SUBSURFACE GEOLOGY OF THE LOWER GENESEE RIVER VALLEY REGION: A PROGRESS REPORT ON THE EVIDENCE FOR MIDDLE WISCONSIN SEDIMENTS AND IMPLICATIONS FOR ICE SHEET EROSION MODELS

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INTRODUCTION

Deep bore holes and gravel pit exposures in the Genesee Valley region north of Avon (Figure 1,9) have provided a number of organic-rich sedimentary horizons that have been dated at the University of Arizona AMS facility in the range between 26,000 and >48,800 years BP. A majority of the finite age determinations lie within the interval from 30,000± to 43,700± years BP, implying a Middle Wisconsin age for the deposits sampled. Most of the materials dated were collected from buried glacial outwash, lacustrine sequences, or overridden peat incorporated in younger till. The stratigraphic relationships suggest that the dated materials are most likely correlative with the Plum Point and/or Port Talbot Interstadials best known from localities in Ontario, Canada.

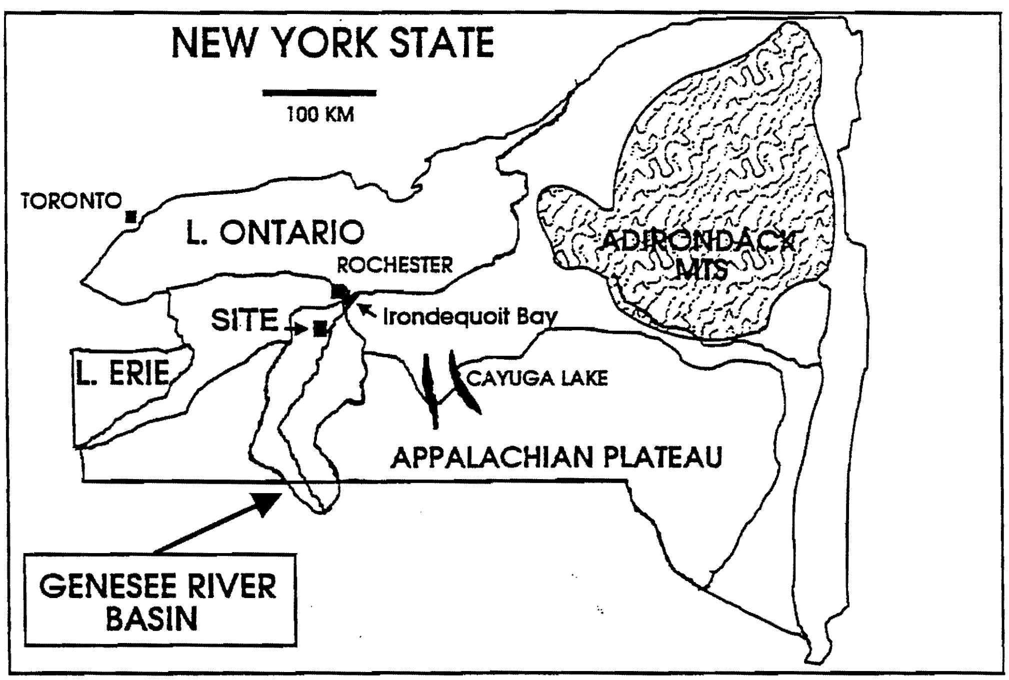

Figure 1: Location of Irondequoit Bay and Middle Wisconsin gravel pit site in Genesee Valley.

The most likely timing of the main ice advance into the Genesee Valley, based on the average of the best constrained ages, is circa 35,000± years BP. This coincides with the recent revision of the H-4 Heinrich event (iceberg discharges) in the north Atlantic and the Dansgaard-Oeschger cycles seen in the Greenland ice cores (Bond and others, 1992, 1993; Taylor and others, 1993).

Whether or not Middle Wisconsin ice crossed the south shore of Lake Ontario and advanced very far into western and central New York has been a controversial subject (Berger and Eyles, 1994; Dreimanis, 1992; Hicock and Dreimanis, 1992). The implied extension of the southern margin of Middle Wisconsin ice sheets into the Genesee Valley and the northern Finger Lakes region has obvious implications for global climatic reconstructions and for additional fine tuning of eustatic sea level curves for the interval represented.

The most diverse and best sampled sections are located in two adjacent sand, gravel, and clay pits in northern Livingston County on the west side of the Genesee River where Middle Wisconsin strata are covered by only 20 to 40 ft (6-12 m) of Late Wisconsin glacial drift. Preservation of these Middle Wisconsin sections in an area of low relief within the broad Genesee Valley raises theoretical questions concerning how the older unconsolidated sediments survived being overridden by the younger, Late Wisconsin ice advance, which extended about 125 km (76 mi) further south into Pennsylvania. The near-surface outcrop location also raises the question of under what circumstances ice sheets scour deep, well-defined bedrock troughs as opposed to advancing across unconsolidated sediments with little apparent erosive impact.

The existence of such a complex Wisconsin section also raises the issue of what similar stratigraphic sequences may be present throughout upstate New York that might not have been recognized due to an absence of preserved or obvious organic horizons. The Middle Wisconsin sediments in the sections studied in the Genesee Valley are very similar in appearance to overlying drift units, and except for the fortuitously preserved organic horizons, the stratigraphy and field relations do not appear very dissimilar from numerous other glacial drift exposures across the region. It is possible that the glacial drift stratigraphy of the region preserves a more complex record of Middle (or older) Wisconsin events than has been assumed, especially within bedrock troughs such as the Genesee Valley and adjacent Finger Lakes.

IRONDEQUOIT BAY SAND BAR SECTION

Deep drill holes to bedrock (continuous split spoon samples) were completed through the Irondequoit Bay sand bar in 1990 to obtain engineering information for anticipated bridge construction across the Bay outlet (Figures 2, 3). R. A. Young was present during much of the drilling and sampling, and carefully collected representative samples for radiocarbon dating. All samples were collected directly from the split-spoon core barrels with stainless steel implements onto aluminum foil and oven dried at 80° C within 24 hours. Depths cited in this paper are relative to lake level elevation, which is approximately 245 feet (ASL). The tops of the drill holes were located 5 feet above lake level along the sand bar (250 ft). English and metric units are both used, depending on the format of references, maps, and data used to compile the results.

