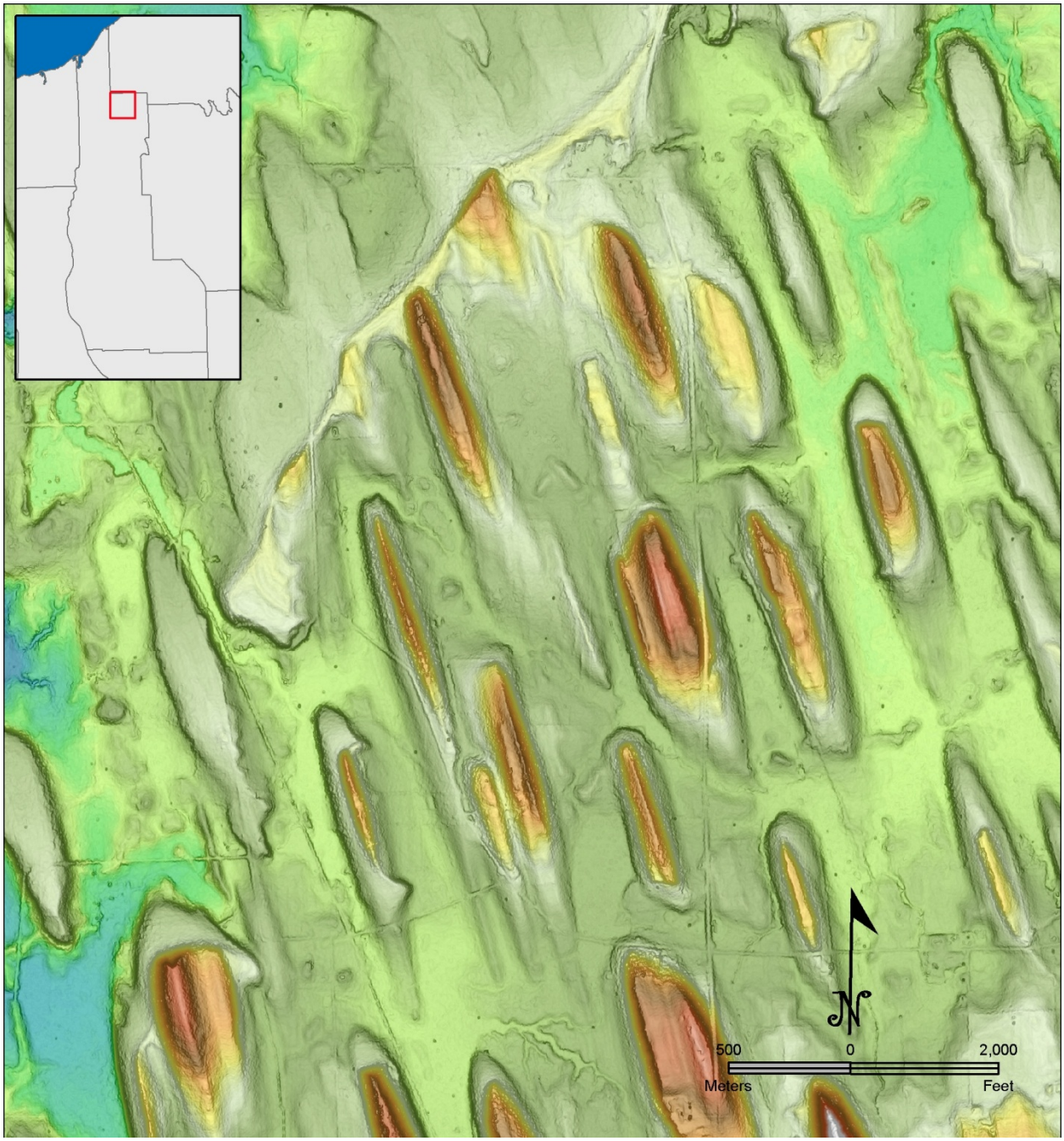


Figure 13: Wave Cut Drumlins

Stop 1:

Locke Ice Contact Delta

Brian Bird

At the southern boundary (42.6171N, 76.3831W), the Locke Delta represents the earliest ice margin in Cayuga County (Figure 14). Meltwater flowing south into a glacial Lake Ithaca deposited this ice contact delta after retreating from the Valley Heads Moraine. Part of the Inner Valley Heads Moraine about 13.2 ^{14}C years (15.7 Calibrated years) ago, this delta is coincident with a moraine in Fillmore Glen State Park (Figure 14). A prominent feature on the landscape, the delta covers approximately 0.5 km^2 and has a vertical relief of nearly 60 m with the upper most topset near 336 meters above mean sea level (mamsl) (Figure 15). This delta was deposited while drainage was south through the White Church Outlet at 307 mamsl into the Susquehanna River basin (Fairchild, 1934).

The delta is composed of mostly stratified sand and gravel with individual beds generally very well sorted and often cross-bedded. Some pockets of well sorted cobbles are occasionally encountered in the mining operation and likely represent a channel deposit that would have winnowed finer material away. Few large boulders and little clay are found in this delta. Sedimentary structures consistent with delta deposits include cross-bedding, climbing ripples and channel fill (Figure 16a & b).

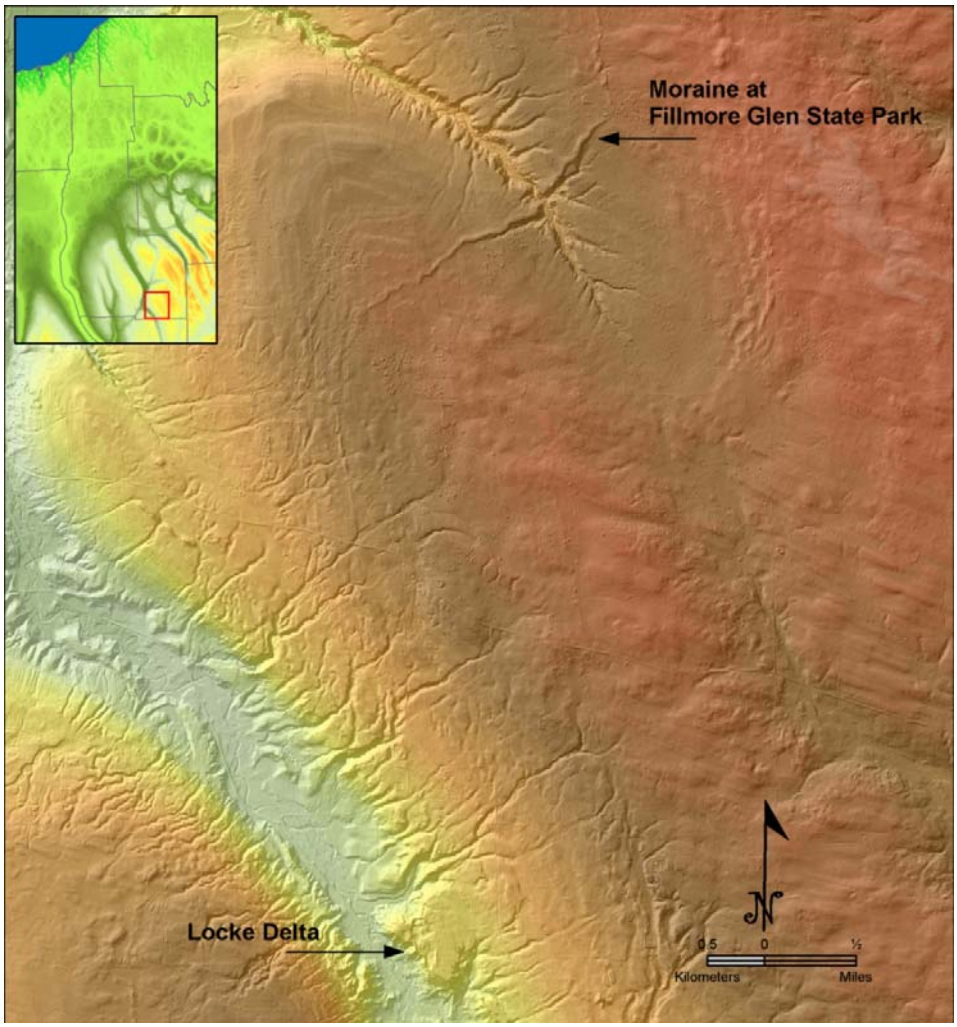


Figure 14: Locke Delta


Figure 15: Topset beds near top of the Locke Delta

Figure 16a: Stratification and Climbing Ripples Common in the Locke Delta

Figure 16b: Channel Fill in Locke Delta

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Poplar Ridge Pock Mark Hills-- Ice walled Lakes?

Andrew Kozlowski

In glaciated terrains closed depressions, sinks and hummocky topography are prevalent. Commonly these landscapes display a pattern of disorganized landforms, in our work in Cayuga County near Poplar Ridge we have identified an anomalous set of landforms that appear as a subtle belt of linear northeast – southwest trending circular shaped landforms. These landforms have nearly continuous rims 1-2 meters in height around them and diameters greater than 250 meters. They parallel ridges to the west and east, and have a unique appearance on the landscape (figure 17).

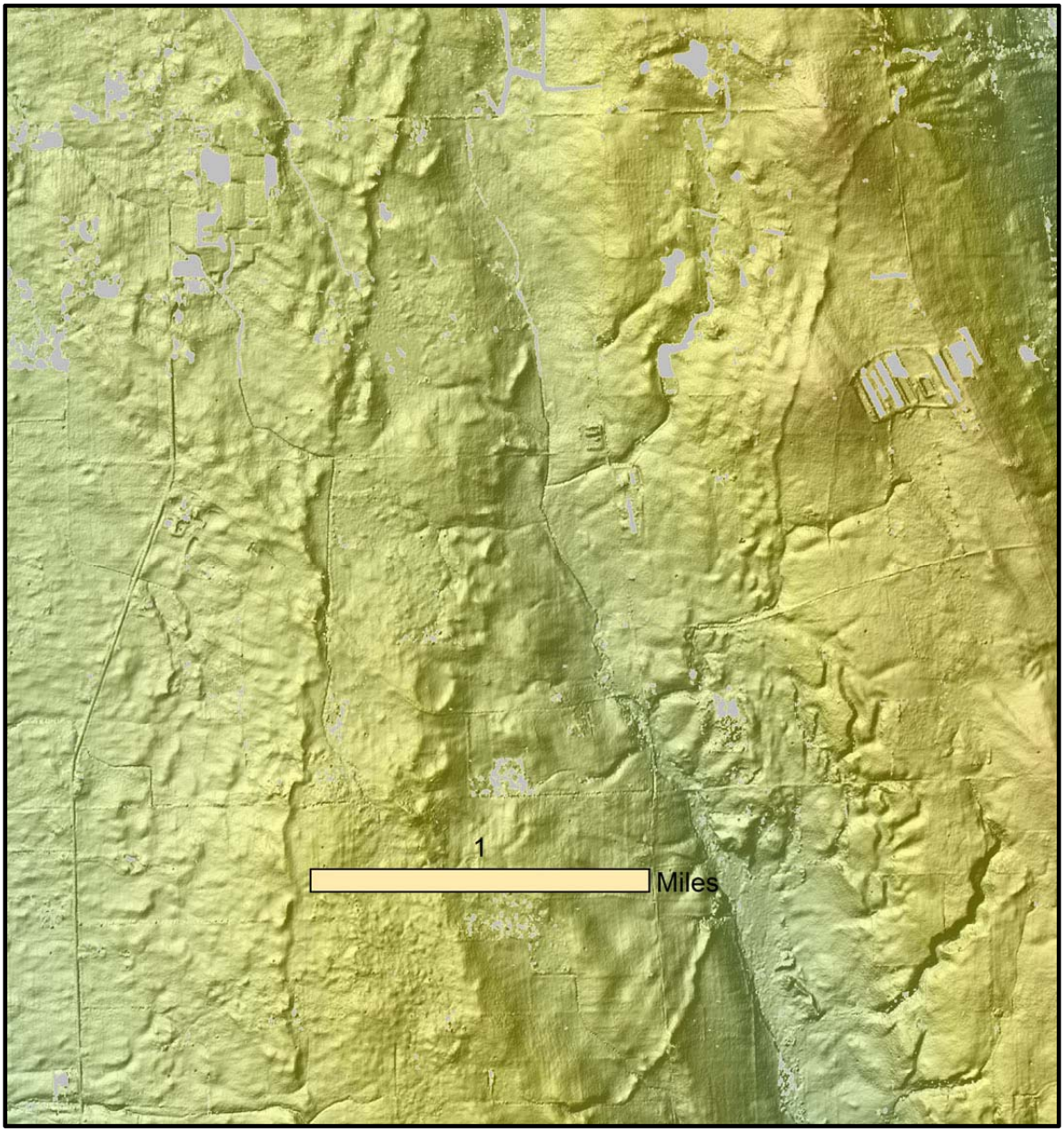

Figure 17. Topographic expression of the Poplar Ridge pock marked hills.

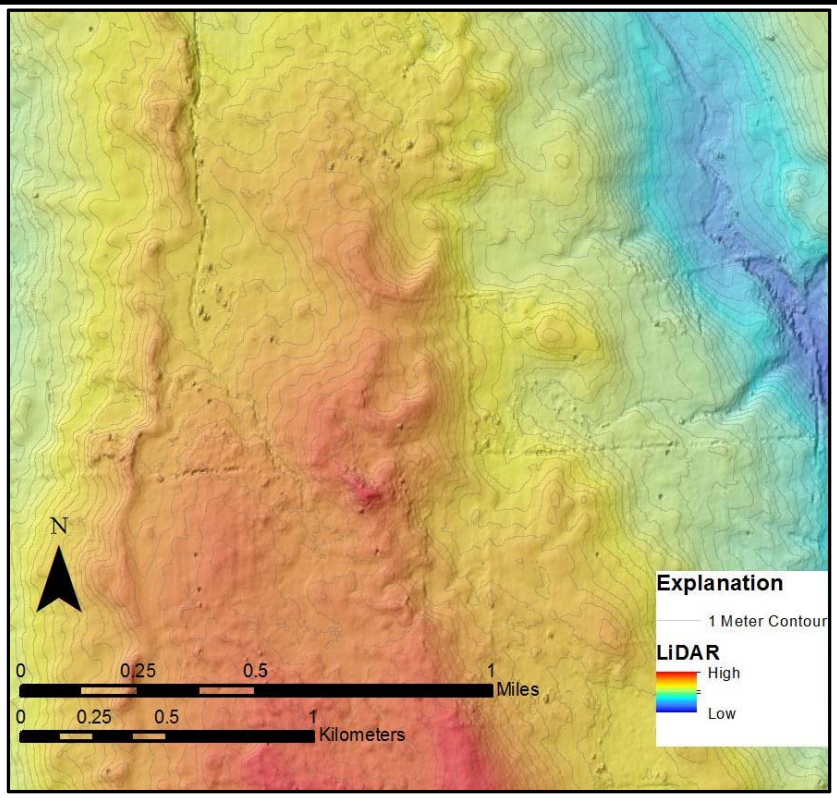

Figure 18. Aerial photo displaying tonal or textural patterns associated with the low-relief rim and high resolution DEM with contours

Flat topped, rimmed hills of varying topographic relief known as ice walled lakes plains are well documented, upland setting and in association with hummocky terrains in the Midwest (Clayton and Cherry, 1967;Ham and Attig, 1996; Clayton et al., 2008; Curry et al., 2010; Curry and Petras, 2011).However, such features are not well documented in the glaciated northeast. At this stage our investigation has only been a geomorphic analysis. Detailed mapping and coring of the features will be part of this field season's (2014) STATEMAP projects.

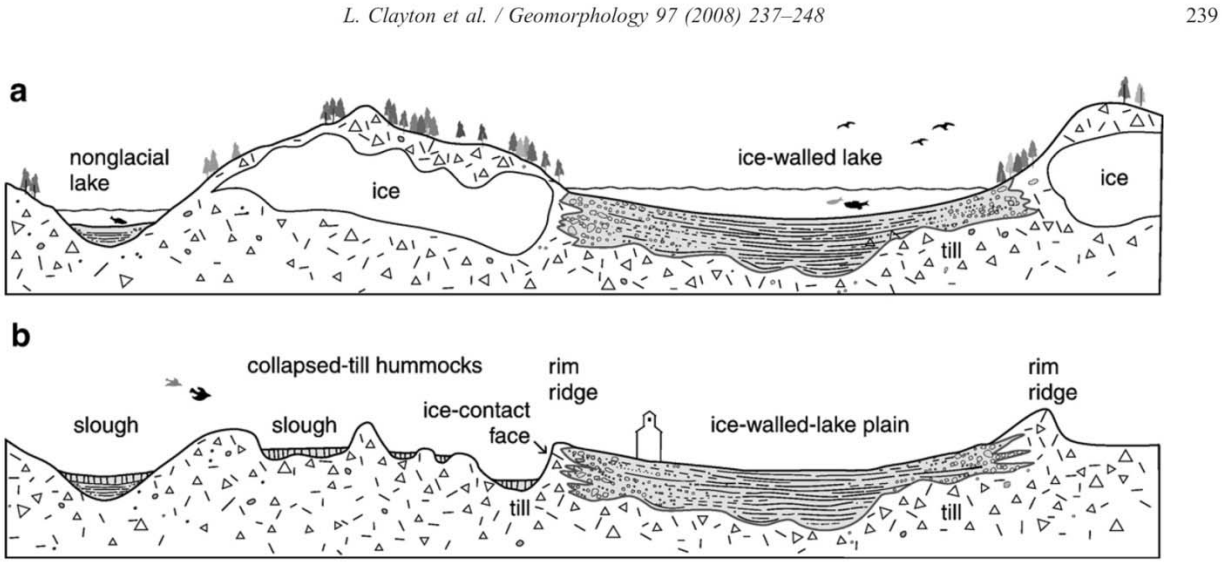

Figure 19. Models of formation proposed by Clayton et al., 2008, (a) confined by ice and (b) and after ice melt.

Carbon Dating of Megafauna and the Colonization of NY State after the Last Glacial Maximum

Robert S. Feranec

Large mammals (megafauna) represent an important component of ancient ecosystems as well as a significant dietary resource for ancient humans. Habitation of the unglaciated areas of New York State after the melting of the Laurentide Ice Sheet (LIS) at the end of the Pleistocene (Dyke et al., 2002) necessitates dispersal of fauna, including humans. The timing of the dispersal of particular species as well as the composition of large mammals present in New York State after the last glacial maximum (LGM) will show when certain species colonized New York, and when they may have been available for the ancient peoples that dispersed and inhabited the newly unglaciated areas of the state.

A few fossil assemblages containing mammalian megafauna of late Pleistocene age are known from NY State, including the western Hiscock locality of Genesee County (Laub papers), the southeastern Dutchess Quarry Caves of Orange County (Funk and Steadman, 1994; Steadman et al., 1997), and the eastern Diddly and Joralemon's Caves of Albany County (Steadman et al., 1993a,b). Although there are only a limited number of assemblage localities, single specimen localities are fairly abundant. For example, mastodons (*Mammut americanum*) represent the most abundant Pleistocene megafaunal species in New York State occurring at over 150 specimens/ localities (Thomson et al., 2008; Hartnagel and Bishop, 1922). Mammoth, including both Columbian Mammoth (i.e., *Mammuthus columbi columbi* and *Mammuthus columbi jeffersoni*) and Woolly Mammoth (*Mammuthus primigenius*) are known from over 20 specimens/localities (Hartnagel and Bishop, 1922; Feranec and Kozlowski, 2010, 2012). Other megafauna, including caribou (*Rangifer tarandus*), muskox (*Ovibos moscatus*), peccary (*Platygonus compressus*), giant beaver (*Castoroides ohioensis*), and giant ground sloth (*Megalonyx jeffersoni*), are known from fewer localities (Thomson et al., 2008; Hartnagel and Bishop, 1922).

Calculating dates and modeling ages for megafauna provides an understanding of the initial habitation of the state by large mammals, and if particular taxa represent the major food source for humans (e.g., proboscideans, caribou), provides a minimum age for when people could also colonize the area. That is, humans would likely not colonize an area without a reliable food source. Besides its implications for human colonization, the dates and modeled ages for megafauna have implications regarding ecosystem assembly and extinction. Because NY was almost entirely glaciated, and thus uninhabited, the dates will show the timing and sequence of mammal species colonization in a previously uninhabited continental environment. The dates and modeled ages also show the latest occurrence for species within the state, providing the timing for regional extinction or extirpation.

We have begun to resolve the timing of occurrence of megafaunal species in NY State through radiocarbon analysis, and have modeled the date of earliest, and for relevant taxa, latest occurrences, for particular species (Feranec and Kozlowski, 2010, 2012). Our dates have been supplemented with

previously published dates that have been vetted for quality using the ranking scale to assess the quality of radiocarbon dates of Meade and Meltzer (1984), which was updated by Barnosky and Lindsay (2010). For our samples, the procedure for collagen extraction generally follow the techniques of Brown et al. (1988) and Bronk Ramsey et al. (2004). Samples were decalcified using 0.5N HCl or EDTA, depending on quality of preservation, to obtain collagen, typically from 24-48 hours. Once decalcified, collagen was gelatinized at 58°C for 17 hours. Subsequently, the gelatin solution was filtered to remove any remaining solids, and then ultra-filtered to remove the 30 kD fraction, which was then lyophilized. In general, lyophilized collagen appeared similar to a white cotton ball. We utilized the National Oceanic Sciences Accelerator Mass Spectrometry (NOSAMS) facility for graphitization and ^{14}C analysis (NOSAMS website).

Two different calibration techniques were utilized. First, calibration of ^{14}C dates from particular individuals was made using the online CALIB 6.0html program (Stuiver et al., 2005). Calibrated dates were obtained at the 2σ age range. We also modelled the calibrated range of dates for the populations of different megafaunal species present in New York after the LGM. Because the individual dates and calibrations likely do not represent the earliest and latest occurrences for each particular species, the models provide ranges for the earliest and latest occurrences of each species within the state.

The modelled calibrated dates show that the megafauna do not appear to colonize NY State all at once, but in a sequence generally following: caribou>mammoth>mastodon=peccary=giant beaver.

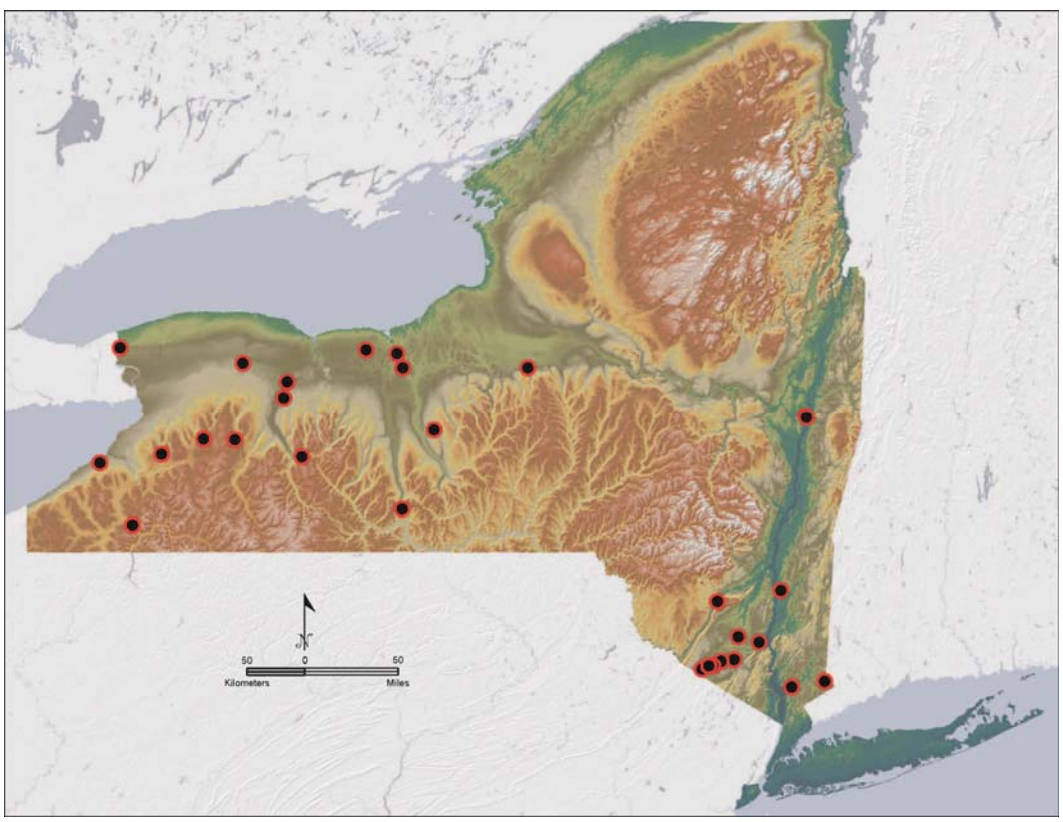

Figure 20. Map of dated megafauna throughout New York State.

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Great Gully – Progress toward the Missing Link –

Andrew Kozlowski, Brian Bird, Brandon Graham

In 1956, Robert Shumaker was a completing his graduate study under the direction of Dr. Ernest Muller, then professor at Cornell. The six weeks of field support was sponsored by Dr. John Broughton, State Geologist and Director of the New York State Geological Survey. His objective at the time was to evaluate moraines south of Auburn and to map and evaluate till textures and lithologic variations of surficial deposits within the Union Springs Quadrangle. At this time there had been an ongoing debate amongst geologists between glacial successions proposed by Fairchild (1932). While Holmes (1953) and others suggested moraines in Central NY represented continuous recession within Cary time (the last ~15,000 years). Previously Chisnell (1951) had described deltas associated with High-standing proglacial lakes (Figure 21).

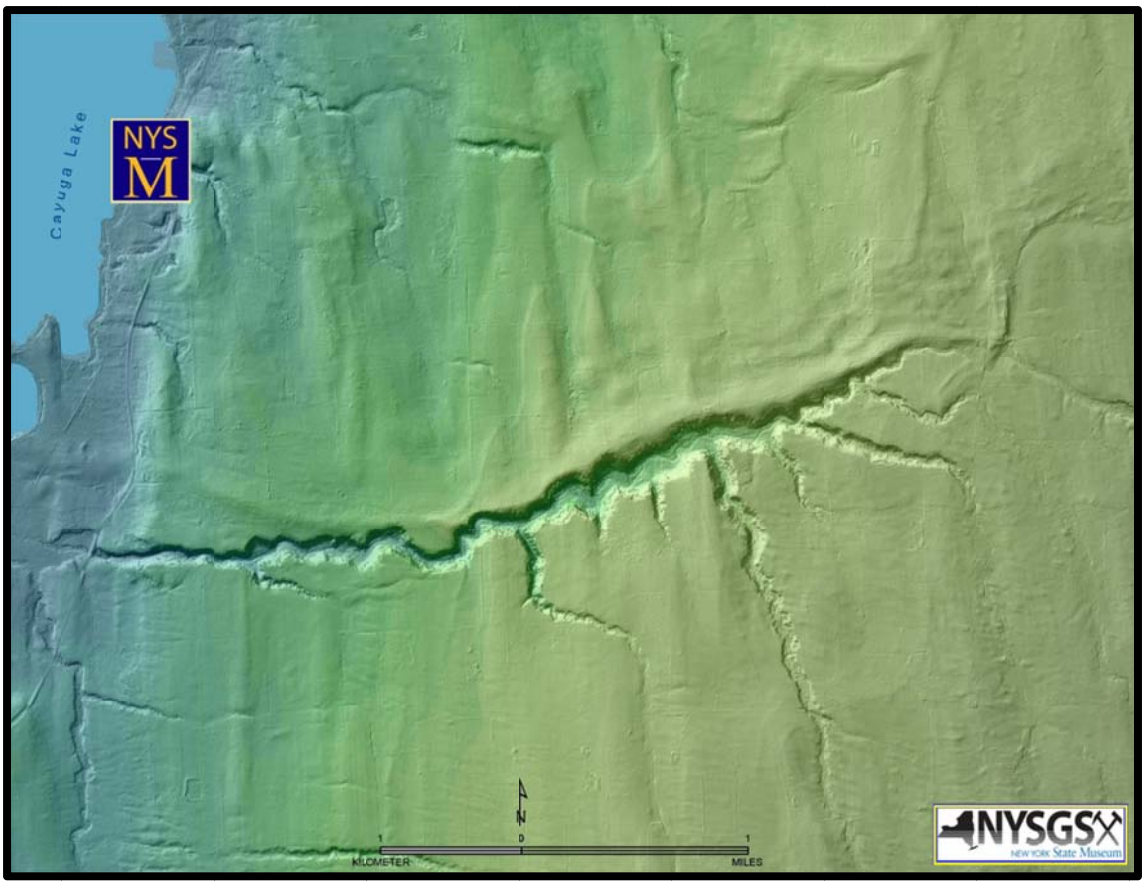


Figure 21. Elevation model of Great Gully and adjacent landforms

While completing this study, he made explorations into Great Gully, an obvious physiographic feature to investigate, and see if glacial deposits and stratigraphy were exposed. There he documented till over sand deposits containing wood and organic debris, underlain by till deposits. Radiocarbon dates on sticks yielded an age of greater than 35,000 years BP, that being the maximum datable range at the time. Shumaker (1957) concluded that till sections in the Great Gully gorge indicate periods of deglaciation. The duration and correlation of the interval is unknown, but he speculated that it may be

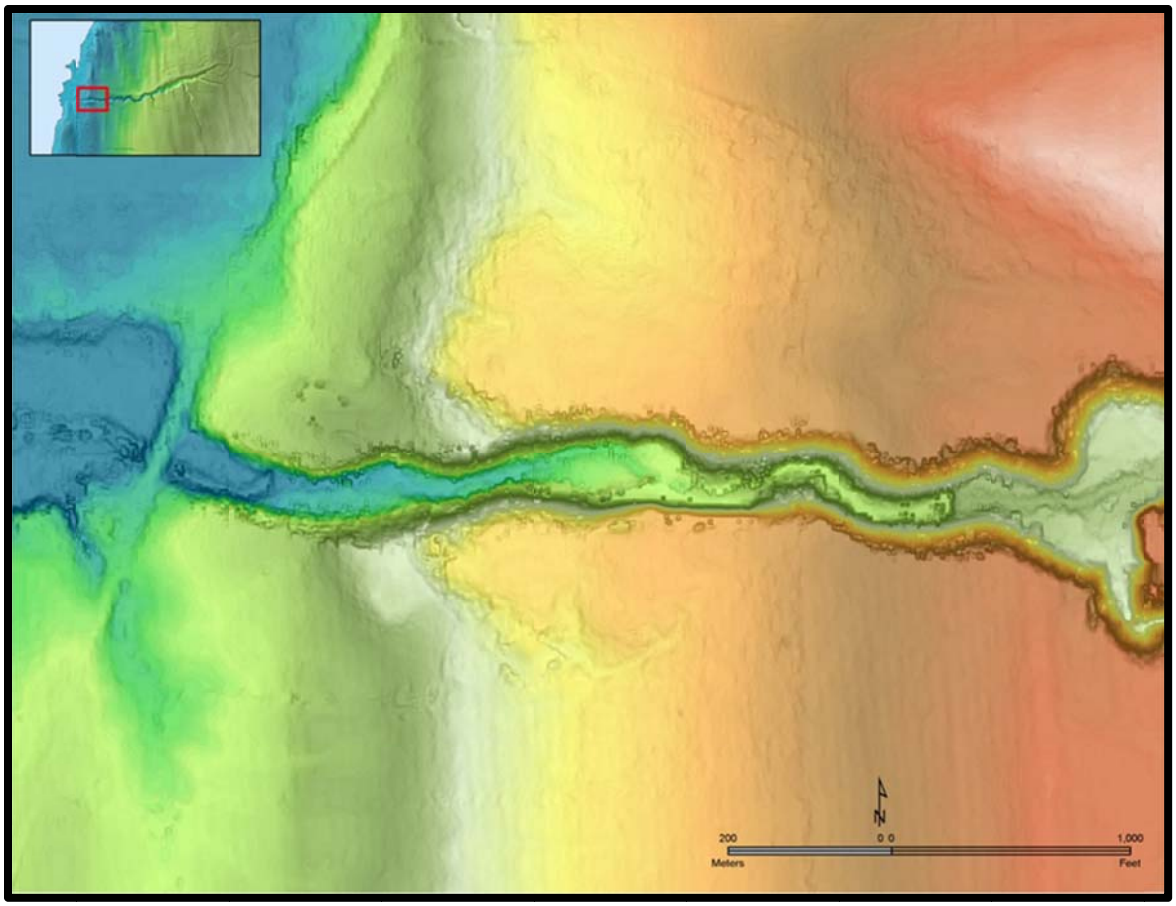
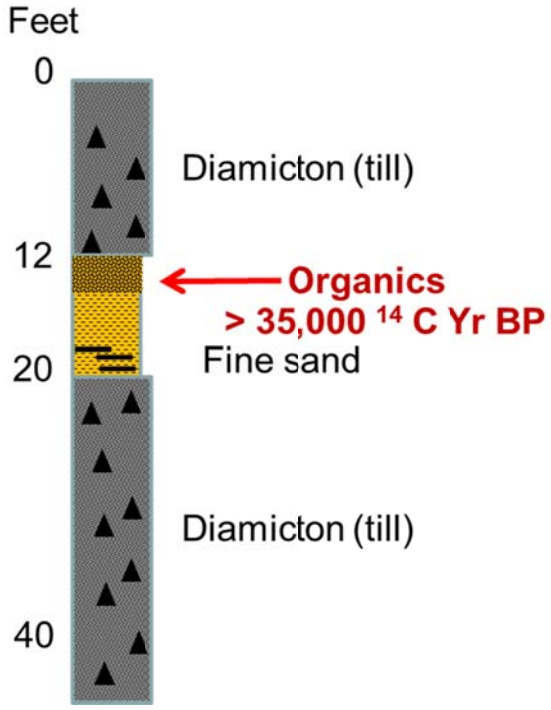


Figure 22. Hanging Deltas along Great Gully



equivalent to Brady Age (Approx Mid Wisconsin). In an earlier correspondence to John Broughton dated May 31, 1957 he further added the time interval represented by lake sediments represents an interval of recession and major fluctuation during the initial advance of Wisconsin Ice into New York.

Muller (1957) briefly discussed the age of the gully and others, and suggested that the gorges may be of Sangamon Age? Yet surprisingly, no one has returned or found enough to report and add to the original observations of Shumaker until now.

Figure 23. Reproduced schematic of Shumaker's original Stratigraphic section.

Rediscovery of Great Gully organic sites

Our initial return and investigation began in February of 2012, at that time Dan Karig, Todd Miller, Bill Kappel and I spent a morning and afternoon trying to relocate the original exposures described by Shumaker. Although we eventually found our way out of bedrock gorges and into glacial sections we left never being convinced that we had found the correct stratigraphy or the exact locations described.

Please note: all points of access we are making within Great Gully are on Private Land and require written permission, all parcels are closed during hunting season. This site currently remains an active research project and we need to maintain access. Please DO NOT return with or without students without securing permission from the land owners. I returned to the area in April, secured permissions from the land owner adjacent to our initial surveys and continued to explore the deep ravine and side tributaries. Shortly thereafter I added the capable assistance of John Wiant and Brandon Graham. They spent a full two weeks scouring every ravine and rill and I would appear late in the week to accompany them and check the stratigraphy of the new sections discovered since the previous week. On Friday June 22nd, I returned with Brandon, that particular day we went to a new tributary he and John had been working in, we proceeded south ward into a shallow (by comparison) ravine and dropped down to stream level and proceeded 400 meters south from where we entered off of Chase Road. Here Brandon began to show me a 2 meter tall exposure of gray sands subcropping a diamicton unit that appeared to be glacial till. While, carving off the exposure with trowels to examine the stratigraphic details, I noticed that there was an abundance of dark laminations and mottled streaks throughout the cross bedded sands, I also noticed the sands were densely packed and some of the dark laminations appeared to be sheared or deformed by the overlying diamicton unit. It was at that point when my trowel snagged a small stick in the sand and snapped. I recall Brandon and I looking at each other and uttering a few energetic expletives. This site would become known as GG04, we spent another hour clearing and photographing the site and collecting dozens of sticks, and compressed organic debris. We then moved southward to a larger exposure 400 meters north of GG04. There, a large (3.5 meter tall exposure was present, after considerable digging to clear off the exposure it became obvious that a similar, yet more complex stratigraphy was present that included three diamicton units (tills) each separated by cross-bedded sands containing abundant organic contents (sticks, plant detritus, organic mats). Again more expletives, and the first realization that these subtill organic bearing units appeared to be laterally continuous. We literally left that day to drive back to Albany with a 5 gallon bucket of samples and wood. All in all we returned numerous times and logged more than 35 days in the Great Gully, a couple of days clearing off exposures in 100 degree heat.

Eventually, a third and fourth locality were discovered that contained a similar stratigraphy with organic bearing units, not all were as abundant bearing sticks, however all were underlying diamicton units. As point of emphasis, sub-till organic sites, in the northeast and particularly NY are rare and the significance of the sites was apparent to us from the start. We focused on quality control and wanted to avoid ambiguity of reporting a single radiocarbon age and where possible submitted two or more (in most cases three) samples from each organic bearing bed for radiocarbon analysis.

Great Gully Site Descriptions – Swayze Farm and The Nature Conservancy

During our analysis and investigation of glacial exposures in Great Gully we investigated numerous exposures and cuts along three deep rills (ravines) located on the south edge of the valley. We present a brief stratigraphic description as an overview and a reference for discussions that follow regarding deformation and chronology. Our first stop descriptions are located along the Swayze Farm Rill, we will spend some time examining the stratigraphy and deformation exposed. Then we will regroup and drive by bus over to the Chase Road Rill.

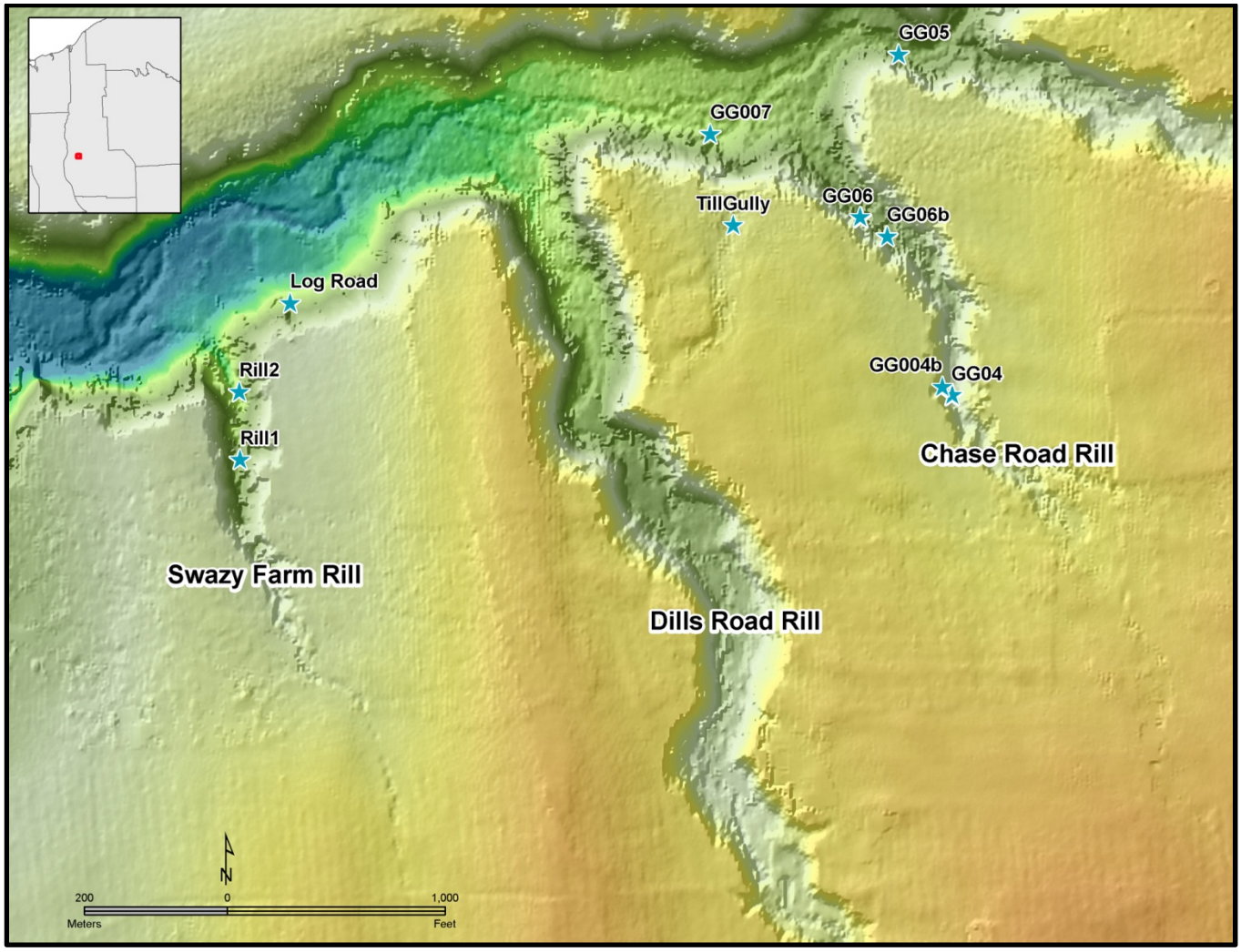
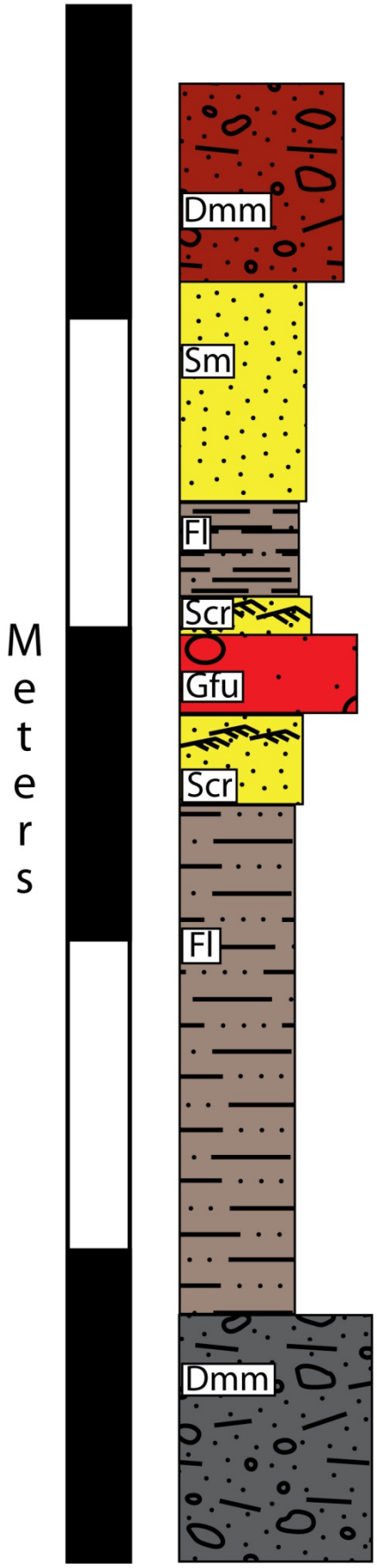
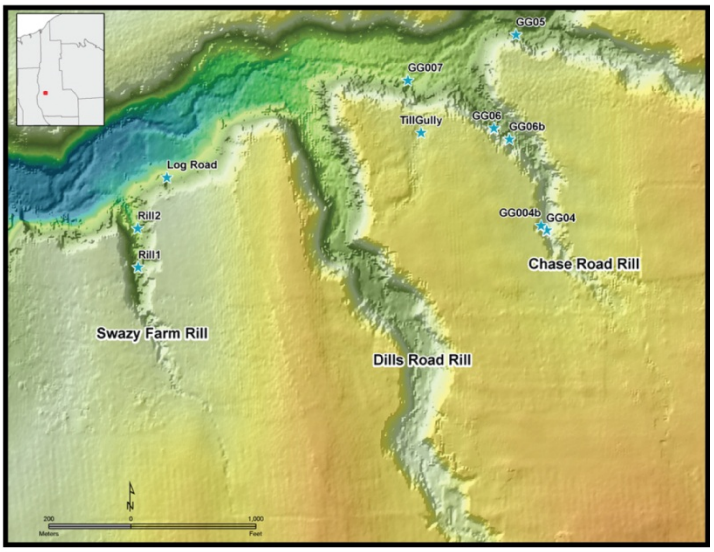


Figure 24. Displaying location of stratigraphic study sites and names of rills referred to herein

Following site descriptions will not be referred to as “Figures”.

GG03 --Log Road Site



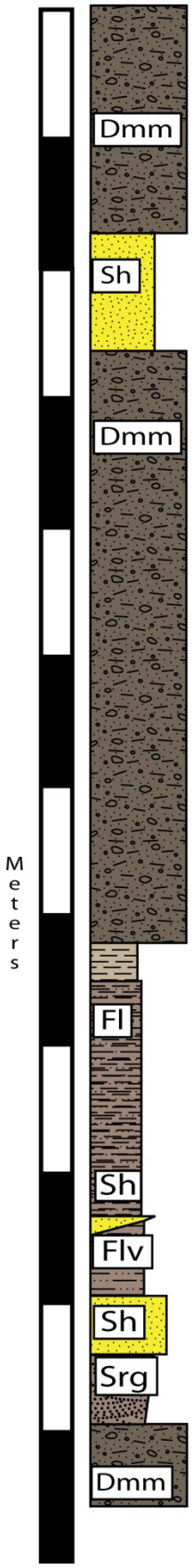


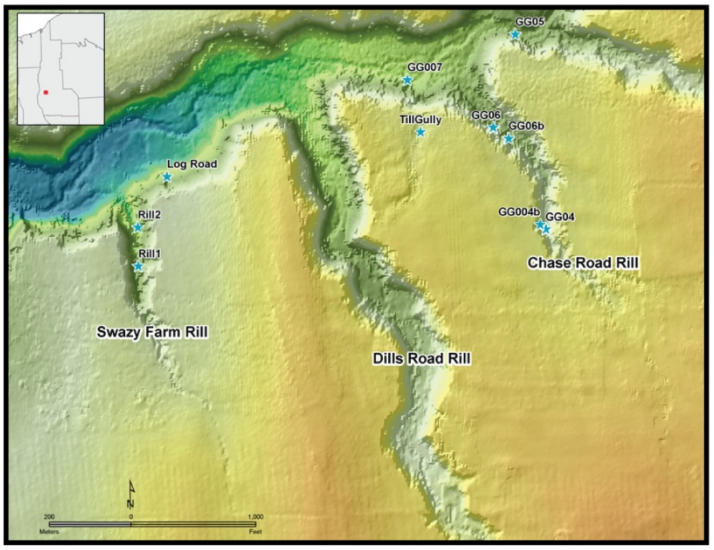
GG03, known as the Log Road site due to its proximity to an old logging road, was one of the first locations to be explored in Great Gully. The site exhibits over compacted, silty, fine sand under a glacial till near the surface of the field. The silty, fine sand exhibited faulting throughout. The layer transitions to dense silt, and then into rippled sands over gravels. The gravels transitioned back into rippled sands and then into a very stiff purple clay. The clay was generally free of gravel in the upper portion. Lower in section, a gradual transition from clay to till was observed.