The upper 139± ft (42.37 m) of section documents the postglacial rise of the lake from the low stand of Early Lake Ontario, and was described by Young (1983, 1988), by Kappel and Young (1989), and by Young and Sirkin (1994). Sampling of the glacial sediments below this depth has allowed a reinterpretation of the data extrapolated by Young (1983) from the less precise record obtained for the Town of Webster water-supply test wells. The section from 139± ft down to bedrock in the deepest hole (380 ft) contains a section of overridden and reworked lacustrine sediments with scattered intervals of sands, gravels, and thin tills (Figures 3,4,5). This entire lower section appears to consist of a variety of overridden ice-contact and proglacial lacustrine sediments, originally deposited further north in the Ontario basin and redeposited by advancing ice as it flowed southward out of the basin.

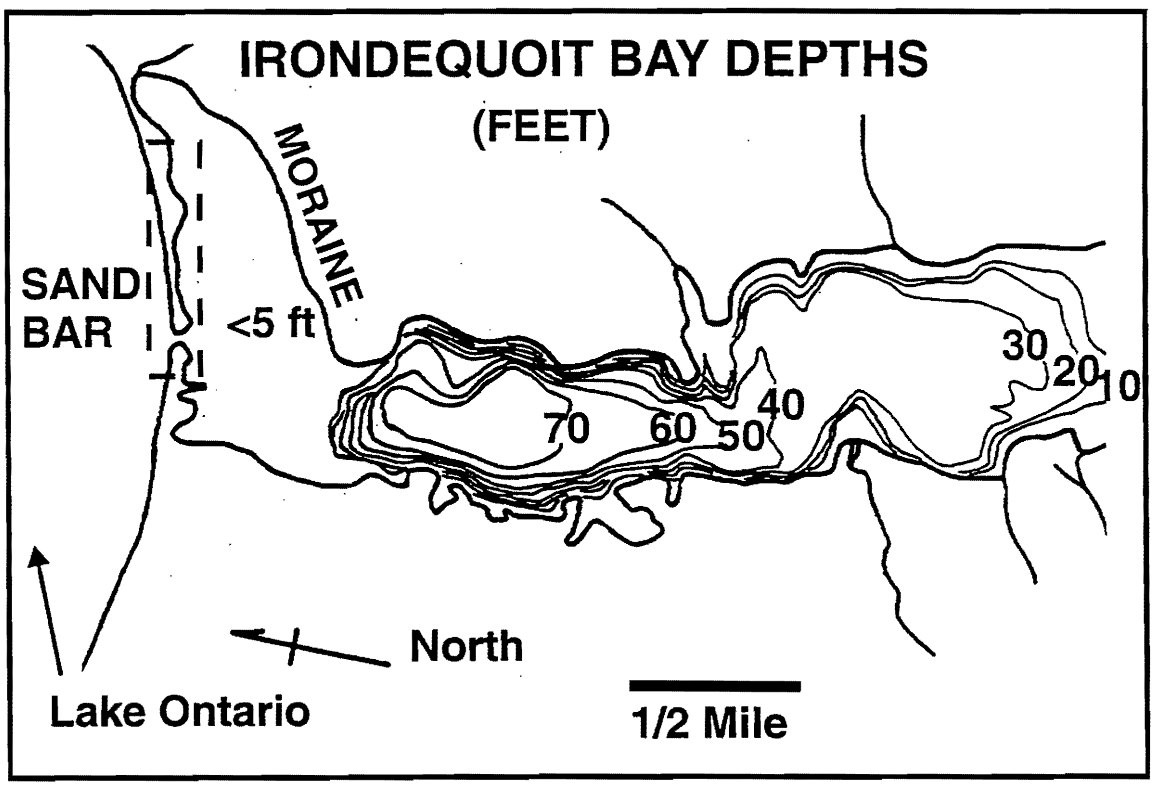

Figure 2: Northern end of Irondequoit Bay with location of sand bar wells (dashed box) and contours showing sediment filling mouth of Bay. Box is also location of Figure 3.

At a depth of 262 ft (79.9 m) a Shelby Tube sample provided a date of 32,000±550 years BP (Table 1). An age of 21,320±170 years BP was obtained at 355 ft (108.2 m) in the same hole. Although both ages may differ slightly from the true age of the surrounding sediments, their inverted age positions support the concept of an ice sheet intermixing younger and older sediments from the lake floor during a readvance. The 32,000 BP age suggests redeposition of Plum Point-agė material from further north in the basin, whereas the lower, younger age would imply a scrambling of the entire lower section (below 137± ft depth) coincident with the Late Wisconsin ice readvance. The true age of this latest advance out of the basin obviously may not be precisely constrained by the single, uncorrected date available from this core. However, the relative ages and positions of the samples, regardless of potential age errors associated with such fine organic sediments, clearly imply an association with Middle and Late Wisconsin depositional events, and both ages are reasonably compatible with the existing record for the Ontario Basin (Dreimanis, 1977; Hicock and Dreimanis, 1992).

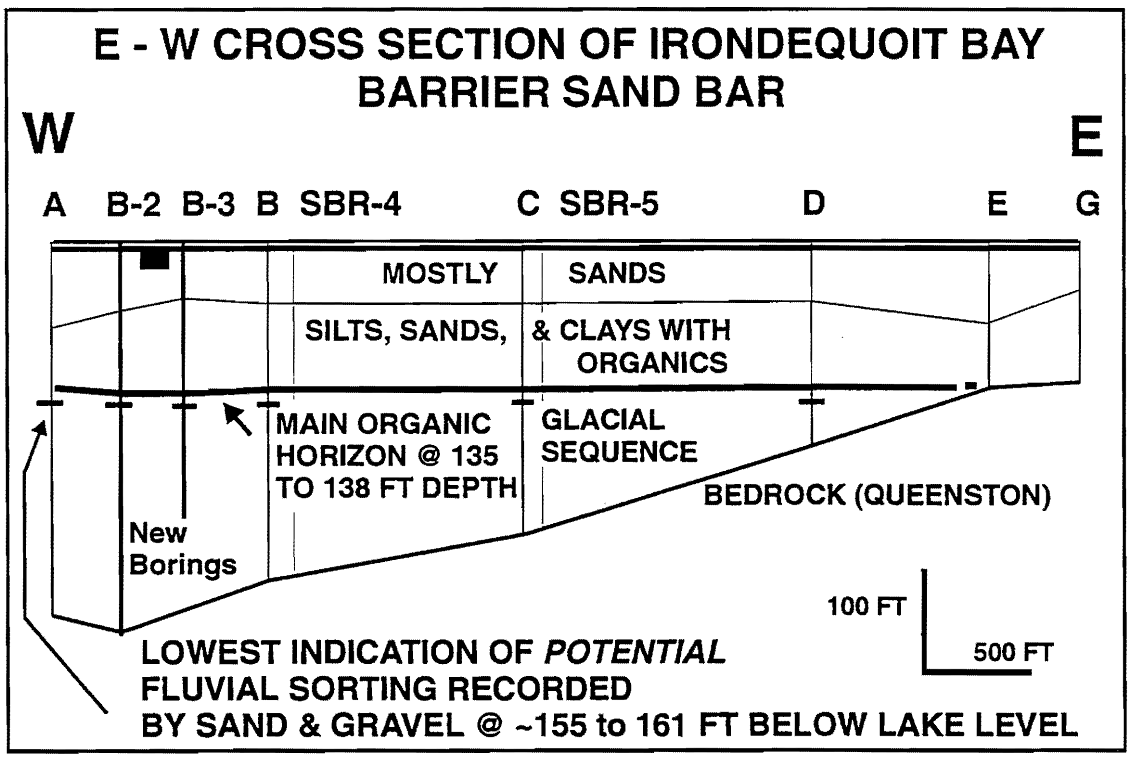

Figure 3: Generalized section through Bar on Figure 2 showing upper, postglacial lacustrine sequence and lower glacial sequence separated by marl and peat horizon near 135 ft depth.

The sequence of lacustrine sediments in the upper 139 feet of sandbar section (Figures 4, 5) records the rise of Early Lake Ontario from a postglacial low stand beginning about 11,500± years BP. The apparent elevation of the lake at that time at the latitude of Rochester appears somewhat inconsistent with published information regarding lake level histories and strandline extrapolations from further to the east and west (Pair and Rodrigues, 1993; Anderson and Lewis, 1985). The paleogeography for the time interval is depicted on Figure 6, which represents the main lake stages bracketing the opening of the St. Lawrence Valley after ice retreated from the northern flank of the Adirondacks.

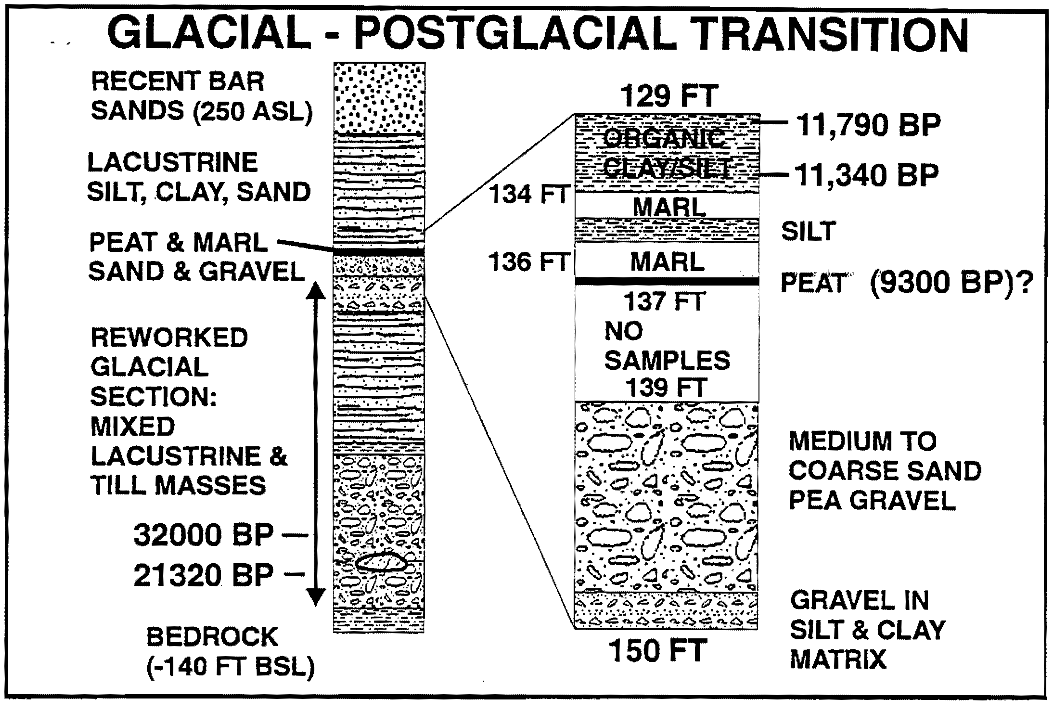

Figure 4: Composite, expanded view of stratigraphy near glacial-postglacial transition in borings B-2 & B-3 (Figure 3) and horizons from which dates were obtained. 9300±BP age is assumed to be in error due to close agreement between two older ages taken from 130 to 134 foot interval. Compare left column with geophysical logs of Figure 5, which demonstrates abrupt change in sediment properties at inferred glacial-postglacial contact.