Deformation was present at the site in the form of sub-meter scale shearing of beds as well as a larger meter scale fold.

Dmm - Matrix-supported, massive
Gfu - Upward-fining (normal grading)
Scr - Climbing ripples
Sm - Massive
Fl - Fine lamination often with minor fine sand and very small ripples

Rill 1

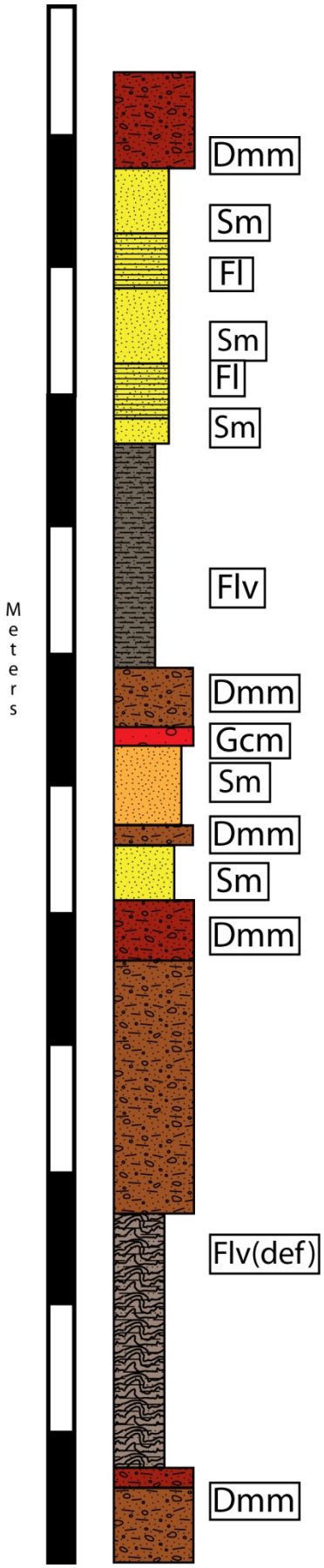


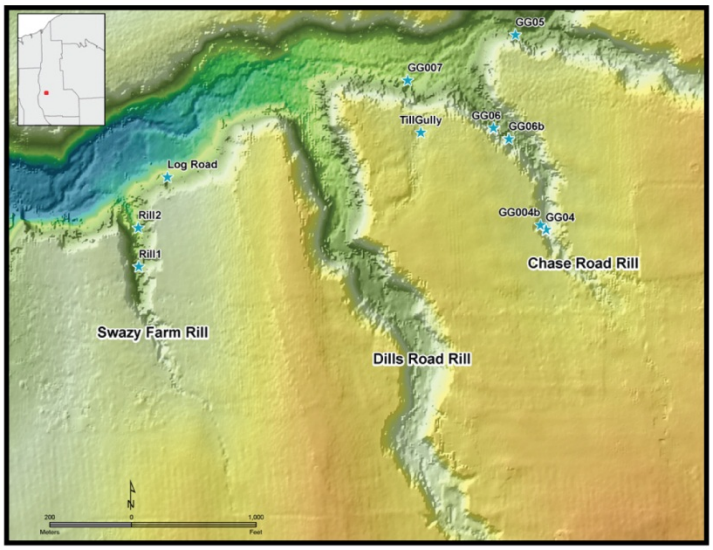


Characterized by two observed tills separated by an approximately 3 meter thick sequence of over-consolidated silt and sand, Rill 1 appears to have no conspicuous deformation in contrast to Rill 2 to the north. The uppermost sand is massive. A thick package of matrix supported till is then underlain by dense, laminated silt and fine sand, marked by a disconformable contact. The silt and sand transition to purple hard clay with a gradual transition to diamicton similar to GG03.

Dmm	-	Matrix supported, massive (till)
Gci	-	Clast supported, imbricated
Gcm	-	Clast supported, massive
Gfu	-	Upward Fining (normal grading)
Scr	-	Climbing Ripples
Sd	-	Deformed bedding
Sh	-	Very fine to medium and horizontally/ plane bedded or low angle cross laminated
Sl	-	Horizontal and draped lamination
Sm	-	Massive sand
Sr(B)		Ripple cross laminated (type B)
Srg	-	Graded cross laminations
Fl	-	Fine Laminations often with minor fine sand and very small ripples
Flv	-	Fine laminations with rhythmites or varves
Fm	-	Massive silty fine sand
Ow	-	Organics, wood and sticks

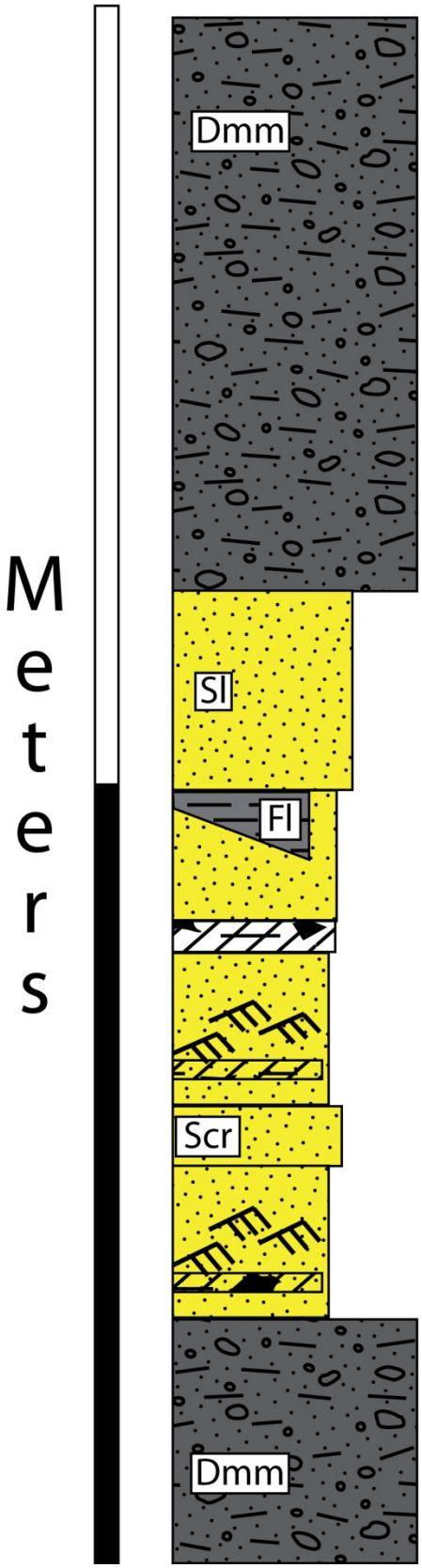
Rill 2

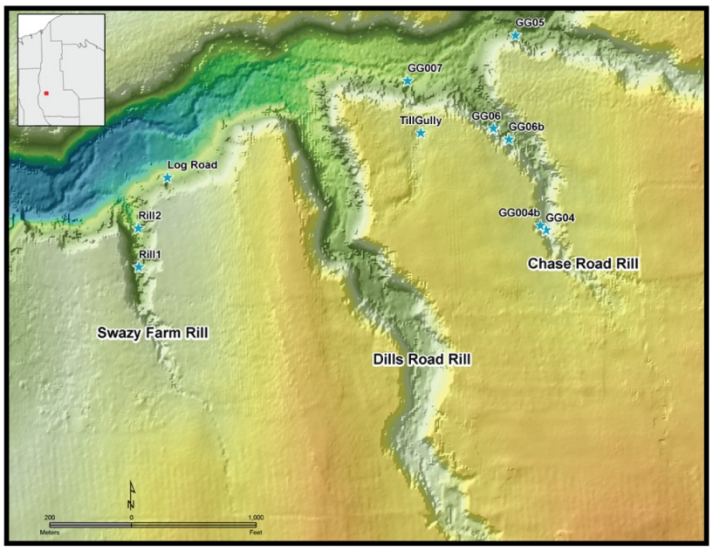




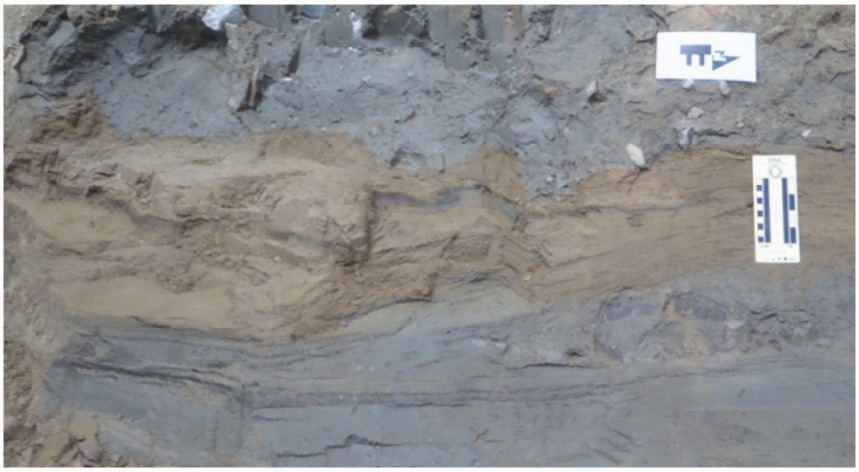
Rill 2, also known as deformation rill, is a steep, narrow rill that was difficult to maneuver due to the slippery access through numerous fallen trees. The characteristic exposure of this site is now completely covered with a large mass of roots and trees. The structural analysis is enclosed in the structural deformation section of this volume. The till color observed in this rill is likely different due to the level of chemical weathering present. The coloration of the matrix is a reddish brown in contrast to the more widely observed blue gray. The numerous joints in the till also exhibit an oxidation rind up to 10 cm thick at times. The non-diamicton sediment layers are laminated silts and sand, with a small layer of gravel.

GG04

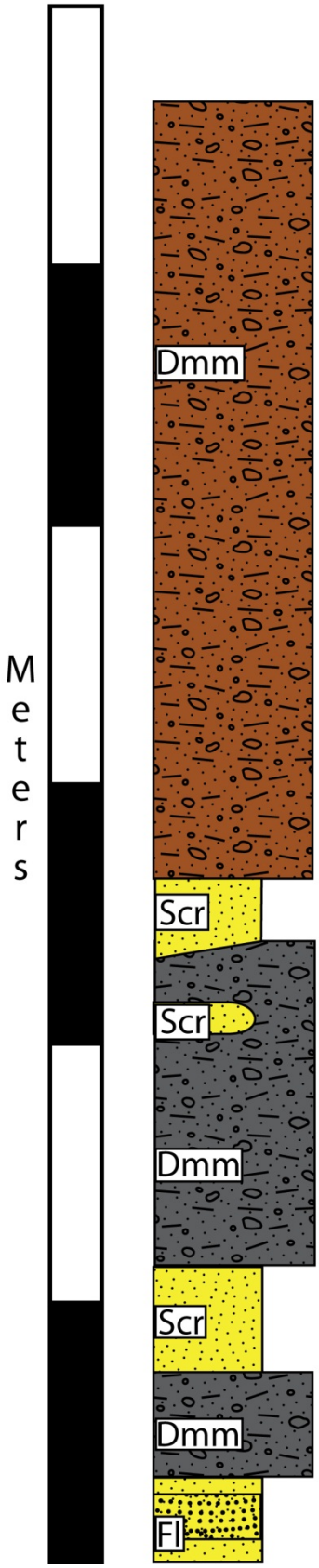


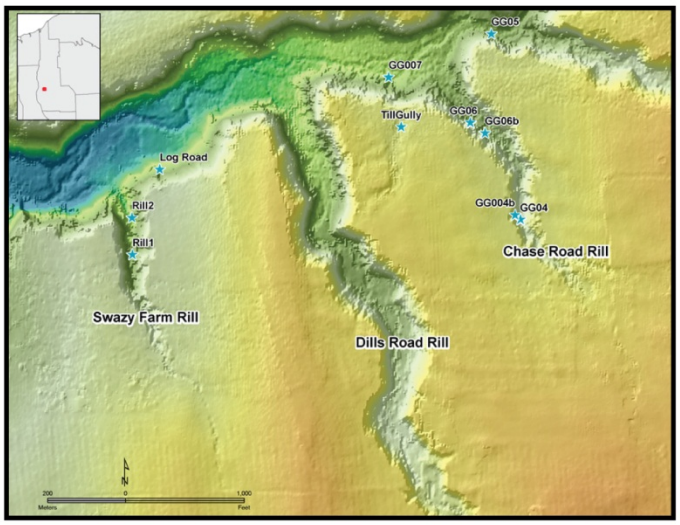


GG04 is found along the streambed of the Chase Rd Rill approximately 300 meters south of GG06. The site, now covered by a slump of tree roots, was approximately 1 meter thick. The contact between the upper till and upper sand is an abrupt, disconformity. There is shearing in the upper organic beds as observed with tight isoclinal folding. Organic layers are present in the exposure with one layer (folded) up to 1 cm thick. Organic matter is observed throughout the exposure, occurring as sheet layers or on the lee side of climbing ripples. The organic matter consists of commutated leaf mats, twigs, needles, and aquatic plants. Up to 2cm clay balls were seen in spots along with veins/dikes of clay, but not observable layers or horizontal lenses of clay were present.



GG06

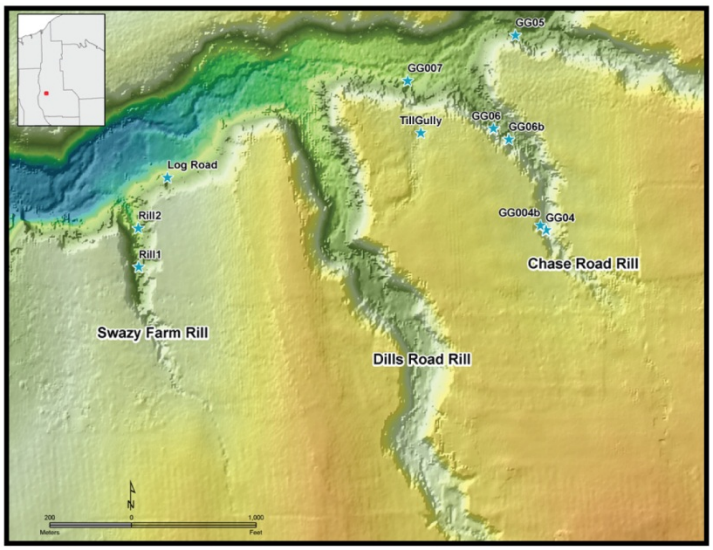


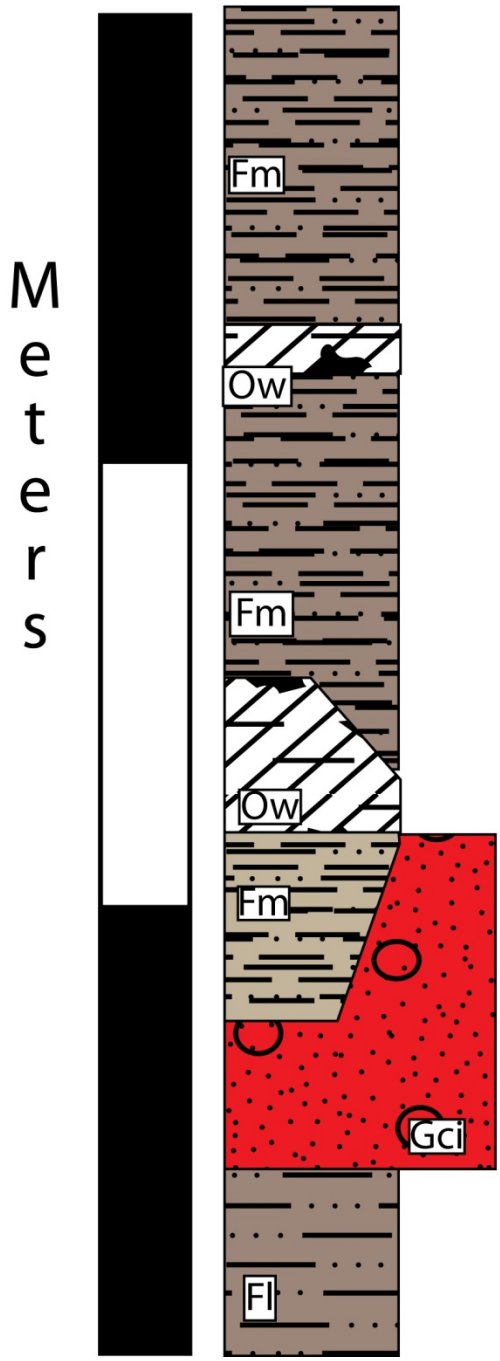


GG06 is found approximately 150 m south of the confluence of the Chase Rd Rill and the main gully. The exposure is fairly large 3.5 meters tall by ~ 15 meters wide with a recent tree root system that slumped over the center, lowest portion. Immediately below the upper-most diamicton, the upper sand contains the most plentiful organic material. The layer is fairly uniform in thickness and grain size, and the till-sand contact is a disconformity. Continuing down, the exposure distinctly becomes till with a sand lens incorporated in the diamicton. A thin layer of sand is encountered, followed by another layer of till, followed again by cross-bedded sands with organics.

Organics were most abundant on the lee side of climbing ripples of the upper most and lower most sand units (Src). The middle sand unit did not provide anything substantial in size to radiocarbon date. OSL samples were collected and the exposure was photographically archived with a Gigapan image. Folding and jointing are present in the exposure,. The base of the exposure is at creek level, and till is not directly observed at this level. Future exploration is planned to look for the sand-till contact.

GG007





This site was recognized by flattened 1-4 cm diameter sticks and branches protruding out of a stream cut. No till was observed at this location as the river bottom consisted of an over-consolidated silt layer. The upper 2 meter mass is an uncharacteristic loose, “gooey” silt with 2-10 cm bands of organics. This massive silt was draped over an imbricated, lightly cemented gravel unit that appeared bare of organic material. This gravel overlies dense, laminated silt that contained charcoal. AMS radiocarbon dates were assessed on the branches and charcoal, and OSL dates were assessed on the laminated and massive silt. The laminated basal silt was over consolidated and difficult to sample by pounding a 1.5 inch, sharpened steel conduit tube, in contrast to the massive silt that could be sampled by simply pushing the tube into the unit

Structural Deformation of Glaciogenic Sediments in Great Gully

Brian Bird

Introduction

Examination of the stream cut exposures within Great Gully as part of an ongoing mapping program revealed numerous sites exhibiting deformation of the glacial sediments. These geologic structures were examined at seven outcrops in Great Gully and included; folds, faults, clast fabric, jointing, ripple crest and clast striations in various sediment assemblages. A Brunton compass was used to gather information about the structures and subsequent analyses were conducted with the program Stereonet version 7.2.0 by Richard Almendinger of Cornell University. Detailed structural analysis was used to characterize the nature and direction of the stresses responsible for the observed deformation. The steep gully terrain is prone to mass wasting. It was of upmost importance to confirm the exposures were in-situ. Upon scrutiny of the exposures and structural analysis of the deformation, the sites do not exhibit structures consistent with mass wasting of the steep slopes into the gully. There is an overall lack of geomorphic forms typical of mass wasting such as slumping, rotation or creeping below the root zone. Stress directions associated with the deformation are consistent with advancing glacial ice and independent of the orientation of the exposure face, as would be expected with mass wasting.

Glacial Till in Great Gully

Three distinct diamicton units are observed in Great Gully. These diamictons are interpreted to be subglacial tills deposited as the Ontario Lobe of the Laurentide Ice Sheet flowed over Great Gully and each likely represents multiple modes of formation. Chronology indicates the Gully was an open, active system prior to the advance to the last glacial maximum; thus ice advanced into and over an open gully. As ice advanced across the open gully some basal debris could have fallen away from the ice into the open cavity (Boulton, 1982). Exemplifying this process is the middle till at GG06, as it forms a disconformable contact with older, stratified sediments. As ice continued to advance, the cavity filled with more debris and ultimately to the point where ice was in direct contact with the bed and forces exerted by the ice deformed the sediment. The initial process of debris falling into the cavity would describe a subglacial melt-out till, however, the subsequent coupling of the till along with the highly deformed sand to the ice would define the middle till as a glacitectonite as defined by Evens et. al.(2006). The upper till in Great Gully can be described as a subglacial traction till (Evens et. al. 2006) where the unit has been largely homogenized by shearing and has incorporated and deformed the underlying sediments. GG06 shows a large deformed inclusion of the lower stratified sediments (Figure 25).

Brief Overview Glaciotectonic Deformation

The deformation of glacial sediment consists of two basic types, brittle or ductile. Structures such as faults and joints are the result of brittle deformation whereas folds are ductile. The orientation of the principal stress direction can be used to reconstruct stress fields or patterns responsible for the

observed strain (Figure 26). According to McCarroll and Rijsdijk (2003), in order to infer the cause of a particular deformation structure, the inference must be based on the structural analysis as well as the reconstruction of the type of stress field necessary to form them.

Brittle deformation occurs when materials fail along a fault or fracture. Normal faults are the result of extensional forces that will generally elongate the overall package of deformed material. Reverse faulting is opposite with the forces involved being compressional and causing an overall shortening of the deformed package. The extensional or compressive nature helps define the overall stress regime of the observed deformation.

Ductile deformation can be divided into two sub-categories: uniform or non-uniform. Non-uniform deformation as a whole is complex, but can be broken down into small, fundamental parts that can be viewed as uniform. Uniform ductile deformation occurs in two forms, pure shear and simple shear. In a ductile material, an imaginary circle will deform into an ellipsoid shape as the result of stresses applied. Pure shear and simple shear will create the ellipsoid by different means. Pure shear is the flattening of the strain ellipse without rotation. In this case, the principal stress direction is perpendicular to the long axis of the strain ellipse. An example of pure shear is loading. The force of gravity acting vertically on a ductile material will cause vertical flattening and horizontal extension. Simple shear is when opposing forces move past each other and results in a rotated strain ellipse. In this case the principal stress direction will be oblique to the strain ellipse long axis. Simple shear will impart a rotational component to the strained material.

Each type of deformation can be expressed as glacial ice passes over an area. Beneath the interior of the glacier the principal stress direction is caused by loading and is straight down. Toward the ice margin, the stresses equalizes to a hydrostatic state where all directions are equal. Near the ice margin sequences are likely to be dominated by large-scale compressional deformation structures with the principal stress axes nearly parallel to the direction of ice movement (Figure 27).

Observed Features

Jointing

The Hamilton Group Shales outcrop in Great Gully Creek as well as isolated locations near the study area. This shale bedrock in the area is heavily jointed. Regional joint orientation has been mapped to trend 344° and 060° (Engelder and Geiser, 1980; Jacobi, 2002). In and around Great Gully the bedrock jointing was measured at six locations. The bedrock jointing planes align with the regional jointing across New York State (Figure 28). Jointing has also been consistently observed in the exposures of glacial sediments along Great Gully. Stratigraphically the jointing typically occurs lower in the section and is observed across a variety of lithostratigraphic units including; diamicton, sand, and silt. At each location the poles to the joint surfaces were plotted (Figure 29).

Folds

Folding of the sediments is observed at all sites in Great Gully. These folds range from centimeter scale to meter scale folds. The folds show a wide range of types from simple upright folds to highly contorted sheath folds (Figure 30). The stress direction calculated from the trend and plunge of the fold axes ranges from and azimuth of 330º to 351º. These data are consistent with forces exerted by the ice moving from a north-northwest direction. Figure 31 shows a plot of all fold axes across the study area. The best fit great circle delineates the shear plane that can account for theses folds. This shear plane would approximate the attitude of the base of the ice and is consistent with the attitude of the base of the till.

Faulting

Evidence of faulting can be found in sites; GG06, Till Gully, Log Road, and Rill 1. These faults are all compressional in nature and can be described as reverse faults or thrust faults when low-angle. Faulted sediments included the Upper Diamicton at GG06, Upper Sand layer at the Log Road Site and Till Gully Site, and highly deformed silt and clay unit beneath a middle diamicton in Rill 1. The faults at the Log Road Site were typical of the area (Figure 32). The poles to the slip surfaces were plotted and the resulting best fit great circle defines a plane that contains the $\sigma 1$ and $\sigma 3$ directions for the faults. This plane is defined as striking 109º dipping 85º SW and indicates the general vector of the principle stresses creating the faults (Figure 33).

Fabric and Striations

Diamicton has been observed at each site in the study area and is inferred to be a basal till. The fabric for each diamicton has been measured across the area. The fabric data was collected by measuring the trend orientation of the elongated axis as well as the plunge from horizontal of that axis. Distributed within the diamictons were numerous large clasts, often black limestone (Ordovician?), and are heavily striated. The fabric and striae data yield information about the flow direction of the glacial till (Figure 34).

Flow indicators

Climbing ripples and imbricated gravel was observed in the eastern portion of the study area. The orientation of these structures were measured at sites; GG04, GG06, and GG07 (Figure 35 and 36) and provide data that allow reconstruction of stream flow conditions at the time these structures were formed.

Interpretation and Discussion

The analysis of structural data demonstrates that Great Gully was affected by dynamic glacial ice and stream flow into a lake. The stratigraphy suggests glacial ice advanced over the gully multiple times, with a fluvial and/or lacustrine environment between the advances. As glaciers advanced over the gully the sediments present were overridden and deformed.

Till clast fabric data indicate the motion of the base of the ice as it crossed the southern rim of the gully. As displayed by the drumlin swarm to the north, ice flow was north-northwest. The till clast fabric of the western sites, Rill 1, Rill 2, and Log Road, indicate a consistent flow direction trending 165°, basically parallel to 163° indicated by the drumlin swarm to the north. These sites, near the southern rim of the gully, would have been influenced by ice traveling nearly perpendicular to the main gully thus stress directions would be directed in line with the advancing ice. Eastern sites, GG06 and GG04 deviate from this with flow trending 257°. This deviation from the general trend of ice flow is attributed to the ice encountering a smaller tributary trending at an angle to ice flow. Ice flow at an angle over the open tributary would impart localized flow oblique to the general ice flow direction. Jointing commonly occurs at the sites in different strata including; till, fluvial, and lacustrine sediments. The jointing is confined mostly in the lower units and is attributed to the advance of ice across the gully, over pre-existing sediments. The general lack of jointing in the uppermost till at the Log Road, Rill 1 and Rill 2 indicates this till was most likely the youngest till and not subjected to the same stress regime as the underlying sediments. When plotted, the poles to joint surfaces for each site consistently align with the bedrock orientation, similar to observations made by Godin et al. (2002).

Fold and fault data across the sites is more consistent across the sites than the clast fabric data. Principal stress directions range from 195° to 123°, with an average trend of 148° for five locations. This consistency with ice flow direction confirms the fold and fault deformation is associated with the advance of ice over the gully and not the result of slumping or sliding of the exposures. The ice advancing over the gully would have coupled with the bed thus shear forces were translated into the substrate. Deformation type would depend upon grain size of the sediment and water content within the sediment. Water content is crucial as if there is free water between the base of the ice and the substrate, shear stress becomes zero and no force is translated to the substrate.

Additional evidence to the preexisting nature of the gully is flow indicators such as climbing ripples. Ripples at GG04 and GG06 are characterized as Type B climbing ripples (Allen, 1973), flow direction is perpendicular to the ripple crest in the direction of the climb angle. At GG07 the flow was inferred from imbricated clast. The strike and dip of the imbrication was measured, with the dip direction indicating upstream. Flow directions were collected on fluvial sediments that were under till. At each site, the paleo-flow is coincident with the modern stream valleys. While this alone is not proof of the gully existing prior to the most recent glaciation, chronology of sand dated by optically stimulated luminescence and radiocarbon dating on wood found in the ripples indicate that these sediments were deposited before the last glacial maximum.

Figure 25: Deformed Inclusion of Sand in Upper Till at Site GG06

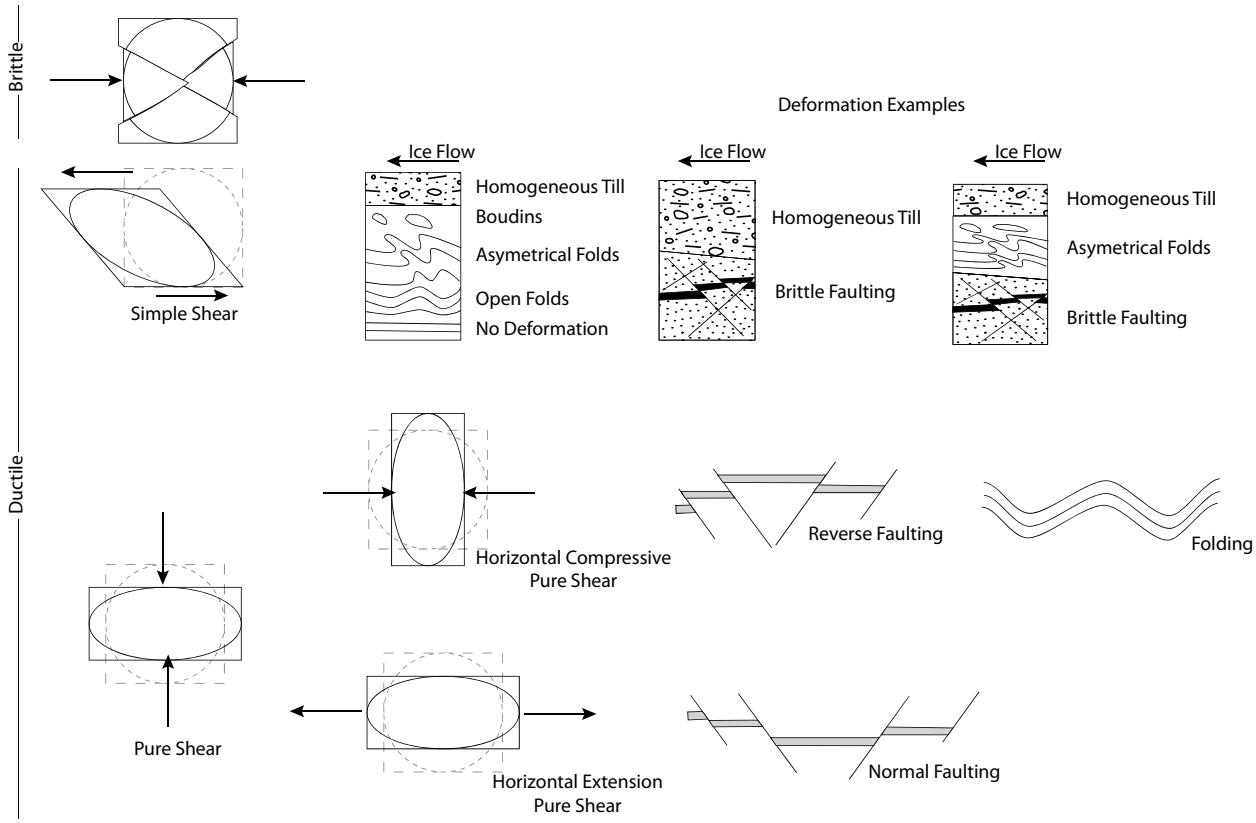


Figure 26: Types of Deformation (modified from McCarroll and Rijsdijk 2003)

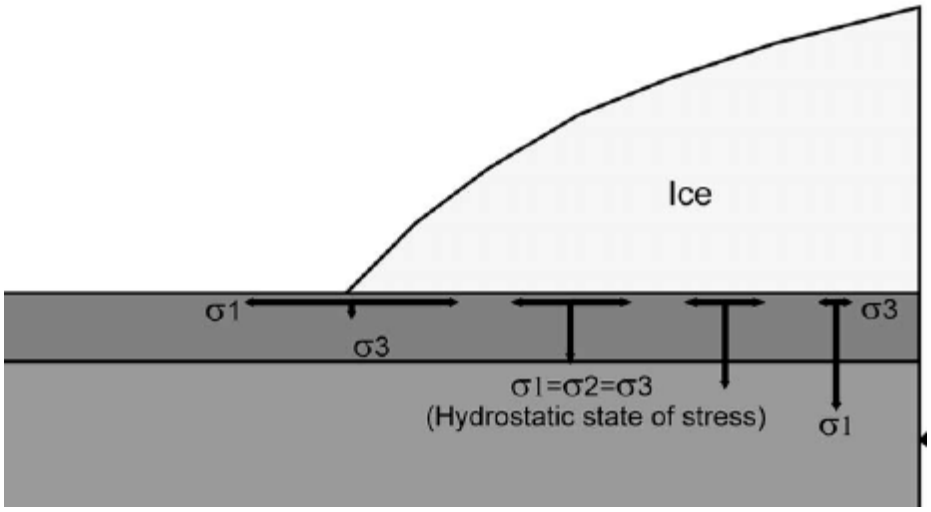


Figure 27: Net Stress Exerted on the Substrate by Loading of a Glacier (Modified from Andersen et al., 2005)

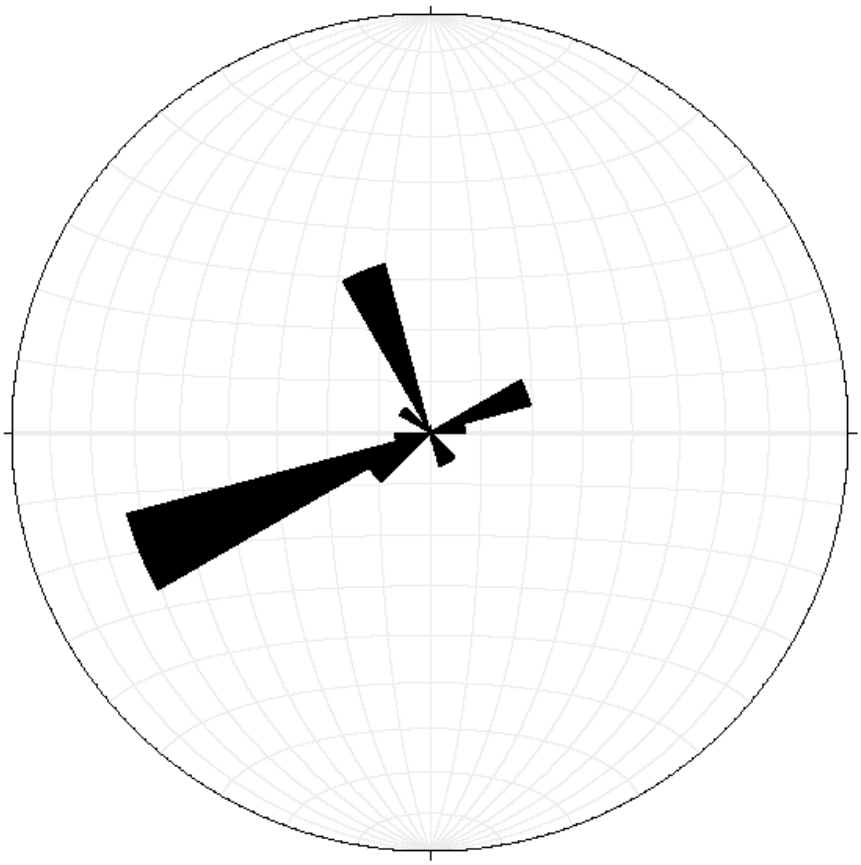


Figure 28: Jointing in the Bedrock near Great Gully

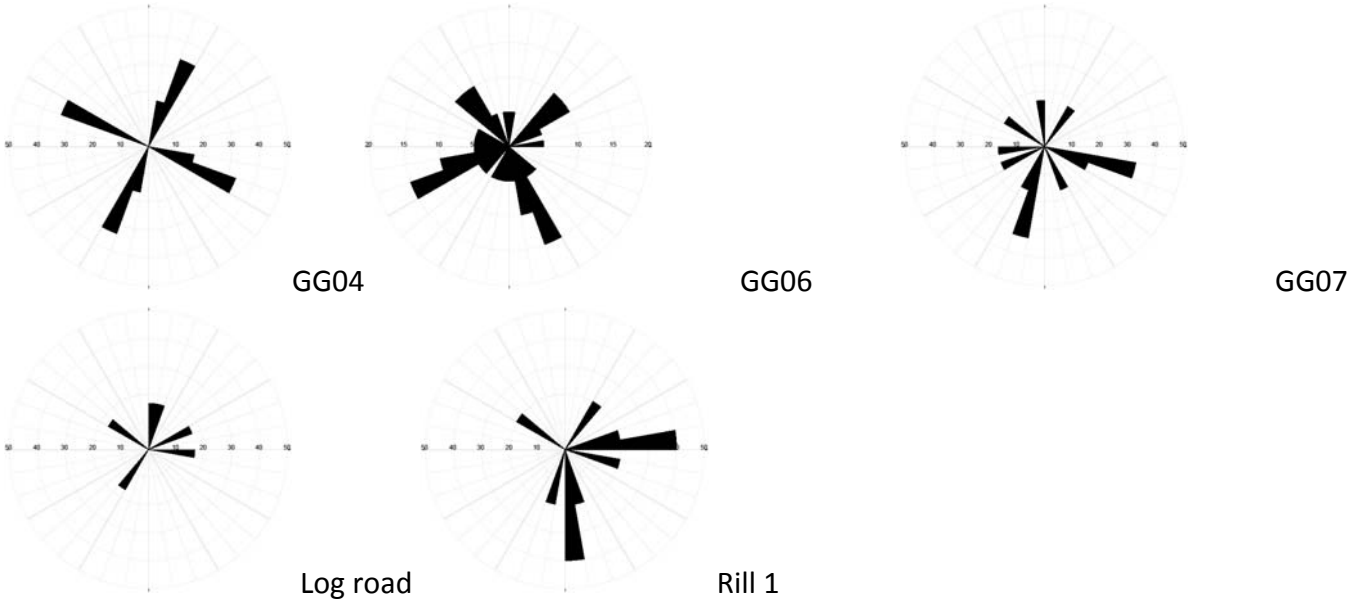


Figure 29: Jointing observed at each site

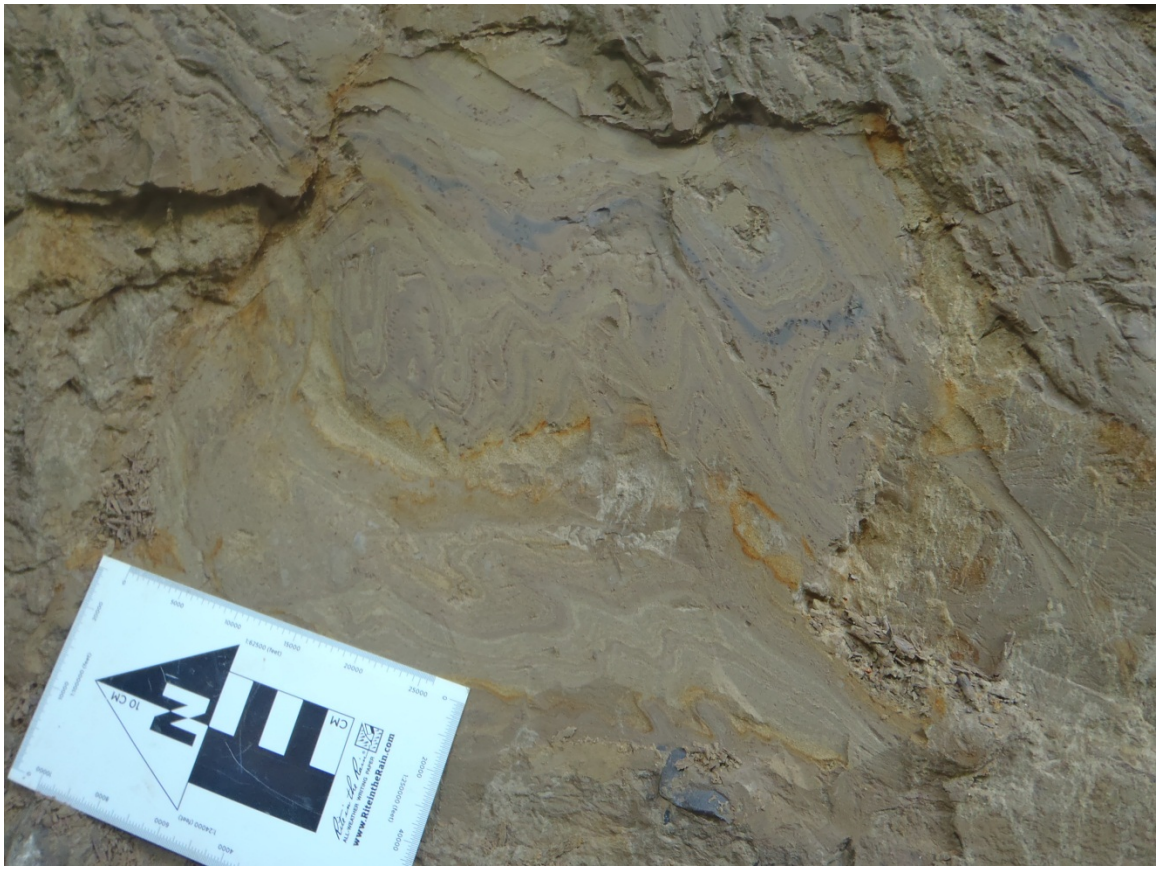


Figure 30: Folding of Lacustrine Sediments Below Till

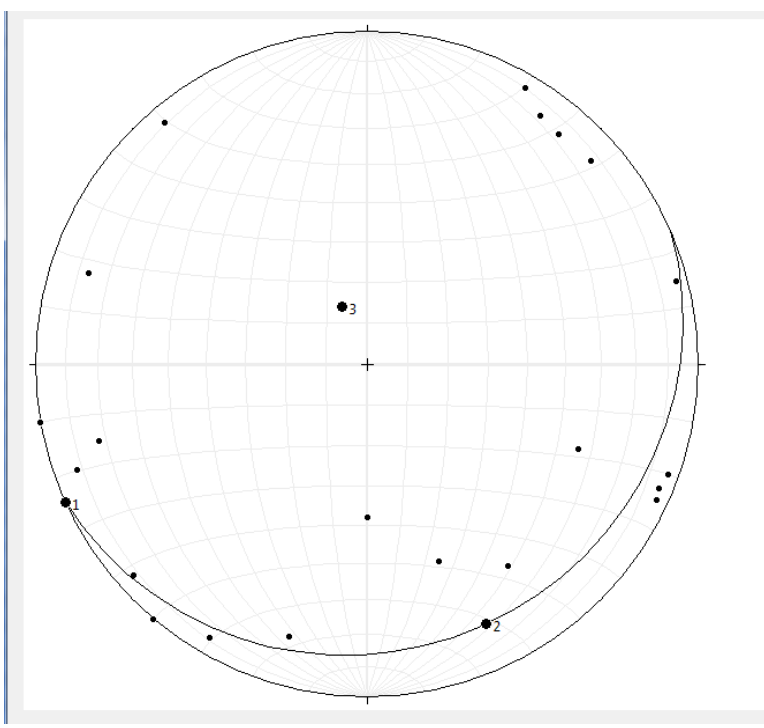


Figure 31: Fold Axes of Folds from all Sites

Figure 32: Faulting of Sand below Upper Till

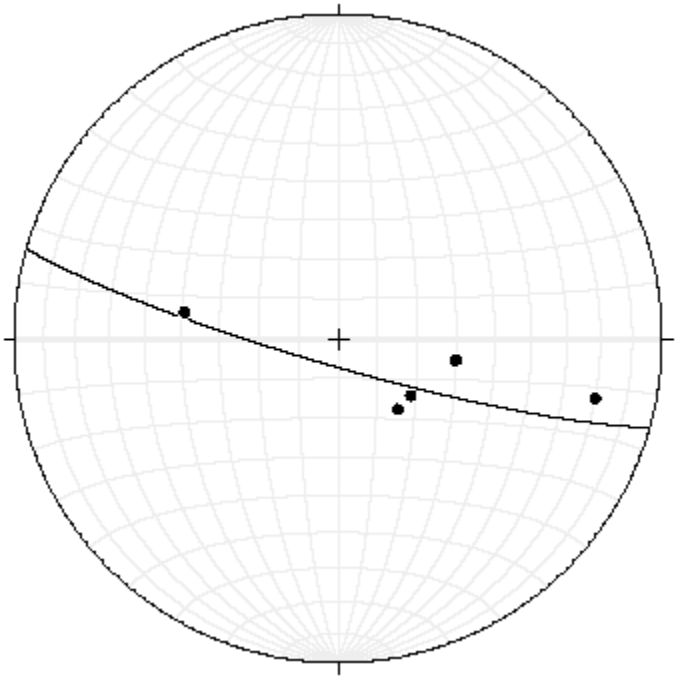


Figure 33: Stereonet of Poles to Faults at Log Road

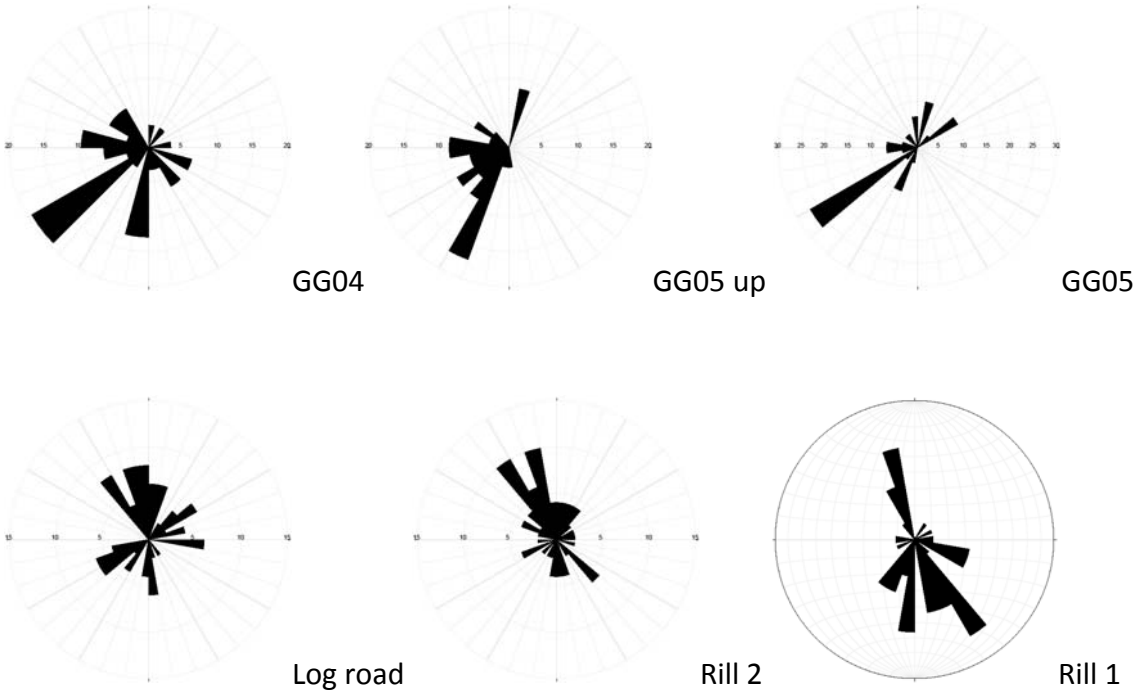


Figure 34: Till Clast Fabric Rose Diagrams

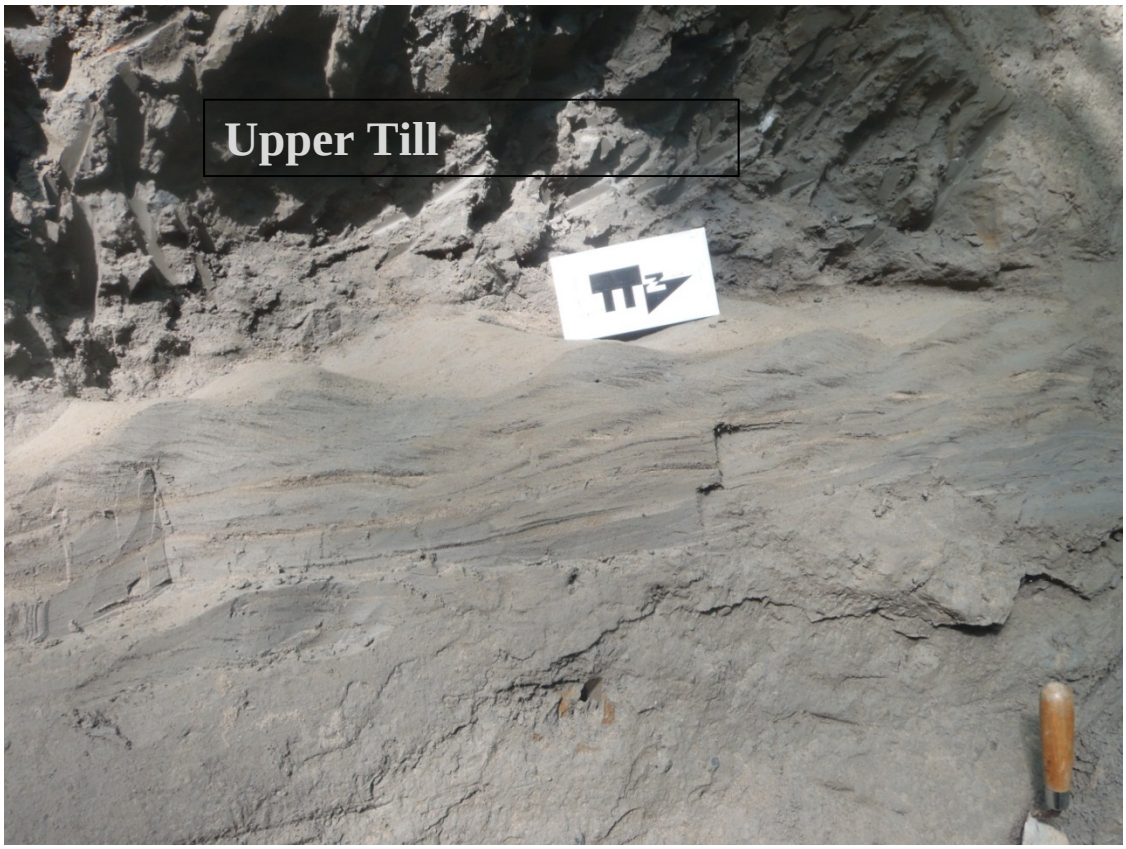


Figure 35: Climbing Ripples at GG06

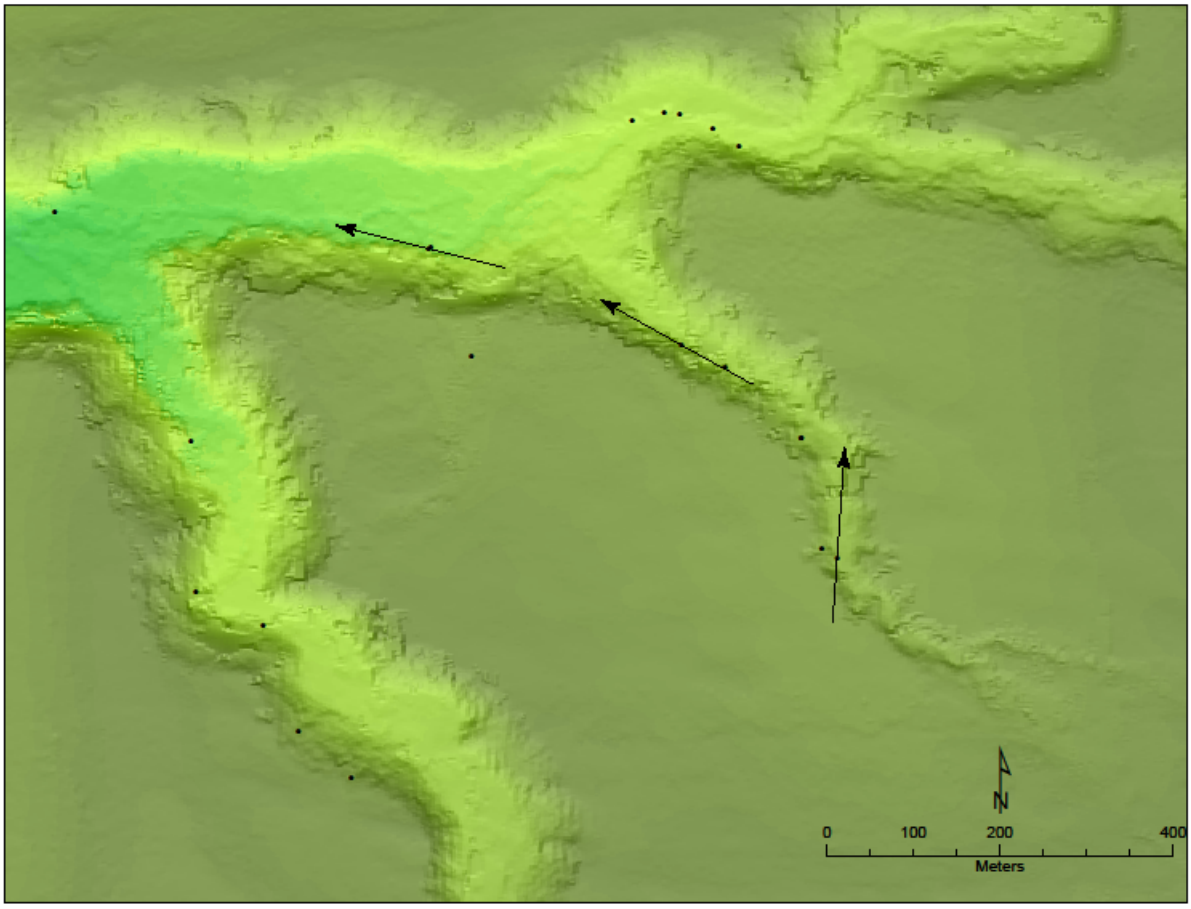


Figure 36: Paleo-flow Directions of sub-till flow indicators (ripples, imbrication)

Great Gully Radiocarbon Results

Andrew Kozlowski, Brian Bird, Brandon Graham

In total we have run more than 20 radiocarbon dates from deposits within Great Gully. The majority of dates are from organics (wood & plant macrofossils) contained within various cross-bedded sand deposits underlying or in between diamicton (till) units. Of these only one date has come back within the range of calibration. All others consistently indicate that deposits in question are greater than 50,000 years old (see figures 39-41, Table 3).

The one radiocarbon age with usable error bars comes from a date on twigs recovered in deformed (folded) sandy silt underlying the upper most diamicton (surface till) at site GG04B, ~1.5 meters below the surface elevation (level of the field). The radiocarbon age of 42,800 ± 2000 yr BP (43,130 - 49,740 Cal yr BP) over laps with another date on sample (twigs) that comes back as beyond calibration, and strongly suggests that at a minimum the sediments are Midwisconsin in age.

Date Reported	Identification	Type	F Modern	Fm Err	Age	Age Err	Δ13C	Δ13C Source
9/20/2012	GG04_Lower sand_S1	Plant/Wood	0.00210	0.00250	> 43300	0	-27.68	MEASURED
9/20/2012	GG04_Lower sand_S2	Plant/Wood	0.00230	0.00250	> 42900	0	-27.37	MEASURED
9/20/2012	GG04_Lower sand_S5	Plant/Wood	0.00430	0.00260	43800	4900	-27.59	MEASURED
9/20/2012	GG04_Upper sand_S1	Plant/Wood	0.00010	0.00250	> 47700	0	-25.02	MEASURED
9/20/2012	GG04_Upper sand_S5	Plant/Wood	0.00160	0.00250	> 44100	0	-28.42	MEASURED
9/20/2012	GG06_LSU1_S1	Plant/Wood	-0.00130	0.00250	> 48000	0	-24.14	MEASURED
9/20/2012	GG06_LSU1_S2	Plant/Wood	-0.00200	0.00250	> 48000	0	-25.74	MEASURED
9/20/2012	GG06_LSU1_S3	Plant/Wood	-0.00160	0.00250	> 48000	0	-27.74	MEASURED
9/17/2012	GG06_LSU3_S1	Plant/Wood	-0.00150	0.00260	> 47800	0	-26.37	MEASURED
9/17/2012	GG06_LSU3_S2	Plant/Wood	-0.00130	0.00260	> 47800	0	-26.3	MEASURED
9/20/2012	GG06_LSU3_S3	Plant/Wood	-0.00140	0.00250	> 48000	0	-27.3	MEASURED
10/18/2012	GG04_Upper Sand_S7	Plant/Wood	0.00230	0.00160	> 48000	5500	-27.53	MEASURED
10/26/2012	GG04B_S1	Twig	0.0048	0.0012	42800	2000	-27.8	MEASURED
10/26/2012	GG04B_S3	Twig	0.0027	0.0012	47500	3600	-27	MEASURED
10/26/2012	GG007_#1 0cm	Charcoal	0.0018	0.0012	>44000	0	-25.2	MEASURED
10/26/2012	GG007_#2 55cm	Wood	0.8086	0.0019	1705	20	-28.7	MEASURED
10/26/2012	GG007_#3 105cm	Wood	0.8172	0.0018	1620	20	-27.3	MEASURED
10/26/2012	GG007_#4 105cm	Wood	0.8159	0.0018	1635	20	-25.9	MEASURED
10/26/2012	GG007_#5 140cm	Wood	0.8142	0.0018	1650	20	-28.5	MEASURED
10/26/2012	GGP-Sol_charcoal_S1	Charcoal	0.8726	0.0019	1095	20	-26.7	MEASURED

Table 3. Radiocarbon Ages obtained for Great Gully

Although the abundance of sub-till radiocarbon ages on reliable materials (wood) are robust, internally consistent and demonstrate that older beds appear as laterally continuous facies throughout the gully system, alone they do not provide a chronologic context for the middle and lower tills units. Hence our frustrations with too many organics (everyone should have this problem) and not enough information provided the catalyst to seek out alternative dating methods that have an extended age range. It was to this end was that we sought out Optically Stimulated Luminescence (OSL) Dating.

Optically Stimulated Luminescence age estimates were obtained for the organic rich fluvial lacustrine sands subcropping tills observed at sites GG04, GG06, massive fine sand silt at GG03 and a basal silt unit at GG07 (OSL analysis by Shannon Mahan Lab Director at the Luminescence Geochronology, USGS, Denver). OSL dating measures the length of time since sediments were last exposed to sunlight (thus providing and age of burial). Upon transport and exposure to sunlight sediment grains “reset” themselves, the luminescence signal accumulates in mineral grains (quartz was used here). The amount of luminescence acquired by the sediment is related to the length of burial and chemistry of the surrounding sediment. We utilized Single Aliquot Regeneration (SAR) methods, measured Quartz OSL ages not feldspar, with emphasis on the lowest dose equivalents. The Quartz samples were not saturated and the limit for these sediments was approximately 180-200 gray.

The OSL ages and additional data are presented in Table 4. The ages presented indicate the fluvial sand beds sampled are 56,000 – 73,000 (Mid-Early Wisconsin) in age. In particular the data and observations from the Chase Road rill strongly suggest that a record of early Wisconsin glacial events is well preserved within Great Gully. The traceable, laterally continuous cross-bedded organic sand beds observed from sites GG04, GG06 and GG07 indicate a subaerial flow system flowing into a high

TABLE 4: Quartz OSL Data and Ages

Sample information	% Water content[a]	K (%)[b]	U (ppm)[b]	Th (ppm)[b]	Cosmic dose[c] (Gy/ka)	Total Dose Rate (Gy/ka)	Equivalent Dose (Gy)	n[d]	Over[e] dispersion (%)	Age (yrs)[f]
GG03 OSL 7	13 (19)	1.22 ± 0.04	1.19 ± 0.13	3.98 ± 0.20	0.17 ± 0.01	1.78 ± 0.07	99.8 ± 2.43	5 (25)	31 ± 3	56,120 ± 2,580
GG04 OSL 6	16 (35)	1.95 ± 0.06	1.61 ± 0.04	6.04 ± 0.30	0.08 ± 0.01	2.33 ± 0.08	155 ± 3.41	8 (25)	19 ± 2	66,640 ± 2,620
GG06 OSL5	15 (29)	1.54 ± 0.05	1.41 ± 0.04	4.70 ± 0.23	0.03 ± 0.01	1.90 ± 0.07	135 ± 3.92	8 (20)	26 ± 2	70,940 ± 3,220
GG07 OSL3	13 (35)	2.07 ± 0.06	1.65 ± 0.04	7.93 ± 0.39	0.02 ± 0.01	2.49 ± 0.08	90.2 ± 2.98	8 (35)	31 ± 3	36,100 ± 1,640
GG07 OSL4	3 (33)	1.74 ± 0.05	1.44 ± 0.04	5.51 ± 0.28	0.02 ± 0.01	2.06 ± 0.07	151 ± 3.47	12 (25)	31 ± 3	73,370 ± 3,080

[a]Field moisture, with figures in parentheses indicating the complete sample saturation %. Ages calculated using approximately 60% of total saturation values.

[b]Analyses obtained using inductively coupled plasma mass spectrometry (ICP-MS). All errors were obtained with calibration standards.

[c]Cosmic doses and attenuation with depth were calculated using the methods of Prescott and Hutton (1994). See text for details.

[d]Number of replicated equivalent dose (De) estimates used to calculate the equivalent dose. Figures in parentheses indicate total number of measurements included in calculating the represented equivalent dose and age using radial plots (weighed mean).

[e]Values are a reflection beyond instrumental error, values below 25% indicate low dispersion, most samples show strong bi-modal distribution.

[f]Dose rate and age for fine-grained 250-90 or 125-90 microns sized quartz. Exponential + linear fit used on equivalent dose, errors to one sigma.

standing lake in the early Wisconsin. Plaeocurrent reconstructions in conjunction with abundant and consistent radiocarbon "dead" wood and plant macrofossils add to validate this hypothesis. An OSL age of 36,100 ± 1,640 (sample GG07 OSL3) represents an outlier to the otherwise older deposits observed, this sample came from an exposure on an active slump, further large wood fragments and branches incorporated into the matrix sampled yielded consistent radiocarbon ages of ~1,600 ^{14}C yr BP indicating that mixing of sediments along the shear plane of the slump occurred. It is our present hypothesis that this Mid-Wisconsin date possibly records the initiation date of the original slope failure.

Another OSL sample on a sub-till sand silt comes from beneath the surface till at the Log Road site from the Swayze Farm Rill locality. Here the OSL estimate yielded an age of 56,120 ± 2,580 (GG03 OSL 7) adding further credence to the presence of Mid-Wisconsin fluvial and lacustrine sediments in the near surface.

Discussion – Implications and Regional Correlation

The internal consistencies of the radiocarbon ages combined with the OSL results strongly suggest that age of the organic rich cross-bedded sands are early Wisconsin in age, particularly in the case of exposures observed in the Chase Road Rill. Early Wisconsin deposits have not been previously identified in the Finger Lakes Region and only a handful of sites (less than 5) in New York State. If this assessment is correct then it implies that the underlying till represents an older ice advance into the Northern Finger Lakes Region of either early Wisconsin or Illinoian age.

Diamicton deposits interpreted as tills representative of early Wisconsin glaciations have been observed in Ontario (Karrow, 1967, 1969; Hicock & Dreimanis, 1989; 1992) and additionally along the north shore of Lake Erie proglacial lake sediments (Member B of the Tyrconnell Formation), were assigned to the early Wisconsin by Dreimanis, (1972, 1992). Calkin et al. (1982), described an anomalous "brown" till near Gowanda, NY (western, NY), Calkin tentatively assigned this "brown" till to a middle Wisconsin age and correlated this to the Titusville Till in northwestern Pennsylvania. Fullerton (1986) while in agreement with the lithostratigraphic correlation between the two tills assigned the age of both till deposits as early Wisconsin. Szabo (1992) later refuted the Middle and Earlier Wisconsin assignments of the Titusville till as did Dreimanis (1992). Dreimanis (1992) hypothesized that the "brown" till at Gowanda Hospital site represents a near terminal position of an early Wisconsin advance of the Ontario Lobe.

Hicock and Dreimanis (1992) reaffirmed earlier interpretations of Karrow (1967, 1969, 1974) and Dreimanis and Karrow (1972) that the Sunybrook drift in the Toronto Region represents an ice advance (Sunnybrook till) of Laurentide Ice through the Ontario Basin during the Guildwood Phase of the Ontario Subepisode in the early Wisconsin (Karrow et al. 2000). Dreimanis (1992) in his reinterpretation of the age of deposits along the north Shore of Lake Erie correlated the "brown" till of

Calkin et al., (1982) to the Sunnybrook drift at Toronto, reinforcing thereby delimiting the western extension of early Wisconsin ice advance.

It is our hope that the age control and stratigraphic relationships observed and presented here will serve as long sought after correlative to convey the geographic extent of Early Wisconsin glaciation in CNY (Figure 37).

Given the age control of the overlying cross-bedded fluvial sands (70,940+\- 3,220 Cal yr BP) at site GG06 the middle till in the sequence must be older than middle Wisconsin. Based on observed stratigraphic relationships and age control we assign the middle till observed in GG06 as being an Early Wisconsin till deposit (Sunnybrook till equivalent?). This assignment is partially based on the observation of a lower till exposed in the floor of the primary (Great Gully) valley that must predate the middle till of GG06. While we do not know the exact age of the lower most till, it seems likely that it must be older than 73,370 ± 3,080 and possibly is correlative to the York Till of Ontario.

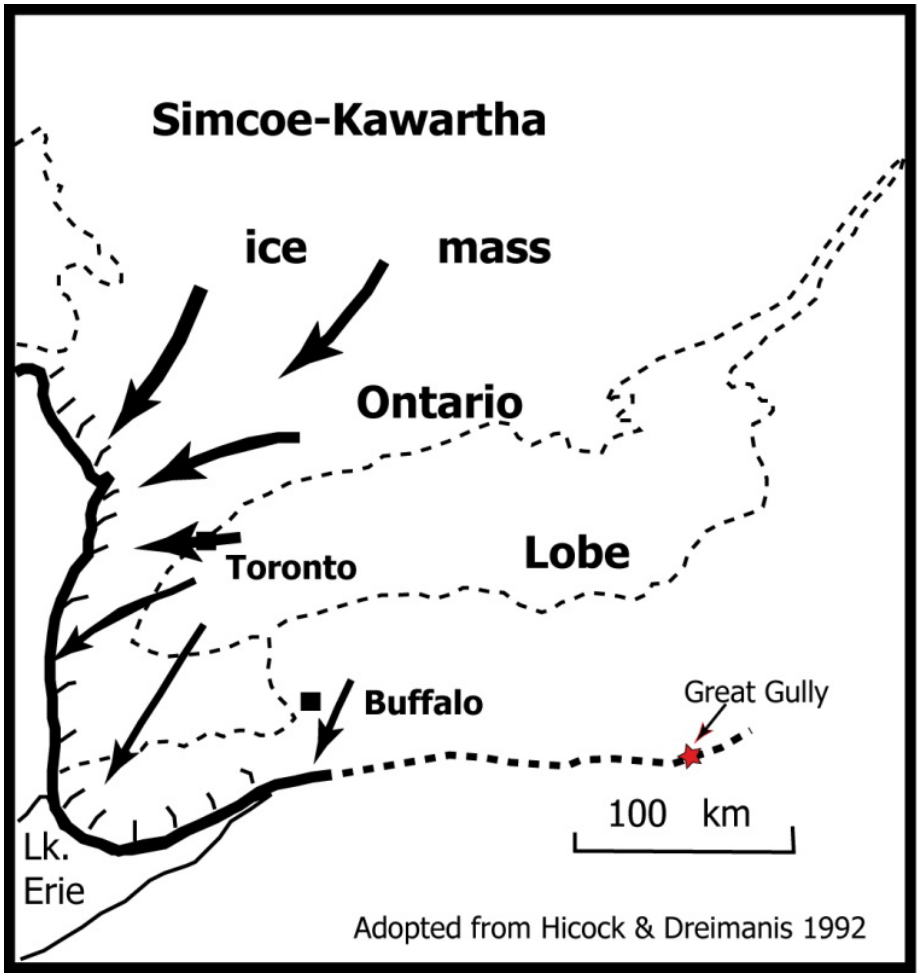


Figure 37. Ice margin map of Eastern Great Lakes depicting extent of Ontario Lobe during deposition of the Early Wisconsin Sunny Point till and proposed correlative in Great Gully.

Other Local Regional Correlations--??

Karrow et al. (2009) completed a thorough study of the Fernbank Site near Ithaca, NY that followed up on earlier studies such as Maury's (1908), Bloom (1967), Bloom and McAndrews (1972 NEFOP), and Karrow (2004). The most recent study (Karrow et al., 2009) thoroughly documents existing chronology, palynology and paleontology (molluscs, ostracodes, insects, vertebrates, plants) and provides supporting evidence to suggest that deposits preserved at Fernbank are equivalent to the well documented interglacial deposits (Don Beds) in Toronto. The previous chronologic constraints of the Fernbank site include three minimum ^{14}C yr BP dates > 50,000 and a single thermoluminescence (TL) date of 81± 11 ka. However, Karrow et al. (2009) suggest that due to the "anomalous fading" effect the TL date may actually be in the range MIS5e (130 ka) and equivalent to the Sangamonian Interglacial. Given the new OSL dates from our sites in Great Gully additional chronologic constraints at Fernbank would be helpful to evaluate any potential correlation between the two sites. Additionally the results of the ongoing (initiating) paleoecologic investigation underway should help to further refine our understanding of the Great Gully site and to develop a more solid correlation between Fernbank.

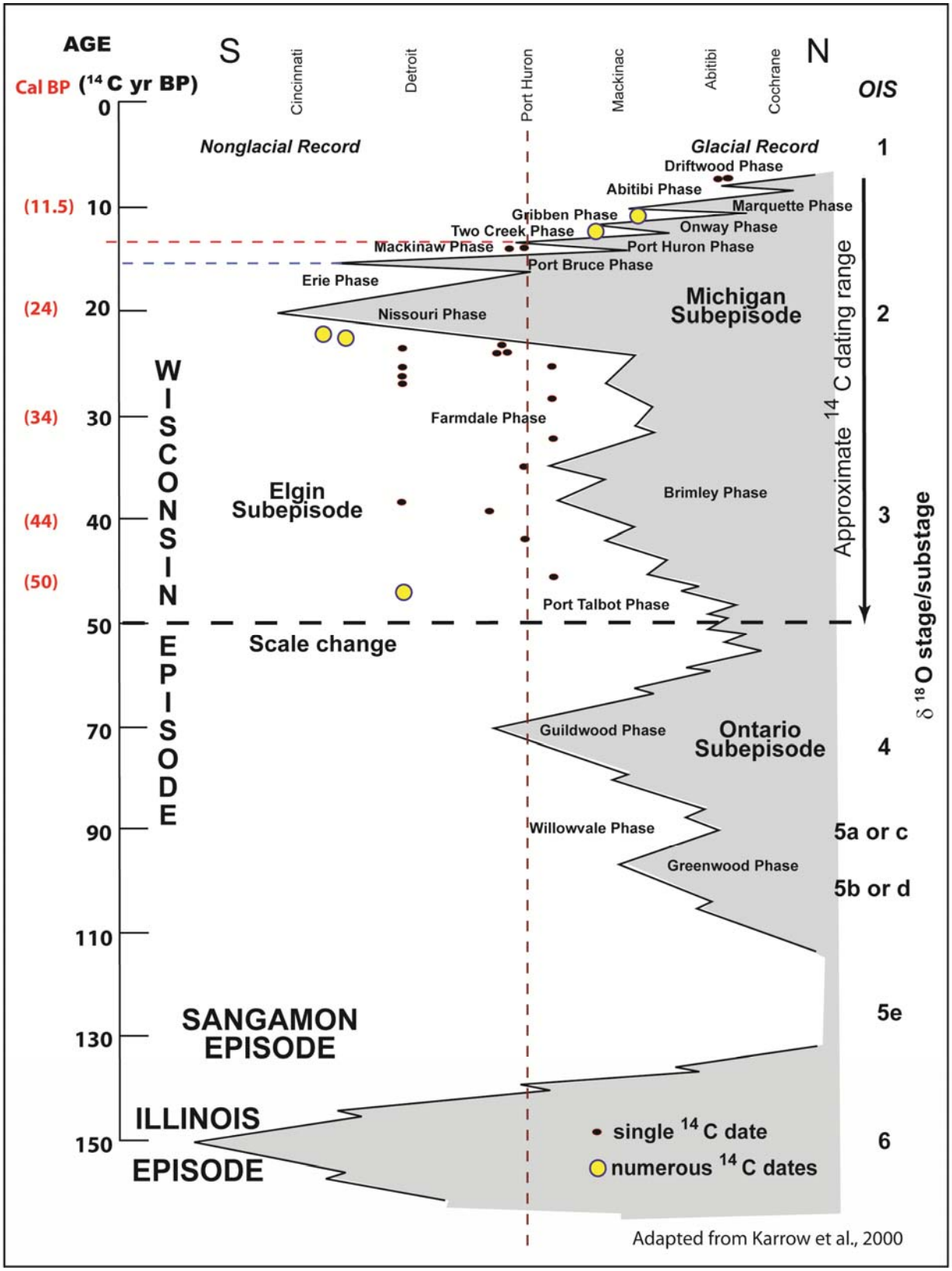


Figure 38. Time-Distance graph outlining Great Gully latitude.

Figure 39. Site Stratigraphy with GG04 OSL age

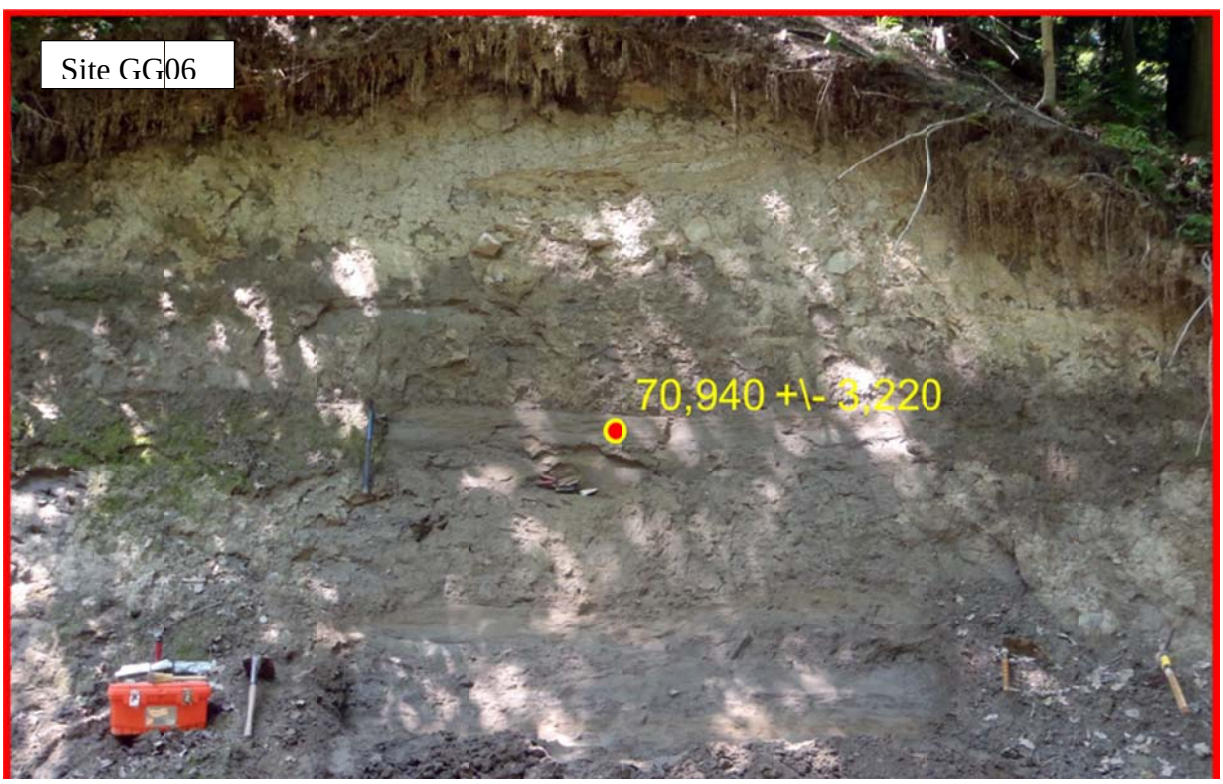


Figure 40. GG06 Site Stratigraphy with OSL age estimates

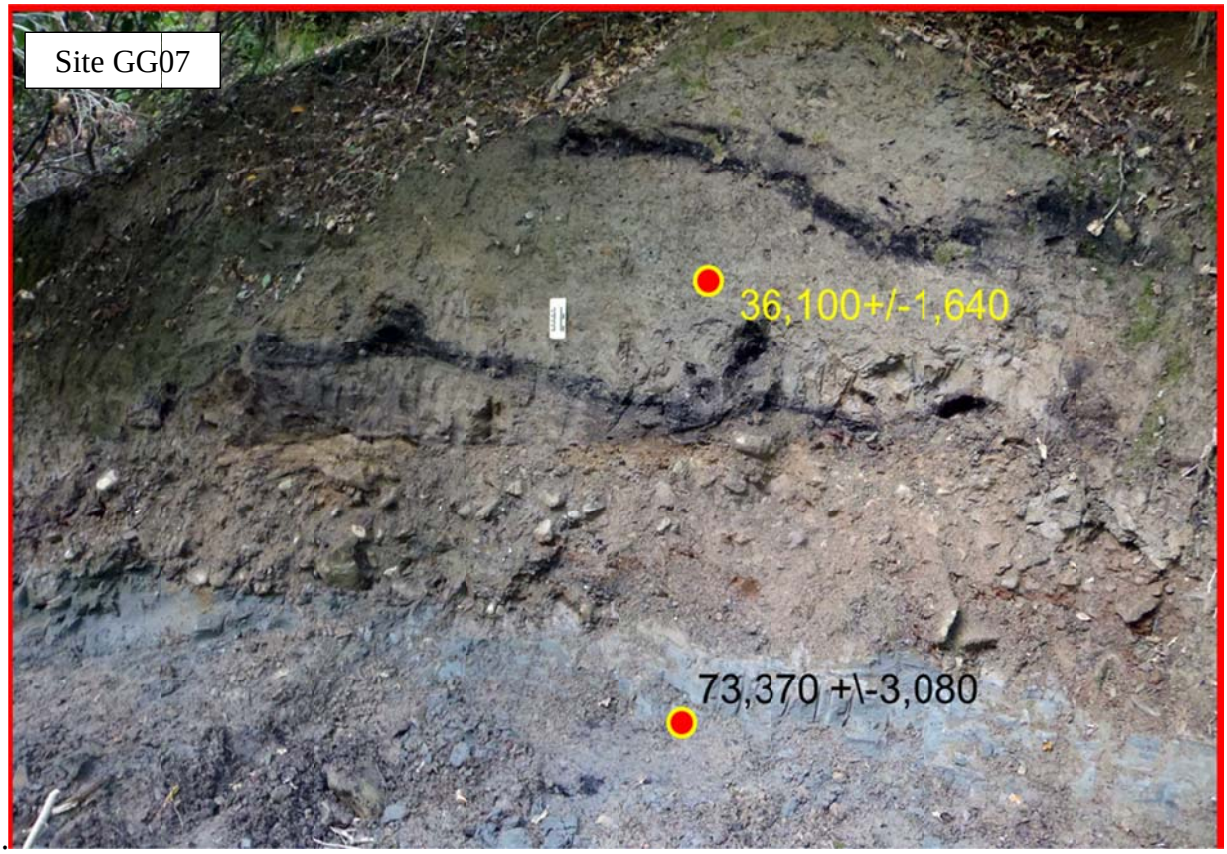


Figure 41. GG07 Site Stratigraphy with OSL age estimates.

---NOTES---

LiDAR and Landscapes

Andrew Kozlowski, Brian Bird, Brandon Graham

With the advantage of LiDAR elevation data for Cayuga County we have had the ability to evaluate the glaciate landscape in unprecedented detail. Prior to this, the surficial geologic maps were largely constructed on 1:24,000 or 1:62,500 scale base map using 10 or 20 foot contour intervals and later compiled at much smaller scales such as 1:250,000 (Muller & Cadwell, 1986). This process served as the template for understanding landforms and spatial patterns until the development of USGS digital elevation models (DEM's) such as the 10 meter data set available for NY. By comparison the LiDAR provides a new prescription that greatly improves our vision.

Although, LiDAR is a phenomenal tool for the glacial geomorphologist, it is not a substitute for ground truthed mapping and stratigraphic investigations to decipher sediment–landform relationships. We have been ardent proponents for the 3D mapping approach developed and adopted by our peers in the Great Lakes Region to map geologic frameworks and address societal needs (Berg et al. 1999, Thorleifson et al., 2010, USGS Circular 2011). Over the last 5 years we have done a considerable amount of drilling and coring to provide robust and reliable stratigraphic data. With gracious help and instruction of Tom Lowell (University of Cincinnati), we have also adopted winter coring of lakes and wetlands. Like all field methods and field techniques, experience is the best teacher and over the last few years we have become more selective or cognizant about determining coring locations and methods.

At first we strove for not only recording stratigraphic details of glacial deposits but were also on the hunt for suitable organic materials upon which we could radiocarbon date and thus help obtain chronologic data about the age of deposits and rates of deposition etc. While this approach proved fruitful, it also became clear that like any good business, our business was dependent on location, location, location. Lakes and low lying wetlands are excellent localities and the depositional environments they foster certainly are conducive to preservation of organic materials. Yet, the context and concerns about reworking of wood and other materials was always present. Amongst the concerns we were aware of is that proglacial low lying areas and lakes settings are subject to fluctuating lake levels and meltwater contributions that can mix materials and that only with robust, internally consistent radiocarbon control can you reasonably obtain results that are interpretable regarding chronology.

It is for this reason that I decided we needed to adopt a higher degree of selectivity in our coring approach. Much like an MRI can serve as a guide to provide lines of converging evidence that assist the interpretation and help focus a surgeon or physician; LiDAR can be used to delineate isolated basins or sinks in a proximity to specific glacial landforms that provide better confidence of their direct association. In our case we focused partially on ice marginal positions, previous geologic maps (Muller

& Cadwell 1986) that cover Cayuga County and the scale they are compiled at depict approximately seven ice marginal positions from south to north. Whereas in our present analysis we have identified as many as 22 ice marginal positions within Cayuga County. The topographic expression and character of these ice margins is highly variable. Some appear as well defined, yet subtle ridges that are continuous across the landscape others appear as discontinuous segments or simply as well defined fans (figure 42). Having this enhanced clarity, we further utilized the elevation models to focus in on a sink or isolated basins of glacial origin (see karst section by Clift for other possibilities) adjacent to ice margins whose association was unquestionable. Such locations serve as the first potential trap, to capture organic materials.

THE FIRST TRAP HYPOTHESIS

As mentioned previously there are many locations to potentially core in an attempt to recover suitable materials worthy of providing age control for chronologic investigations. Our "First Trap Hypothesis" "focuses" on locations with emphasis on the following criteria: 1) Locations containing isolated, upland basins proximal to ice margins; 2) Locations that would likely serve as a basin to trap and potentially preserve early colonizing vegetation; 3) Perhaps most importantly, locations, that would have experienced minimal modification and sediment input since deglaciation. As our mapping and experience working on the landscape has improved, we have been able to test this First Trap Hypothesis starting with the Mapleton Moraine.

Mapleton Moraine

The Mapleton Moraine was named and assigned by Shumaker (1957) and consists of an easily traceable and fairly continuous ridge first observed near Mapleton. The moraine exists as a subtle rising ridge on the northern rim of Great Gully and extends for more than 50 kilometers trending northeastward to Fleming and Auburn. Although Shumaker was uncertain, the LiDAR and DEM data clearly indicate its association with constructional topography northeast of Owasco Lake. On the north side of the moraine margin subtle undulating ridges parallel the principle margin creating a hummocky appearance (figure 43), whereas the distal edge appears to have well defined border. Meltwater channels are traceable back to the margin and Shumaker (1957) and Chisnell (1951) described open packed, clean and poorly sorted exotic (different from underlying bedrock) gravel outwash deposits present over a distance of two miles along the margin. Shumaker measured flow indicators in sand deposits near the head of Great Gully indicating westward flow from the Mapleton outflow channel (Chisnell, 1951). Aside from these descriptions there are few outwash facies described and we have not identified any well-defined fans associated with this margin. Approximately 1 kilometer northeast of the head of Great Gully is the Dumond Wetland (stop # 4), an isolated basin immediately adjacent to the Mapleton Moraine.

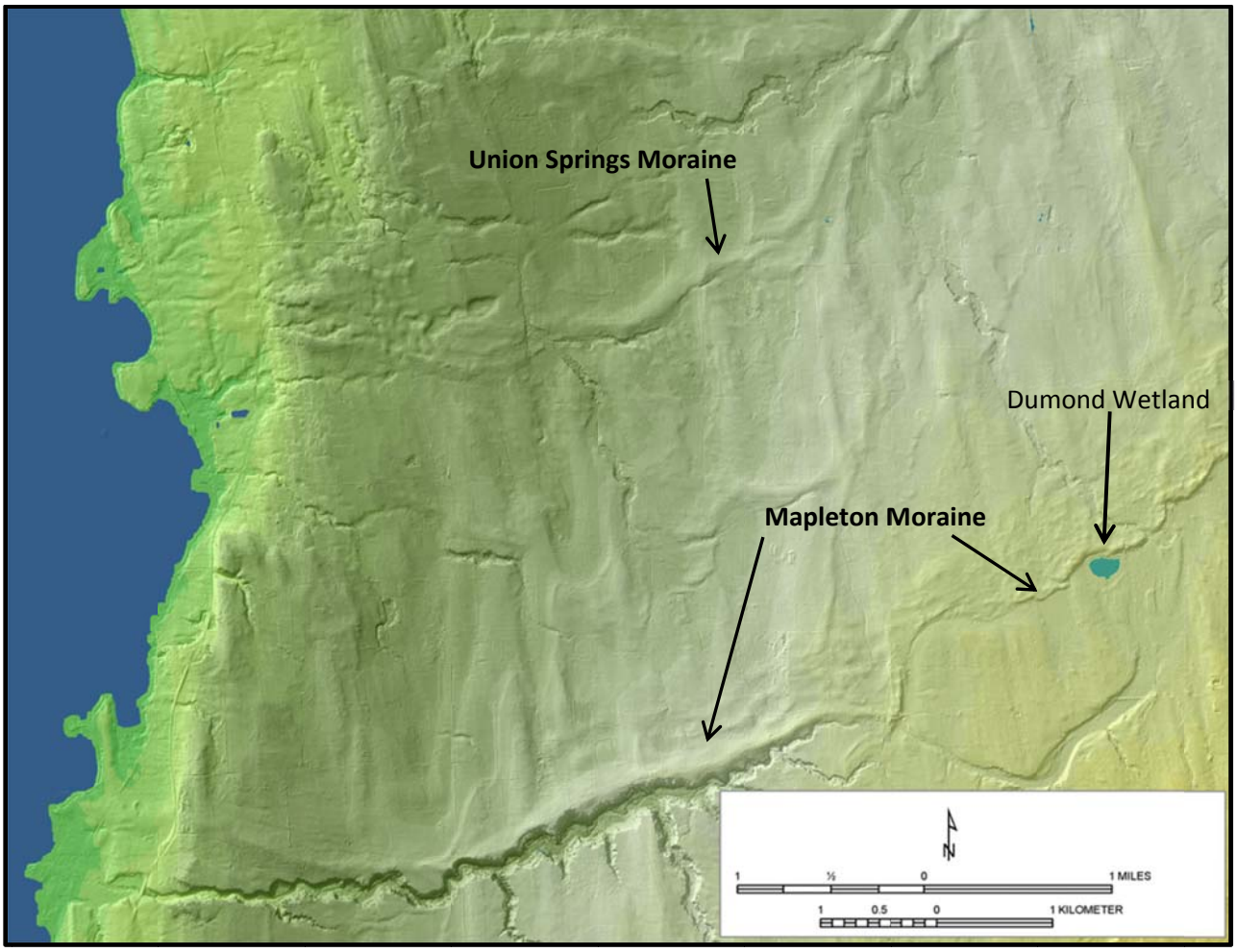


Figure 42. Illustrating the topography and spatial relationship between the Mapleton Moraine, Great Gully and the Union Springs Moraine.

Dumond Wetland

The Dumond wetland is a closed, oval shaped basin approximately 300 meters in diameter. This location was identified December of 2011 as a possible coring location and was prioritized as a locality based on the criteria outlined in our First Trap hypothesis (See Previous Section). Our initial reconnaissance in June of 2012 with tile probes indicated ~ 2.5 meters of soft substrate present. Coring occurred in March of 2013, at that time we literally "broke the ice", and using a Wink Vibracore system and Livingstone corer, we cored seven locations (Figure 43) to investigate the nature of the stratigraphy and deposits contained within this basin. While thickness of stratigraphic units varies within the wetlands, generally the stratigraphy consists (from top to bottom) of ~ 0.35 metres thick fibric peat that grades into a 1.0 meter thick package of organic rich marl below. The marl grades into a laminated silt that tends to be about 1.3 meters in thickness. The laminated silt terminates abruptly into a coarse sand containing gravel that tends to be 0.1 meters thick and overlies a dense gravel rich matrix supported diamicton of undetermined thickness interpreted as till.