It is possible that the two, similar radiocarbon dates (11,340± and 11,790±BP) in the Irondequoit Bay core do not, in fact, record the lowest postglacial position of the lake (allowing several feet for normal wave base fluctuations). However, the simplest interpretation of the stratigraphy and the cross section profile of the Bay outlet imply that postglacial erosion by Irondequoit Creek would have rapidly incised through the unconsolidated sediments and would have closely coincided with the minimum elevation to which the lake fell (Figure 7). Given the 1 to 10 m^3/sec average flow of Irondequoit Creek today, it would appear reasonable that a similar stream discharge across the glacial sediment threshold beneath the present-day bar would have rapidly eroded down to the base level of Early Lake Ontario.

INTERPRETATION

Lake Levels

The assumed depositional environment associated with the southward progradation of the sand bar and the accumulation of organic sediments behind such a bar in a rising lake is diagrammed in Figure 7B. In such a sand bar environment, with a laterally shifting outlet and occasional storm wave resedimentation, it is likely that undisturbed, vertical sediment accumulation would be the exception. Erratic incision and redeposition of organic-bearing sediments during lateral migration of the bar outlet could have produced disruption of the normal stratigraphic order as sampled in isolated borings. The relief across the modern Bay outlet and bar is greater than the total thickness of the dated interval between 130 and 138 feet. The elevation differences between the older and younger postglacial samples is small and within the range of wave-base disturbances found on modern bars. The 9300± BP peat sample from near 142 ft rested on a sand and gravel interval in boring B-3, whereas the samples dated at 11,340± (B-3) and 11,790± (B-2) came from between 130 and 134 ft in their respective borings on either side of the modern outlet (Figures 3, 4).

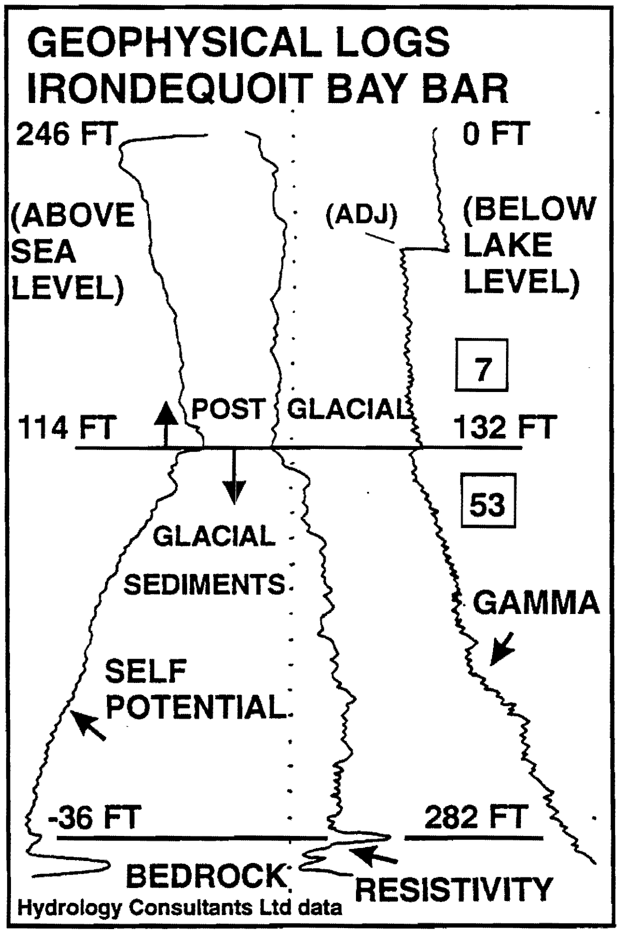

Figure 5: Geophysical logs from boring C (Figure 3) showing contrast between postglacaial lacustrine sequence (top) and denser sediments below. Numbers in boxes indicate average split-spoon blow counts (140 lb hammer) for 6-inch sampler penetration. Change in sediment properties with depth on geophysical traces are approximate because sediment properties are averaged by downhole probes. Adj.= adjustment of recorder trace (not change in sediments). Measurement units not available (graphs included for relative changes in properties only).

Pollen Record and Radiocarbon Dates

The pollen data near 11,500 BP from the Irondequoit Bay bar show mainly a pine, hemlock, and oak forest (see Appendix) with other minor deciduous species plus marsh and field (nonarboreal) flora. There is an apparent absence of spruce at a time (11,500 BP) when spruce was present in the northern and northeastern Ontario Basin and the Ottawa Valley region (Anderson, 1988; Anderson and Lewis, 1985). The effect of a potential pine pollen influx from the upstream (southern) Genesee River basin headwaters is difficult to assess adequately. It could be argued that the postglacial vegetation succession from spruce- to pine-dominated forest expanded more rapidly northward along the Genesee Basin, with its lower elevations and direct connection to more southerly forests in Pennsylvania. This atypical basin geography might have allowed more rapid expansion of forest zone succession or delivery of pollen to the Irondequoit Bay area atypical of the postglacial succession characterizing the more northern Lake Ontario shore at that time. However, a total lack of spruce pollen in the two small samples (Appendix, Dia. 5), compared with the north and northeastern lake shores, is somewhat troubling if the circa 11,500± BP ages are reliable.

Data from elsewhere in the Genesee River basin (south of Geneseo) may further clarify this apparent contradiction. When Interstate route I-390 exploratory borings in the Genesee Valley were drilled through the alluvial section into the glacial sediments two miles south of Mt. Morris (see accompanying field trip route maps), two dates were obtained near the glacial-postglacial transition from lacustrine clay to peat (near 32 ft depth). A date of 10,730 BP was obtained from the organic residue washed from clays cored between depths of 31 and 35 feet (below postglacial peat transition). In an adjacent hole, a piece of wood at a depth of 32± ft (0.8 ft above the peat-clay transition) was dated at 11,160 BP (assumed stratigraphically higher than 10,730 age). Examination of pollen from the latter hole, two inches above the wood date, indicated pine with minor spruce pollen (3 grains; Appendix, Dia. 4). Pine with traces (single grains) of hemlock, spruce, and birch were present about 6 inches above the wood (Appendix, Diagram 4).