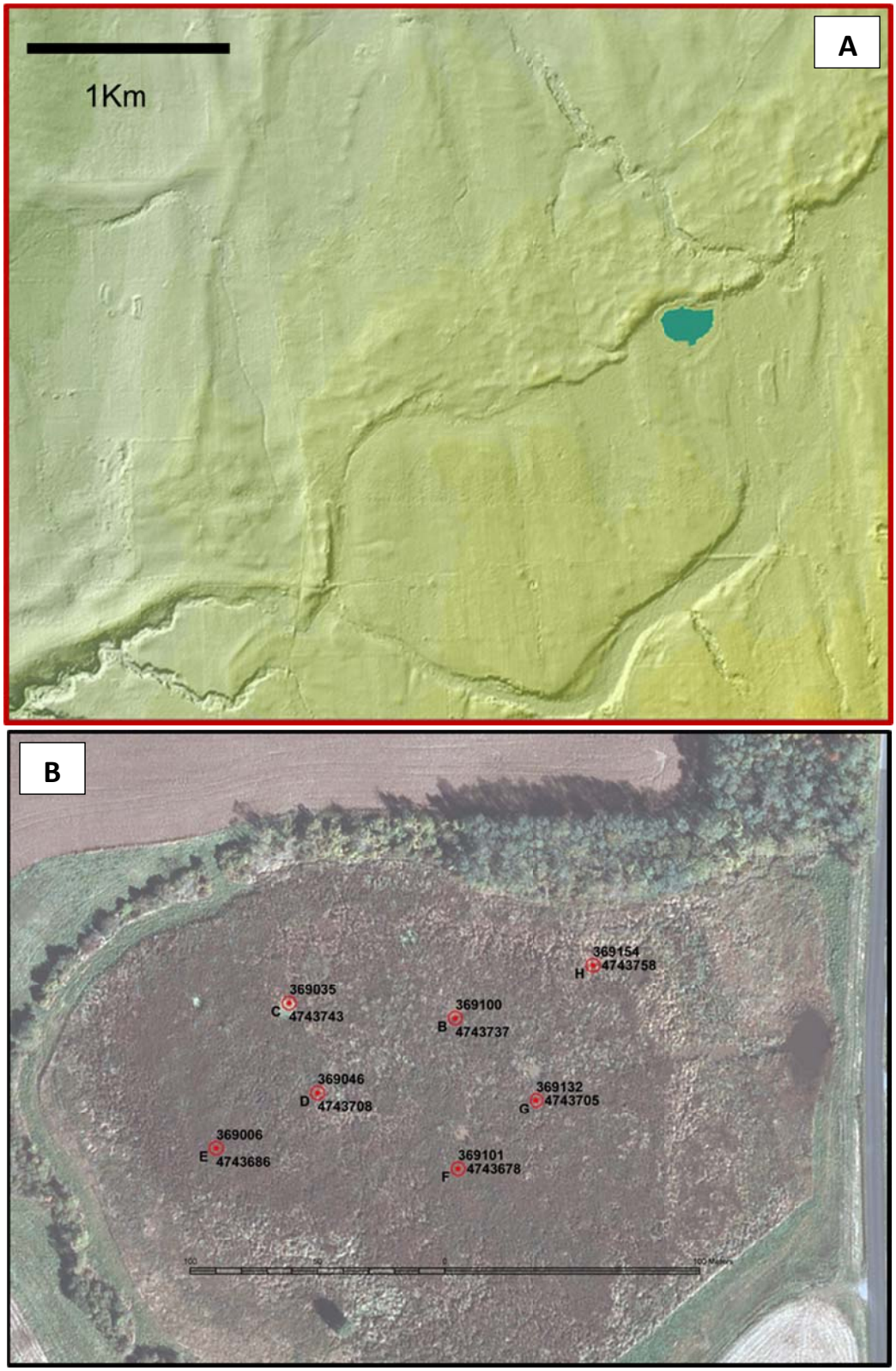


Figure 43. Topography (A.) showing Mapleton Moraine, Dumond Wetland and Meltwater Channels . (B) Coring locations from initial surveys.

Our initial survey and cores yielded some promising organics. We were able to isolate a few *Dryas integrifolia* leaves following recommendations from Brandon Curry (ISGS). These organics gave us a baseline from which to evaluate time-stratigraphic information regarding the depositional history of the basin. The initial radiocarbon dates came back to yield ages that supported the criteria #2 & #3 of our hypothesis. This indicates the locality served as a trap and suggested that sediment input had been minimal over the last ~10,000 years. Two of our deepest recovered plant macrofossils did not have enough mass to yield datable results, so in October of 2013 we again returned and completed another 20 cores. During our initial visit in March coring with the Livingstone halted when we hit the coarse sand unit and the Vibracore unit became plugged with coarse gravel and cobbles or simply refused to penetrate into the substrate. In our October return visit, we borrowed a hand driven Geoprobe unit from Parratt & Wolff Drilling in Syracuse, NY in an attempt to achieve/obtain maximum sampling depth. We found that utilizing a sharp barreled Livingstone Corer to recover the laminated silt portion of the cores worked very well and the hand Geoprobe was an outstanding choice to recover the silt/sand contact as well as consistent recovery of 0.1m of diamicton (till).

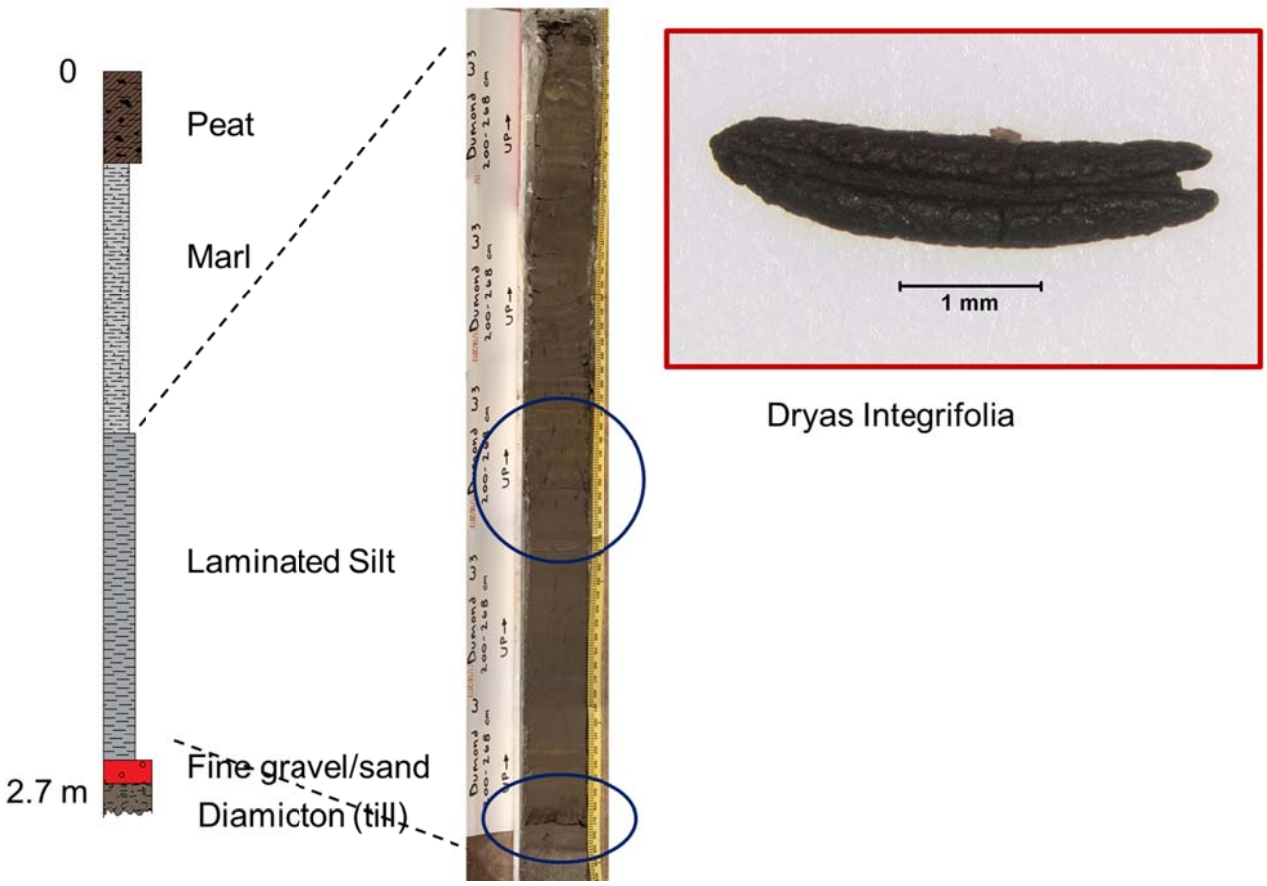


Figure 44. Generalized stratigraphy of the Dumond Wetland with photo of laminated Silt core black specs within core (blue circles) are organics predominantly consisting of leaves of *Dryas integrifolia*.

Radiocarbon Results

With the additional cores we recovered an abundance of organics including some twigs, needles and wood, however the dominant materials recovered deep in section within the laminated silts are *Dryas* leaves. To date we have obtained more than 20 radiocarbon dates on the stratigraphy of the Dumond Basin. The dates are internally consistent, do not indicate any mixing or inversion and appear to provide a robust record of deglaciation. It is important to note that while conducting our investigation we were informed that the wetland was and always has been a shallow body of water in historic times and has never dried up. Our radiocarbon based chronology indicates sedimentation has been continuous since our earliest radiocarbon date of 9,700 ± 25 ^{14}C yr BP through ~ 12,650 ± 45 ^{14}C yr BP (figure 45.) During collection we fortuitously recovered a piece of wood within the basal coarse sand that provided a single radiocarbon date of 35,490 ± 490 ^{14}C yr BP.

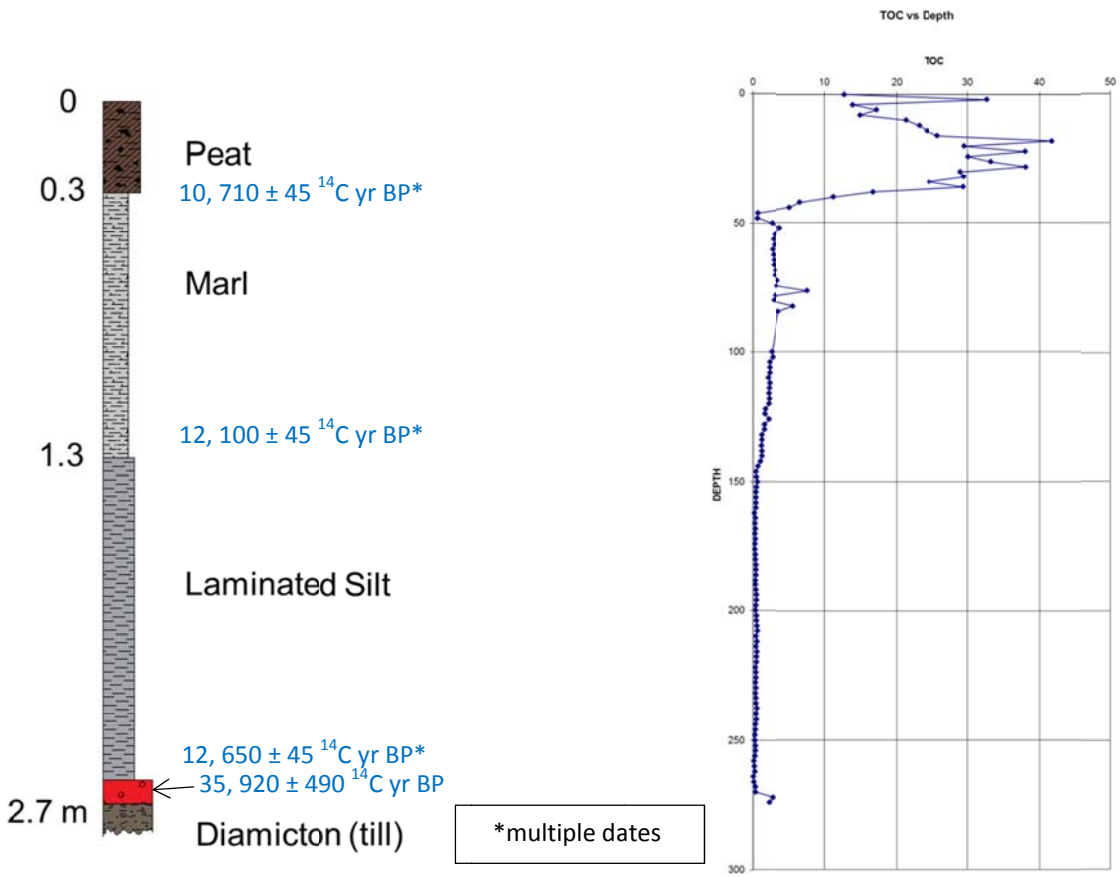


Figure 45. Stratigraphy, radiocarbon chronology and LOI/TOC Curve for the Dumond Wetland

Discussion

The robust radiocarbon chronology and consistent basal dates on *Dryas* leaves gives us the information necessary to confidently suggest that formation of this basin and by association the Mapleton Moraine formed about 12,650 ^{14}C BP . This interpretation seems logical not only based on the radiocarbon age but also based on the type of material we are dating. *Dryas* leaves are early colonizers, and although their leaves can be tough, they likely would not survive in such pristine condition if they were being reworked by abrasive meltwater like wood is able to. The aspect of the slope (facing south) and carbonate rich substrates are particularly suited to support development of *Dryas* plants as an earlier colonizer (D.M. Peteet pers. Comm). This basal age of 12,650 ^{14}C BP is the approximate age suggested for the Port Huron Readvance in Michigan (Blewett et al. 1993) and the Rome Readvance (Muller and Calkin, 1993) and Nine Mile Readvance of Ridge and Franzi (1992). Geographically, the Mapleton Moraine is only slightly south of Port Huron, MI and the low lying Ontario Lowlands to the north would have allowed a strong readvance to easily reach this latitudinal position. If our assignment of the Mapleton Moraine as representing the Port Huron Phase (Karrow et al., 2000) is correct then it provides a long sought after anchor point for regional correlation between the upper Midwest and Central NY.

While the basal age dates provided by the *Dryas* leaves provide long sought after clarity on deglacial chronologies, the same cannot be said for the single radiocarbon date obtained from the coarse sand deposits. At first glance the Mid Wisconsin date (Brimley Phase) would seem to indicate the underlying till must be older than ~36,000 Yr BP and thereby suggests a Mid Wisconsin Ice Advance. While this is plausible and recent work by Karig and Miller (2013) and Young and Burr (2006) advocate for an advance, this would suggest that the landscape at this latitudinal position has no record of either the Nissouri Advance or the Port Bruce Readvance. While this is plausible it seems unlikely to us that such erosional efficacy would occur, particularly in light of the ancient glacial deposits preserved less than 1 km away in Great Gully. An alternative interpretation might be that perhaps the older wood represents a recycled and reworked piece of Mid Wisconsin material entrained subglacially up-ice and later ejected into outwash. The anomalously old date aside, the radiocarbon chronology, plant macrofossils and stratigraphy provide encouraging results that support our First Trap Hypothesis and we are anxious to employ this technique elsewhere to see if similar success may result.

---End of Day 1---

The Montezuma Wetlands Complex (MWC)

Andrew Kozlowski, Brian Bird, Brandon Graham

The Montezuma Wetlands Complex (MWC) Project is an effort by the U.S. Fish and Wildlife Service, the New York State Department of Environmental Conservation, and Ducks Unlimited, Inc. to protect, restore and enhance wildlife habitat. It encompasses approximately 50,000 acres of freshwater wetlands, grasslands, shrub fields and forested tracts in Seneca, Cayuga and Wayne counties in upstate New York and is located in the middle of one of the most active flight lanes in the Atlantic Flyway. The MWC includes the federal Montezuma National Wildlife Refuge, the state Northern Montezuma Wildlife Management Area, and lands owned by conservation groups, farmers, and other private landowners. Public lands and some private lands are managed to provide habitat for wildlife and recreation and education for people*.

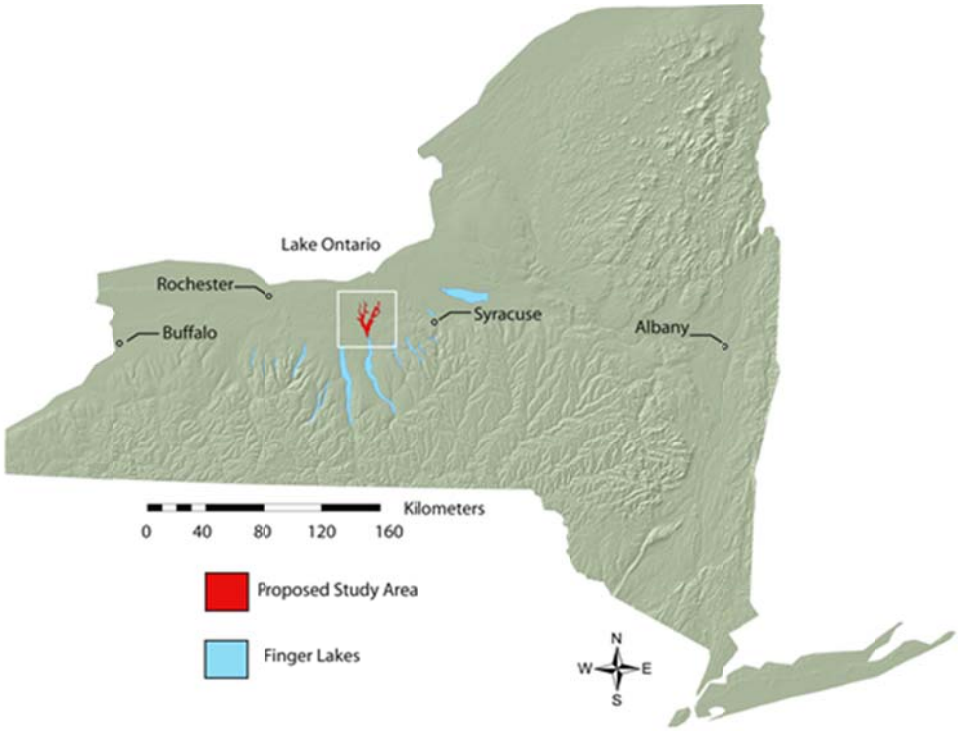


Figure 46. Location of Montezuma Wetlands Complex North of Cayuga Lake.

The MWC is located in the Ontario Lowlands north of the Cayuga Lake Basin, and occupies a series of broad dendritic channel forms in the heart of the Ontario Drumlin field. During multiple phases of glacial retreat and advance, the region was flooded by proglacial lakes. Preliminary data indicates a wide array of lithologies at the surface with a high abundance of diamicton associated with the drumlins. The thickness of glacial drift is between 25-60 meters. The complex interaction of proglacial lakes during deglaciation also deposited fine-grained lacustrine sediments that may serve as impermeable barriers or aquicludes between surface water and groundwater. Overlying and interbedded within the glacial till and lacustrine units are multiple peat units showing the existence of forest ecosystems, substantial water level fluctuations, and evidence of abrupt climate change in the late Quaternary.

*Adapted from Friends of The Montezuma Wetlands Complex http://www.friendsofmontezuma.org/

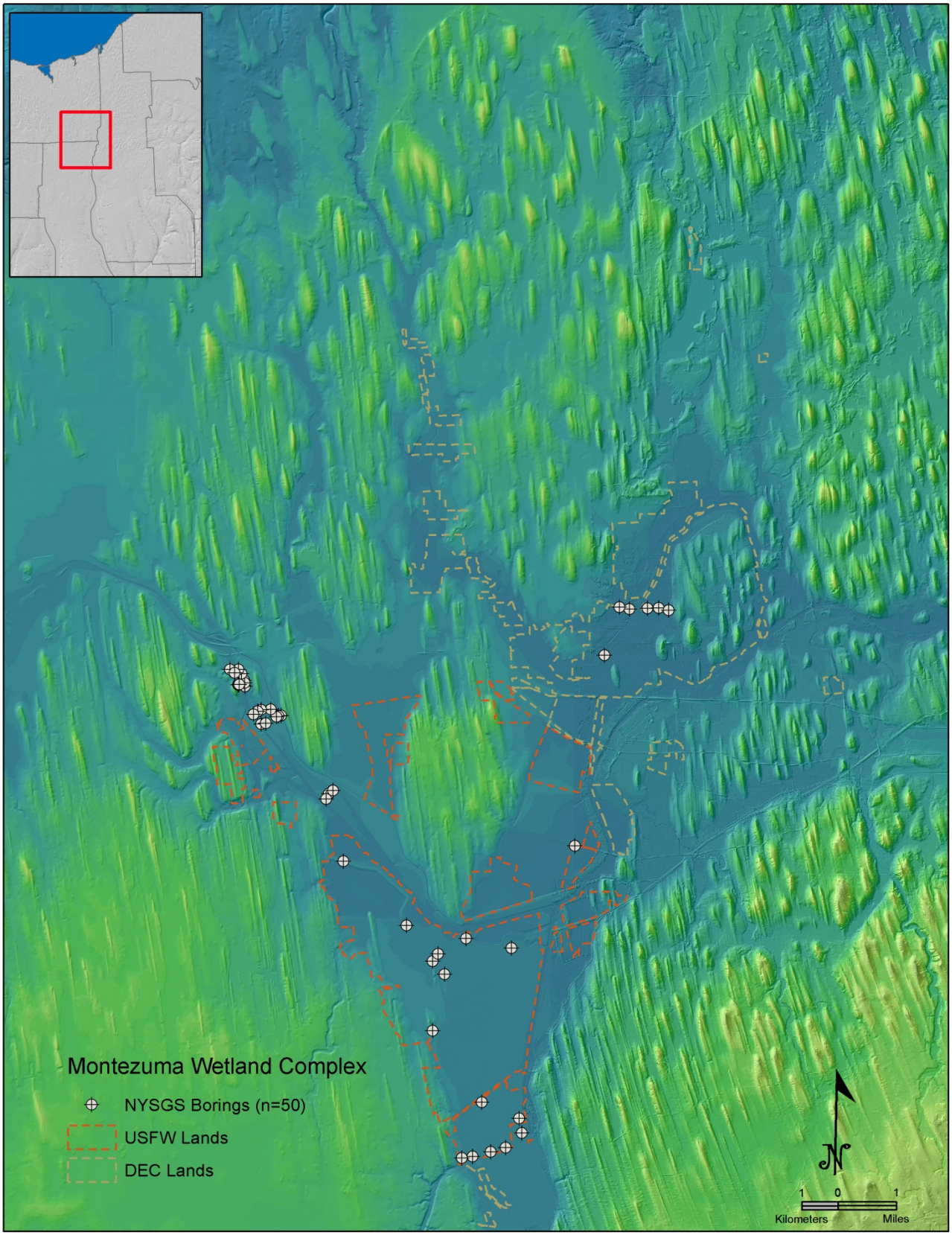


Figure 47. Dendritic channelized topography of the MWC displaying administrative boundaries and NYSGS Borehole Locations.

In 2008 I made the first reconnaissance visits with Bill Kappel to investigate reports of buried trees in Unit 17 of the USFW management area at the southern end of the MWC along the north shores of Cayuga Lake. Several excavations yielded multiple peats separated by deposits of marl, the lower most peat unit around 7 meters in depth consistently produced abundant wood fragments that dated around 10,350 ^{14}C YR BP. In 2009 we began mapping and working on the 3D framework as a formal member of the Great Lakes Geologic Mapping Coalition. Return trips utilizing sonic drilling and the USGS drill crew provided deeper stratigraphy that yielded a substantial thickness of varved lake clays from 12-24 meters depth on top of a 5 meter thick unit of dense diamicton interpreted to be till. Beneath the till we continued in fine-medium grained sands to a depth of 56 meters without reaching bedrock. Since that initial investigation we have completed more than 50 exploration boreholes principally utilizing Geoprobe coring methods.

We returned to Lock 26 in Clyde, NY along the northwestern arm of the MWC for secondary confirmation on early reports regarding the stratigraphy of recovered faunal remains during the 1910 construction of Lock 26 on the Erie Canal. While digging out the canal, workers found a mammoth tooth lying on gravel overlain by layered sand and silt deposits. The mammoth tooth was submitted to the New York State Museum and accessioned into the collections. A radiocarbon age on the tooth reported in 2010 (Feranec and Kozlowski) yielded an age of 11, 750 ± 65 ^{14}C YR BP (13,419 -13,770 Cal BP). Numerous Geoprobe cores collected on the Lock 26 side (west side) and east side yielded an abundance of datable materials

Figure 48. Excavation and concrete looking west.

Our subsurface investigations in the MWC reveal a rugged, deeply incised bedrock topography as deep or deeper than 65 meters and is in harmony with an earlier seismic analysis completed by Petrucionne et al. 1996. The MWC can be divided into an eastern and western arm with varying sedimentological properties. Generally the predominant materials present in the subsurface comprise glaciofluvial /glaciolacustrine sand, silt and clay with rythmites common. Typically within the upper 10 meters we encounter wood fragments preserved in peat layers that consistently date to ~10,350 ^{14}C YR BP; consistent with our initial discoveries in Unit 17. This indicates a wide spread stratigraphic horizon representative of the Younger Dryas interval. The eastern arm tends to be finer grained and becomes coarser to the north as it approaches eskers and ice contact fans around Duck Lake. Diamicton (till) units up to 32 meters thick lie beneath the sand and silt facies and are traceable southward to the confluence of both buried channels that funnel into the north end of the Cayuga Basin. Alternatively, coarse gravel facies are abundant in the western arm particularly in the vicinity of the Lock 26 near Clyde, NY.

Figure 49. View of spoil bank from hydraulic dredge Clyde.

These coarse gravel units are associated with a series of meltwater channels related to the Fairport Channels (Fairchild, 1909). These channels feeding into the western arm represent lowering spillways migrating north as ice retreated. Some gravel deposits observed in our collected cores likely originate as proximal outwash deposits. For example southeast of Lock 26 ground penetrating radar profiles collected display thrust faulting within gravel deposits (figure 50.) that indicate compressional stress originating from the north. Elsewhere gravel deposits appear to represent flood event(s) associated with the meltwater channels. The gravels associated with flood pulse(s) near lock 26 are traceable as a facies change that fine to a 3 meter thick medium sand flood pulse within a 12 meter

thick section of rhythmites (varves) in Unit 17 and are observed in multiple cores. Wood fragments >44,000 ^{14}C YR BP collected within the sand unit support the hypothesis that the sand unit represents a flood pulse reworking older material from the west delivered through the meltwater spillways near Clyde.

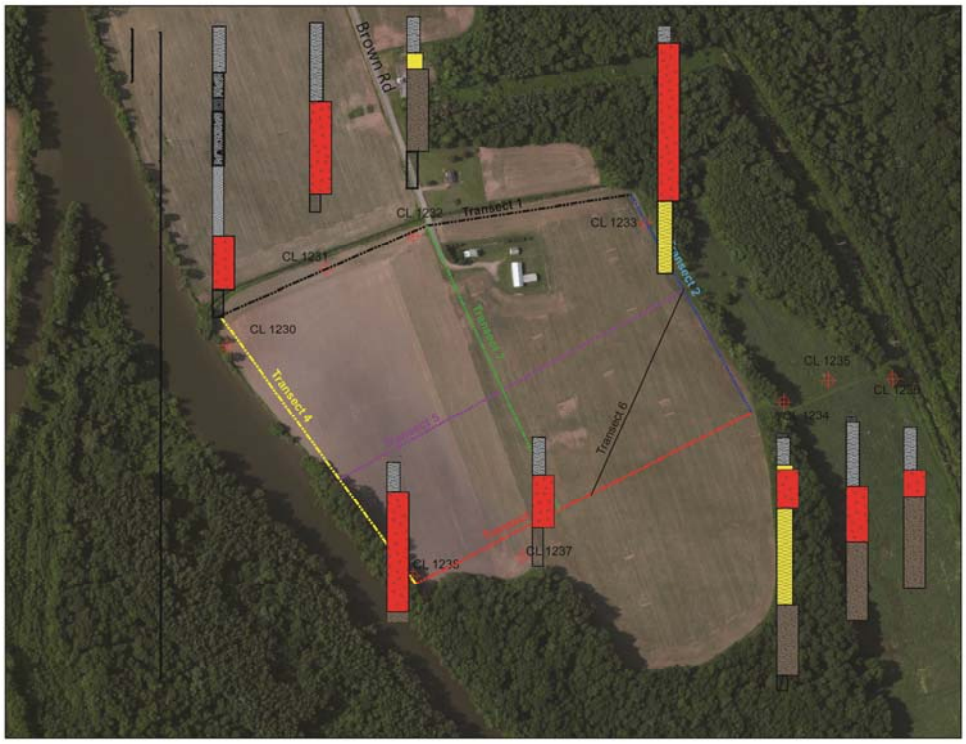


Figure 50. Southeast of Lock 26 boring locations and GPR transects.

Our subsurface analysis in the MWC supports suggestions of Mullins et al., 1996 (and references therein) that initial development of the large western and eastern arms as subglacial tunnel channels possibly over multiple glacial cycles. The thick sand and silt deposits observed are the result of copious sediment influx into a proglacial basin during retreat. The detailed stratigraphic and chronologic data collected near Lock 26 provides the spatial and temporal resolution to evaluate base level fluctuations recorded as incision and fill events. The reported deposits associated with the age of the mammoth tooth and newly collected radiocarbon dates on wood from the same stratigraphic interval indicate coarse-grained deposition in a proglacial setting ~12,300 ^{14}C YR BP; thereby the meltwater channels must predate this time. Internally consistent ages of 10,300 ^{14}C YR BP inset within overlying deposits indicates a significant base level drop at this time as corroborated by the observed peat horizons.

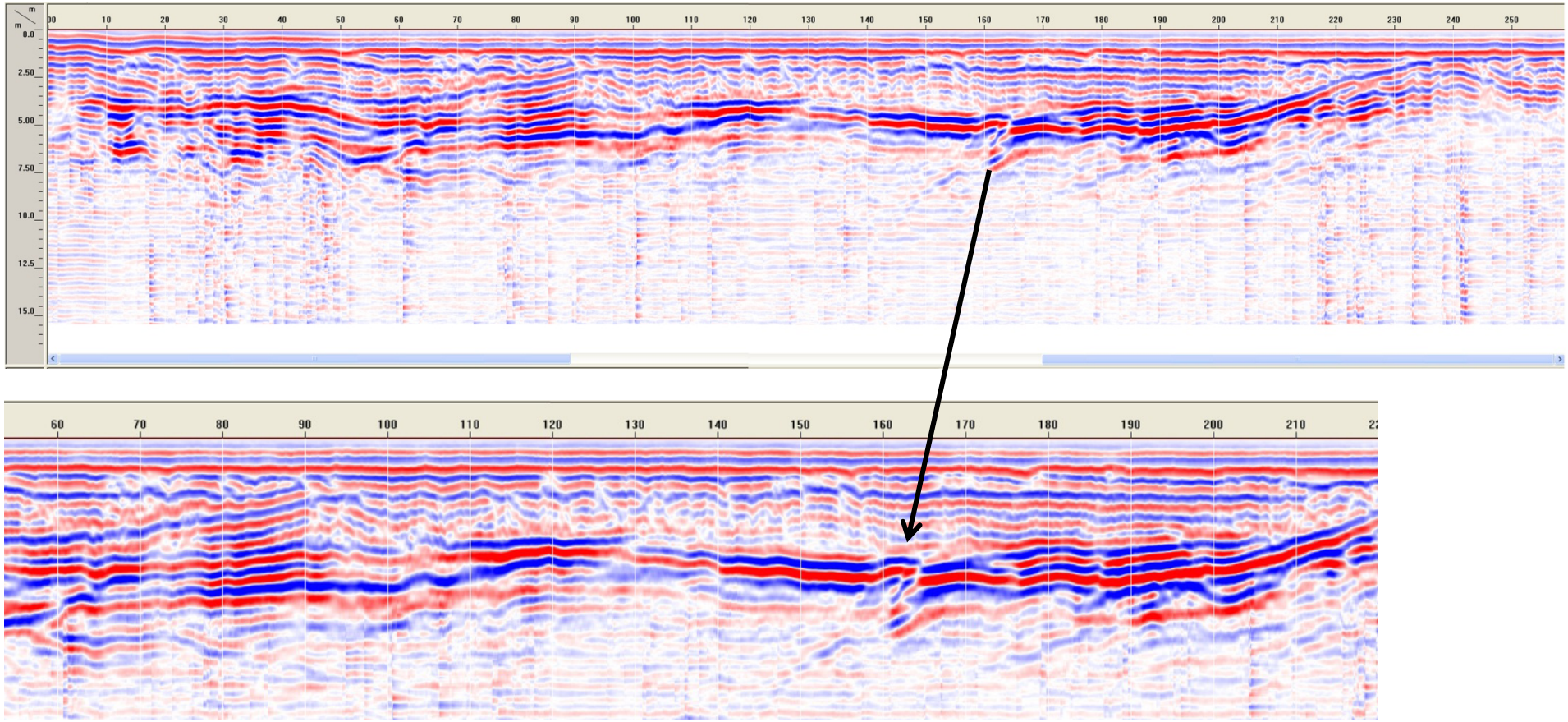


Figure 51. North to south 100 MHZ bi-static GPR transect collected southeast of Lock 26

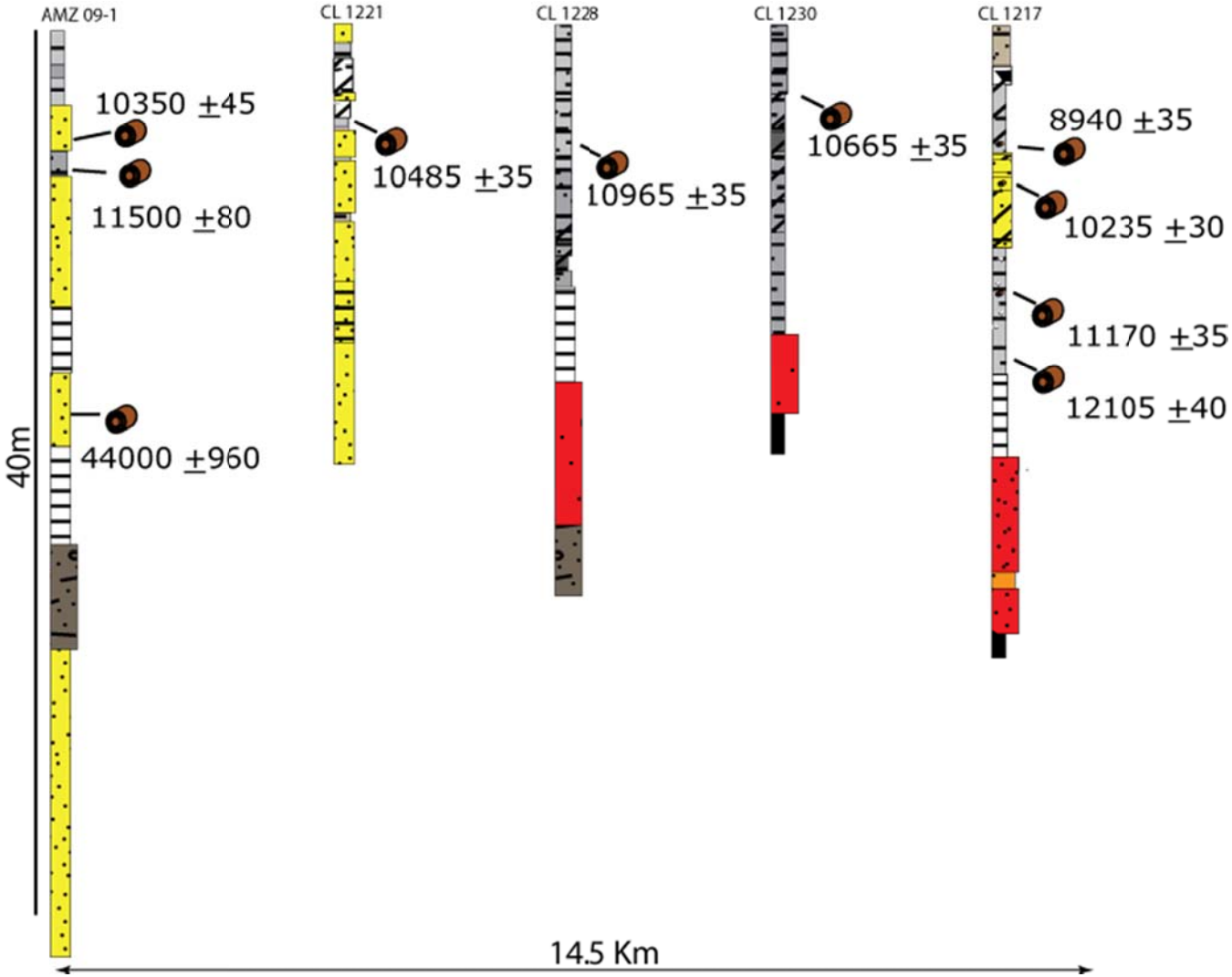


Figure 52 Western arm stratigraphic cross section from south to north.

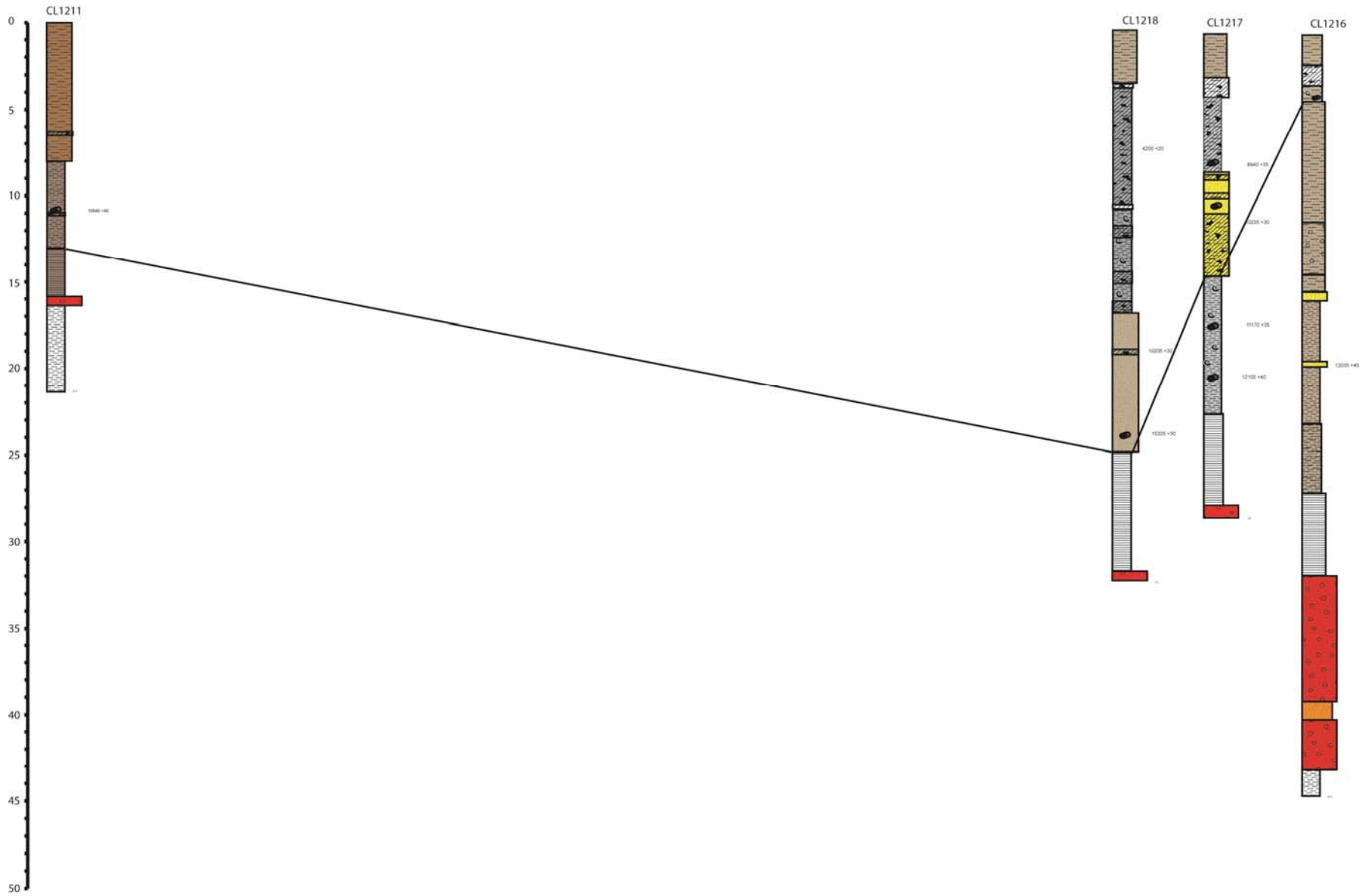


Figure 53. East to west cross section at Lock 26

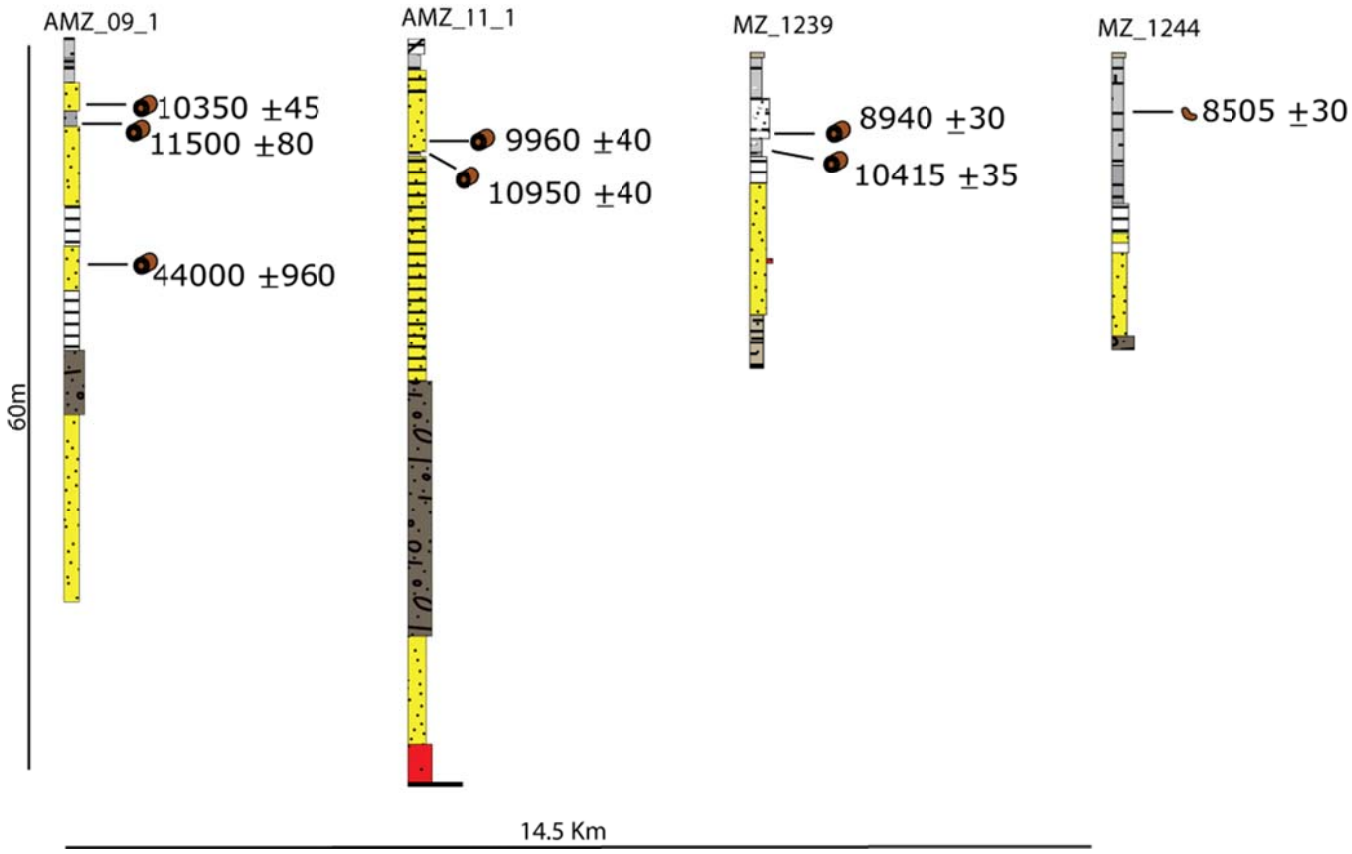


Figure 54. Eastern arm cross section from south to north

Figure 55. 100 MHz bi-static GPR survey Transect 2. Southeast of Lock 26. Left to right: Andrew Kozlowski, Andrew Clift (front), Brian Bird (back).

Paleoindian Occupations in Central New York

Jonathan C. Lothrop, James W. Bradley, Susan Winchell-Sweeney and Meredith H. Younge

Introduction

The purpose of this section is to discuss current evidence for earliest human occupation of Central New York during the Late Pleistocene and Early Holocene. Known to archaeologists as Paleoindians, these Native American populations are believed to have colonized Central New York and the broader glacial Northeast from regions to the west and south, beginning about 13,000 calendar years before present (Cal BP). This research is presented in the context of the Central New York region's evolving deglacial landscapes, as well as paleontological research on the appearance of possible Paleoindian prey species and other Ice Age fauna that also colonized the region during the Late Pleistocene.

The provisional data presented here derive largely from an ongoing research project known as the New York Paleoindian Database Project (NYPID) (Lothrop 2009). As a counterpoint to investigations of individual Paleoindian archaeological sites, the goal of the NYPID project is to collect data on the distribution of one artifact class – fluted and unfluted Paleoindian stone weapons tips – as a complementary data set for modeling the human colonization and settlement of the region.

Our geographic frame for this research is a seven-county study area in Central New York (CNY), consisting (from east to west) of Oneida, Madison, Oswego, Onondaga, Cayuga, Seneca, and Wayne counties (Figures 56 and 57). In sections below, we provide (1) background on this CNY study area and on Paleoindian occupation of the broader Northeast, (2) description of NYPID research methods and preliminary results for this CNY study area, and (3) discussion and conclusions.

Background

Physical Setting, Deglacial Sequence, and Regional Resources

The northern portion of this seven-county CNY study area occupies the Ontario Lowlands (or lake plain) south of Lake Ontario, while the southern sector encompasses higher elevation settings on the Appalachian plateau (see Figures 56 and 57). To the east, other areas of higher elevation include the Tug Hill Plateau and the Adirondack Highlands.

Major drainages on the Ontario Lowlands of CNY include the Seneca River, flowing east from its outlet at the north end of Seneca Lake, and the Oneida River, flowing west from its outlet at the west end of Oneida Lake (see Figure 57). These two drainages meet at Three Rivers in northern Onondaga County, and form the Oswego River, which flows northwesterly to Lake Ontario. Eastern Oneida County encompasses the headwaters of the east-flowing Mohawk River, while the southern portions of

Onondaga and Madison Counties (on the Appalachian Plateau) are drained by south-flowing tributaries of the North Branch of the Susquehanna River.

Deglacial events in the Ontario basin set the stage for human colonization of CNY and much of the broader New York region. As reported in this guidebook by Kozlowski and Bird, the Mapleton Moraine (indicative of the Port Huron readvance in CNY) has been dated 12,650 +/- 60 radiocarbon years before present (RCYBP), resulting in a two-sigma calibrated calendar age of 14,749-15,249 Cal BP. After ice retreat from this position in the Ontario basin, glacial melt waters pooled in the Oneida basin, forming the first footprint of proglacial Lake Iroquois (estimated by Anderson and Lewis [2012:523] at circa 14,500 Cal BP). With further ice retreat, this proglacial lake expanded its footprint across much of the Ontario basin. Sometime between about 14,600-13,800 Cal BP, final ice retreat from the Mohawk Valley created an outlet for proglacial Lake Iroquois near Rome, routing huge quantities of melt water down the Mohawk Valley and into the Hudson Valley (Ridge 2003; Stanford 2009: 12; Wall 2008).

Anderson and Lewis (2012:522-523) propose that Lake Iroquois reached its maximum footprint at circa 13,500 Cal BP (11,700 RCYBP), inundating most or all of the Ontario Lowlands portion of the seven-county study area (see Figure 57). Following this date, northward retreat of Laurentide ice from the northern Adirondacks in New York opened a series of lower drainage outlets via the St. Lawrence, leading to a sequence of progressively lower lake levels in the Ontario basin, and culminating with Early Lake Ontario (circa 110 m lower than Lake Iroquois). Anderson and Lewis (2012: 522-523) date the smaller footprint and lower level of Early Lake Ontario to between about 12,900 and 12,300 Cal BP (11,000-10,400 RCYBP), when this water body was likely confluent with the newly formed Champlain Sea. In CNY, the shoreline of Early Lake Ontario lay 5-10 km north of the shoreline of modern Lake Ontario (see Figure 57). Thereafter, the combination of rising basin outlet sills (due to isostatic rebound) and an Early Holocene dry climate (beginning circa 11,600 Cal BP), resulted in a closed-basin, continued lake low-stand in the Ontario basin between 12,300 and 8,300 Cal BP (10,400-7500 RCYBP). Thus, during the entire sequence of Paleoindian occupation in the CNY study area, lake footprints in the Ontario basin were significantly smaller and lower compared to the modern footprint.

Toolstone would have been a critical resource for Paleoindian populations colonizing CNY in the Late Pleistocene. The Onondaga escarpment, marking the northern front of the Appalachian Plateau, extends east-west through the middle of the study area (see Figure 56). Colonizing Paleoindian populations moving across the Ontario Lowlands likely learned early on of chert-bearing Devonian formations exposed along the north-facing Onondaga escarpment (Funk and Wellman 1984; Lothrop 1989). Further to the east in the Hudson Valley, Paleoindian groups also exploited other Devonian exposures of chert-bearing limestone and chert-bearing shales Ordovician Normanskill group (Funk 2004). Paleoindian groups also obtained Cambro-Ordovician jasper from the Hardyston formation in the Reading Prong district of the middle Delaware Valley and cherts from the upper Mercer and Vanport limestones in east-central Ohio (Lothrop and Bradley 2012: 14-15).

Late Pleistocene recolonization of New York's deglacial landscapes by potential vertebrate prey species figures prominently in discussions of earliest human occupations. Based in part on a number of recent radiocarbon determinations for Pleistocene fauna (Feranec and Kozlowski 2010, 2012), mammoth and caribou were likely present by circa 16,000 Cal BP, while mastodon appear circa 15,000-14,000 Cal BP. Megafaunal extinction, including mammoth and mastodon occurs by 10,000 Cal BP (or perhaps earlier); review of radiocarbon dates for mastodon remains at the Hiscock site in Genesee County suggest several centuries of overlap between mastodon and Paleoindian populations in New York (Lothrop and Bradley 2012:14). Studies at a limited number of locations indicate freshwater fish species recolonized streams and water bodies from southern refugia after 14,000 Cal BP (Daniels and Peteet 1998), and the Champlain Sea harbored a diverse cold water marine fauna that could have also provided subsistence resources for Paleoindians (Lothrop and Bradley 2012:13-14).

Across the broader Northeast, Early Paleoindian groups colonized and then adapted to a relatively cold and dry, largely subarctic environment, ranging with latitude and elevation from tundra to boreal parkland and forest. For the New England Maritimes (NEM), this environmental setting persisted through much of the Younger Dryas, circa 12,900-11,600 Cal BP (Newby et al. 2005), while in the Eastern Great Lakes (EGL), a transition to closed pine forests seems to have occurred gradually over this same time span (Ellis et al. 2011).

For the CNY study area, early paleoenvironmental studies based on pollen stratigraphy document transitions from spruce to pine-dominated taxa (Cox 1959; Durkee 1960). However, minimal radiocarbon dating based on bulk samples and other shortcomings limit the utility of these early studies for reconstructing the ancient environments encountered by colonizing and subsequent Paleoindian populations in the CNY study area. This in turn impacts our ability to project the local presence of potential terrestrial vertebrate prey species and other animal and plant resources from the Late Pleistocene into the Early Holocene.

Overview of Previous Paleoindian Research – New York Region

Systematic research on the first peoples of the New York region began with William A. Ritchie's 1957 publication of "Traces of Early Man in the Northeast." By the time of publication, the association of human stone tools and extinct Ice Age fauna, including mammoth and bison, was well-established on the High Plains, but little evidence was available for the human occupation of the Late Pleistocene Northeast. Ritchie (1957) presented data on find spots of fluted points and locations of Paleoindian sites in relation to deglacial landscapes, demonstrating human occupation of the Northeast during the Late Pleistocene. In 1969, Ritchie provided additional data, including mapping that showed that the highest density of fluted points in New York was found in the CNY study area reported here (Ritchie 1969:3-8). In subsequent years, investigations of individual Paleoindian sites in New York and the broader Northeast (e.g., Funk 1973, 2004; Gramly and Lothrop 1984; Gramly 1998, 1999; Lothrop

1989; Robinson et al. 2009) and continued collection of distributional data on fluted and unfluted Paleoindian bifaces (Wellman 1982; Bradley and Funk, N.D.) led to changing models of earliest human occupation of the New York region (Gramly 1988; Lothrop and Bradley 2012; Newby and Bradley 2007; Petersen 2004). Underlying themes in this research have included questions of chronology for early human occupation, subsistence and vertebrate prey species, stone tool technology, post-colonization seasonal mobility and landscape use, and connections with Paleoindian populations in adjoining regions (Lothrop and Bradley 2012).

Current Models of Paleoindian Lifeways in New York and the glacial Northeast

Based on limited radiocarbon dating and geochronological associations, most researchers generally agree that Paleoindian populations colonized the glacial Northeast circa 13,000 Cal BP (11,000 RCYBP) from the adjacent Midwest and Mid-Atlantic regions (Ellis et al. 2011; Lothrop et al. 2011; Lothrop and Bradley 2012; Newby and Bradley 2007).

Based on the current archaeological record, the small number and size of residential Paleoindian sites suggest they were living in very low population densities across both the EGL and NEM regions (Ellis et al., 2011; Lothrop et al. 2011). The typically small size of these sites and low density of stone artifacts that mark these ancient encampments also indicate these sites mostly represent short-term, small-group seasonal Paleoindian encampments.

At most Paleoindian residential sites, archaeological excavations typically recover discarded stone tools and flaking debris from tool manufacture. Organizational studies of these artifact assemblages suggest that Paleoindians conducted most tool and tool blank manufacture close to quarry sites, creating a highly portable stone toolkit to accommodate a mobile lifestyle. Common tool forms consisted of both hafted and hand-held implements, including bifacial points and preforms (unfinished points) and hafted knives for hunting and butchery, and unifacial tools such as triangular and narrow-bit endscrapers (hafted) and sidescrapers and flake gravers (hand-held) for working hide, wood and bone (Ellis 2008; Ellis and Deller 1988).

Residual cortical surfaces on flaked stone artifacts found at northeastern Paleoindian sites suggest that these peoples were routinely procuring better-grade toolstone from primary bedrock sources, and most researchers believe Paleoindians typically obtained most of this toolstone directly by quarrying at the source outcrop (rather than by exchange) (Burke 2006; Ellis 2008, 2011; Lothrop and Bradley 2012; Petersen 2004). Moreover, distances between some Paleoindian sites and the geologic sources where toolstone was quarried and fashioned into tools (approaching 4-500 km), suggest extensive seasonal movements of Paleoindian residential groups that were not matched by Native Americans in later periods of prehistory (Ellis 2011).

Given the regional evidence for subarctic environments and the occasional recovery of caribou or more generalized cervid faunal remains at some Paleoindian sites (Storck and Spiess 1994), most researchers believe that northeastern fluted point groups seasonally hunted caribou at certain times of the year, but that they also relied on other animal species, as well as limited gathered plant foods (Lothrop and Bradley 2012: 35-37). Mastodon appear to have overlapped with early Paleoindian groups for several centuries in New York, and the discoveries of fluted points with mastodon remains at the Hiscock site in western New York suggest that while Paleoindian groups may not have hunted mastodon, they may well have scavenged dead or dying individuals for hide, meat, bone or ivory (Laub and Spiess 1993). Following the Younger Dryas, late Paleoindian groups (circa 11,600-10,000 Cal BP) adapted to warmer and drier Early Holocene environments dominated by close pine forests over most of the Northeast. In New York, these peoples likely hunted solitary cervids such as deer and moose (Newby et al. 2005). Finally, the association of both fluted point and late Paleoindian sites with the Champlain Sea in northeastern New York and Western Vermont raises the possibility that Paleoindians may have also exploited marine resources of this inland sea (Loring 1980; Robinson 2012).

The New York Paleoindian Database Project: Central New York Study Area

Methods

Paleoindian biface sequences for the time span of approximately 13,000 to 10,000 Cal BP – developed by working in the EGL and NEM – provide the chronological framework for this research (Deller & Ellis 1990, 1992; Bradley et al. 2008). For the EGL, this sequence includes (1) the fluted point forms of Gainey, Barnes and Crowfield, (2) the transitional, poorly fluted Holcombe point, and (3) late Paleoindian (post-Younger Dryas) unfluted Hi-Lo and parallel-flaked Plano (Agate Basin and Eden) forms (see Figure 58). Defined primarily by researchers in Ontario and Michigan, these types are polythetic, based on constellations of metric and nominal attributes. Further, this sequence is viewed as a time series, so that each individual point type represents "an arbitrary segment in a temporal continuum of morphological and technological change…" (Deller and Ellis 1992:36). Geochronological associations in the EGL region, and radiocarbon dating of analogous biface forms in the adjoining NEM region, together support the view that this biface sequence represents a temporal series. This EGL biface sequence provides the relative chronological control for the research summarized here.

The NYPID project relies on some basic assumptions about the behavioral implications of Paleoindian points found across New York. First, these bifaces represent flaked stone weapons tips that were hafted on shafts of wood or bone for use as hunting weaponry, either as thrusting spears, javelins, or atlatl-propelled darts. Also, depending on design, and because of their probable hafting to foreshafts, these stone weapons tips were likely also sometimes used as butchering tools for dismembering prey. Where discovered, the find spots of these artifacts may reflect past locations of (1)

hunting or butchering of prey, (2) discard at seasonal encampments, or rarely, (3) stone tool caches. As such, data collected on the distribution, toolstone, form and relative age of these points can provide evidence of regional Paleoindian colonization and post-colonization settlement.

Our methods for data collection are straightforward and are modeled after Paleoindian point surveys employed in other states and provinces, calibrated to the established biface sequences for the EGL and NEM. Specifically, for individual cases, we record data on provenience, lithic raw material, metrics, and technological and stylistic attributes, coupled with digital photography. Stylistic and technological data and digital images are then used for relating each point to the regional temporal sequence. Provenience data are used to map distributional trends. To date, we have recorded individual Paleoindian points with New York provenience curated in museums and other institutions (both in and outside New York), as well as specimens in private collections of avocational archaeologists.

NYPID Results for CNY

Our collection and analysis of data for this CNY study area is ongoing, and the summary statements below are provisional and will be revised as the research proceeds.

1. Site Association and Chronology. As part of the NYPID project, to date, we have recorded 80 cases in the seven-county, CNY study area (see Figure 57). These include 68 fluted points (85% of the sample), while 12 specimens consist of unfluted, late Paleoindian points. Based on data recorded thus far, about half of the total sample consists of largely unbroken and/or not heavily resharpened or reworked points that can be tentatively assigned to the EGL biface sequence. The fluted point inventory includes good examples of Gainey, Barnes, and Crowfield point forms, but thus far, we have no clear cases of Holcombe points. Unfluted bifaces in the sample include the parallel-flaked "Plano" forms, including Agate Basin and Eden-like varieties, but no examples of Hi-Lo bifaces. With ongoing analysis of metrics and nominal attributes, additional specimens in this current data set may be assignable to types of the EGL sequence. We also note [as have other researchers (Deller and Ellis 1992:36)] that several examples in the current data set display combinations of attributes, suggesting they may represent transitional forms within this EGL Paleoindian biface sequence.

2. Sites versus Isolates. Of the 80 cases, five discrete localities have produced 17 bifaces (two or more points at each location); these represent known or probable Paleoindian residential encampment sites. The remaining 63 specimens in the current sample are isolated finds, but we suspect that several of these find spots likely represent residential encampments that could perhaps be confirmed with field investigation.

3. Sites and Isolated Finds. In this CNY study area, the Potts site in Oswego County represents an early Paleoindian site with probable Gainey points (Ritchie 1969; Gramly and Lothrop 1984; Lothrop

1989). Potential middle Paleoindian sites include an unreported site in Seneca County between Cayuga and Seneca Lakes; recovered artifacts include possible middle Paleoindian Barnes points and endscrapers. Further east in Madison County and south of Oneida Lake, the Glass Factory site (recently discovered by Mike Beardsley and Mark Clymer) has produced four unfinished fluted bifaces, including what may represent two or three unfinished Crowfield points. Finally, in the 1930s, former state archaeologist William A. Ritchie, conducted excavations at the Oberlander #1 site in Oswego County on the north shore of the outlet of Oneida Lake (Ritchie 1940); in addition to recovery of a major Late Archaic Brewerton point component dating to circa 5000 Cal BP, Ritchie also excavated unfluted late Paleoindian points, including parallel-flaked "Plano" Agate Basin and Eden-like forms.

4. Toolstone. Ellis (2011) recently reviewed evidence for trends in toolstone acquisition through time for Paleoindian sites in the Northeast. He found that earlier Paleoindian sites (with Clovis-like or Gainey points) yielded flaked stone artifacts made of toolstone from geologic sources were located up to 500 km straight line distance from site to toolstone source. For later fluted point Paleoindian sites, the maximum distance from toolstone source to site decreased, suggesting that over time, there was a reduction in the annual geographic ranges of these peoples. In the CNY study area, our limited sample to date may or may not reflect this larger regional trend. For example, based on macroscopic criteria, the DeAngelo point (a transitional Gainey-Barnes form) from Cayuga County, appears to have been manufactured from upper Mercer chert – a toolstone that outcrops in east-central Ohio, circa 500 km to the southwest. The unreported Paleoindian site in Seneca County includes fluted points and endscrapers made from what appears to be high-grade jasper, likely from the Reading Prong district in the middle Delaware Valley of eastern Pennsylvania (350 km to the southeast), as well as endscrapers of Normanskill chert from the mid-Hudson Valley (250 km to the east-southeast) and possible Vanport chert from east-central Ohio (500 km to the Southwest). Conversely, bifaces from the presumed middle Paleoindian Glass Factory site and late Paleoindian Oberlander #1 site have both yielded artifacts made on gray cherts that likely derive from Devonian limestone formations outcropping just a few miles south along the Onondaga escarpment.

5. Paleoindian Point Distributions. Figures 59 and 60 illustrate GIS mapping of Paleoindian biface distributions in the CNY study area. In reviewing this mapping, the reader should bear in mind the relative positions of the maximum footprint of proglacial Lake Iroquois at circa 13,500 Cal BP and Early Lake Ontario at circa 12,900 Cal BP. At its greatest extent, proglacial Lake Iroquois covered most of the northern portion of the CNY study area (primarily the Ontario lake plain), but did not extend onto higher landscapes to the south and east. Assuming human colonization of CNY at circa 13,000-12,900 Cal BP, this mapping demonstrates that a major portion of the study area had been drained of proglacial lake waters only a handful of centuries earlier. Conversely, mapping also shows how the shoreline of Early Lake Ontario lay north of the shoreline of modern Lake Ontario, and how the intervening 5-10-km-wide strip of terrain that fringed Early Lake Ontario, is now submerged. Thus, we have a significant sampling bias in our data, since we have no information on Paleoindian artifact

distributions for the now-inundated landscapes that fronted Early Lake Ontario. We are nevertheless left with intriguing distributional patterns for the remaining landscapes in the CNY study area.

Figure 59 illustrates the density of Paleoindian points by county (per 1000 km²) for the CNY study area. Sorted in this fashion, counties in longitudinal center of the study area display the highest point density. When these density data are projected at the township level (per 100 km²), however, a very different pattern emerges (see Figure 60). By township, highest point density appears primarily on the lake plain portion of the study area, largely within the footprint of proglacial Lake Iroquois, with far fewer cases south of the escarpment on higher elevation terrain of the Appalachian Plateau to the south. Within the Ontario Lowland, Paleoindian biface density is highest in townships that front Oneida Lake and the east-west trace of the Oneida and Seneca rivers. When broken down further by relative age, peak densities for early and middle (fluted), and late (unfluted) Paleoindian bifaces all show essentially the same distribution, with highest densities mostly in townships fronting Oneida Lake and the Oneida and Seneca drainages.

Discussion and Conclusions

GIS mapping of Paleoindian point distributions for this CNY study area suggest land-use strategies that may have focused on interior settings of the Ontario Lake plain, including along the Seneca and Oneida drainages. Conversely, Paleoindian settlement on higher elevation settings south of the Onondaga escarpment and northeast of Oneida Lake may have been more limited. In considering this patterning, we also have to acknowledge at least two types of sampling bias that surely affect these data. The first of these is geologic: the absence of archaeological evidence for Paleoindian occupations on the now-submerged landscapes that fronted Early Lake Ontario, so early Native American use of these areas is unknown.

A second potential source of bias relates to collection of archaeological data in this CNY study area. Antiquarians in the 19th century and professional and avocational archaeologists in the 20th century excavated and surface collected at major Archaic and Woodland-era sites (dating to the middle and Late Holocene) that were found along these same Seneca and Oneida drainages. The incredibly rich archaeological assemblages recovered at these sites (stone tools, pottery, animal bone and other food refuse) testify to the intensive re-occupation of these locations by later prehistoric Native Americans (e.g., Ritchie 1940, 1969; Ritchie and Funk 1973). This clearly reflects the high biomass productivity of this CNY Ontario Lowlands region in the Holocene, including fisheries in the Seneca/Oneida drainages and waterfowl associated with the Montezuma wetlands (Bradley 1987:11-14). At these very same locations, archaeologists also fortuitously discovered several of the Paleoindian artifacts that comprise parts of the current NYPID CNY data set. Hence, the high density of Paleoindian material recorded in the Seneca and Oneida drainages partially reflects the intensive investigation of later Native American sites in these areas by professional and avocational archaeologists since the 19th century.

Nevertheless, it is noteworthy that this CNY study area – reported by Ritchie (1969) as the highest density cluster of Paleoindian artifacts in New York – still maintains that distinction with this enhanced data set. In other regions of the glacial Northeast, archaeological evidence indicates that Paleoindian groups in the Late Pleistocene and Early Holocene often repeatedly reoccupied specific geomorphic landscapes (dune fields, relic proglacial lake strandlines, etc.), perhaps because of proximity to valued seasonal resources (Boisvert 2012; Ellis et al. 2011:540; Lothrop et al. 2011: 555-560; Spiess et al. 2012). These preliminary NYPID CNY data may provide evidence of the same strategic practice of Paleoindian reuse of certain landscapes. The relative chronology of this CNY data set is also noteworthy, because it indicates that this Ontario Lowlands landscape along the Oneida and Seneca drainages attracted both early, middle, and late Paleoindian populations from the late Pleistocene into the Early Holocene. Looking forward, future advances in our understanding of the Paleoindian occupation of the CNY region will rely on (1) targeted studies of Paleoindian sites and artifacts, (2) ongoing research on postglacial landscapes and associated fauna, and (3) perhaps most importantly, new research on the paleoenvironments of these first New Yorkers.

Acknowledgments

We wish to thank the many individuals thus far who have contributed data and provided access to their collections for the NYPID research project. In the CNY study area, these individuals include Mike Beardsley, Charles “Billy” Breen, and Mark Clymer. In addition to the NYSM, institutions that provided access to collections include the Rock Foundation, and the Rochester Museum and Science Center; we thank George Hammel of the Rock Foundation, and Sarah LeCount and Kathryn Murano of the RMSC for their gracious assistance. Thanks also to Christopher Ellis and Michael Gramly for insights and data on Paleoindian occupations in the EGL; and to Andrew Kozlowski, Brian Bird, and Robert Feranec for their insights on the CNY study area. Finally, thanks to John Hart, NYSM Director of Research and Collections, and Mark Schaming, NYSM Director, for their support of this research.

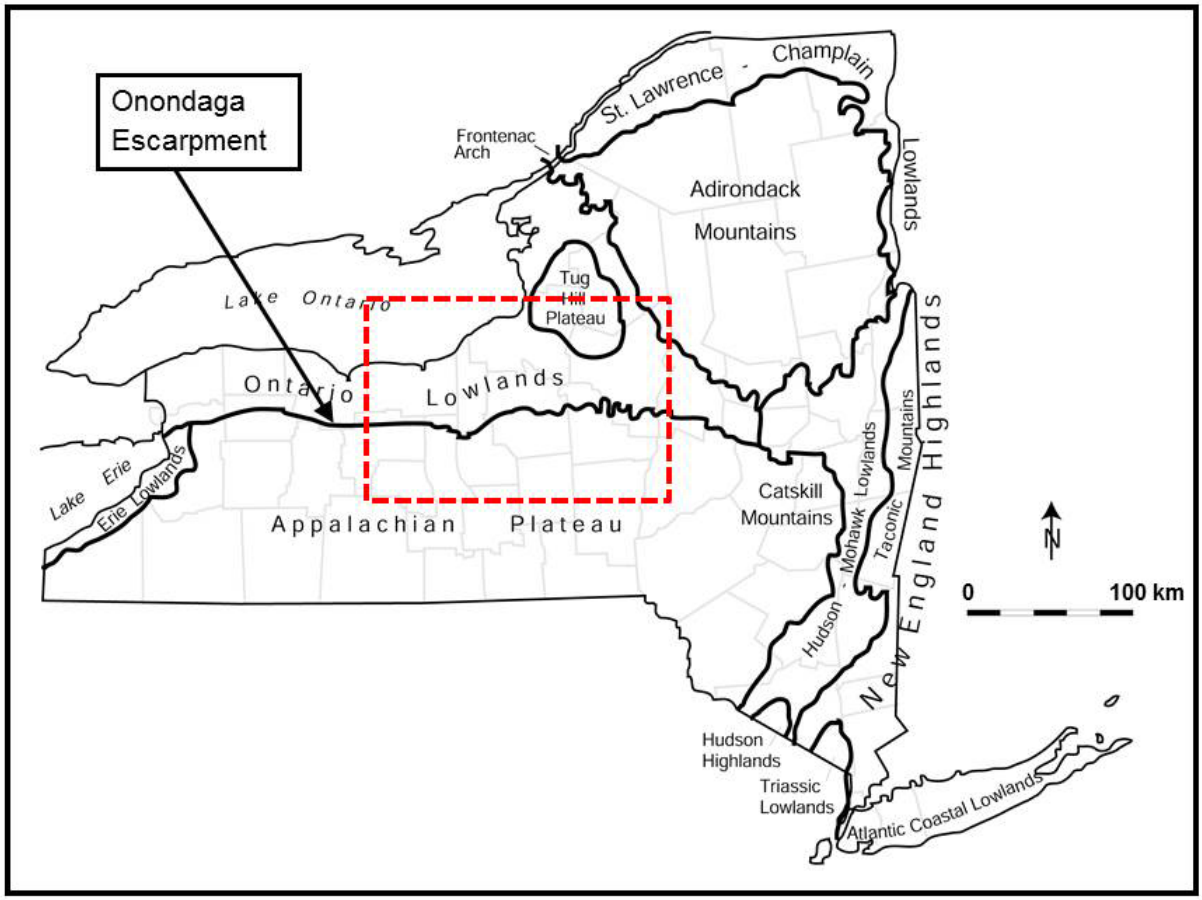


Figure 56. Location of NYPID Central New York study area in relation to physiographic regions of New York (after Cadwell et al. 2003).

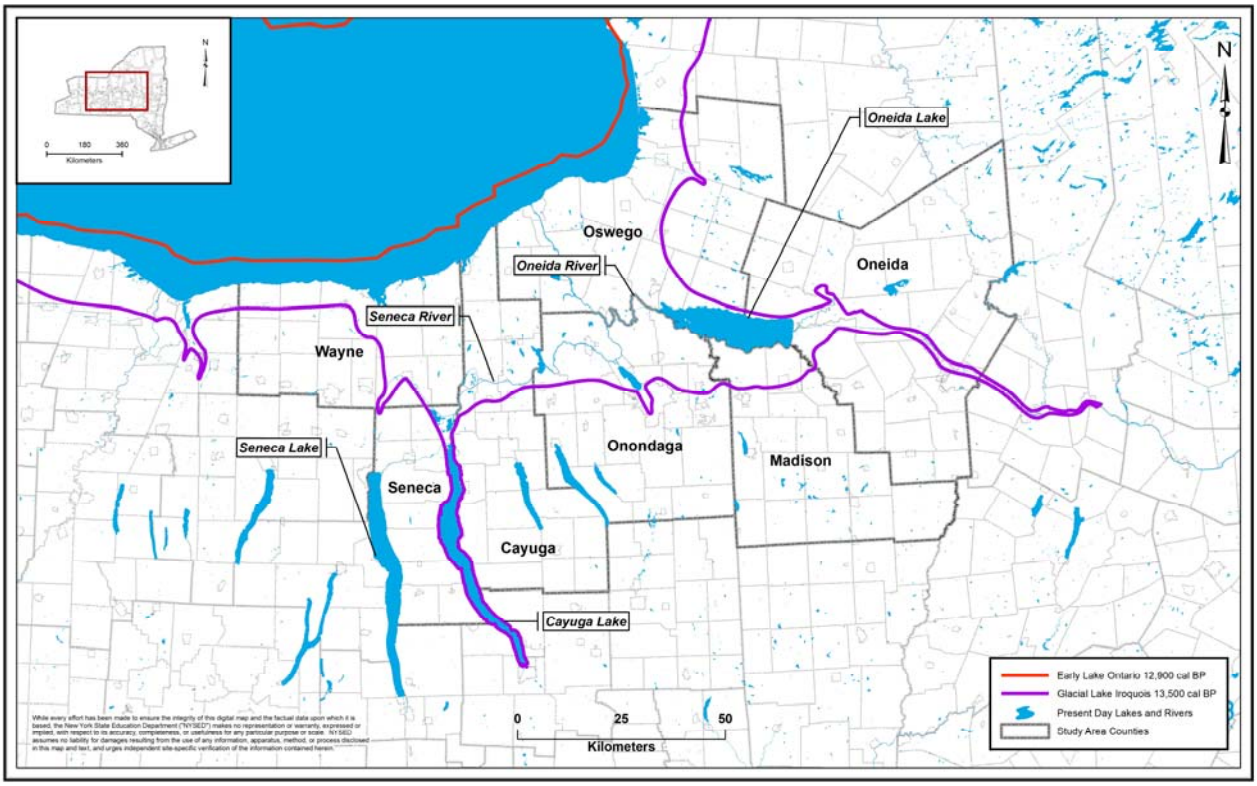


Figure 57. NYPID CNY seven-County study area, south of Lake Ontario, in relation to modern water bodies and streams, maximum footprint of proglacial Lake Iroquois (circa 13,500 Cal BP), and Early Lake Ontario (circa 12,900 Cal BP). (Note: Lake Iroquois footprint, after Kozlowski and Bird [this volume]; Early Lake Ontario footprint, after Dyke 2004).

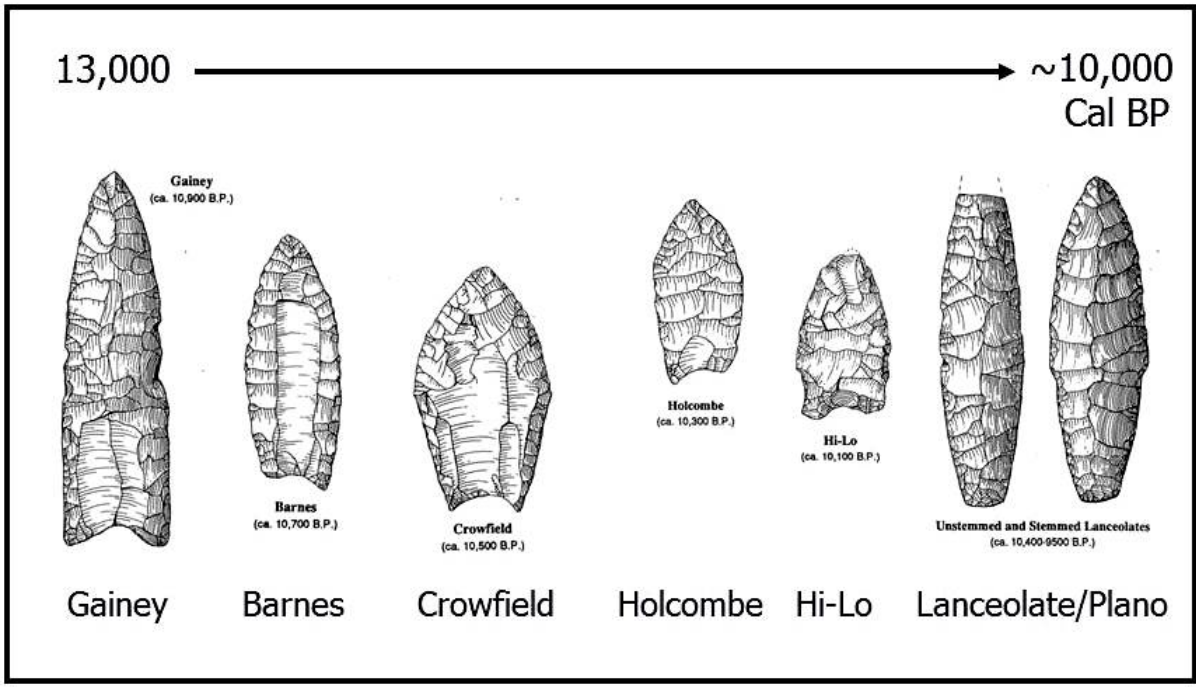


Figure 58. Paleoindian biface sequence for eastern Great Lakes region, circa 13,000-10,000 Cal BP (after Ellis and Deller 1990).

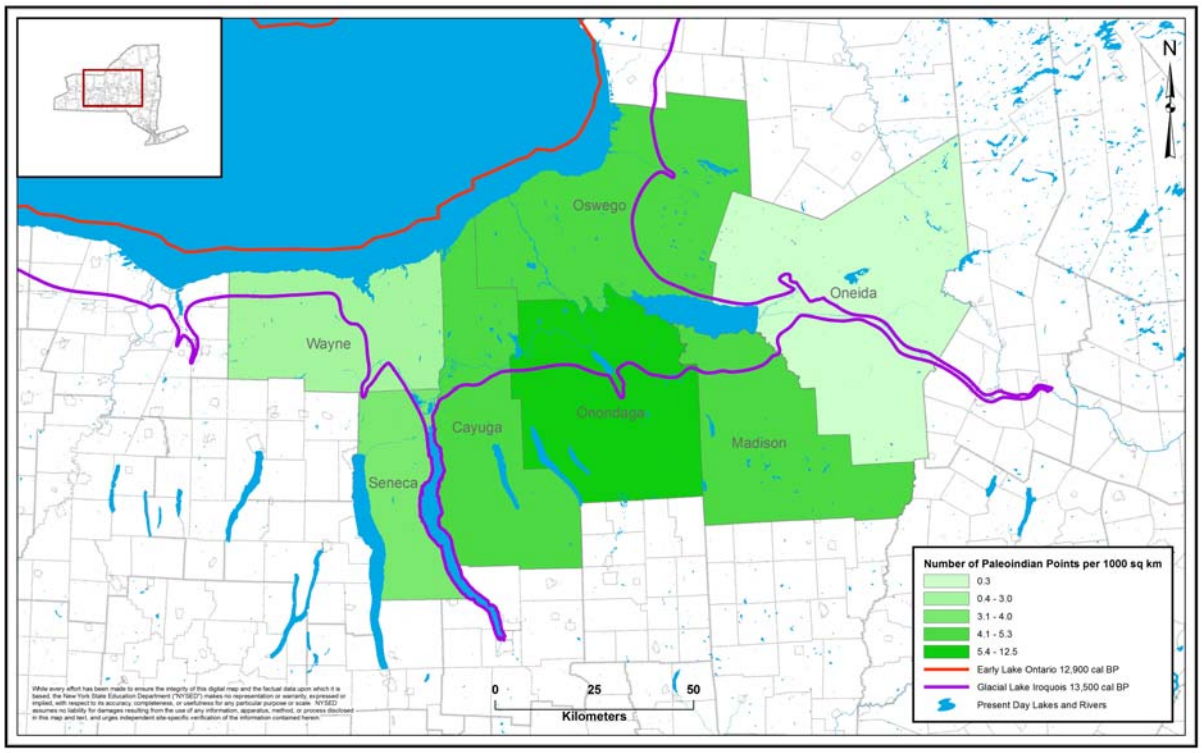


Figure 59. Density of Paleoindian points by county, per 1000 km², NYPID CNY study area (as of May, 2014).

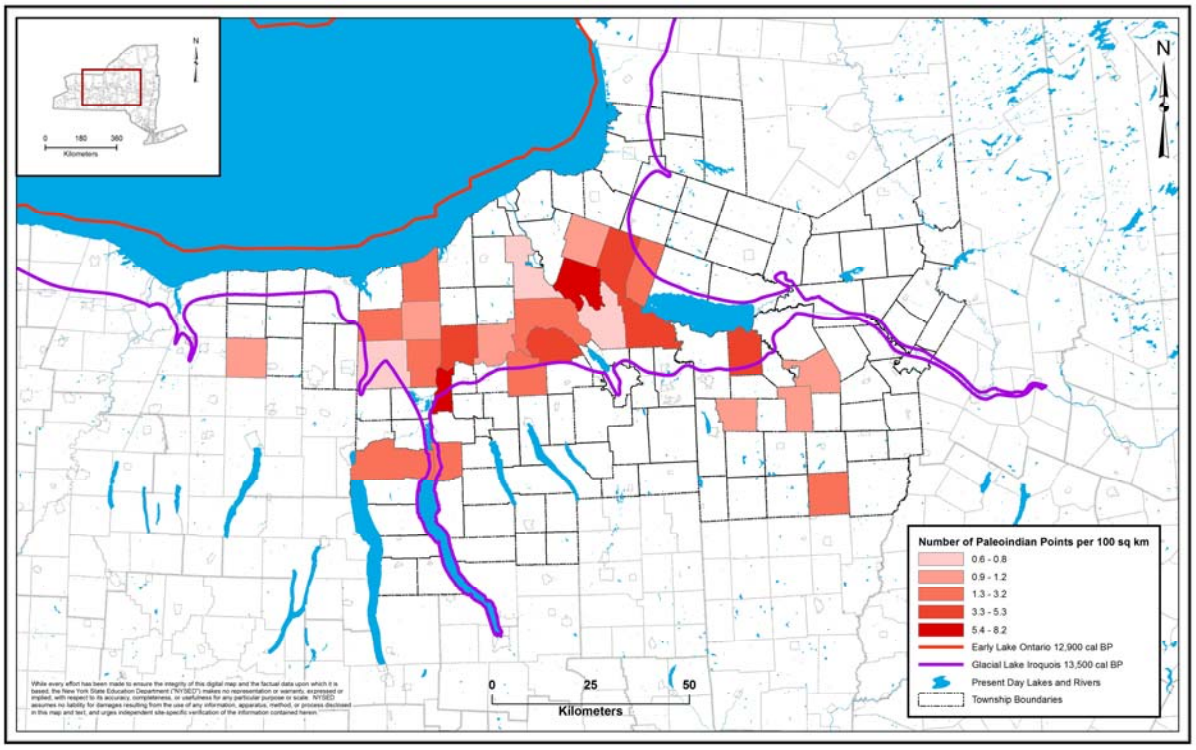


Figure 60. Density of Paleoindian points by township, per 100 km², NYPID CNY study area (as of May, 2014).

THE MONTEZUMA BRINE AQUIFER

Saltwater Springs in the Northern Cayuga Lake Trough

William Kappel – U.S. Geological Survey Emeritus, Ithaca, NY 14850
Dr. William Goodman – RESPEC Consulting and Services, Oviedo, FL 32766

Introduction

The Montezuma Wetlands Complex comprises disconnected ponds and marshes over a 50,000-acre area at the north end of Cayuga Lake in central New York (Friends of the Montezuma Wetlands Complex, 2014). Locally referred to as the Montezuma Marsh, the southern part of the wetlands is home to the Montezuma National Wildlife Refuge. The Marsh has a long association with the socio-economic development of the northern Finger Lakes region. The west-east flowing rivers (and New York State Barge Canal) that traverse the region, and their association with the north-south oriented Finger Lakes made for the initial transportation routes that spurred regional settlement, first by the Native Americans and then by European colonists. The presence of salt springs in the area north of Cayuga Lake further enhanced westward regional development by providing a much needed food preservative (salt) whose value was readily traded for other goods and services.

While the history of salt development in New York is primarily centered around Onondaga Lake and Syracuse, New York (Kappel, 2000), the salt springs along the east side of the Montezuma marshes were first noted by Jesuit missionary Father Raffeix in 1671 (http://www.newyorkroots.org/bookarchive/historyofnewyorkstate/bk7/ch4.html).

The development of the salt industry along the eastern side of the Montezuma Marsh in the early 1800s is explained in detail below, however these are but a few of the salt springs found in the Montezuma Marsh area. The Montezuma springs, much like the Syracuse salt springs, are sourced from the Syracuse Salt (Upper Silurian Salina Group). The glacial geology of central New York, which modified the landscape to create the Finger Lakes and their bedrock valleys created the connections between the buried halite deposits and the saltwater springs, which discharge from the edges of the glacial aquifer in the Cayuga Lake (Montezuma) and Onondaga Creek (Syracuse) valleys.

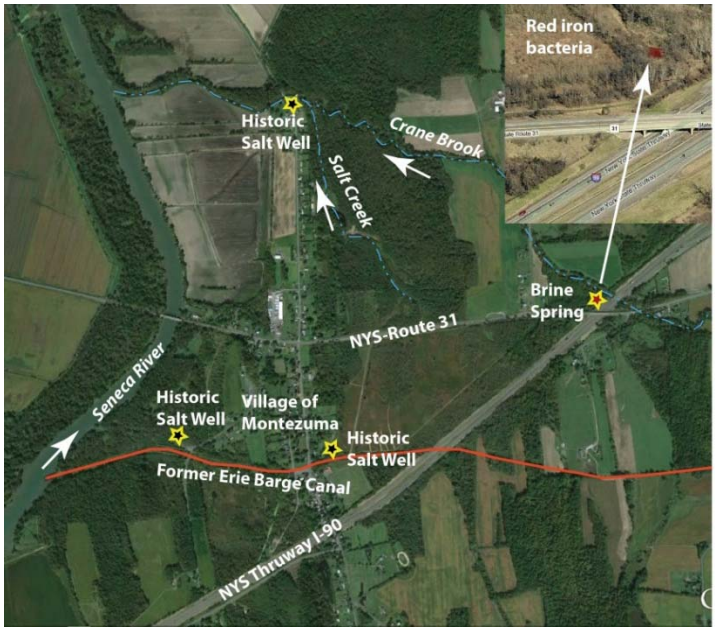


Figure 61. Village of Montezuma and associated brine wells and springs.

History of the Montezuma Salt Springs

Native Americans reportedly produced salt from holes dug into the Montezuma marsh at the foot of the high ground where the first white settlers established the Village of Montezuma in 1798 (Van Rensselaer, 1823). The first person to commercially produce salt at Montezuma was Comfort Tyler in 1798 (Storke, 1879). Tyler previously produced salt at the Syracuse salt works and used that experience to develop the salt springs at Montezuma. In 1807, General John Swartwout also set up salt works north of the Village of Montezuma near the junction of Crane Brook, Salt Creek, and the Seneca River.

Wells were dug up to 50 feet deep for the early production of salt [Van Rensselaer, 1823]. In about 1810, the first deep well was dug at Montezuma by the Cayuga Manufacturing Company on the west side of the village. The well extended to about 100 feet in unconsolidated glacial deposits and intercepted three discrete brine-bearing zones at depth. A second well, sunk to 80 feet in 1812–1813 and deepened to 120 feet in 1824, intercepted the same high-strength, brine-bearing zones (Beck, 1842).

The Montezuma brines were compared to the early Syracuse brines and some of the salt was produced by solar evaporation while the remaining was produced by boiling off the water and separating out the 'bitterns' (calcium sulfate and iron). Production from the wells and springs in 1822 amounted to between 16,000 and 20,000 bushels (Van Rensselaer, 1823). Montezuma salt operations continued for decades, but the early facilities were poorly constructed (Beck, 1842). Facilities were upgraded by New York State in 1840, and in 1858 the State took ownership of the salt manufactory properties and appropriated funds to redevelop the salt works (Storke, 1879). Wells were drilled east and west of the Village of Montezuma but the brine was not sufficiently strong enough to warrant further exploration.

Figure 62 – New York State (NYS) historic marker "Salt Springs" found along NYS Route 90, north of NYS Routes 5 & 20, and south of the NYS Thruway (I-90) and the Village of Montezuma.

A third well was drilled 1.5 miles southwest of the village (currently near NY Route 90 at Salt Block Road – see Figure 61) and reportedly yielded brine, which was equal in strength to that at Syracuse. This well produced for about 3 years until mechanical difficulties caused production to stop. An attempt to deepen the well to 1,000 feet was made but did not yield any additional brine or salt. Finally, in 1872, a company was formed to manufacture salt by solar evaporation at Montezuma (Storke, 1879), but the project was soon abandoned due to a high percentage of calcium and magnesium chlorides in the brine. The failure of these enterprises marked the end of the commercial salt production at Montezuma.

Cayuga Lake Trough Hydrogeology

A bedrock surface map (fig. 63) developed by the U.S. Geological Survey and the NYS Geological Survey shows the northward extension of the Cayuga Lake Trough beyond the northern end of Cayuga Lake. The trough trends northward and splits into two arms – one which trends northwest toward Savannah and Clyde, NY, and the larger eastern arm that trends both northward into the Town of Victory and eastward into Cayuga County north of Weedsport, N.Y. This latter branch over which the Seneca River flows is deeper than the northern branch. The northwestern arm may also have a bedrock valley extension into northern Seneca County and farther west into Wayne County. The Clyde River flows through the Montezuma Marsh from the west and thence to the Seneca River which flows east from the Montezuma Marsh.

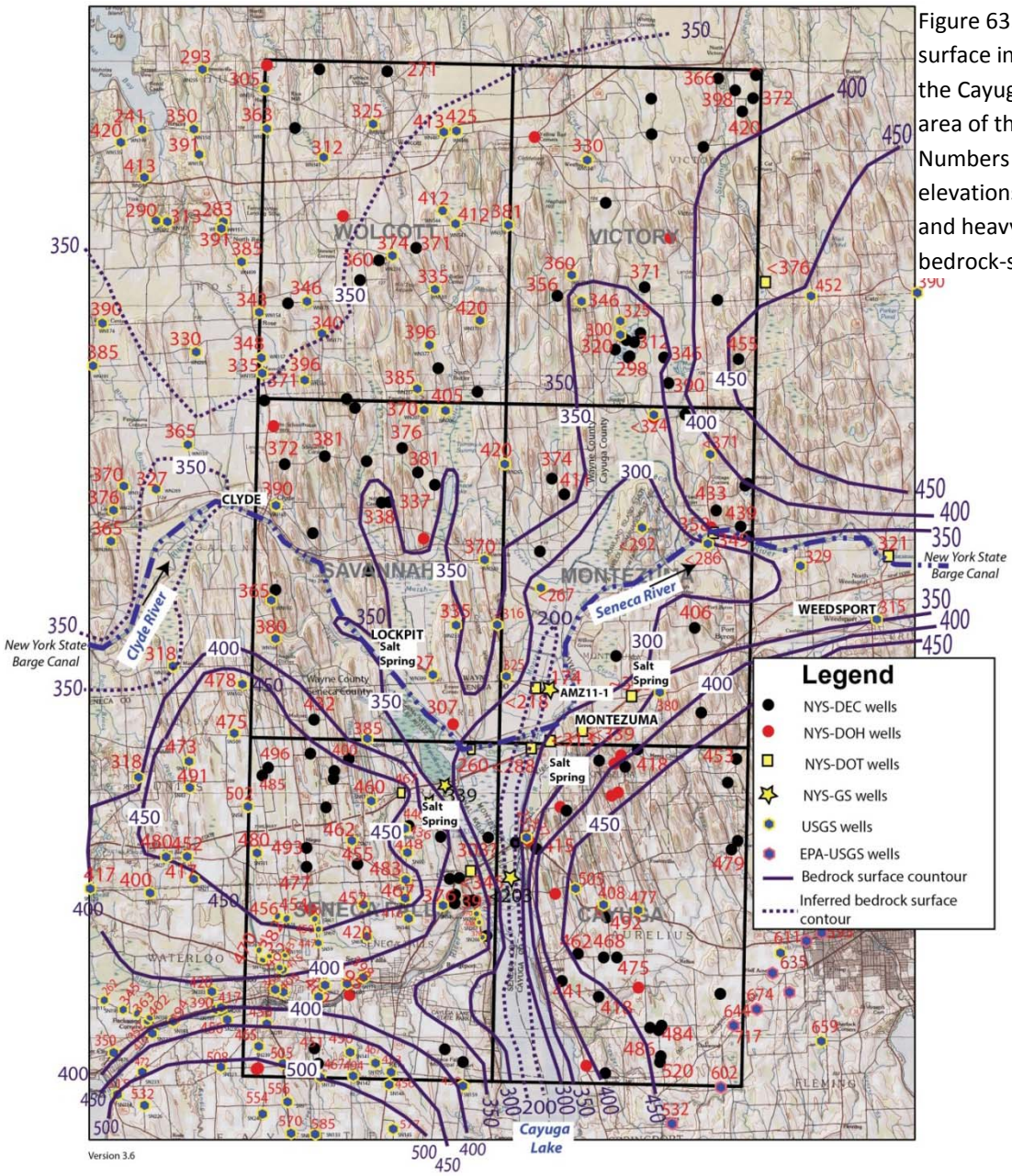


Figure 63. Map of the bedrock surface in the northern end of the Cayuga Lake Trough, in the area of the Montezuma Marsh. Numbers are bedrock surface elevations for individual wells and heavy black lines are bedrock-surface contours.