In a third hole, immediately above the peat-clay transition (depth 32 ft), an undated sample contained pine, hemlock, and grass pollen with minor oak and nonarboreal pollen, but no spruce (Appendix). These closely spaced sites (presumably all near 11,160 BP in age) would have been collecting pollen from around a shallow postglacial lake as well as from upstream (south) in the Genesee and Canaseraga Basins (Muller and others, 1988). Assuming the wood date (11,160 BP) is the more reliable, and that the sample closest to the peat-clay transition (no spruce) is at least as old as the wood horizon, these data lend support to the possibility that spruce-dominated forest had been largely replaced by pine at this time in a major portion of the lower (northern) Genesee Valley, somewhat earlier than is documented for the northeastern end of Lake Ontario (Anderson, 1988). These additional data, from about 25 miles (40 km) south of Rochester, lend support to the possibility that the 11,500±BP ages for the "spruce-free" horizon at Irondequoit Bay are not unreasonable for that latitude and time.

Alternatively, if the <u>younger</u> peat age (9300±BP) obtained at a depth near 136 feet at Irondequoit Bay (Figure 5) is the more accurate date, the age discrepancy could be due to contamination of the two higher (older) samples with older carbon contamination from the lake water. This choice would still leave unresolved the issue of why fluvial erosion by ancestral Irondequoit Creek (area: >395 km^2; >153 mi^2) would not have been capable of incising the glacial sediments underlying the sand bar down to the lowest contemporaneous lake level following deglaciation at a time inferred elsewhere to be close to 11,500±BP (Pair and Rodrigues, 1993). The glacially reworked (overridden) lacustrine sediments are neither exceptionally dense nor strongly consolidated (blow counts, Fig. 5), and thick, stony till sections are absent. The average modern stream discharge (1 to 10 m^3/sec) from the basin appears to be more than adequate to erode completely down through such weakly consolidated materials in the period of several hundred years that probably separated the Lake Iroquois and Early Lake Ontario stages (Pair and Rodrigues, 1993).

<u>Relation of Lake Stage to Sand Bar</u>

The documented stages of Lake Iroquois and Early Lake Ontario and their approximate ages, as discussed by Pair and Rodrigues (1993), are shown in Figure 6. Incision of the glacial section by ancestral Irondequoit Creek had to occur after the drop to Early Lake Ontario, as shown in Figure 7. The subsequent rise of Lake Ontario caused the southward (onshore) migration of the sandbar and trapped organic materials on the shallow, protected, south side of the bar (similar to the modern bar setting).

This scenario probably means that the deeper borings penetrate the base of the postglacial sand bar sequence at a point somewhat landward (south) of the initial position of the ancestral bar, which presumably began to form near the lowest stand of the lake. However, any Irondequoit Creek channel that would have been incised through the center of the modern Irondequoit Bay valley and sand bar should still be detectable at the present location of the bar. The Creek had to erode through up to 300 ft (91 m) of sediment within the confines of the modern Bay, based on the thickness of glaciolacustrine sediments remaining in the modern Bay bluffs (Kappel and Young, 1989). As the lake continued to rise, the gap in the sand bar associated with the Irondequoit Creek outlet would have become gradually broader and less incised, but the outlet is likely to have maintained a position closely aligned with the incised channel reach immediately to the south. The Creek alignment appears to have been slightly closer to the east side of the Bay at its narrowest, northernmost projection (Figure 2), but on a nearly direct line from the Bay into Lake Ontario (Kappel and Young, 1989).

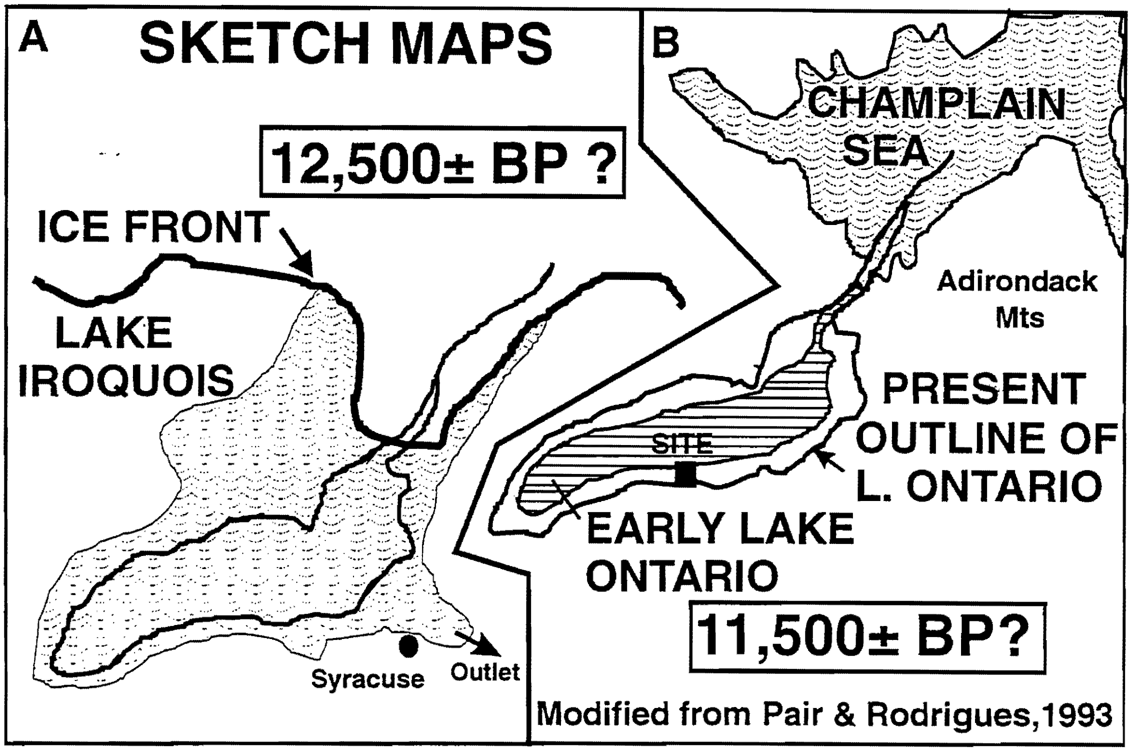

Figure 6: Recession of the ice sheet from the St. Lawrence Valley led to the low Early Lake Ontario stage coincident with the near incursion of the Champlain Sea into the basin while sea level was low and postglacial rebound less complete than at present.

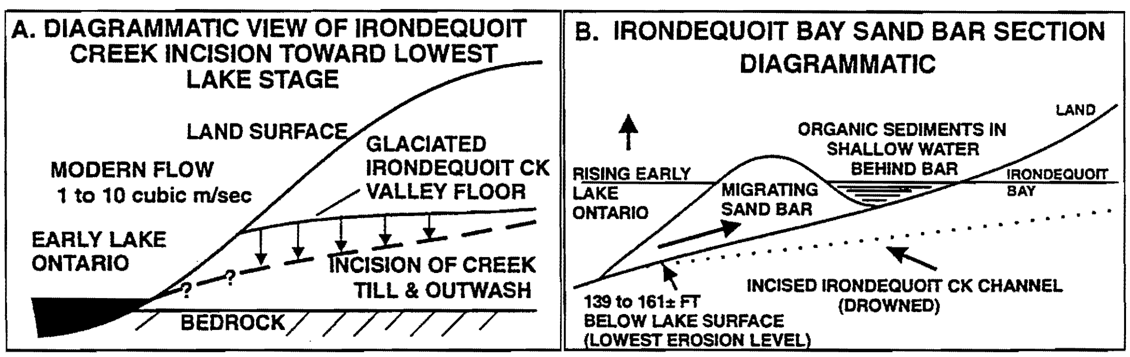

Figure 7A: Diagrammatic cross section of incision of Irondequoit Creek through glacial section following establishment of Early Lake Ontario. 7B: Progradation of sand bar southward with rising lake stages. Organic sediments become buried by bar migration.

Figure 8 illustrates the discrepancy between the Rochester data for the lowest elevation of Early Lake Ontario recorded at the Irondequoit Bay sand bar and the published data of Anderson and Lewis (1985). It should be noted, however, that the data from the east end of the Lake are much better constrained and more numerous than the data available for the region to the west. Dates from the western end of Lake Ontario generally involve estimates of water depths for organic horizons in lake-bottom cores. Work in progress by

T.W. Anderson (personal communication, 1994) on the Oakville-Grimsby bar may resolve some of these issues.

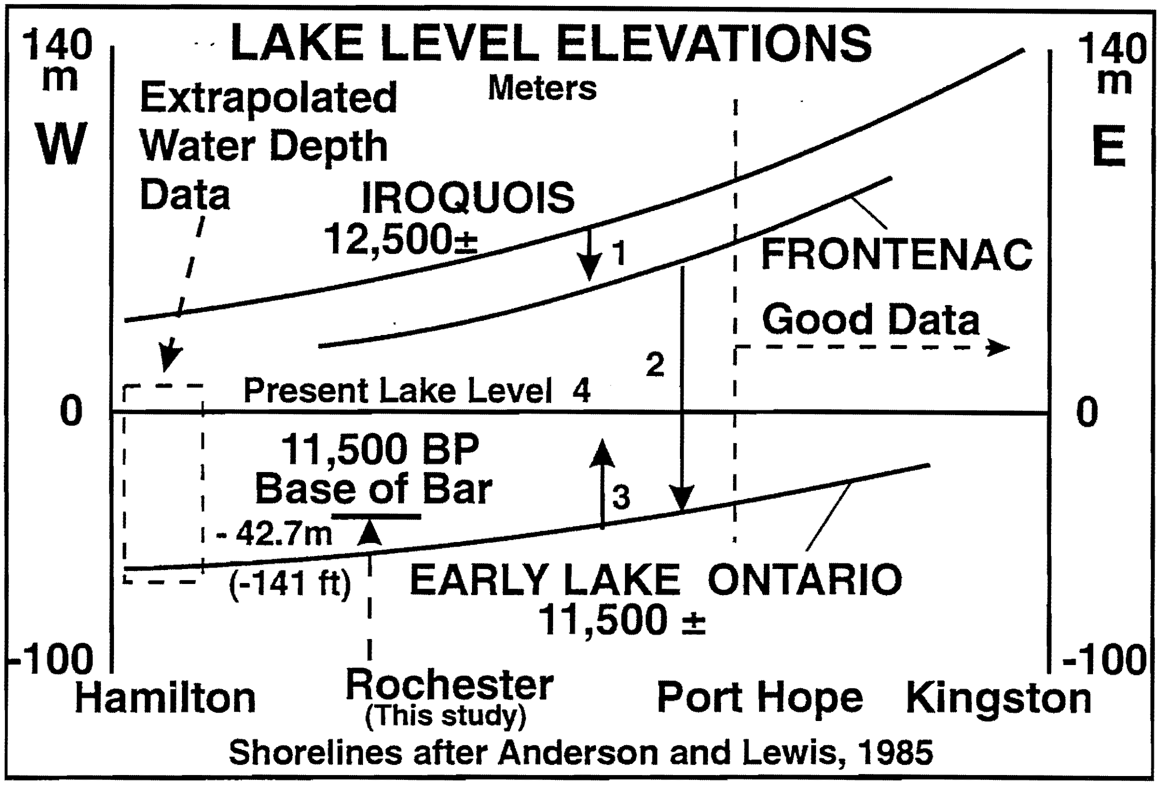

Figure 8: Sequence of falling and rising lake levels (numerical order) illustrating discrepency between earlier work and Rochester data. Dashed box is area of uncertain water depth data.