The brine springs of the Cayuga Lake Trough, which supported local salt production as described above, primarily occur along the eastern and western edges of the Trough (fig. 63). Beyond the documented brine springs near Montezuma, saltwater springs and the manufacture of salt was also reported along the western side of the Cayuga Lake Trough in the Town of Galen since the early 1800s (Spafford, 1824; Hall, 1843; French, 1860 and Merrill, 1893). Additionally, a deep well drilled near Lockpit, NY (fig. 62; east of Clyde, NY) in 1842, found saltwater at 400 feet (Transactions of the New York State Agricultural Society, 1854), and a salt spring exists in Black Lake, along the western side of the Montezuma Wildlife Refuge's main pool between the NYS Thruway and NYS Routes 5 & 20 (Linda Ziemba, U.S. Fish and Wildlife Service, oral commun., 2014). In 2011 the NYS Geological Survey drilled a deep hole (AMZ11-1, fig. 63) northwest of the Village of Montezuma along NYS Route 31 and found saltwater of increasing concentration below a marl layer at 30 feet down to bedrock at 208 feet below land surface.

Saltwater springs on the east and west sides of the Cayuga Lake Trough serve as a discharge zone for what is termed the Montezuma Brine Aquifer (Goodman and others, 2010). This brine reservoir is analogous to that of the Onondaga Brine Aquifer at Onondaga Lake in Syracuse, N. Y. (Kappel and Miller, 2005). Both aquifer systems are similar in that the halite (salt) beds in the Salina Group that underlie each north-south trending bedrock valley were exposed during glacial erosion of their respective bedrock troughs. The dissolution of the halite during glacial erosion created the high salinity brines, which eventually flowed northward in confined aquifers situated near the bedrock floor of each valley. These brines then discharged locally as salt springs in marshy environments in their respective valleys. The halite beds themselves are not present in the northern part of Cayuga Lake, nor in the northern part of the Onondaga Creek valley.

Montezuma Brine Aquifer Boundaries

(From Goodman and others, 2010)

In the region around Cayuga Lake and farther north throughout the Montezuma Marsh, freshwater in drinking water wells can occur to depths of 300–400 feet below land surface, with a shorter distance to the freshwater-saltwater interface being found on the valley floors. The bedrock below the Montezuma Marsh is generally the Hamilton Group shales, which are underlain by carbonates – the Onondaga and Helderberg Limestones. Between these carbonates and those of the Lockport Group, the Salina Group shales are found.

	Group	Unit
DEVONIAN	Hamilton	Tully Ls. Moscow Sh. Ludlowville Sh. Skaneateles Sh. Marcellus Sh.
		Onondaga Ls.
	Tristates	Oriskany Ss.
	Helderberg	Manlius Ls. Rondout Dol.
SILURIAN		Akron Dol.
	Salina	Camillus Sh. Syracuse Sh.(Salt) Vernon Sh.
	Lockport	Lockport
	Clinton	Rochester Sh. Irondequoit Ls. etc.....

The vertical boundaries of the brine aquifer system in bedrock are stratigraphically controlled. Low-permeability shale zones and bedding-plane aquifers are present within the Salina Group. These low-permeability shales serve as aquitards, which isolate several bedding-plane flow zones within the lower and middle part of the Salina Group. The lowest part of the Salina Group apparently isolates any fluids from the underlying Lockport Group carbonates from the brine flow system within the Salina

Group. The Camillus Shale appears to serve as the upper confining unit for the brine aquifer and may separate the saline waters in the Salina Group below, from mixed-formation water compositions in the carbonate-dominated Onondaga-Helderberg interval above. The low-permeability Hamilton Group shales, which overlie the Onondaga-Helderberg carbonates, appear to serve as a barrier to freshwater flow into deeper units, thereby precluding significant dilution of brine in the underlying Salina Group. Beneath the shale cover, the Onondaga and Helderberg formation waters are also mostly brackish, with some hydrogen sulfide occurring in the brackish water zone, while saltwater is prevalent to the east of the Cayuga Lake Trough.

Where salt beds are found, downward, cross-formational flow to recharge the brine aquifer is limited. Salt-bed preservation, alternating shale aquitards and bedding-plane brine-flow zones, and the hydrogeochemistry of the upper 1,500 feet of the subsurface in this area are more consistent with Pleistocene-age, down-dip, high-pressure, subglacial fluid injection through the scarp slope of the basin margin (the Onondaga Limestone escarpment) as the recharge mechanism for the deep, subsurface bedrock portion of the Montezuma Brine Aquifer system. The current driving force for high-strength brine that discharges at the edge of the Montezuma Marsh may simply be ongoing pressure relief (i.e., “brine squeeze”) of bedding-plane flow zones that were overpressurized during the Pleistocene. If true, there may be little cross-formation communication between brine-bearing zones south of the Onondaga escarpment which trends in an east-west fashion across upstate New York.

Down-dip glacial injection of fluids is evident along the margins of the Alberta and Williston Basins in western Canada and the Upper Midwest of the United States where they faced advancing glaciers (Bekele et al., 2003; Person et al., 2007; Grasby and Chen, 2005). So, it is reasonable to expect that the Appalachian Basin, in which the Montezuma (and Syracuse) brine aquifers are found, shares a similar chapter in its fluid history (Bense and Person, 2008). Additional geochemical analyses for major ions and dissolved solids content, as well as tritium and stable isotopic analyses on fluids in the Montezuma Brine Aquifer would certainly aid in defining the boundaries between Pleistocene-age brines and mixed fluids containing a modern water component.

APPLICATION OF GEOPHYSICAL SURVEYS IN THE MONTEZUMA WETLANDS COMPLEX: Monitoring the changes in sediment and locations of subsurface waters

Dea Musa, Laura Sherrod

INTRODUCTION AND BACKGROUND OF SITE

As described earlier the landscape present today at the Montezuma Wetlands Complex (MWC) of Central New York has been developed by the advance and retreat of the Laurentide Ice Sheet. The resulting landforms and deposits now control subsurface water flow. The development of the wetland complex, which covers more than 50,000 acres, occurred when topographically low areas flooded and filled with sediments and organic muck in the eutrified proglacial lake basins.

Remnant ice-contact deposits such as eskers and fans near Duck Lake (Figure 64, Location A) rise above the flat topography of the wetland and are seen throughout the complex. Eskers, formed from a subglacial stream flowing south under the Ontario Lobe, consists of coarse grained gravel deposits that onlap a drumlin of glacial till in the east and terminate at the Duck Lake Ice Margin Fan. The southern edge of the wetland (Figure 64, Location B) has many interesting hydrogeological features. Though many different species of wildlife and vegetation have been observed at the MWC, the presence of certain salt marsh species of terrapin and shrubs are of particular interest (Knightly, 1993). Today these marshes are rejuvenated by rain and runoff but the MWC has long been known to have brine pockets present. The source of this brine is difficult to trace to its source. The geologic three dimensional (3-D) framework of sediments in the MWC greatly impacts the flow of groundwater. Thus by mapping the subsurface glacial deposits, the presence and pathway of brines may be identified or better understood.

Geophysical techniques have become an important tool for 3-D mapping of glacial deposits (Gerber et al., 2007; Sherrod et al., *in press*). The contrast of resistivity between typical glacial deposits saturated by fresh groundwater (greater than 10 Ωm) and glacial deposits saturated by brine (less than 1 Ωm) facilitates the detection of the boundary of the brine plume (Carter et al., 2007), (Evgeny and Ozorovich, 2006), (Shagar et al., 2006), (Taniguchi et al., 2007).

Within the MWC during the summers of 2012 and 2013, resistivity and ground penetrating radar (GPR) surveys were performed to identify geophysical responses of glacial features at the Duck Lake site in 2012 (Figure 64, Location A) and to detect the flow path of brine at the southern site in 2013 (Figure 64, location A). The goal of the work was to obtain 3-D views of the sediments in the MWC through resistivity and GPR surveying. While the GPR results are particularly useful to identify lateral changes in sediment composition and texture at the Duck Lake site, the resistivity images at the brine spring site show the spatial variation of the fresh/salt water interface in the subsurface. These images reveal details of the glacial framework and geologic controls on the hydrogeology.

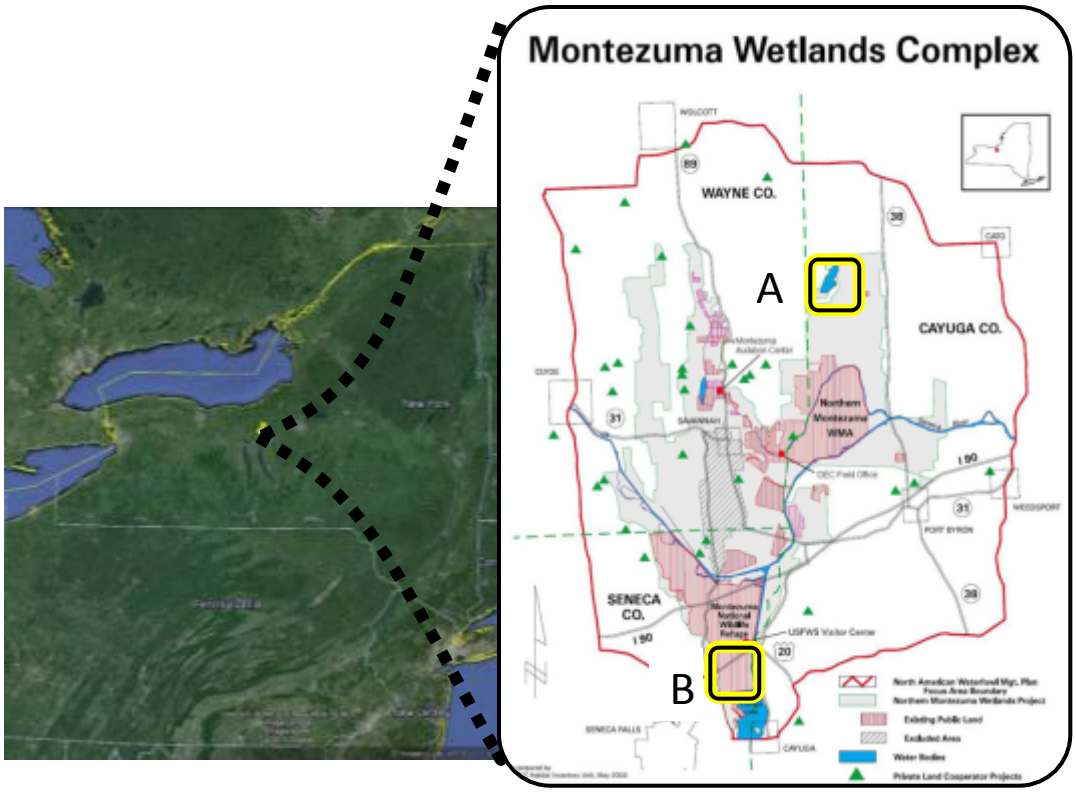


Figure 64. Location of Montezuma Wetlands Complex in Central New York, just north of the Finger Lakes. Site A represents the location of the surveys at Duck Lake (2012), and site B represents the location of the brine spring mapping (2013).

(Source: http://www.dec.ny.gov/images/regions_images/nmwmamap.jpg**)**

THEORY

Ground Penetrating Radar (GPR)

Ground penetrating radar (GPR) is a geophysical technique that uses pulses of electromagnetic energy to image the subsurface. Reflections are caused by changes in the electrical properties of materials in the subsurface. Thus, subsurface stratigraphy with layers of differing electrical properties can be imaged through a GPR survey. High resistivity materials tend to have a stronger GPR response while low resistivity materials tend to cause the signal to become faded or attenuated. Sedimentary structures such as channels, faults, and cross bedding can be imaged as the edge of these structures produce a nice reflection. Discrete objects in the subsurface, such as boulders or pipes, produce hyperbolic reflections in the GPR profiles as the electromagnetic waves diffract off the edges of the object.

The center frequency used in the GPR surveys were 100 MHz, 200 MHz, and 400 MHz. Figure 65 shows the 200 MHz GPR antenna, as used in the 2012 GPR survey at the Duck Lake site. The depth of penetration and resolution are impacted by the frequency and there is an inverse relationship

between these two factors; therefore, higher frequency has a higher resolution, but a lower depth of penetration.

GPR has several limitations. The quality of the signal is impacted by the electrical conductivity of materials. As the subsurface conductivity increases, the penetration depth and resolution decreases, often producing profiles that appear washed out or faded. Another limitation of GPR is that it requires a flat, smooth surface to produce high quality results. The electromagnetic energy is slowed by the presence of water, thus giving a lower quality and shallower result in saturated environments when compared to results in similar unsaturated sediments. The research performed in 2013 using GPR was in an area of relatively saturated soils, causing a shallower profile penetration depth and poorer resolution.

Electrical Resistivity

Electrical resistivity surveys are performed to measure the apparent resistivity (inverse of conductivity) of the subsurface in Ohm*meters (Ωm). Resistivity measurements employ two electrode pairs. The first pair is used to inject a direct current into the subsurface and the second pair is used to measure the resulting change of potential, or the voltage. To change the depth of measurement, increasing the electrode spacing will result in a deeper resistivity measurement.

Resistivity surveys can be used to identify transitions in subsurface sediments (Crook et al., 2008), between unsaturated and saturated zones (Oliverira Braga, 2006), and between freshwater and saltwater conditions (Shagar et al., 2006). Sediments have a wide range of resistivity values, with lower resistivity values in clays (4 - 35 Ωm) while sands and gravels range in the upper resistivity values (30 - 1000 Ωm) (Reynolds, 2011). This large contrast in resistivity of sediments provides a basis for differentiating subsurface stratigraphy based upon resistivity surveys.

The saturation state plays a large role in the overall resistivity. Electrolytic conduction through saturated sediments tends to lower the bulk resistivity compared to unsaturated conditions. Thus the resistivity will decrease significantly as the measurements pass from the vadose to the saturated zone. Also, brine saturated sediments will have a much lower resistivity than freshwater saturated sediments.

Figure 65. Bistatic 200 MHz antenna pulled behind a field vehicle at the Duck Lake Site 2012.

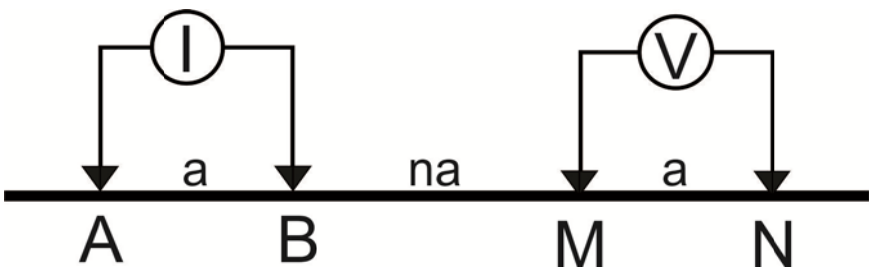


Figure 66. Dipole-dipole array for the electrical resistivity surveys performed at the brine water site.

FIELD METHODS

Site A: Duck Lake

GPR

The chosen survey path followed a residential road that passed over glacial features of interest (Figure 67). Resistivity surveys require a straight path and direct contact with the ground by the stainless steel electrode stakes used for this technique. Due to this and the aforementioned requirements of straight line surveying, the winding asphalt roads at Duck Lake prohibited the use of resistivity surveying. The GPR surveys were performed with a GSSI SIR 3000 system using the 100, 200, and 400 MHz antennas with ranges of 460 ns, 100 ns and 45 ns respectively. The 100 and 200 MHz antennas were operated in bistatic mode with continuous data collection as they were pulled behind a field vehicle. The 400 MHz antenna was pushed in the standard GSSI cart and was only performed over the esker deposit, approximately 1.7 km. The GPR survey across the fan shaped deposit was approximately 1.9 km and across the esker and drumlin was 2.7 km. The distance was recorded using a survey wheel pulled behind the GPR. The data were later corrected more precisely to the GPS start/stop locations.

Site B: Subsurface Brine

Several field methods were performed to map the changes in conductivity of the soils and identify the fresh marsh water and brine boundary. Borehole logs, resistivity profiles and GPR profiles were collected along two perpendicular transects (Figure 68). The transects ran EW and NS in the Montezuma National Wetlands Reserve on service roads present prior to this survey allowing for a convenient, obstacle free path, with little change in topography. When constructed, this path had been leveled with fill material ranging from 1.2 to 3.2 m in thickness.

Borehole Logging

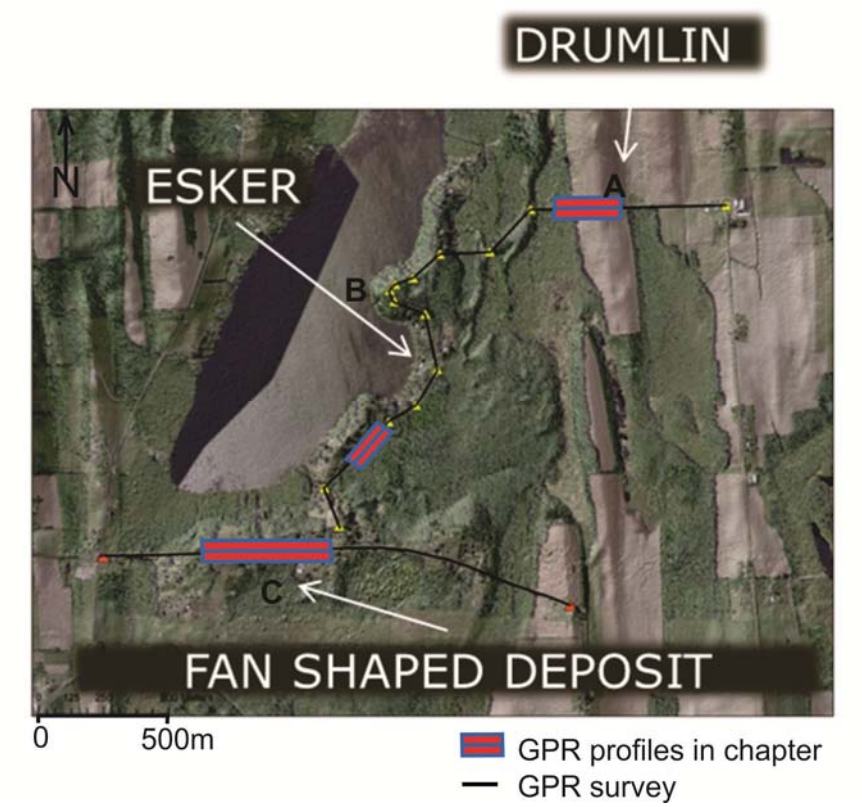
Six boreholes were drilled by the NYS Geological Survey using a track mounted Geoprobe with continuous core sampling techniques (Figure 68). Four boreholes were located at approximately equal spacing along the EW transect and the final borehole was located at the northernmost edge of the NS transect. The bore holes ranged from 9.8 to 30.5 m in depth.

Electrical Resistivity

Resistivity data were collected along 21 consecutive overlapping resistivity lines oriented in the EW direction going westward for a total of 1.8 km. 17 consecutive overlapping resistivity lines were collected along the NS transect moving northward 1.4 km (red lines in Figure 68). All the NS lines overlapped each other by one half of the survey length. The EW lines also overlapped each other, but the distance of overlap varied due to the curvature of the road. An MPT DAS-1 Electrical Impedance Tomography System with 32 electrodes was used to survey the site. Profiles were 155 m in length and a dipole-dipole array was used with a base electrode 'a' spacing of 5 m. This spacing was expanded to spacings of 10, 15, and 20 m, and a maximum 'n' value of 6 to achieve the maximum depth of penetration possible with the available equipment.

GPR

GPR was performed along the same transects as the resistivity lines, for a total of 5.41 km of GPR profiles (yellow lines in Figure 68). Some segments of these profiles were surveyed more than once at differing frequencies to test the depth of penetration and quality of resolution. Data were collected using a GSSI SIR 3000 system with 100 and 200 MHz antennas in continuous collection mode with a record length set between 90 and 120 ns. The antennas were pulled manually and with a field vehicle, as site accessibility allowed. The distance was recorded using a survey wheel and later corrected more precisely to the GPS locations using 40 points that mapped the start/stop of the files recorded with the GPR.



.Figure 67. Duck Lake, site A from Figure 64, with glacial features labeled, black lines showing the location of GPR surveys, and red/blue lines highlighting the profiles of the GPR lines which are presented in the results.

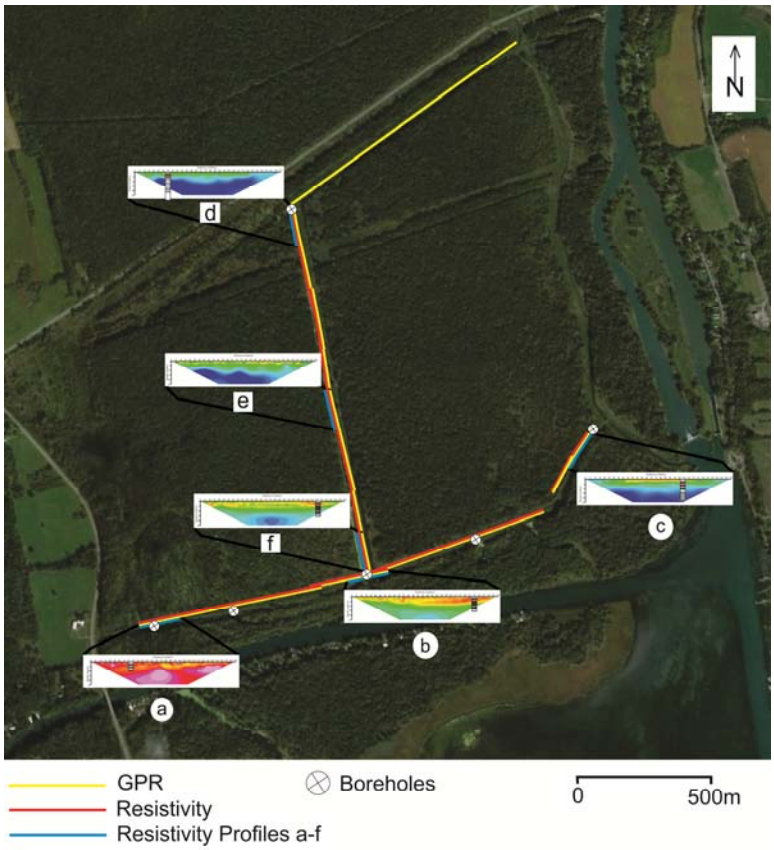


Figure 68. Brine springs mapping area, site B from Figure 65. Yellow lines represent the GPR surveys performed and red lines represent the resistivity performed at the site. Six representative resistivity profiles are shown on the map with their locations represented by blue lines in the survey area.

RESULTS

Site A: Duck Lake

Three GPR profiles from the Duck Lake site are presented in Figure 69, with the location of each profile labeled in Figure 67. The most notable feature in the first profile (A), which passes over the drumlin, is the bright ringing reflector directly under the bridge at 80 m. This series of reflections is an artifact of the metal within the bridge and may mask true subsurface anomalies at depth. Also visible at this location, on either side of the bridge, is a channel-shaped reflection which extends to a depth of approximately 3 m at its center and reaches a width of 80 m. There are numerous changes in the strength of the reflections over the length of this 260 m profile. Zones of high attenuation can be seen from 20 m to 40 m and 235 m to 260 m. Stronger reflections are apparent at several other locations, including 0 m to 20 m, 125 m to 140 m with dipping reflectors, and 215 m to 235 m with numerous intersecting hyperbolic reflectors.

Within the esker the signal is highly attenuated within the top 2.5 m of the profile, however, the lower 3 m exhibit very well-defined hyperbolic reflections. These hyperbolas range in horizontal expanse, with the widest and most distinct centered at 13 m and 112 m. Between these hyperbolas, which are highlighted in Figure 69B, the subsurface hyperbolic reflections are less clear as the tails of the hyperbolas cross and interfere with each other.

The final survey profile from the southern part of this site is of the fan shaped deposit (Figure 69C). The profile was not corrected for topography as there was not a significant change in elevation. Although the other locations at the Duck Lake site have a depth of 5.2 m with good data quality throughout the profile, to get a good resolution at this location with a 200 MHz antenna, the depth of the signal was decreased to 2.5 m. This profile shows strong sloping reflectors in the middle (20 to 290 m), but is highly attenuated throughout the rest of the profile, with a slight increase in signal strength between 360 and 420 m.

Site B: Subsurface Brine

Changes in resistivity were observed both in the EW (Figure 70) and NS (Figure 71) transects with the resistivity ranging from less than 1 Ωm to greater than 100 Ωm. Overall, the resistivity gradually decreased moving northward and eastward along the respective transects. The depth of subsurface material mapped with resistivity was 25m. High resistivity layers (greater than 20 Ωm) are present in the upper few meters of most of the profiles, but as depth increases the resistivity decreases. The western-most profile is dominated by resistivity values greater than 50 Ωm. The eastern-most profile and the profiles along the NS transect show resistivity values of less than 10 Ωm across most of the profile length.

The boreholes ranged in depth from 6.9 m for the borehole at resistivity profile **a**, to 30.5 m for the borehole at resistivity profile **d**. On average the borehole data showed from 1.5 to 3.2 m of fill

material closest to the surface, followed by an organic layer of peat and marl ranging from 0 to 4.7 m. This was followed by a 0.2 to 1.0 m layer of clay, and the final layer was composed of mostly sands and gravels until the end of the bore hole where a diamicton layer or rhythmic silts and clays were encountered.

Observing the GPR profiles, although the lower frequency antennas provided a greater depth of penetration, the 100 MHz data from this site did not provide any significant information due to the low resistivity nature of the subsurface. The 200 MHz data, however, produced a profile 3.2 m in depth. Unfortunately, this depth range was mostly confined to the fill material at the site and did not expand the understanding of the overall hydrogeology. Figure 70 displays GPR profiles that correspond to the resistivity profiles at locations **a** and **c** at this site. A noticeable change in signal of the sediments was observed at the transition between fill materials, as logged in the boreholes, to the more natural sediments of the wetland in the deeper sections of the profile.

DISCUSSION

Site A: Duck Lake

The channel-shaped feature beneath the bridge in Figure 69A is interpreted and labeled as the base of the swamp. The large scale of the depression, 80 m wide and 3 m deep, show the stratigraphic succession of sediments in the swampy area under the bridge. These sediments are likely fine grained and may represent an area of restricted groundwater flow in this glacial environment. This basin may have a large impact upon the subsurface water flow. The drumlin was crossed perpendicular to its long axis. The area of higher topography represented by this transect, shows jumbled reflections representative of diamicton. This poorly sorted glacial sediment produces discontinuous GPR reflections from the numerous diffraction hyperbolae which reflect off the discrete larger sediments.

The survey represented in Figure 69B was performed parallel to the long axis of the esker. Although the upper few meters of the profile are attenuated, likely due to the asphalt road on which these surveys were made, the lower half of the profile shows clear reflections from the coarse-grained sediments of the esker. The large hyperbolic signatures most likely represent cobbles and boulders. The GPR reflections at this location indicate braided gravel deposits interspersed with these larger cobbles and boulders.

The third feature, the fan shaped deposit, was interpreted on a LIDAR image as a glacial fan marking an ice margin. The signals from the GPR were of interest to determine the types of sediments present within this feature. Unfortunately, the results of this survey were highly attenuated and the depth of penetration was much shallower than the previous two survey profiles. These characteristics indicate a subsurface with a low resistivity, which may be caused by the asphalt road, clay-rich sediments, the presence of road salt, John Wiant driving too fast, or some other low resistivity anomaly. The only features of interest found within the GPR profile across the fan-shaped deposit

were the dipping reflectors between 200 m and 290 m. As the survey was performed along a road and the depth of these reflectors extends only 1.5 m below the ground surface, a likely interpretation of these reflectors is that it may be fill material from the construction of the road. Borehole logging to a depth of 2 m could help to confirm the interpretations of the sediment and fill at this location.

Site B: Subsurface Brine

The resistivity profiles show the transition to lower resistivity values towards the northeast area of the site. Resistivity values of less than 1 Ωm are indicative of salt water. Thus the dark blue zones of the resistivity profiles are interpreted to contain brine. Clastic sediments such as clays, tills, sands, and gravels are expected to have a resistivity range of 4 to over 100 Ωm when saturated with freshwater. As the resistivity values obtained in these eastern and northern profiles match those expected from brine-saturated sediments, the path of the brine can be tracked across the survey area. The low resistivity anomaly in the EW transect begins to become apparent along the base of the profile (20 m deep) in profile **b**. It becomes increasingly visible toward the east and by the eastern edge of the survey can be seen at a depth of only 10 m. Likewise, the low resistivity anomaly in the NS transect is apparent at depth in the southern reaches of the survey line (15m depth) as at profile **f**, but becomes increasingly shallow, reaching a depth of only 5m in the northern-most profile (d).

The diamicton layer is shallowest to the west. The resistivity is negatively correlated with the depth of the diamicton layer; as the depth to the diamicton layer increases the low resistivity anomaly becomes increasingly apparent. Thus, the sediments present between the land surface and the diamicton layer may contain brine. By comparing the borehole logs with the resistivity profiles, it is apparent that the low resistivity zones interpreted as brine saturated sediments occur in finer grained sands mixed with gravels.

The GPR data profiles were too shallow or too attenuated to identify the brine-saturated sediments at this site. There is a distinct interface observed in the GPR profiles of **a** and **c** at a depth of 2 m. This interface corresponds to the transition between fill material and the underlying wetland sediments. Additionally, sloping reflections are apparent in the upper 2m of the profile. These sloping reflectors are interpreted as fill material put in place when the service roads were constructed. The GPR results are useful for identifying the depth of fill material along the roadway, but not for locating brine-saturated sediments at depth.

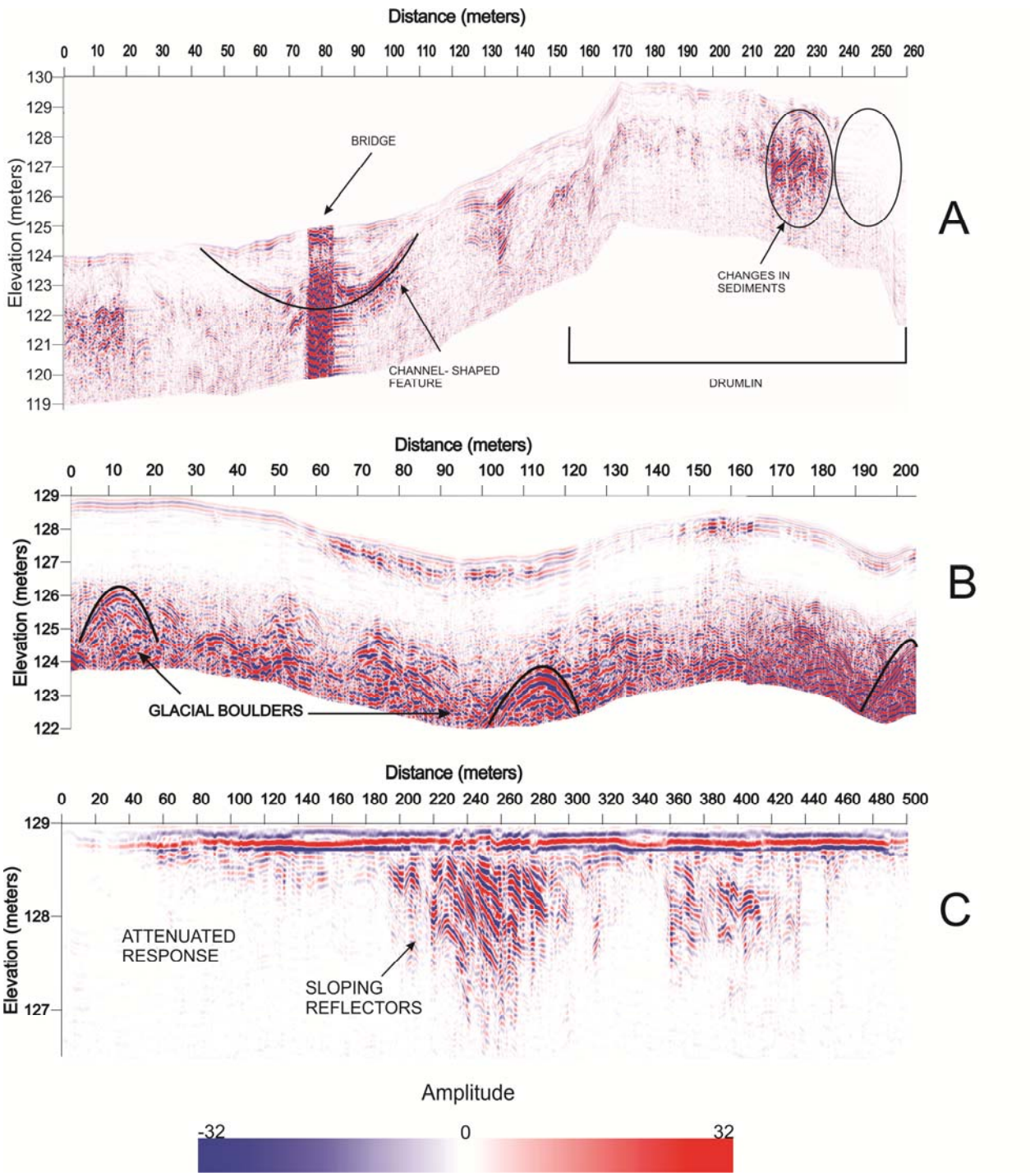


Figure 69. GPR profiles from the Duck Lake Site. The top profile (A) is across the drumlin, the middle profile (B) is across a part of the esker, and the bottom profile (C) crosses the fan shaped deposit.

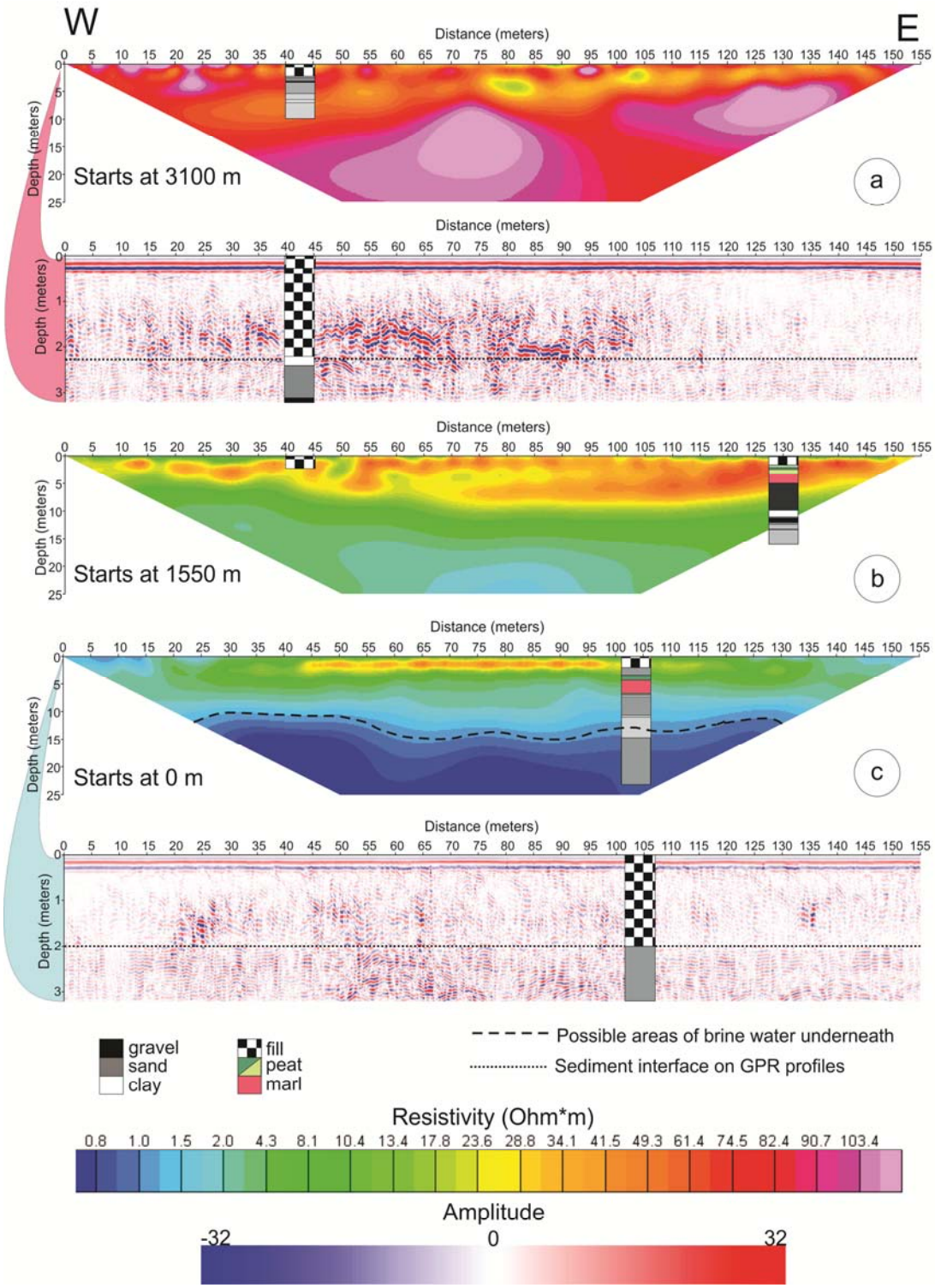


Figure 70. Resistivity and GPR profiles along the EW transect, with borehole logs overlain at the appropriate locations. The top resistivity line (a) is on the far west side, the middle resistivity line (b) is mid-way along the EW transect, and the bottom resistivity line (c) is on the far east side of the site. The corresponding GPR profiles are included for locations a and c. Note: the depth scale differs between the GPR and resistivity profiles. The possible areas of brine water saturated sediments are underneath the dashed lines and the sediment interface of the GPR profiles is indicated with a dotted line.

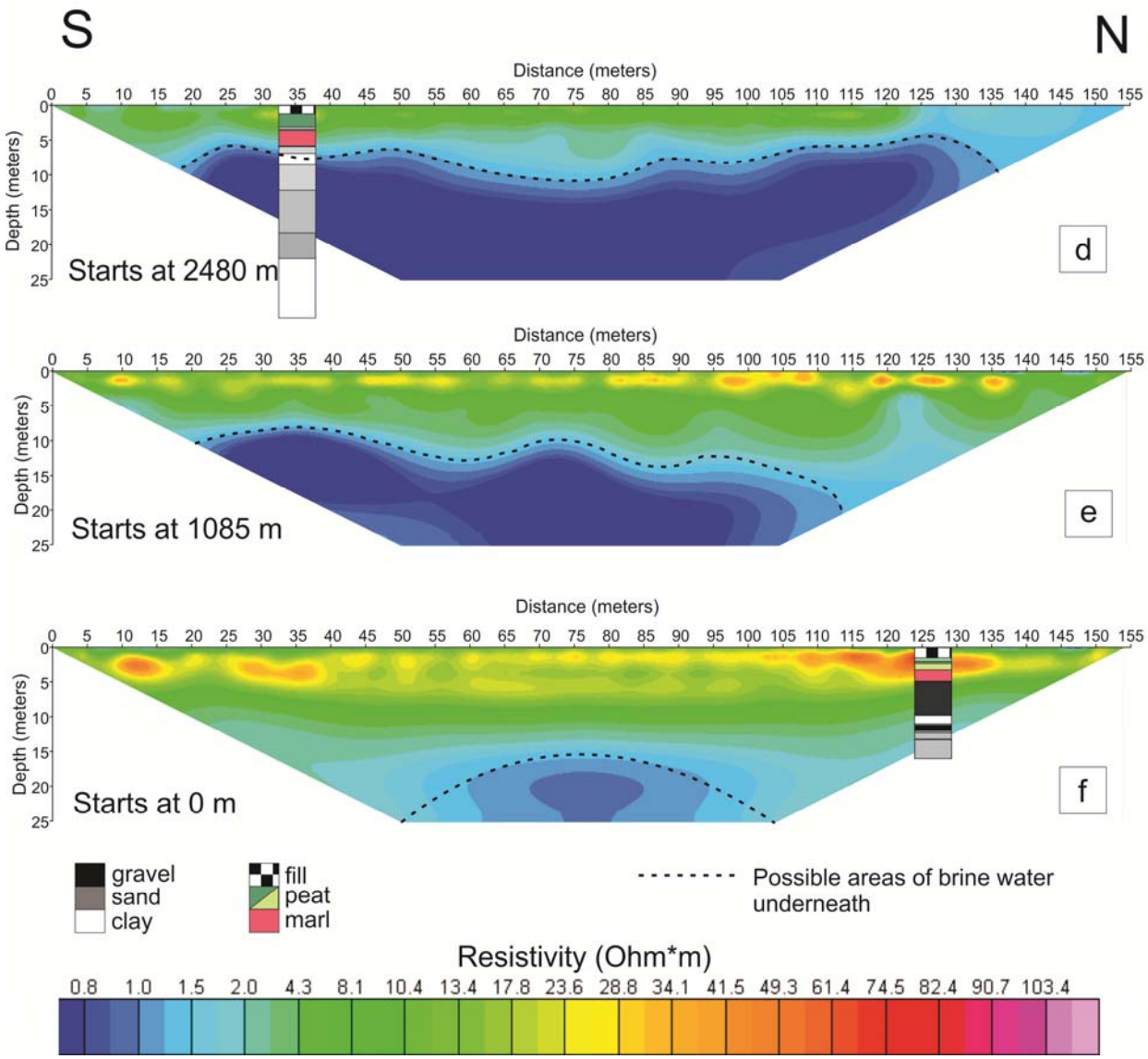


Figure 71. Resistivity profiles along the NS transect, with borehole logs overlain at the appropriate locations. The top resistivity line (d) is on the north-most end, the middle resistivity line (e) is mid-way along the NS transect, and the bottom resistivity line (f) is on the far east side of the site. The possible areas of brine water saturated sediments are underneath the dashed lines.

CONCLUSION

Changes in sediment were determined using a combination of different geophysical methods and the physical constraints of boreholes. The measured properties of the sediments as determined from the resistivity and GPR surveys provide information about the subsurface glacial deposits within the MWC. At Duck Lake, the GPR profiles of different glacial sediments yield insight into the formation of these features and how the sediments within them may impact groundwater flow. At the brine springs site, brine water was interpreted within the sediments at a depth of 5 to 25 m where resistivities of < 1.0 Ωm were recorded. These surveys show a greater concentration of brine-saturated sediments in the eastern and northern sections of the transects and provide a novel look at mapping saltwater springs in wetland environments. This work has provided a new and unique way of looking at glacial sediments within the MWC.

Using LiDAR to Map Karst Features in Cayuga County, New York

Andrew Clift

Introduction

Commonly within glaciated landscapes are notable topographic depressions. Often these depressions are caused by buried blocks of ice "melt out" and result in a hummocky topography associated with kettle and kame landscapes. In this current mapping study for Cayuga County, as part of our background research, it was noted by Shumaker (1957) that while completing a study of till composition and related surficial landforms, some depressions around the Union Springs quadrangle may actually be the product of karst processes partially masked by glacial deposits and not true kettles of glacial origin. Similarly, recent mapping projects conducted by the Ohio Geological Survey have utilized LiDAR to identify karst features buried by glacial drift (Aden 2013). We took this opportunity to incorporate existing LiDAR topography to reevaluate the earlier suggestions of Shumaker (1957).

Background

Karst landscapes are prevalent throughout New York State, and occur naturally along bedrock with characteristics conducive to their development (Figure 72). Innovative computational methods for identifying karst topography have been effective in delineating some of these features, based on high resolution airborne LiDAR (Light Detection and Ranging) surface elevation data (Aden 2013, Doctor and Young 2013, Shrake et al. 2010). Various processing techniques, including the application of geographic information systems (GIS), have allowed researchers to program functions that can isolate potential karst features based on specific geospatial criteria. This approach is useful as a preliminary method of assessing karst landscapes, and requires subsequent field verification in order to confirm the classification of such features.

In this study, Cayuga County was evaluated for the presence of karst topography, specifically sinkholes, due to the extensive underlying presence of carbonate rocks. Sinkholes are defined as "the surface and near-surface expressions of the internal drainage and erosion process in karst terrain, usually characterized by depressions in the land surface" (Beck 2005). The two primary processes that are responsible for sinkhole development are (1) the downward transport of surface material along solutionally enlarged channels; and, (2) the collapse of the rock roof over large bedrock cavities (Weight 2008). These process-oriented interpretations allow sinkholes to be classified into five major groups: solution sinkholes, cover-collapse sinkholes, bedrock or cave collapse sinkholes, cover-subsidence sinkholes, and buried sinkholes. The surface expression of each of these varies in size, shape and depth, and can form gradually over time or develop suddenly. Sinkholes can pose significant risks to human safety, infrastructure, and groundwater pathways, and as such, are of primary concern to the New York State Geological Survey's mission to map and document geologic hazards in New York State.