DISCUSSION

The presence of sediments with ages between 32000± and 21320± years BP near the base of the Irondequoit bar section demonstrates that organic-rich lacustrine sediments in this approximate age range were overridden by late Wisconsin ice, and probably brought to their present latitude near Rochester from an uncertain position beneath the modern Lake Ontario basin. The precise age of the sediments may be in doubt because of potential contamination from older carbon likely to be present in lake water with rivers draining such a large region. However, the 21,320± age could represent a maximum age limit for the passage of Late Wisconsin ice across the south shore of Lake Ontario and into central New York, assuming it represents contemporaneous lake-bottom sediment incorporated in the ice sheet as it advanced southward out of the basin. The age is in reasonable agreement with the data of Miller and Calkin (1992) and Muller and Calkin (1993).

The depth of maximum postglacial erosion beneath the sand bar is estimated from the position of the lowest lacustrine beds and immediately underlying, loosely consolidated sediments with textures compatible with fluvial or wave deposition. The postglacial-glacial transition is at least 137 to 139 ft in depth and possibly as deep as 161 ft in boring C, where coarse sands and gravels are described (older water test well logs). The most prominent, continuous horizon in all the borings is the organic horizon at depths between 135 and 138 ft. This depth is assumed to represent the earliest lake level with a conspicuous organic facies located in the lee of the sand bar (Figure 7B). The slightly older (lower) sandy to gravelly (wave scour?) zone may mark the actual low stand of the lake at the site. The contrast in blow counts on Figure 5 suggests that the level is near the organic horizon. In any event the level of Early Lake Ontario had to be at least 137 to 161 feet lower than at present near Rochester at some postglacial time. It seems reasonable to assume that a creek with mean flows somewhere between 1 and 10 m^3/sec would have easily cut through relatively unconsolidated sediments and become graded to the lowest level of Early Lake Ontario in a short period. Thus, the elevation of the glacial-postglacial transition observed at the bar must be close to the lowest level of the lake (regardless of age uncertainties).

There is one other (unlikely) possibility, which could allow for a lower lake, not recorded in the existing borings. The distribution of existing borings could have missed the deepest portion of the incised outlet channel across the bar, and the dated sediments might represent a slightly younger (higher) interval, containing sediments deposited during the southerly migration of the bar toward its present location. The only logical location for such a hypothetically deeper outlet channel would be between borings SBR-4 and C (Figure 3), which are separated by about 1000 feet. The stratigraphic reconstruction from available borings, combined with the bay-bottom contours and local physiography (Figure 2), argue for an outlet close to the position of the modern one. Any outlet displaced further to the east would require the ancestral creek to have veered sharply eastward, parallel to the adjacent moraine (Figure 2), just as the creek descended the steeper slope toward the lake shore. Such an abrupt course change on a steepening slope appears very unlikely. The most likely outlet is presumed to have been located near its present position, somewhere between A and SBR-4 (Figure 3).

The sand bar data may better constrain the elevation of Early Lake Ontario at one of the few well-documented, subsurface stratigraphic localities in the western half of Lake Ontario (Anderson and Lewis, 1985) . However, the inconsistent distribution of radiometric ages and the lack of spruce pollen at the dated ~11,500± BP level pose uncertainties that need to be better resolved.

MIDDLE WISCONSIN SITE IN NORTHERN LIVINGSTON COUNTY

The glacial geologic framework of the Genesee Valley was described by Muller and others (1988). Figure 9 records the location of a shallow gravel pit section studied between 1991 and 1994 by R.A. Young with pollen studies completed by Les Sirkin. The general stratigraphy (Figure 10) contains the following sequence of units beginning at the base of the exposed section:

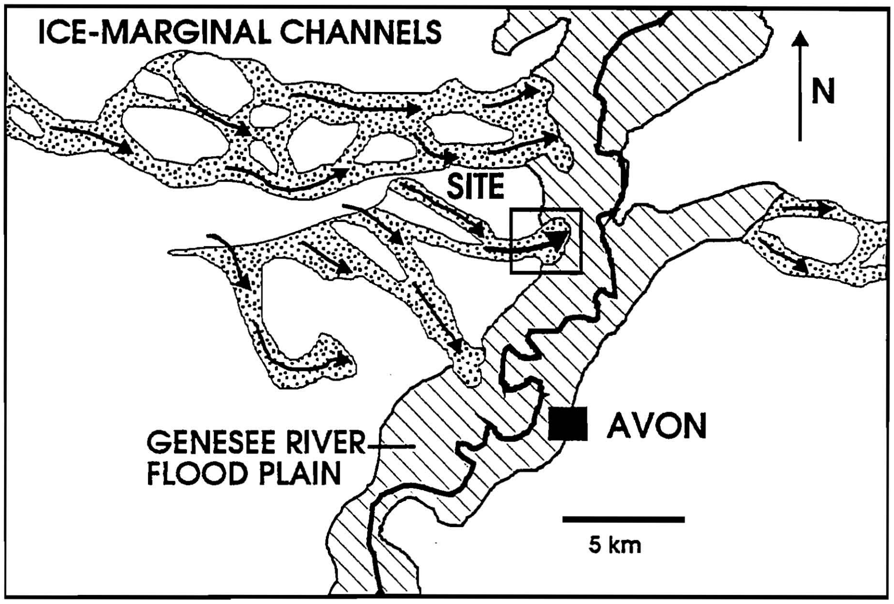

Figure 9. Location of Middle Wisconsin site in northern Livingston County.