Methods

Digital mapping involving a conjunctive interpretation of 3-meter LiDAR data, aerial photography, bedrock maps and well log data, has been effective in identifying potential sinkholes in Cayuga County. The basic software platform used in this investigation was Esri's ArcMap (v. 10.1), which offers a variety of geostatistical tools fundamental to this type of study. These processing tools, combined with peripheral data and aerial photography, allowed for the development of computational and interpretive methods useful for delineating sink features and differentiating between natural and artificial depressions. Different analytical methods including LiDAR preparation and processing may be necessary or useful based on the quality and completeness of the data, and should be respective of the scale and extent of the study area. For example, small gaps in LiDAR data may need to be resolved by using interpolation techniques. In other instances, artificial features such as man-made structures can thwart certain geostatistical operations and require manual modification of these structures from the LiDAR. The latter issue is particularly important to consider when creating hydrologically-correct digital elevation models (DEM), which are essential for delineating depression features.

Bare-earth LiDAR data was used to create a DEM that was subsequently used to produce hillshade and slope rasters. The next step was to use the "fill" spatial analyst function on the LiDAR DEM to produce a raster in which all the sinks are filled to their spill point. The original unfilled DEM was then subtracted from the filled DEM in order to generate a raster that represents the depth of depressions on the original surface. Various raster calculations based on establishing geometric criteria were implemented for the purpose of removing anomalous sinks caused by negligible surface resolution errors in LiDAR data, and artificial features. The remaining sinks were then converted to polygon features and manually filtered using peripheral data including high resolution aerial photography.

Results

The results of the DEM analysis found over 150 potential sinkholes distributed throughout Cayuga County (Figure 73, 74) that were identified through prescreening geostatistical processing and subjective filtering; the latter necessitated due to limitations of both automatic delineation techniques and DEM modification. The visual context of sink raster data by use of aerial photography, hillshade and slope rasters aided in distinguishing between natural and artificial sinks, the latter commonly caused by unnatural infrastructure relief (such as bridges and culverts) that obstruct drainage on the LiDAR. Due to time and processing constraints, LiDAR preparation was minimal in respect to modifying these features beforehand, on a county-wide scale. As a result, manual filtering was applied in order to remove erroneous depressions. Overall, the geospatial distribution of these results seems agreeable with the geography of underlying carbonate bedrock units in Cayuga County. Future fieldwork will seek to verify the nature of these sinks, as well as differentiate karst sinkholes from glacially-derived landform features characteristic of this region, such as kettles, ice-block depressions, etc.

Conclusions

The methodology used in this study demonstrates the practicality of applying high-resolution LiDAR to evaluate karst terrain prior to field mapping, and sets the framework for continued exploration into applied LiDAR techniques for evaluating karst terrains as a preliminary method of landscape assessment prior to fieldwork. This can be a time efficient approach for pre-screening areas of interest depending on the scope and scale of the research goals involved. Subsequent fieldwork can also help improve the ability of discerning sinkholes on LiDAR by comparing known sinkholes in the field, with their geospatial signatures observed on a DEM. This may help constrain the geostatistical parameters associated with computational processing, resulting in more accurate automated delineation techniques with less emphasis on manual filtering. Similarly, this may limit any unnecessary subjectivity and bias associated with the tediousness of manual sinkhole recognition. Currently the NYSGS is applying this method to evaluate karst terrain in Albany County (Figure 75).

Figure 72: Map of New York State outlining Cayuga County (red) and showing distribution of karsts topography (blue).

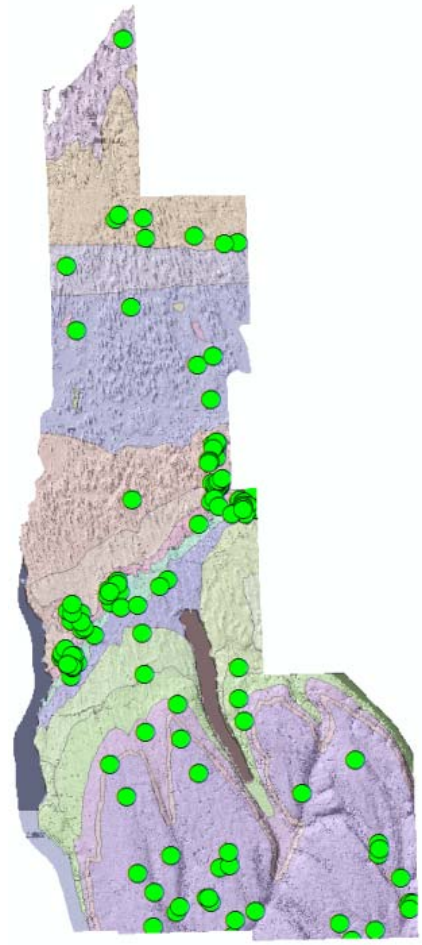


Figure 73: Map of Cayuga County, NY, showing distribution of potential sinkhole features (represented by points) identified in this study, overlaying hillshade and bedrock layers.

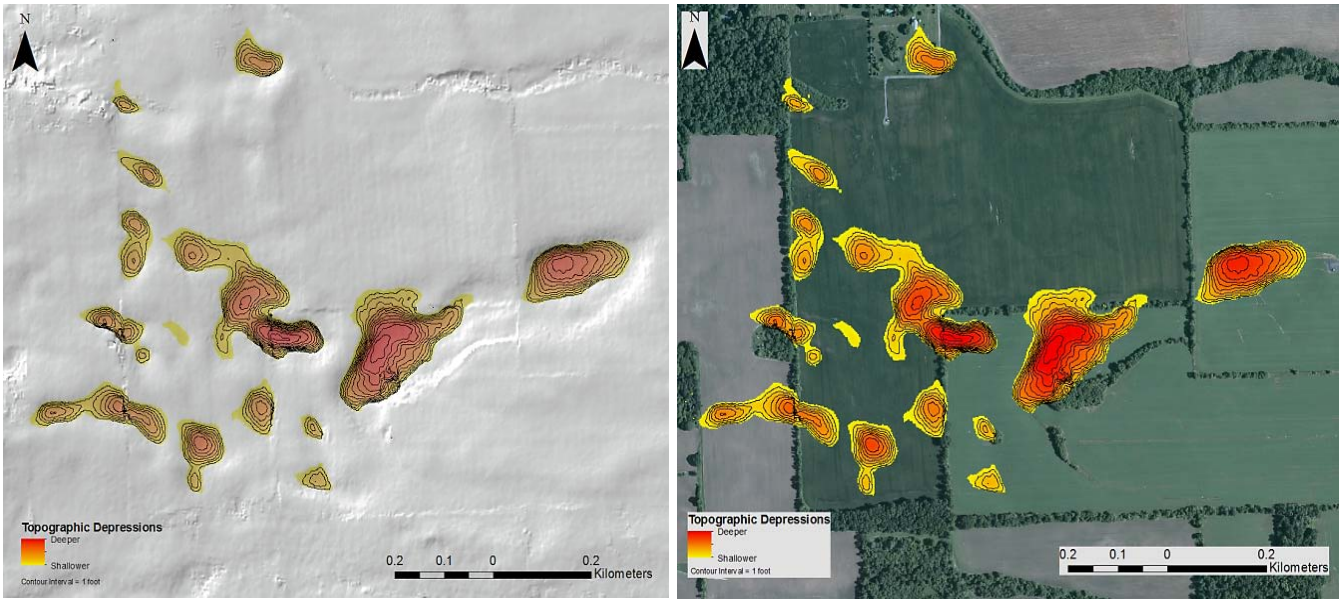


Figure 74: Examples of topographic depressions in the Union Springs quadrangle, Cayuga County, New York, showing hillshade (left image) and aerial photography (right image).

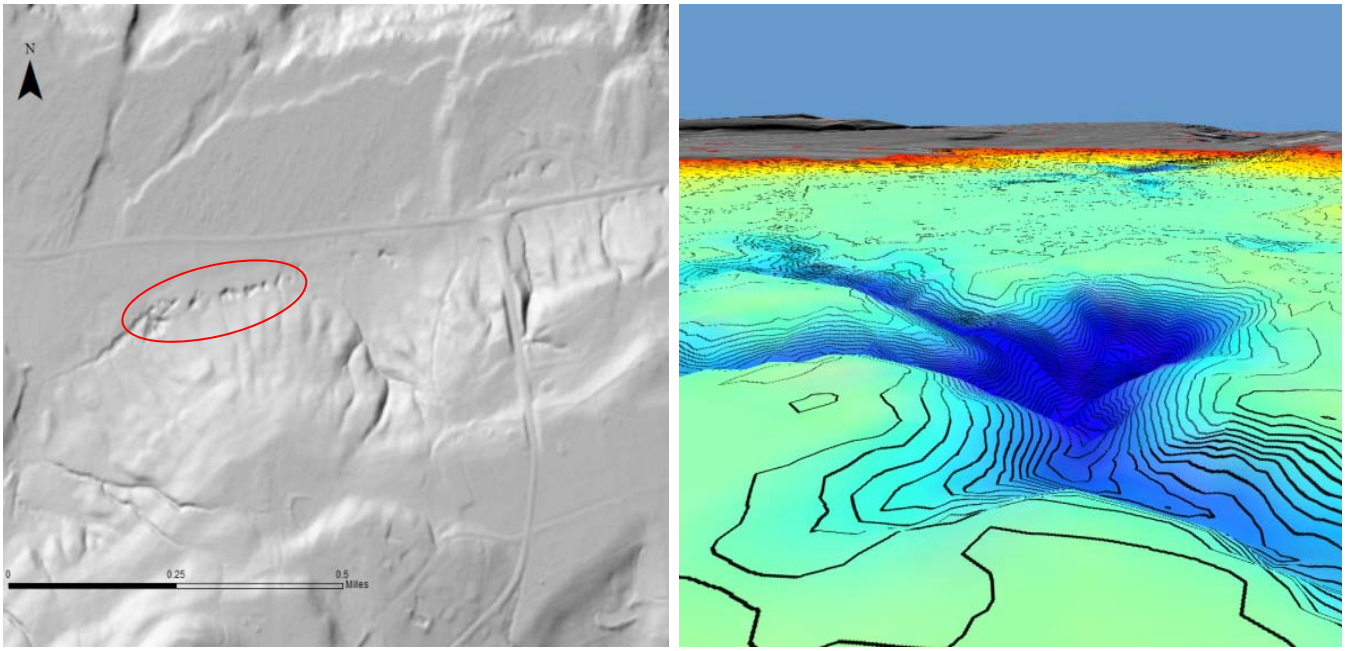


Figure 75: Karst features in Albany County, showing sinkholes on LiDAR (left; circled in red) and 3D rendering of karst-induced internal drainage system with 1-foot contours (right).

Thoughts on Subglacial Till Kinematics and the Origin of Drumlins

Nathan R Hopkins*, Edward B Evenson*, Kenneth P Kodama*, Matthew Gentoso, Andrew L Kozlowski**

*Dept. of Earth and Environmental Sciences, Lehigh University, Bethlehem, PA
** Geologic Survey, New York State Museum, Albany, NY

Introduction

Landscapes dominated by subglacial bedforms are among the most interesting and complex terrestrial landscapes. Because most subglacial bedforms cannot be viewed during active formation like their fluvial or eolian counterparts, determining the specific mechanisms leading to their formation is complicated. Drumlins – elongate, streamlined deposits of subglacial till – are arguably the most controversial subglacial bedform, and have confounded researchers since the inception of modern geology (Davis, 1884). Detailed reviews of drumlin characteristics are numerous (Muller, 1964; Menzies, 1987; Clark, 2010), and it is clear that there exists great variety both between and within individual drumlin fields, including characteristics of their composition, internal structures, and geospatial patterns. For this discussion, we would like to follow the approach taken in recent review by Clark (2010); the commonalities of drumlins are the key to identifying the mechanisms of their genesis. These commonalities include, but are not limited to, (1) their longitudinal asymmetry; (2) their occurrence within vast drumlin fields; (3) a systematic distribution and evolution within drumlin fields; (3) a dominantly till composition; (4) the long axes of drumlins parallel regional ice flow directions.

The Weedsport Drumlin Field (WDF) of Upstate New York is among the most impressive in the world (Figure 76). These features were developed during the Wisconsinian glaciation beneath the Ontario Lobe of the Laurentide Ice Sheet at it extended south through New York. Formation of these features likely occurred near the time of the deposition of the Valley Heads moraine. The landscape was subsequently modified during deglaciation by glacio-fluvial and lacustrine action. The drumlins of the WDF exemplify the key common characteristics of drumlins deposited beneath large ice sheets. Though drumlin composition has been shown to vary considerably worldwide, the drumlin of New York State are composed predominantly of thick, overconsolidated lodgement till of the Furnaceville and Somerset members (Calkin and Muller, 1992).

Numerous researchers have investigated the drumlins of the WDF and have come to disparate conclusions. Early fabric studies within the New York and Wisconsin drumlin fields have suggested an accretionary model of drumlin formation (Andrews and King, 1968; Savage, 1968; Evenson, 1971). In such a model, till is deposited locally and streamlined into the drumlin form. Alternatively, based upon interpreted stratigraphy exposed along lake shore drumlin cuts, Kerr and Eyles (2007) have proposed a

two-stage erosional model for the generation of the WDF. This model first requires the accretion of a thick till sheet followed by intense, localized erosion of the interdrumlin low coincident with deposition of glacier terminal moraines (Kerr and Eyles, 2007). Based upon local drill cores, this accreted till layer would need to approach 40 meters in thickness. This model has been echoed by previous and subsequent articles (Whittecar and Mickelson, 1977; Boyce and Eyles, 1991; Boulton et al., 2001; Vreeland and Iverson, 2009). It is worth noting that the role of bed deformation in the erosion of the deposited till and streamlining of the drumlin form is not resolved (e.g., Boyce and Eyles, 1991; Vreeland and Iverson, 2009). Another group of models, such as the 'till dilatancy' model of Smalley and Unwin (1968) and the flow instability models (Clark, 2010; Stokes et al., 2013), call upon the rheological properties of the subglacial material. These models are not discussed in detail here.

An erosional model for the genesis of drumlins makes the specific prediction that the fabric of the pre-existing till unit should be genetically unrelated to the flow responsible for drumlin formation. If bed deformation is occurring during this period of erosion, a thin layer of till mantling the drumlin surface may record an interaction between localized ice flow and the drumlin shape. For any till layer beneath the deforming bed, the measured fabric should show no relationship to the three-dimensional form of the drumlin. In a series of studies summarized here, we investigated the anisotropy of magnetic susceptibility (AMS) till fabrics of numerous drumlins in the WDF. These results indicate (1) bed deformation was pervasive during deposition of the WDF tills, (2) that AMS fabrics faithfully record regional ice flow directions as interpreted from bedform long axes, and (3) that the internal fabrics measured within one drumlin show systematic variations indicating interaction between the drumlin form and ice flow. These results are inconsistent with an erosional model of drumlin formation for the WDF and favor a model wherein till is locally accreted at the ice-till interface and simultaneously streamlined.

Evolution of Till Fabrics

The action of the over-riding ice and resulting deformation of subglacial sediments has a profound effect on the orientations of rock and mineral clasts embedded beneath the glacier sole. Even under relatively small amount of shear, non-equant clasts will become rotated into a preferred orientation and develop a measurable till fabric. The specific characteristics of a given fabric reflect the subglacial conditions during deformation (e.g., pore pressures, overburden, etc.) and the sedimentologic characteristics of the till itself. The earliest recognitions of till fabrics focused on the preferred orientations of macroscopic clasts – pebble and larger-sized grains embedded in a matrix of fines (Holmes, 1941; Harrison, 1957; Andrews and King, 1968; Ham and Mickelson, 1994; Gentoso *et al*., 2011). These macrofabrics, or clast fabrics, have become a powerful and commonly used tool for geologic and glaciologic investigations. Evenson (1970) developed a technique to measure the fabric of sand grains and later showed that clast fabric and sand fabric analysis yield consistent results that faithfully record the direction of ice flow (Evenson, 1971). As discussed later, fabrics from the elongate silt sized particles have also been shown to yield similar directions of inferred ice flow, indicating pervasive shear of all grain sizes (Gentoso, 2011).

Another means of measuring till fabrics relies on the magnetic properties of the till itself. Magnetic susceptibility has been used infrequently in early till fabric studies with varying success (Fuller, 1964; Stupavsky *et al*., 1974; Boulton, 1976). However, in a series of recent papers (Hooyer & Iverson 2000; Thomason & Iverson 2006; Hooyer *et al.,* 2008; Iverson *et al*. 2008), a laboratory basis for the development, interpretation and application of anisotropy of magnetic susceptibility (AMS) till fabrics was established using a ring-shear device and natural tills. In these experiments, controlled levels of uni-directional shear was shown to generate shear parallel fabrics that vary in strength as a function of shear strain (Hooyer &Iverson 2000; Thomason & Iverson, 2006; Hooyer *et al.*, 2008; Iverson *et al.*, 2008). Anisotropy of magnetic susceptibility (AMS) is a measurement of the orientations and magnitudes of the maximum, intermediate and minimum axes of preferential magnetization of a sample. These 'easy' axes are the k_1, k_2, and k_3 eigenvectors, respectively, and define an ellipsoid (Figure 77). Magnetic susceptibility is the proportionality tensor between an applied and induced magnetic field, and for non-equant mineral crystals is controlled dominantly by shape, wherein susceptibility is at a maximum along the long axis of a mineral grain (Tarling & Hrouda 1993). Anisotropy arises when the 'easy' axes of mineral within a sample become preferentially aligned (Tarling & Hrouda 1993). In a deformed till that experienced shear and where the magnetic signature is dominated by an elongate ferromagnetic grain, the ellipsoid should be elongate in the direction of flow (*i.e.*, k_1 oriented parallel to flow).

AMS Fabric Analysis

Our investigation of the subglacial till kinematics and the genesis of drumlins in the WDF occurred in two related studies. The first study – published in a 2011 issue of Boreas – sought to demonstrate the effectiveness of AMS as an indicator of ice flow direction and evaluate the possibility of bed deformation within streamlined tills (Gentoso et al., 2011). Regional ice flow direction in the WDF is known based upon the long axis orientations of streamlined bedforms. Five drumlins exposed in lake cuts and five flutes exposed in road cuts were sampled using standard paleomagnetic sampling techniques (Figure 76). Each of these features was sampled near their centerline to avoid any complications arising due to interactions between ice flow and the form of the feature. At each site, a standard macrofabric was also collected for comparison. The resulting AMS k_1 fabrics were typically strong and consistent with traditional macrofabrics collected at the same localities (Figure 78). The mean orientations of all AMS k_1 fabrics and pebble long axes are statistically identical. Additionally, AMS k_1 fabric orientations were parallel to drumlin and flute long axes (Figure 79). Collectively, these results indicate that AMS fabric faithfully records ice flow directions in the WDF; additionally, it is clear that bed deformation occurred within the WDF and that deformation was pervasive (i.e., affecting all grain sizes).

Following the previous proof of bed deformation and AMS fabric reliability, we undertook an investigation to evaluate the possible interactions between ice flow and drumlin form. The end-member models for drumlin genesis of erosion and deposition or accretion make distinct predictions

for internal AMS fabrics. As stated earlier, a purely erosional genesis for the drumlin form would predict internal AMS fabrics with no relationship to the surface morphology. The contrary model of till accretion would lead to internal fabrics yielding some relationship to the drumlin form at the time of till deposition. To test these predictions, one drumlin of the WDF in the vicinity of the village of Cato, NY, was selected for systematic sampling of AMS fabric and rock magnetic analysis (Figure 80). This drumlin was chosen due to its ideal representation of the drumlins in the WDF, accessibility, and aid from local landowners.

The selected drumlin and adjacent interdrumlin low was sampled in 19 locations in an attempt to accurately capture any potential interaction between the form of the drumlin and localized ice flow (i.e., divergence or convergence around the drumlin form). At each site, a pit approximately 1 meter wide and 2.5 meters deep was excavated via backhoe. At a depth of approximately 2 meters, 25 oriented 8 cm^3 till samples were collected. AMS fabric and rock magnetic analyses were conducted at Lehigh University using a KLY-3S Kappa Bridge, and 2G Enterprises 755 superconducting rock magnetometer.

The spatial patterns of localized flow indicated by the AMS k_1 fabric orientations indicate potentially complex interactions between the flow of overriding ice and a disparity at the ice-till interface (Figure 81). In general, fabrics are strong (alignment values greater than 0.85), with the mean fabric plunging north and parallel to the drumlin long axis. Anisotropy ellipsoids are triaxial to weakly oblate, reflecting the presence of both elongate ferrimagnetics and platy, paramagnetic clays. Spatially, fabrics show southwesterly oriented k_1 axes on the up-ice end of the drumlin and south-southeast oriented fabrics along drumlin flanks. Down-ice fabrics are dominated by oblique down-slope (SSW) directions. Fabrics along the drumlin centerline crest are all strong, parallel to drumlin long axis, and plunging north. Interdrumlin fabrics are typically weak (alignment values less than 0.8) and parallel drumlin long axis. When all principal long axes are evaluated without respect to location, the resulting fabric is quite strong with a mean orientation parallel to drumlin long axis (Figure 81). The principle axis orientations plot an asymmetric bimodal distribution around the drumlin long axis, with the most points deflected eastward (Figure 82). This arises due to the southwesterly-oriented divergent fabrics at the drumlin head and tail.

Interpreting coherent, longitudinal flow lines is difficult given the scatter in the data; however, the results are suggestive of divergent flow around the up-ice end of the drumlin. Similar fabrics have been observed previously (Savage, 1968; Andrews and King, 1968). Our observation of continued divergent fabrics on the down ice portion of drumlins has not to our knowledge been previously noted, and may appear contradictory to some predictions. This observation may be explained however if we consider the common observation of the en echelon formation of drumlins within a field. The consistent oblique down-slope flow suggested by our fabric results would suggest flow in the direction of the adjacent drumlin.

In addition to fabric orientation, parameters describing the shape of the anisotropy ellipsoid also yield interesting relationships to drumlin form. The degree of alignment of the maximum susceptibility axes (k_1) is a useful metric for gauging the degree of shear experienced by the till. Alignment values (S_1) are normalized such that alignment of 0.33 indicates isotropic orientations of k_1, whereas alignment values of 1 indicates perfect alignment. Alignment values show a strong relationship to the elevation-dimension of the drumlin form (Figure 83). The interdrumlin low possesses typically weak fabrics. Peak values are reached at the drumlin crest. A similar pattern is observed for percent anisotropy, calculated as the percent difference between the k_1 and k_3 principal eigenvalues.

Conclusion

The recent development of AMS till fabrics has allowed for a renewed interest and ability to investigate the kinematics and behavior of subglacial tills. As this article summarizes, AMS till fabrics have also provided a new tool to study old problems. Collectively, these results indicate (1) bed deformation was pervasive during deposition of the WDF tills, (2) that AMS fabrics faithfully record regional ice flow directions, and (3) that the internal fabrics measured within numerous drumlins show systematic variations indicating interaction between the drumlin form and ice flow. These results are inconsistent with an erosional model of drumlin formation for the WDF and favor a model wherein till is locally accreted at the ice-till interface and simultaneously streamlined. Alternatively, the relationships observed here could be explained by a thick deforming bed; however, the approximately 3 meter thick deforming bed required in this location exceeds existing estimates of deforming bed thicknesses (Alley, 2000). Reconciling the fabric data presented here with the rheological models of drumlin formation will hopefully be undertaken in future work. However, it is clear that any model of drumlin genesis must provide an explanation for the fabrics recorded in these till-cored drumlins.

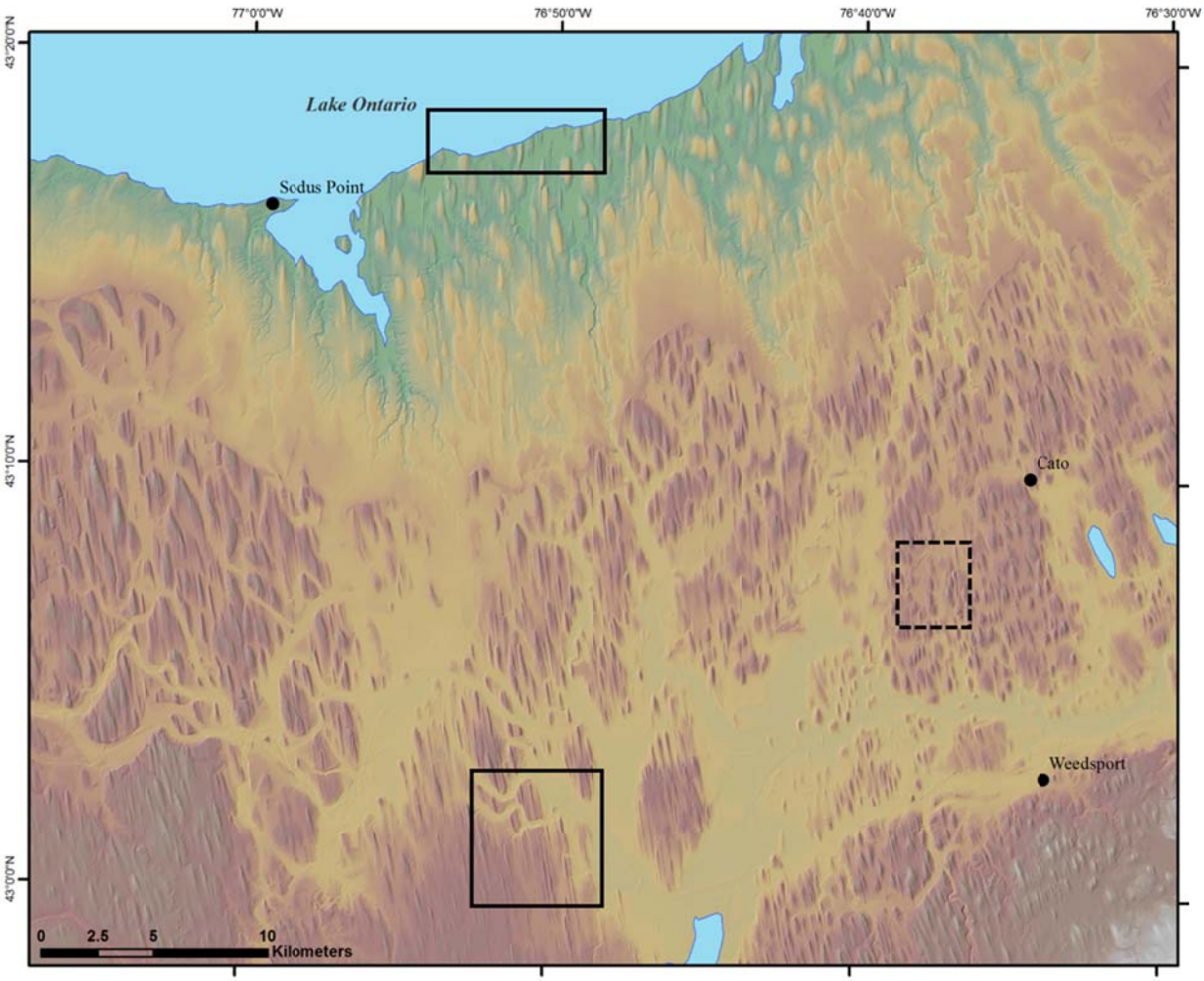


Figure 76. Digital elevation model (10 meter) and hillshade of the Weedsport Drumlin Field, New York State. Sampling localities of Gentoso *et al*., (2011) are shown in solid squares. Sampling locality of recent work is shown in dashed square. The WDF exhibits the classic characteristics of drumlin fields. Note the systematic variation to more elongate bedforms in the down-ice direction (south). A noticeable lake-shore parallel band of muted and planed drumlins resulting from wave base erosion in glacial Lake Iroquois is clearly visible.

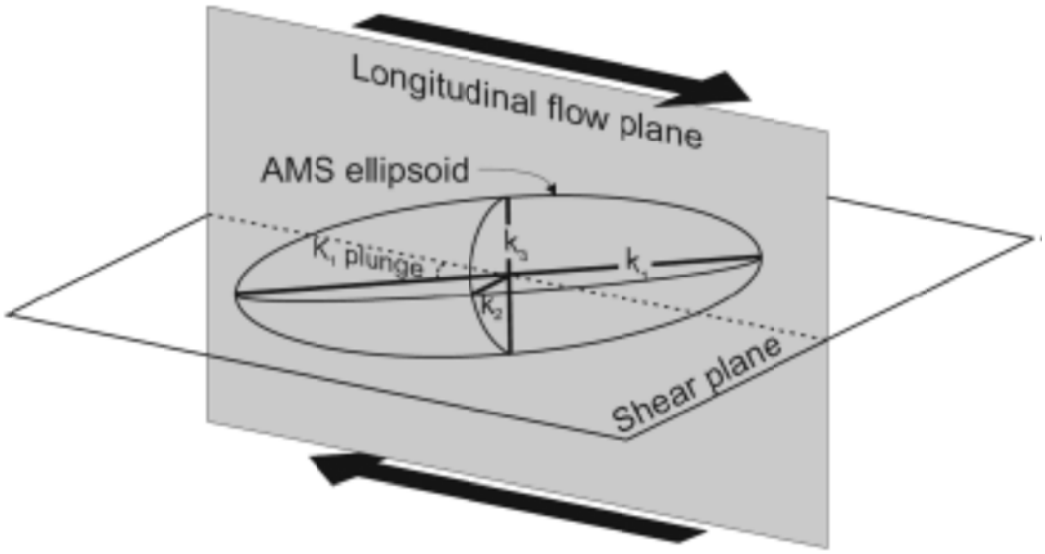


Figure 77. Graphical representation of the anisotropy of magnetic susceptibility ellipsoid (from Shumway & Iverson, 2009). Materials subjected to uni-directional shear develop shear parallel fabrics that plunge up-shear as elongate magnetic grains are rotated such that their long axes parallel the direction of shear. The volume-average susceptibility ellipsoid defined by the k_1, k_2, and k_3 eigenvalues, or 'easy' axes', reflects the orientation of these particles.

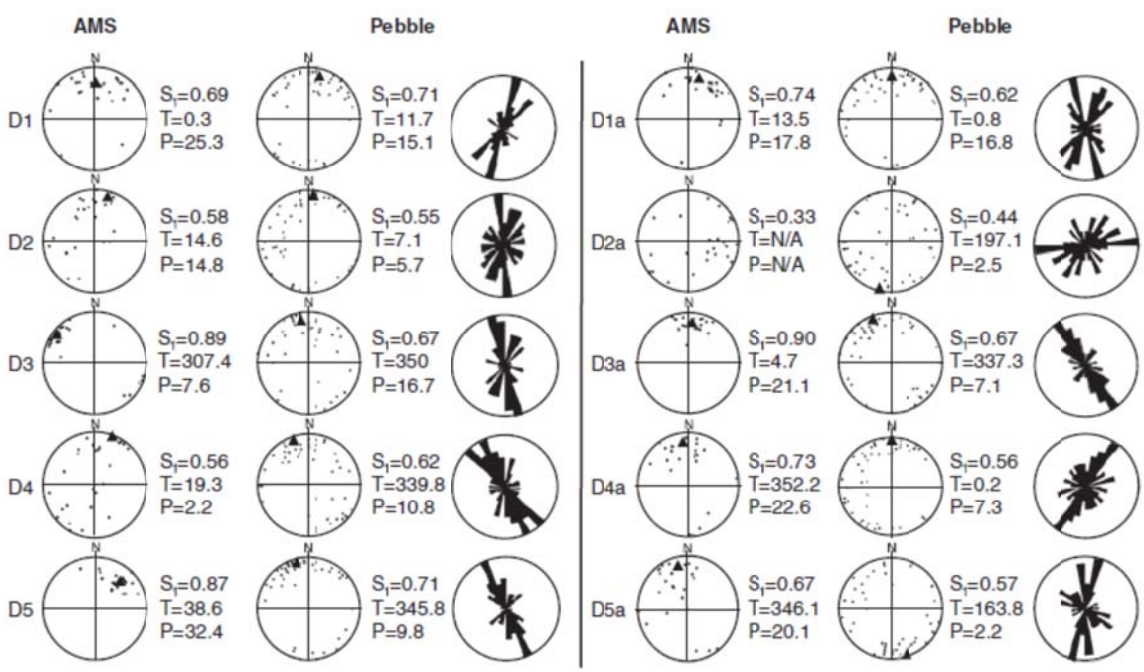


Figure 78. AMS k1 orientations and pebble long axis orientations for five drumlins from Gentoso *et al*., (2011), plotted on lower hemisphere stereoplots. Pebble fabric rose diagrams are shown for comparison. Fabric mean orientation (black triangles) or AMS and pebble fabrics show strong agreement, though individual fabrics vary by as much as 70 degrees. Fabric alignment values (S_1) and AMS shape factors (T and P) are also reported.

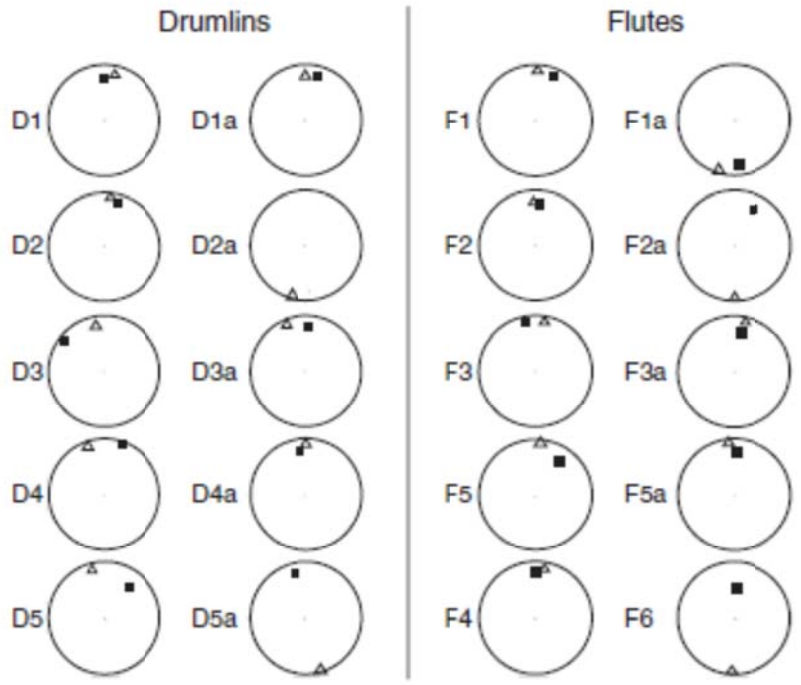


Figure 79. Summary of AMS fabrics and macrofabrics for drumlins and flutes from Gentoso *et al*. (2011). Fabric orientations based upon AMS k1 orientations (black squares) and pebble long-axis orientations (open triangles) on lower hemisphere stereoplots for five drumlins and five flutes samples in the WDF. Drumlin and flute orientation is north-south.

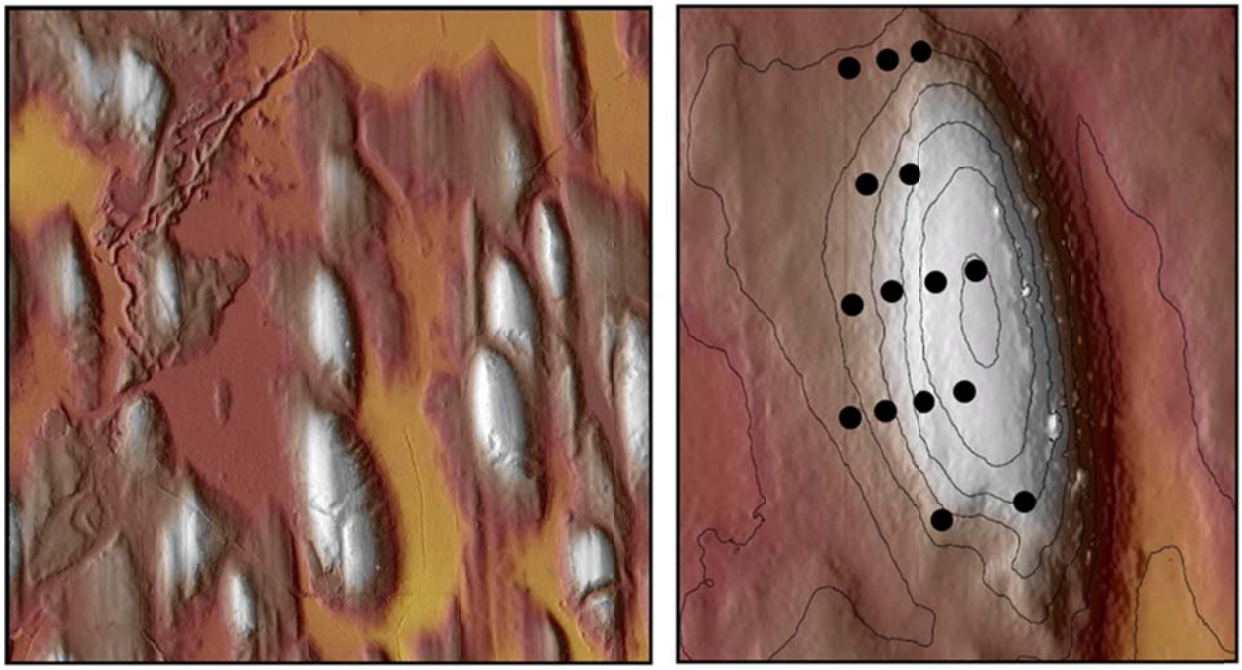


Figure 80. LIDAR map of sampling locality near the Village of Cato. The drumlin sampled (center left) is positioned within an en echelon array of nearby drumlins, which is typical. Note the sinuous esker complex in the northwest corner of the map. The drumlin was systematically sampled (dark circles, right) to capture potential interaction between the drumlin form and ice flow.

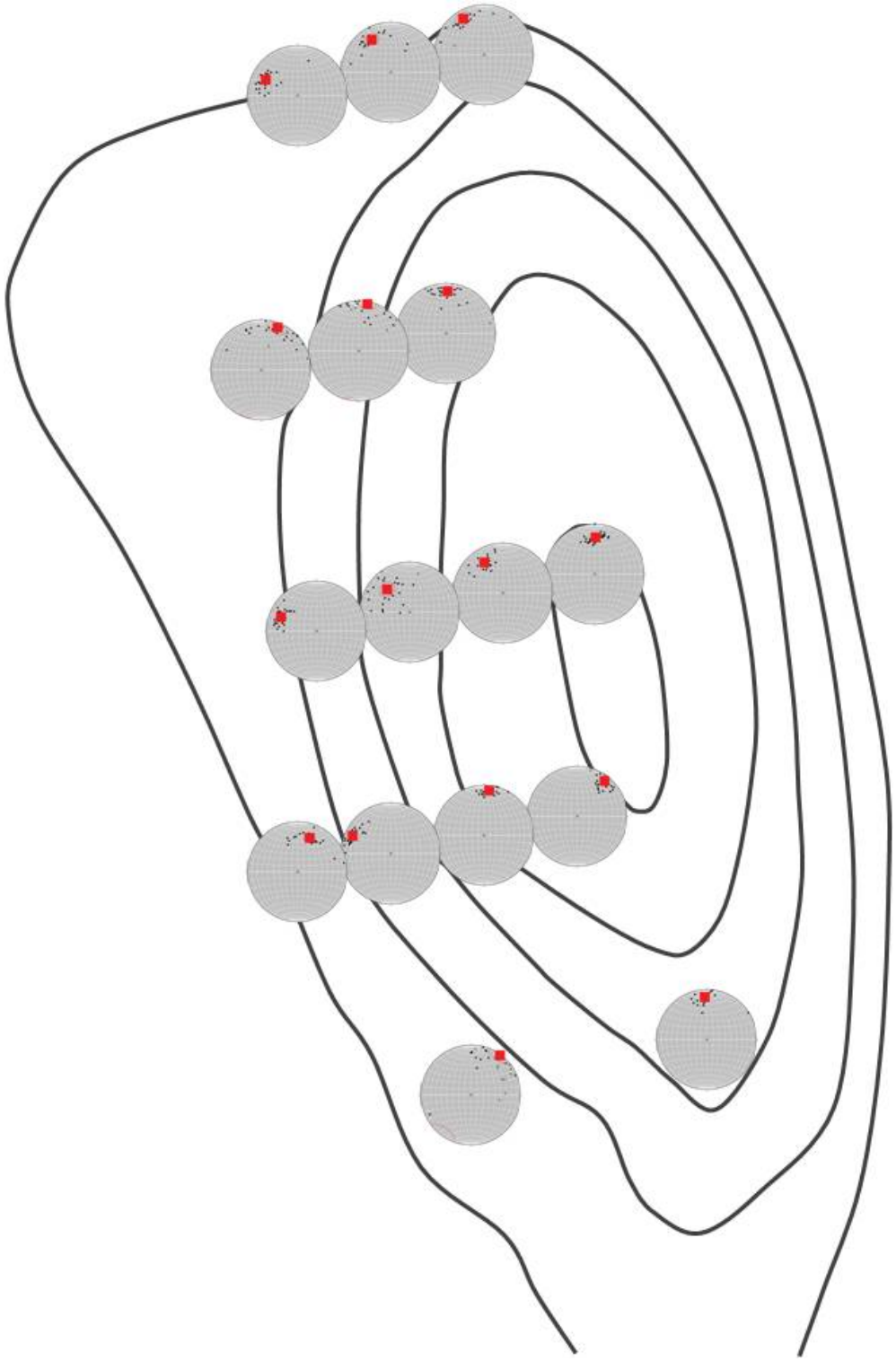


Figure 81. Simplified 20-foot elevation contours of sampled drumlin with individual AMS k_1 orientations (black dots) and the mean (red squares) for each of 25 samples taken at 16 sampled pits plotted on lower hemisphere stereonets.

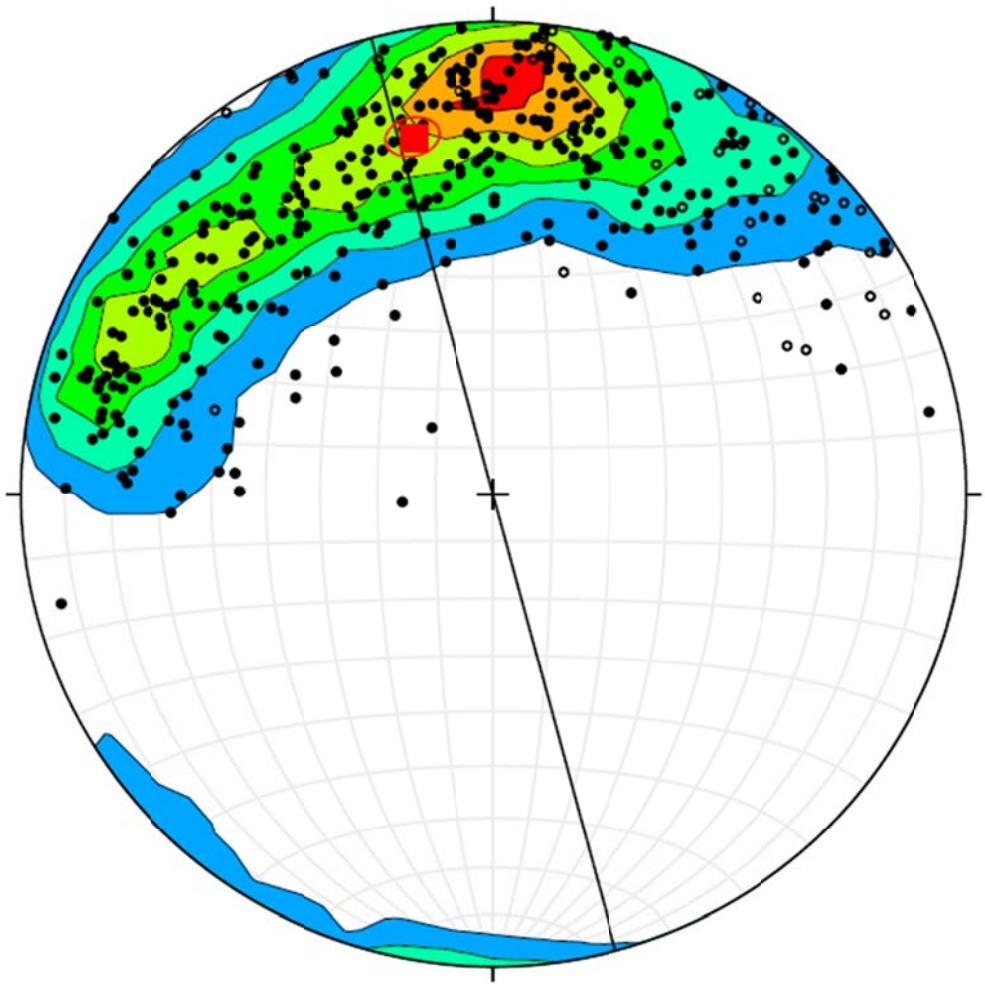


Figure 82. All AMS k1 orientations from 16 sample sites with the mean k1 orientation (red square) and 95% confidence ellipse (red circle). Drumlin long axis (345°) is shown as the black line. Closed (open) circles represent lower (upper) hemisphere k1 orientations.

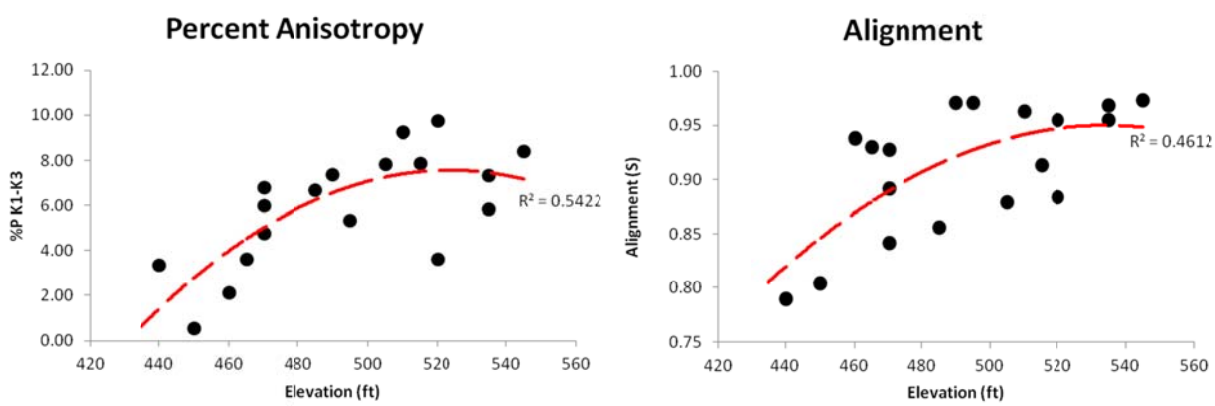


Figure 83. Scatter plots of site mean percent anisotropy (left) and fabric alignment (right) plotted as a function of elevation of each site. Low elevations represent in the interdrumlin low, and high elevations are the drumlin crest. Each dataset is best fit by a second order polynomial.

Aden, D.J., 2013, Mapping karst in the north-central Ohio region: Geological Society of America Abstracts with Programs, v. 45, no. 4, p. 18.

Agassiz, L., 1864, Ice period in America. Atlantic Monthly 14: 86-93.

Allen, J.R., 1973, A classification of climbing-ripple cross-lamination. Journal of the Geological Society, 129(5), 537-541.

Alley, R. B., 2000, Continuity comes first: recent progress in understanding subglacial deformation. In Maltman, A. J. & Hambrey M. J. (eds.): Deformation of Glacial Materials, 171-179. Geological Society, London, Special Publication 176.

Andersen, L.T., Hqansen, D.L., Huuse, M., 2005, Numerical modelling of thrust structures in unconsolidated sediments: Implications for glaciotectonic deformation Journal of Structural Geology, 27, 587–596

Anderson, T.W., and Lewis C.F.M., 2012, A new water-level history for Lake Ontario Basin: evidence for a climate-driven early Holocene sow stand: Journal of Paleolimnology, 47:513-530.

Andrews, J. T. and King, C. A. M., 1968, Comparative till fabrics and till fabric variability in a till sheet and a drumlin: a small-scale study. Proceedings of the Yorkshire Geological Society 36, 435-461.

Apfel, E.T., 1946, Origin of the New York State Finger Lakes (Abstract). Geol. Soc. Am. Bull. 57:1175-1176.

Barnosky, A.D., and Lindsey, E.L., 2010, Timing of Quaternary megafaunal extinction in South America in relation to human arrival and climate change. Quaternary International 217, no. 1: 10-29.

Beck, B.F., 2005, Soil piping and sinkhole failures. In: Encyclopedia of Caves, Culver, D.C. and W.B. White (eds.). Elsevier. p. 521-526.

Beck, L. C., 1842, Mineralogy of New York – comprising detailed descriptions of the minerals hitherto found in the state of New York, and notices of their uses in the arts and agriculture, W and A White and J. Visscher, New York, New York.

Bekele, E. B., Rostron, B. J., and Person, M. A., 2003, Fluid pressure implications of erosional unloading, basin hydrodynamics and glaciation in the Alberta Basin, Western Canada: Journal of Geochemical Exploration, Vol. 78–79, p. 143–147.

Bense, V. F. and Person, M. A., 2008, Transient hydrodynamics with intercratonic sedimentary basins during glacial cycles: Journal of Geophysical Research, vol. 113, 17 p.

Berg, R.C., (and others), 2000, Mapping the glacial geology of the Central Great Lakes region in three dimensions – a model for state-federal cooperation. U.S. Geol. Surv. Open-file Report 99-349, 65p.

Blewett, W., Winters, H., and Rieck, R., 1993, New Age Control on the Port Huron Moraine in Northern Michigan. Physical Geography 14:131-138.

Bloom, A.L., 1967, "Fernbank": A rediscovered Pleistocene interglacial deposit near Ithaca, New York; Geological Society of America Northeastern Section, Abstracts for 1967, Special Paper 115, p. 251.

Bloom, A.L., 1984, Unanswered questions about Finger Lakes geomorphology: Cornell Quarterly, v. 19, p. 57-61.

Bloom, A.L., 1986, Geomorphology of the Cayuga Lake basin. N.Y. State Geol. Assoc. Field Trip Guidebook, pp. 261-279.

Bloom, A.L., McAndrews, J.H., 1972, Schedule and guidebook. Friends of the Pleistocene 35th Annual Reunion, Cornell University. 20.

Boisvert, R.A., 2012, The Paleoindian Period in New Hampshire. In: Late Pleistocene Archaeology and Ecology in the Far Northeast, edited by Claude Chapdelaine, Texas A&M University Press, College Station, pp. 77-94.

Boulton, G. S., 1976, The origin of glacially fluted surfaces – observations and theory. Journal of Glaciology 17, 287-309.

Boulton, G.S., 1982, Subglacial processes and the development of glacial bedforms. In: Research in Glacial, Glacio-Fluvial and Glacio-Lacustrine Systems. Proceedings of the 6th Guelph Symposium on Geomorphology, 1980. 1982. P 1-31, 12 fig, 33 ref.

Boulton, G. S., Dobbie, K.E., and Zatsepin, S., 2001, Sediment deformation beneath glaciers and its coupling to the subglacial hydraulic system. Quaternary International 86, 3-28.

Boyce, J. I., and Eyles, N., 1991, Drumlins carved by deforming till streams below the Laurentide Ice Sheet. Geology 19, 787-790.

Bradley, J.W., 1987, Evolution of the Onondaga Iroquois: Accommodating Change, 1500-1655: University of Nebraska Press, Lincoln and London.

Bradley, J.W., Spiess A., Boisvert R., and Boudreau, J., 2008, What's the point?: modal forms and attributes of Paleoindian bifaces and the New England-Maritimes Region: Archaeology of Eastern North America, 36:119-172.

Bradley, J.W., and Funk, R.E., N.D., Paleoindian findspots in Central New York State: An Update: Unpublished manuscript in possession of senior author.

Brigham, A.P., 1893, The Finger Lakes of New York. Am. Geog. Soc. Bull. 25:203-223.

Bronk-Ramsey C., T. Higham, A. Bowles, and R. Hedges, 2004, Improvements to the pre-treatment of bone at Oxford. Radiocarbon 46:155-163.

Brown, T.A., D. E. Nelson, J.S. Vogel, and J.R. Southon, 1988, Improved collagen extraction by modified longin method. Radiocarbon 30:171-177.

Burke, A., 2006, Paleoindian ranges in Northeastern North America based on lithic raw materials sourcing. In: Notions De Territoire Et De Mobilité: Exemples De L'Europe Et Des Premières Nations En Amérique Du Nord Avant Le Contact Européen, edited by C. Bressy, A. Burke, P. Chalard, H. Martin, Études et Recherches Archéologiques de l'Université de Liège, Liège. pp. 77-89.

Cadwell, D.H., Muller, E.H., and Fleischer P.J., 2003 Geomorphic history of New York State. In: Geoarchaeology of Landscapes in the Glaciated Northeast, edited by David L. Cremeens and John P. Hart, New York State Museum Bulletin 497., pp. 7-14.

Calkin, P. E., Muller, E. H., & Barnes, J. H., 1982, The Gowanda Hospital interstadial site, New York. American Journal of Science, 282(7), 1110-1142.

Calkin, P. E. and Muller, E. H., 1992, Pleistocene stratigraphy of the Erie and Ontario Lake Bluffs in New York. In Fletcher, C. H. and Wehmiller, J. F. (eds): Quaternary Coasts of the United States: Moraine and Lacustrine Systems, 385-396. SEPM Special Publication 48.

Carney, F., 1904, Direction of proglacial flow in central New York. Am. Geol. 33:196-198.

Carney, F., 1907a, Glacial erosion in longitudinal valley [Owasco Lake, N.Y.]. Jour. Geol. 15:722-730.

Carney, F., 1907b, Pre-Wisconsin drift in the Finger Lake region of New York. Jour. Geol. 15:571-585. Also: Denison Univ. Sci. Lab. Bull. 14: 3-18.

Carney, F., 1909, The Pleistocene geology of the Moravia quadrangle. Denison Univ. Sci. Lab. Bull. 14: 335-342.

Carter, E.S., White, S.M., and Wilson, A.M., Variation in groundwater salinity in a tidal salt marsh basin, North Inlet Estuary, South Carolina: Estuarine, Coastal and Shelf Science, v. 76, p. 543-552.

Central Great Lakes Geologic Mapping Coalition, 1999, Sustainable growth in America heartland – 3D geologic maps as the foundation. U.S. Geol. Surv. Circular 1190, 17p.

Chisnell, T.C., 1951, Recognition and interpretation of proglacial strand lines in Cayuga basin: Cornell Univ., Ph. D. dissert.

Clark, C. D., 2010, Emergent drumlins and their clones: from till dilatancy to flow instabilities. Journal of Glaciology 51, 1011-1025.

Clayton, L., Cherry, J.A., 1967, Pleistocene superglacial and ice-walled lakes of west-central North America. North Dakota Geological Survey, Miscellaneous Series 30, 47-52.

Clayton, L., Attig, J.W., Ham, N.R., Johnson, M.D., Jennings, C.E., Syverson, K.M., 2008, Ice-walled-lake plains: implications for the origin of hummocky glacial topography in middle North America. Geomorphology 97, 237-248.

Coates, D.R., 1968, Finger Lakes, in Fairbridge, R.W., ed., Encyclopedia of geomorphology: New York, Reinholds Books, p. 351-357.

Coates, D.R., 1974, Reappraisal of the glaciated Appalachian Plateau, in Coates, D.R., ed., Glacial geomorphology. Publications in Geomorphology: Binghamton, State University of New York. P. 205-243.

Cox, D.D., 1959, Some Postglacial Forests in Central and Eastern New York State Is Determined by the Method of Pollen Analysis. New York State Museum and Science Service, Bulletin 377.

Crook, N., Binley, A., Knight, R., Robinson, D.A., Zarnetske, J., and Haggerty, R., 2008, Electrical resistivity imaging of the architecture of substream sediments: Water Resources Research, v.44.

Curry, B. B., Konen, M. E., Larson, T. H., Yansa, C. H., Hackley, K. C., Alexanderson, H., & Lowell, T. V., 2010, The DeKalb mounds of northeastern Illinois as archives of deglacial history and postglacial environments. Quaternary Research, 74(1), 82-90.

Curry, B., and Petras, J., 2011, Chronological framework for the deglaciation of the Lake Michigan lobe of the Laurentide Ice Sheet from ice-walled lake deposits. Journal of Quaternary Science, 26(4), 402-410.

Daniels, R., and Peteet, D., 1998, Fish scale evidence for rapid post-glacial colonization of an Atlantic coastal pond. Global Ecology and Biogeography Letters 7:467-476.

Davis, W. M., 1884, Drumlins. Science 4, 418-420.

Deller, D.B., and Ellis, C.J., 1988, Early Paleoindian complexes in Southwestern Ontario, In: Late Pleistocene and Early Holocene Paleoecology and Archaeology of the Eastern Great Lakes Region, edited by R. S. Laub, N.G. Miller, and D.W. Steadman, Bulletin of the Buffalo Society of Natural Sciences 33, pp. 251-263.

Deller, 1992, Thedford II, A Paleo-indian site in the Ausable River Watershed of Southwestern Ontario, Memoirs Museum of Anthropology, University of Michigan No. 24, Ann Arbor, Michigan.

Doctor, D.H., and Young, J.A., 2013, An evaluation of automated GIS tools for delineating karst sinkholes and closed depressions from 1-meter LiDAR-derived digital elevation data, in Proceedings, National Cave and Karst Research Institute Symposium 2., p. 449-458.

Dreimanis, A., and Karrow, P.F., 1972, Glacial history of the Great Lakes-St. Lawrence region, the classification of the Wisconsin(an) Stage, and its correlatives. In: 24th International Geological Congress, Section 12, pp. 5-15.

Dreimanis, A., 1992. Early Wisconsinan in the north-central part of the Lake Erie basin: A new interpretation, in Clark, P.U., and LeaP.D., eds., The Last Interglacial-Glacial Transition in North America; Boulder Colorado, Geological Society of America Special Paper 270, 109.

Dryer, C.R., 1904, Finger Lakes region of western New York. Geol. Soc. Am. Bull. 15:499-460.

Durkee, L.H., 1960, Pollen Profiles from Five Bog Lakes in New York State. Unpublished PhD dissertation, Syracuse University.

Dyke, A. S., 2004, An outline of North American deglaciation with emphasis on central and northern Canada. In: Quaternary Glaciations - Extent and Chronology, Part II: North America, Developments in Quaternary Science 2, edited by J. Ehlers, and P.L. Gibbard, pp.201-207. Elsevier: Amsterdam.

Dyke, A.S., J.T. Andrews, P.U. Clark, J.H. England, G.H. Miller, J. Shaw, and J.J. Veillette, 2002, The Laurentide and Innuitian ice sheets during the last glacial maximum. Quaternary Science Reviews 21: 9-31.

Ellis, C.J., 2011, Measuring Paleoindian range mobility and land use in the Great Lakes/Northeast. Journal of Anthropological Archaeology 30:385-401

Ellis, C.J., 2008, The fluted -point tradition and the Arctic small tool tradition: What's the connection? Journal of Anthropological Archaeology 27:298-314.

Ellis, C.J., and Deller, D.B., 1997, Variability in the archaeological record of Northeastern Early Paleoindians: A View from Southern Ontario. Archaeology of Eastern North America 25:1-30.

Ellis, C.J., and Deller, D.B., 1990, Paleoindians. In: The Archaeology of Southern Ontario to A.D. 1650, edited by Christopher Ellis and Neil Ferris, Occasional Publications, London Chapter, Ontario Archaeological Society, Number 5.pp. 37-63.

Ellis, C.J., and Deller, D.B., 1988, Some distinctive paleoindian tool types from the Lower Great Lakes Region. Midcontinental Journal of Archaeology 13(2): 111-158.

Ellis, C. J., Carr, D.H., and Loebel, T.J., 2011 The Younger Dryas and Late Pleistocene peoples of the Great Lakes Region. Quaternary International 242:534-545.

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. Journal of Geophysical Research: Solid Earth (1978-2013), 85(B11), 6319-6341.

Engeln, O.D. von, 1918, Glacial erosion of rock basins (Abstract). Assoc. Am. Geog. Annals 7:83-85.

Engeln, O.D. von, 1926, The geography of the Ithaca, New York, region. Assoc. Am. Geog. Annals 16:124-150.

Engeln, O.D. von, 1929, Interglacial deposit in central New York: Geological Society of America Bulletin. v. 40, p. 469-479.

Engeln, O.D. von, 1931, A proglacial or interglacial gorge near Seneca Lake, New York. N.Y. State Mus. Bull. 286:127-131.

Engeln, O.D. von, 1961, The Finger Lakes region: Its origin and nature: Ithaca, New York, Cornell University Press. 130p.

Evans, D.J.A., Phillips, E.R., Hiemstra, J.F., Auton, C.A., 2006, Subglacial till: formation, sedimentary characteristics and classification. Earth-Science Reviews 78, 115-176.

Evenson, E. B., 1970, A method for 3-dimensional microfabric analysis of tills obtained from exposures of cores. Journal of Sedimentary Research 40, 762-764.

Evenson, E. B., 1971, The relationship of macro- and microfabric of Till and the genesis of glacial landforms in Jefferson County, Wisconsin. In Goldthwait, R. P. (ed.) Till: A Symposeum, 345-364, Ohio State University Press.

Evgeny, A.K., and Ozorovich, Y.R., 2006, Geo-electromagnetic survey of the fresh/salt water interface in the coastal southeastern Sicily: Continental Shelf Research, v. 26, p. 843-851.

Fairchild, H.L., 1895, Lake Newberry, the probable successor of Lake Warren. Geol. Soc. Am. Bull. 6:462-466. Also: Science 1:61; Am. Geologist 15:202-203.

Fairchild, H.L., 1896, Glacial Genesee lakes. Geol. Soc. Am. Bull. 7:432-452.

Fairchild, H.L., 1897, Lake Warren shore lines in western New York and the Geneva Beach. Geol. Soc. Am. Bull. 8:269-284.

Fairchild, H.L., 1899a, Glacial Lakes Newberry, Warren and Dana in central New York. Am. Jour. Sci. 7:249-263.

Fairchild, H.L., 1899b, Glacial waters in the Finger Lakes region of New York. Geol. Soc. Am. Bull. 10:27-68.

Fairchild, H.L., 1900a, A channeled drumlin (Abstract). Science 11:104.

Fairchild, H.L., 1900b, Iroquois Beach. New York State Geologist Ann. Rep. 54: 103-139.

Fairchild, H.L., 1900c, Kame areas of western New York (Abstract). Rochester Acad. Sci. 3:190.

Fairchild, H.L., 1907, Drumlins of central western New York. N.Y. State Mus. Bull. 111:391-443.

Fairchild, H.L., 1909, Glacial waters in central New York. N.Y. State Mus. Bull. 127. 66p.

Fairchild, H.L., 1911, Radiation of glacial flow as a factor in drumlin formation (Abstract). Geol. Soc. Am. Bull. 22:734.

Fairchild, H.L., 1923, The Pinnacle Hills or the Rochester kame moraine. Rochester Acad. Sci. Proc. 6:141-194.

Fairchild, H.L., 1929, New York drumlins. Rochester Acad. Sci. Proc. 7:1-37.

Fairchild, H.L., 1932, New York moraines. Geol. Soc. Am. Bull. 43:627-662.

Fairchild, H.L., 1934, Cayuga Valley lake history. Geol. Soc. Am. Bull. 45: 233-280.

Feranec, R.S., and Kozlowski, A.L., 2010, AMS radiocarbon dates from Pleistocene and Holocene mammals housed in the New York State Museum, Albany, New York, USA. Radiocarbon 52(1): 205-208.

Feranec, R.S., and Kozlowski, A.L., 2012, New AMS radiocarbon dates from Late Pleistocene mastodons and mammoths in New York State, USA. Radiocarbon 54(1): 275-279.

Fisher, D.W., Isachsen, Y.W., Rickard, L.V., 1970, Geologic map of New York State, consisting of 5 sheets: Niagara, Finger Lakes, Hudson-Mohawk, Adirondack, and Lower Hudson, New York State Museum and Science Service, Map and Chart Series No. 15, scale 1:250,000.

French, J.H., 1860, Gazetteer of the State of New York - Embracing a comprehensive view of the geography, geology, and general history of the State, and a complete history and description of every county, city, town, village, and locality. With full tables of statistics: Syracuse, New York, R.P. Smith, 739 p.

Friends of the Montezuma Wetlands Complex, 2014, accessed May, 2014 at http://friendsofmontezuma.org/index.html

Fuller, M.D., 1964, On the magnetic fabrics of certain rocks. Journal of Geology 72, 368-376.

Fullerton, D. S., 1986. Stratigraphy and correlation of glacial deposits from Indiana to New York and New Jersey. Quaternary Science Reviews, 5, 23-37.

Funk, R.E., 1973, The West Athens Hills Site (Cox 7). In: Aboriginal Settlement Patterns in the Northeast, by William A. Ritchie and Robert E. Funk, New York State Museum and Service Memoir 20., pp. 9-36.

Funk, R.E., 2004, An Ice Age quarry-workshop: the West Athens Hill Site revisited. New York State Museum Bulletin 504.

Funk, R.E., and Steadman, D.W., 1994, Archaeological and Paleoenvironmental Investigation in the Dutchess Quarry Caves, Orange County. Persimmon Press, Buffalo. 128 pp.

Funk, R.E., and Wellman, B., 1984, The Corditaipe Site: a small isolated Paleoindian camp in the Upper Mohawk Valley. Archaeology of Eastern North America 12:81-109.

Gentoso, M.J., Evenson, EB, Kodama, K.P., Iverson, N. R. , Alley, R. B., Berti, C. and A. Kozlowski, 2011, Exploring till bed kinematics using AMS magnetic fabrics and pebble fabrics: the Weedsport drumlin field, New York State, USA. Boreas, 10.1111/j.1502-3885.2011.00221.x. ISSN 0300-9483.

Geostudi Astier srl and Multi-Phase Technologies LLC, 2006, ERTLab 3D ELECTRICAL RESISTIVITY TOMOGRAPHY INVERSION SOFTWARE User Manual, 89pp.

Gerber, R., Salat, C., Junge, A., and Felix-Henningsen, P., 2007, GPR-based detection of Pleistocene periglacial slope deposits at a shallow-depth test site: Geoderma, v. 139, p. 346-356.

Godin, L., Brown, R.L., Dreimanis, A., Atkinson, G.M., and Armstrong, D.K., 2002, Analysis and reinterpretation of deformation features in the Rouge River Valley, Scarborough, Ontario. Canadian Journal of Earth Sciences, 39(9), 1373-1391.

Goodman, W.M., Gnage, D.J., and Smith, P.H., 2010, Historical review of early salt manufacturing from natural salines in New York State – implications for present-day challenges: Clarks Summit, Pennsylvania, Solution Mining Research Institute, SMRI Spring 2010 Technical Conference, Grand Junction, Colorado, 30 p.

Gramly, R.M., 1988, Paleoindian sites south of Lake Ontario, Western and Central New York State. In: Late Pleistocene and Early Holocene Paleoecology and Archaeology of the Eastern Great Lakes Region, edited by R. S. Laub, N.G. Miller, and D.W. Steadman, Bulletin of the Buffalo Society of Natural Sciences 33., pp. 265-280.

Gramly, R.M., 1998, The Sugarloaf Site: Paleoamericans on the Connecticut River. Persimmon Press, Buffalo.

Gramly, R.M., 1999, The Lamb Site: a Pioneering Clovis Encampment. Persimmon Press, Buffalo.

Gramly, R.M., and Lothrop, J.C., 1984, Archaeological investigations of the Potts Site, Oswego County, New York. Archaeology of Eastern North America 12: 122-158.

Grasby, S., and Z. Chen, 2005, Subglacial recharge into the western Canada Sedimentary Basin—impact of Pleistocene glaciation on basin hydrodynamics" vol. 117, p. 500–514.

Hall, J., 1843, Geology of New York, Part 4, comprising the survey of the Fourth Geological District, Albany, New York: Carroll and Cook, Printers to the Assembly, 693p.

Ham, N. R. and Mickelson, D. M., 1994, Basal till fabric and deposition at Burroughs Glacier, Glacier Bay, Alaska. Geological Society of America Bulletin 106, 1552-1559.

Ham, N.R., Attig, J.W., 1996, Ice wastage and landscape evolution along the southern margin of the Laurentide Ice Sheet, north-central Wisconsin. Boreas 25, 171-186.

Hand, B.M., 1978, Syracuse melt-water channels. In: New York State Geol. Assoc. Guidebook, 50th Annual Meeting (pp. 23-24).

Hand, B.M., and Muller, E. H., 1972, Syracuse channels: evidence of a catastrophic flood: New York State Geological Association Fieldtrip Guidebook 44. P. I-1-12.

Harrison, P. W., 1957, A clay-till fabric: its character and origin. The Journal of Geology 65, 275-308.

Hartnagel C.A., and Bishop, S.C., 1922, The mastodons, mammoths and other Pleistocene mammals of New York State. New York State Museum Bulletin 241 242.

Hickock, S.R., and Dreimanis, A., 1989, Sunnybrook drift indicates a grounded early Wisconsin glacier in the Lake Ontario basin: Geology, v. 17, p. 160-172.

Hickock, S.R., and Dreimanis, A., 1992, Sunnybrook Drift in the Toronto area, Canada: Reinvestigation and reinterpretation. In: The Last Interglacial-Glacial Transition in North America. (P.U. Clark and P.D. Lea, Eds.), Geological Society of America Special Paper 270, pp. 139-161.

Holmes, C. D., 1941, Till fabric. Geological Society of America Bulletin 52, 1299-1354.

Holmes, C.D., 1953, Drift dispersion in west-central New York. Geological Society of America Bulletin 63:993-1010.

Hooyer, T.S., and Iverson, N. R., 2000, Clast-fabric development in a shearing granular material: implications for subglacial till and fault gouge. Geological Society of America Bulletin 112, 683-692

Hooyer, T. S., Iverson, N. R., Lagroix, F., and Thomason, J. F., 2008, Magnetic fabric of sheared till: A strain indicator for evaluating the bed deformation model of glacier flow. Journal of Geophysical Research 113, F02002, doi: 10.1029/2007JF000757.

Isachsen, Y.W., Landing, E., Lauber, J.M., Rickard L.V., and Rogers, W.B., ed., 2000, Geology of New York, a simplified account, Second Edition, New York State Museum Education Leaflet 28, N.Y. State Museum/Geological Survey, State Education Department, Albany, NY.

Iverson, N. R., Hooyer, T. S., Thomason, J. F., Graesch, M. and Shumway, J. R., 2008, The experimental bases for interpreting particle and magnetic fabrics in sheared till. Earth Surface Processes and Landforms 33, 627-645.

Jacobi, R.D., 2002, Basement faults and seismicity in the Appalachian Basin of New York State. Tectonophysics, 353(1), 75-113.

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 139-00, 8 p. http://ny.water.usgs.gov/pubs/fs/fs13900/

Kappel, W. M. and T. Miller, 2005, Hydrogeology of the valley-fill aquifer in the Onondaga Trough, Onondaga County, New York: United States Geological Survey, Scientific Investigations Report 2005–5007, 14 p. http://ny.water.usgs.gov/pubs/wri/sir055007/

Karig, D. E., Miller, N. G., 2013, Middle Wisconsin glacial advance into the Appalachian Plateau, Sixmile Creek, Tompkins Co., NY. Quaternary Research, 80, 522-533.

Karrow, P. F., Bloom, A. L., Haas, J. N., Heiss, A. G., McAndrews, J. H., Miller, B. B., ... & Seymour, K. L. (2009). The Fernbank interglacial site near Ithaca, New York, USA. Quaternary Research, 72(1), 132-142.

Karrow, P.F., 1967, Pleistocene geology of the Scarborough area: Ontario Department of Mines Geological Report 46.

Karrow, P.F., 1969, Stratigraphic studies in the Toronto Pleistocene. Geological Association of Canada, Proceedings, 20, 4-16.

Karrow, P.F., 1974, Till stratigraphy in parts of southwestern Ontario. Geological Society of America Bulletin 85, 761-768.

Karrow, P.F., 2004, Late Quaternary stratigraphic comparisons in south-central Ontario and western New York and the OIS 5E to early 3 interval. Northeastern Geology and Environmental Sciences 26, 202-210.

Karrow, P.F., Dreimanis, A., Barnett, P.J., 2000, A proposed diachronic revision of Late Quaternary time-stratigraphic classification in the eastern and northern Great Lakes Area. Quaternary Research 54, 1-12.

Kerr, M. and Eyles, N., 2007, Origin of drumlins on the floor of Lake Ontario and in upper New York State. Sedimentary Geology 193, 7-20.

Kerr-Lawson, L. J., Karrow, P. F., Edwards, T. W. D. & Mackie, G. L. 1992, A paleoenvironmental study of the mollusks from the Don Formation (Sangamonian?) Don Valley Brickyard, Toronto, Ontario. Canadian Journal of Earth Science 29, 2406-2417.

Knight, F., 1993, The marsh: A New York state wildlife habitat: Conservationist, v. 47, issue 4, pp. 26.

Kozlowski, A.L., et. al., 2011, Glacial land systems and stratigraphy of the Montezuma Wetlands Complex: implications for late quaternary meltwater discharge events in Central New York. Geological Society of America Abstracts with Programs, Vol. 43, No. 1, p. 58.

Laub, R.S., Miller, N.G., and Steadman, D.W, 1988, Late Pleistocene and early Holocene paleoecology and archeology of the eastern Great Lakes region: proceedings of the Smith Symposium, held at the Buffalo Museum of Science, October 24-25, 1986. Bulletin of the Buffalo Society of Natural Sciences 33: 1 316.

Laub, R.S. (Ed.), 2003, The Hiscock Site: Late Pleistocene and Holocene Paleoecology and Archaeology of Western New York State: Proceedings of the Second Smith Symposium, Held at the Buffalo Museum of Science, October 14-15, 2001. Bulletin of the Buffalo Society of Natural Sciences 37: 1 327.

Laub, R.S., and Spiess, A.E., 2003, What were Paleoindians doing at the Hiscock Site? In: The Hiscock Site: Late Pleistocene and Holocene Paleoecology and Archaeology of Western New York State, edited by R.S. Laub, Bulletin of the Buffalo Society of Natural Sciences 37, pp. 261-271.

Lincoln, D.F., 1892, Glaciation in the Finger Lake region of New York. Am. Jour. Sci. (3) 44:290-331.

Lincoln, D.F., 1894, The amount of glacial erosion in the Finger Lake region of New York. Am. Jour. Sci. (3) 47:105-113.

Lincoln, D.F., 1895a, Delta terraces of Seneca County. N.Y. State Mus. Geol. Rep. 14:69-71.

Lincoln, D.F., 1895b, Drumlin belt of Seneca County. N.Y. State Mus. Geol. Rep. 14:69-71. Also: N.Y. State Mus. Rep. 51:69-71.

Loring, S., 1980, Paleoindian hunters and the Champlain Sea: A Presumed Association. Man in the Northeast 19:15-42.

Lothrop, J.C., 2009, The New York Paleoindian Database Project: a call for data. New York State Archaeological Association Newsletter 5(3): 6.

Lothrop, J.C., 1989, The organization of Paleoindian Lithic technology at the Potts Site. In: Eastern Paleoindian Lithic Resource Use, edited by Christopher J. Ellis and Jonathan C. Lothrop, Westview Press, Boulder., pp. 99-138.

Lothrop, J.C., and Bradley, J., 2012, Paleoindian occupations in the Hudson Valley, New York. In: Late Pleistocene Archaeology and Ecology in the Far Northeast, edited by Claude Chapdelaine, Texas A&M University Press, College Station, pp. 9-47.

Lothrop, J.C., Newby, P., Spiess, A., and Bradley, J., 2011, Paleoindian responses to the Younger Dryas in the New England-Maritimes Region. Quaternary International, 242:546-569.

Maury, C.J., 1908, An interglacial fauna found in Cayuga Calley and its relation to the Pleistocene of Toronto. Jour. Geol. 16:565-567.

McCarroll, D., and Rijsdijk, K.F., 2003, Deformation styles as a key for interpreting glacial depositional environments. Journal of Quaternary Science, Vol. 18, pp. 473-489.

Mead, J.I., and D.J. Meltzer, 1984, North American late Quaternary extinctions and the radiocarbon record. Pp. 440-450. In Martin, P. S. and Klein, R. M. (Eds.), Quaternary extinctions: a prehistoric revolution. Arizona University Press, Tucson.

Menzies, J., 1987, Towards a general hypothesis on the formation of drumlins. In Menzies, J., Rose, J. (eds.) Drumlin Symposium, 9-24, A.A. Balkema, Rotterdam, Boston.

Merrill, F.J.H., 1893, Salt and gypsum industries of New York: Albany, N.Y., Bulletin of the New York State Museum, vol. 3, no. 11, 587 p.

Merritt, R.S., and Muller, E.H., 1959, Depth of leaching in relation to carbonate content of till in central New York State. Am. Jour. Sci. 257:465-480.

Milkovich, S.M., and Head, J.W., 2003, Olympus Mons fan shaped deposit morphology: Evidence for debris glaciers: Sixth International Conference on Mars.

Miller, N., 2008a, Contemporary and prior environments of the Hyde Park, New York, mastodon, on the basis of associated plant macro fossils. In: Mastodon Paleobiology, Taphonomy, and Paleoenvironment in the Late Pleistocene of New York State: Studies of the Hyde Park, Chemung, and North Java Sites, edited by Warren D. Allmon and Peter L. Nestor, pp. 151-181.

Miller, N., 2008b The Cohoes Mastodon and Younger Dryas in Eastern New York. In: Glacial Geology of the Northern Hudson Through Southern Champlain Lowlands, by David DeSimone, Gary Wall, Norton Miller, John Rayburn, and Andrew Kozlowski, Guidebook to Field Trips for the 71st Annual Reunion, Northeastern Friends of the Pleistocene, May 30-31 and June 1, 2008. pp 19-25.

Muller, E.H., 1957, Filled bedrock gorges in the drainage basin of Cayuga Lake, New York (Abstract). Geol. Soc. Am. Bull. 68:1771.

Muller, E.H., 1965, Bibliography of New York Quaternary Geology. N.Y. State Mus. Bull. 398, 116p.

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

Muller, E.H., Franzi, D.A., and Ridge, J.C., 1986, Pleistocene geology of the western Mohawk Valley, New York. In: The Wisconsinan Stage of the First Geological District, Eastern New York, edited by D.H. Cadwell, pp. 143-157. New York State Museum Bulletin 455. The University of the State of New York, Albany.

Muller, E. H., 1974, Origins of Drumlins. In Coates, D. R. (ed.) Glacial Geomorphology, 187-204, Binghamton: State University of New York.

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

Muller E.H., and Prest, V.K., 1985, Glacial lakes in the Ontario Basin, in Karrow, P.F., and Calkin, P.E., Quaternary evolution of the Great Lakes: Geological Association of Canada Special Paper 30, p. 213-229.

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. 633-625.

Mullins, H.T., Hinchey, E.J., Wellner, R.W., Stephens, D.B., Anderson, Jr., W.T., Dwyer, T.R., Hine, A.C., 1996. Seismic stratigraphy of the Finger Lakes: a continental record of Heinrich event H-1 and Laurentide Ice Sheet instability. In: Mullins, H.T., Eyles, N. (Eds.), Subsurface Geologic Investigations of the New York Finger Lakes: Implications for Lake Quaternary Deglaciation and Environmental Change. Geological Society of America Special Paper, 311, pp. 1-36.

Newby, P., Bradley, J., Spiess, A., Shuman B., Leduc, P., 2005, A Paleoindian response to Younger Dryas climate change. Quaternary Science Reviews 24:141-154.

Newby, P., and Bradley, J., 2007, Post-glacial landscape changes and Paleoamericans in the New England maritimes regions. Current Research in the Pleistocene 24:16-18.

Oldenburg, D.W. and Li, Y., 1999, Estimating depth of investigation in dc resistivity and IP surveys: Geophysics, v.64, p.403-416.

Oliveira Braga, A.C., Filho, W.M., and Dourado, J.C., 2006, Resistivity(DC) method applied to aquifer protection studies: Revista Brasileira de Geofisica, v. 24, n. 4, p. 573-581.

Person, M., J. McIntosh, V. Bense, and V. H. Remenda, 2007, Pleistocene hydrology of North America: the role of ice sheets in reorganizing groundwater flow systems: Reviews of Geophysics, vol. 45, 28 p.

Petersen, J., 2004, Foreword: West Athens Hill, the Paleoindian Period, and Robert E. Funk in Northeastern Perspective. In: An Ice Age Quarry-Workshop: the West Athens Hill Site Revisited, By Robert E. Funk, pp. ix-xlix. New York State Museum Bulletin 504.

Reynolds, J. M., 2011, An Introduction to Applied and Environmental Geophysics, 2nd ed: John Wiley & Sons, p.

Ridge, J.C., Franzi, D.A., and Muller, E.H., 1991, Late Wisconsinan, pre-Valley Heads glaciation in the western Mohawk Valley, central New York, and its regional implications. Geological Society of America Bulletin 103:1032-1048.

Ridge, J.C., 2003, The last deglaciation of the Northeastern United States: a combined varve, paleomagnetic, and calibrated 14C chronology. In: Geoarchaeology of Landscapes in the Glaciated Northeast, edited by David L. Cremeens and John P. Hart, New York State Museum Bulletin 497, pp. 15-48.

Ridge, J.C., 2004, The Quaternary glaciation of western New England with correlations to surrounding areas, in Ehlers, J. & Gibbard, P.L. (eds.), Quaternary Glaciations – Extent and Chronology, Part II: North America. Developments in Quaternary Science, vol. 2B, Amsterdam, Elsevier, p.163-193.

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.

Ritchie, W.A., 1969, The archaeology of New York State (revised). The Natural History Press, Garden City, New York.

Ritchie, W.A., 1957, Traces of early man in the Northeast. New York State Museum and Service, Bulletin 358.

Ritchie, W.A., 1940, Two prehistoric village sites at Brewerton, New York. Research Records of the Rochester Museum of Arts and Sciences No. 5.

Ritchie, W.A., and Funk, R.E., 1973, Aboriginal Settlement Patterns in the Northeast. New York State Museum Memoir 20.

Robinson, B.S., Ort, J.C., Eldridge W.A., Burke, A.L., and Pelletier, B.G., 2009, Paleoindian aggregation and social context at Bull Brook. American Antiquity 74(3)423-447.

Robinson, F.W., IV, 2012, Between the mountains and the sea: an exploration of the Champlain Sea, and Paleoindian land use in the Champlain Basin. In: Late Pleistocene Archaeology and Ecology in the Far Northeast, edited by Claude Chapdelaine, Texas A&M University Press, College Station, pp. 191-217.

Rockman, M., and Steele, J., 2003, Colonization of unfamiliar landscapes. Routledge, London and New York.

Savage, W. Z., 1968, Application of plastic flow analysis to drumlin formation. M. S. thesis, Syracuse University, 60p.

Schmidt, V.E., 1947, Varves in the Finger Lakes Region of New York State. Ph.D. thesis Department of Geology, Cornell University.

Schrott, L., and Sass, O., 2008, Application of field geophysics in geomorphology: Advances and limitation exemplified by case studies: Geomorphology, v. 93, p. 55-73.

Shagar, S.A., El-Filky, A., and Eid, S., 2006, Delineation of the fresh/salt water zone at a coastal Mediterranean area, West of Alexandria using electrical resistivity investigation: International Journal of Oceans and Oceanography, v. 1, n. 2, p. 299-310.

Sherrod, L., Schlosser, K., Kozlowski, A., Bird, B., Werkema Jr, D.D., and Swiontek, J., 2014, Geophysical characterization of the Keene Valley landslide in New York state, (in press).

Shrake, D.L., Powers, D., Pavey, R.R., Venteris, E.R., Angle, M.P., Larsen, G.E., and Miller, E., 2010, Using LiDAR and high-resolution aerial photography to map karst in Delaware County, Ohio: Geological Society of America Abstracts with Programs, v. 42, no. 2, p. 77.

Shumaker, R.C., 1957, Till texture variation and Pleistocene deposits of the Union Springs and Scipio Quadrangles, Cayuga County, New York: Cornell University M.S. Thesis, 44 p.

Shumway, J.R., and N.R. Iverson, 2009. Magnetic fabrics of the Douglas Till of the Superior lobe: exploring bed-deformation kinematics. Quaternary Science Reviews, 28, 107-119.

Smalley, I. J. & Unwin, D. J., 1968, The formation and shape of drumlins and their distribution and orientation in drumlin fields. Journal of Glaciology, 7, 377-390.

Spafford, H.G., 1824, Gazetteer of the State of New York, an ample history and description of every county, city, town, village, and locality, canals, etc.:, Albany, New York, B.D. Packard, 620 p.

Spencer, J.W.W., 1890, Ancient shores, boulder pavements, and high level gravel deposits in the region of the Great Lakes. Geol. Soc. Am. Bull. 1:71-86.

Spencer, J.W.W., 1894, The rock basin of Cayuga Lake. Am. Geologist 14:134-135.

Spencer, J.W.W., 1912, Hanging valleys and their proglacial equivalents in New York. Geol. Soc. Am. Bull. 23:477-486.

Spiess, A. E., Cowie, E., and Barton, R., 2012, Geographic clusters of Fluted Point Sites in the Far Northeast. In: Late Pleistocene Archaeology and Ecology in the Far Northeast, edited by Claude Chapdelaine, Texas A&M University Press, College Station, pp. 95-112.

Stanford, S.D., 2009, Onshore record of the Hudson River drainage to the continental shelf from the Late Miocene through the Late Wisconsinan Deglaciation, USA: Synthesis and Revision. Boreas 39(1): 1-17.

Steadman, D.W., Craig, L.J., and Bopp, J., 1993a, Diddly Cave: a new late Quaternary vertebrate fauna from New York State. Current Research in the Pleistocene 9: 110-112.

Steadman, D.W., Craig, L.J., and Engel, T., 1993b, Late Pleistocene and Holocene Vertebrates from Joralemon's (Fish Club) Cave, Albany County, New York. The Bulletin of the New York State Archaeological Association 105: 9-15.

Steadman, D.W., Stafford Jr, T.W., and Funk, R.E., 1997, Nonassociation of paleoindians with AMS-dated late Pleistocene mammals from the Dutchess Quarry Caves, New York. Quaternary Research 47:105-116.

Stokes, C. R., Fowler, A. C., Clark, C. D., Hindmarsh, R. C.A., & Spagnolo, M., 2013, The instability theory of drumlin formation and its explanation for their varied composition and internal structure. Quaternary Science Reviews, 62, 77-96.

Storck, P.L., and Spiess, A.E., 1994, The significance of new faunal identifications attributed to an early Paleoindian (Gainey Complex) occupation at the Udora Site, Ontario, Canada. American Antiquity, 59(1):121-142.

Storke, E. G., 1879, History of Cayuga County, New York, with illustrations and biographic sketches of some of its prominent men and pioneers: Syracuse, New York, D. Mason & Company.

Stuiver M, Reimer, P.J., and Reimer, R.W., 2005, CALIB 6.0. http://calib.qub.ac.uk/calib/. Accessed 2012 February 28.

Stupavsky, M., Gravenor, C. P., & Symons, D.T.A., 1974, Paleomagnetism and magnetic fabric of the Leaside and Sunnybrook tills near Toronto, Ontario. Geological Society of America Bulletin, 85, 1233-1236.

Szabo, J. P. (1992). Reevaluation of early Wisconsinan stratigraphy of northern Ohio. Special Papers-Geological Society of America, 99-99.

Taniguchi, M., Ishitobi, T., Burnett, W.C., and Wattayakorn, G., 2007, Evaluating ground water - sea water interactions via resistivity and seepage meters: Ground Water, v. 45, n. 6, p. 629-735.

Tarling, D. H. & Hrouda, F., 1993, The Magnetic Anisotropy of Rocks. 217 pp. Chapman & Hall, London.

Tarr, R.S., 1893, Glacial erosion. Am. Geologist 12:147-152.

Tarr, R.S., 1894, Lake Cayuga, a rock basin. Geol. Soc. Am. Bull. 5:339-356.

Tarr, R.S., 1904a, Artesian well sections at Ithaca, New York. Jour. Geol. 12:69-82.

Tarr, R.S., 1904b, Hanging valleys in the Finger Lake region of central New York. Am. Geologist 33:271-291.

Tarr, R.S., 1905a, Drainage features of central New York. Geol. Soc. Am. Bull. 16:215-228.

Tarr, R.S., 1905b, The gorges and waterfalls of central New York. Am. Geog. Soc. Bull. 37:193-212.

Tarr, R.S., 1905c, Moraines of the Seneca and Cayuga Lake valleys. Geol. Soc. Am. Bull. 16:215-228.

Tarr, R.S., 1906, Glacial erosion in the Finger Lake region of central New York. Jour. Geol. 14:18-21.

Taylor, F.B., 1897, Moraines of recession and their significance in glacial theory. Jour. Geol. 5:421-466.

Taylor, F.B., 1907, Distribution of drumlins and its bearing on their origin (Abstract). Geol. Soc. Am. Bull. 17:726.

Taylor, F.B., 1925, Reconnaissance of the recessional moraines of the Finger Lakes region (Abstract). Geol. Soc. Am. Bull. 36:142.

Thomason, J. F., & Iverson, N. R., 2006, Microfabric and microshear evolution in deformed till. Quaternary Science Reviews, 25, 1027–1038. doi:10.1016/j.quascirev.2005.09.006

Thomson, L.M., McIntosh, G.C., Allmon, W.D., 2008, Discoveries of the American mastodon (Mammut americanum) in New York State: 1922-2007. Paleontographica Americana 61:25-41.

Thorleifson, H., Berg, Richard, C.B., and Russell, H.A.J., 2010, Geological mapping goes 3-D in response to social needs. GSA Today (20) 8:27-28.

Transactions of the New York State Agricultural Society, 1854, Albany, New York, C. Van Benthuysen, Printer to the Legislature: Report of Professor George H. Cook to Harvey Rhoades Esq. – Superintendent of Onondaga Salt Springs, N.Y., vol. 13, p. 137-149.

Van Rensselaer, J., 1823, Essay on salt, containing notices of its origin, formation, geological position and principal localities, embracing a particular description of the American salines; with a view of its uses in the arts, manufactures, and agriculture: Wilder and Campbell, New York, New York.

Vanuxem, L., 1842, Geology of New York. Part III, Survey of 3d Geol. District, New York State. Albany, N.Y., pp 306.

Vreeland, N. P., 2009, Magnetic fabrics of drumlins of the Green Bay Lobe: Implications for the bed-deformation model of drumlin formation. Graduate Theses and Dissertations. Paper 11117. http://lib.dr.iastate.edu/etd/11117

Wall, G., 2008, IroMohawk River discharge and origin of the Cohoes Falls. In: Glacial Geology of the Northern Hudson Through Southern Champlain Lowlands, by David DeSimone, Gary Wall, Norton Miller, John Rayburn, and Andrew Kozlowski, Guidebook to Field Trips for the 71st Annual Reunion, Northeastern Friends of the Pleistocene, May 30-31 and June 1, 2008, pp 17-19.

Weight, W.D., 2008, Hydrogeology field manual: New York, McGraw-Hill, 751 p.

Wellman, B., 1982, A survey of New York fluted points. Archaeology of Eastern North America 10:39-40.

Wellner R.J. and Dwyer T.R. 1996. Late Pleistocene– Holocene lake-level fluctuations and paleo-climates at Canandaigua Lake, New York. Geol. Soc. Amer. Spec. Pap. 311: 65–76.

White, W.A., 1985, Drumlins carved by rapid water-rich surges: Northeastern Geology, v. 7. P. 161-166.

Whittecar, G. R. and Mickelson, D. M., 1977, Sequence of till deposition and erosion in drumlins. Boreas 6, 213-217.

Young, R.A., Burr, G.S., 2006, Middle Wisconsin glaciation in the Genesee Valley, N.Y.: a stratigraphic record contemporaneous with Heinrich Event H4. Geomorphology 75, 226-247.