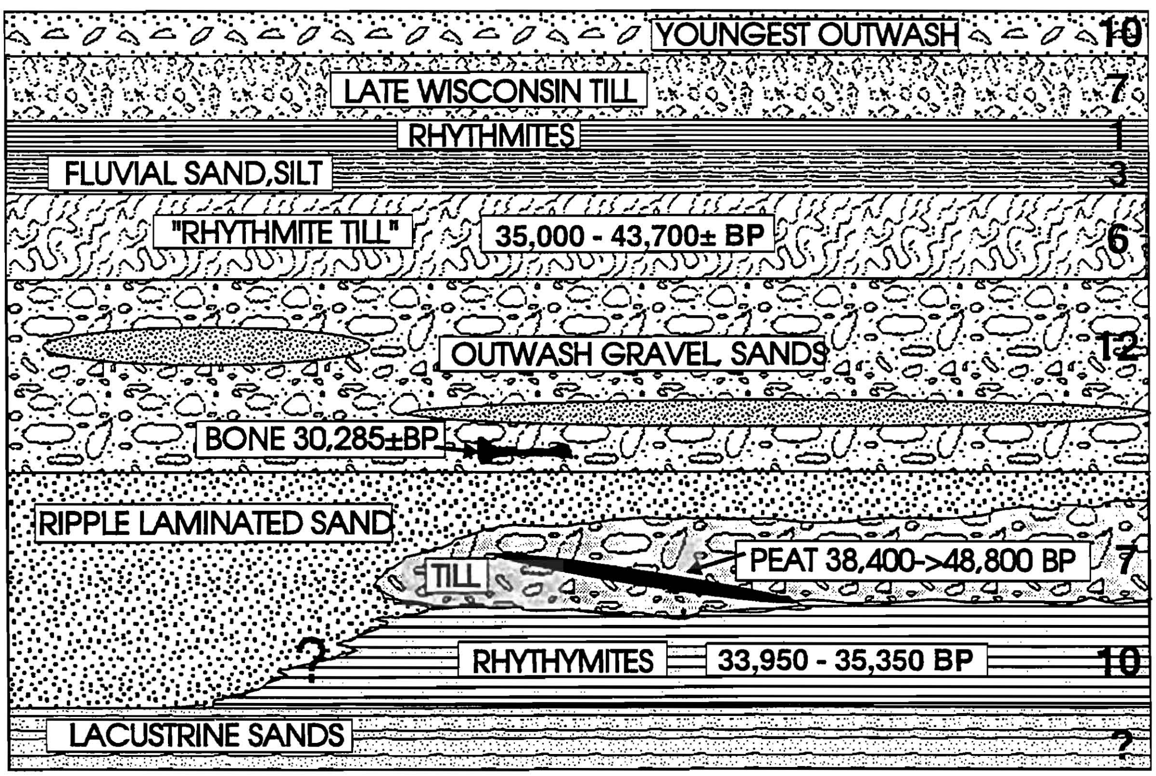

Figure 10. Diagrammatic section of Middle Wisconsin site of Figure 9. Average thicknesses of units at right side in feet. There is no conspicuous evidence of weathering profiles between units except irregular oxidation front (brown) extending from surface(?) down into gray "rhythmite till."

1) Undated lacustrine(?) sands and silts of unknown thickness underlying the lowest rhythmically bedded silty clays, which are being excavated for landfill construction. The existing pits are water-filled, and the base of the section is poorly exposed. These basal sands and silts were penetrated in older test borings (pit foreman, oral communication) and the uppermost beds were briefly examined near the base of one exposed section just above the pit water level. The contact with the overlying rhythmites appeared sharp and conformable.

2) A lower rhythmite sequence about 10 feet (3 m) thick with darker (finer-grained) and lighter (coarser-grained) beds with individual couplets averaging from 1 to 3 cm. thick, but with both thicker- and thinner-bedded intervals. This unit shows evidence of ice deformation and folding of beds in portions of the pit, but the stratigraphy is generally coherent, near horizontal, and not overturned.

3) A clast-poor, gray till (7 ft; 2 m) with small fragments and larger beds of compressed peat scattered throughout. One extensive, 6- to 12-inch-thick bed of peat was sampled for the initial radiocarbon analysis near the lower till contact. This peat bed dipped to the northeast into the excavated section and its lower limit was not visible. On the easternmost exposed edge of the excavations the "peat till" was observed to thin and terminate in a drainage trench, where it was surrounded by a sequence of deformed, ripple laminated sands.

4) A relatively thick sequence of outwash sands and gravels (12 ft; 2.5 m). The gravel contains thick, cross-bedded, poorly sorted gravel units, as well as prominent sand layers, which are generally thin but relatively extensive (traceable for tens of meters). A fragment of an ice-overridden mastodon or mammoth rib (proximal end) was recovered from the floor of a drainage trench dug through this unit. The bone fragment was slightly compressed, cracked, encased in coarse sand and pebbles cemented to its exterior, and impregnated with clay from the pressure of overriding ice.

5) A section of fine-grained, gray sediments, grading laterally from undeformed rhythmites, to folded and contorted beds of the same materials, to a massive silt and clay with a till-like appearance (6 ft; 2m). In less adequate exposures these sediments might not have been recognized as correlative with each other, and would probably have been described as separate units (clay till and lacustrine beds). The massive part of the unit appears very similar to "clay tills" occasionally seen in restricted outcrops or penetrated in engineering borings. The pressure of overriding ice appears to have produced selective, spontaneous liquefaction of the rhythmite texture, converting the beds to a massive, structureless "till" in some sections (Young, 1993). Some of the fine couplets in the preserved rhythmites contained thin black, organic laminae about 2 mm thick (single date of 26,000 ±BP). Subsequent ages on thicker, more organic-rich beds provided the generally older ages plotted on Figure 10.

6) A thin sequence of red-brown fluvial (outwash) sands and silts grading upward into reddish-brown rhythmites (3 ft; 1 m).

7) An uppermost, stony, red-brown till, representing the last (Late Wisconsin) ice advance across the region (7 ft; 2m).

8) Outwash sands and gravels (largely removed in older excavations) poorly exposed across the disturbed work areas under spoil piles (10 ft; 3m?). These outwash gravels may originally have been covered by an unknown thickness of lacustrine sands, silts, and clays associated with the last proglacial lake stage in the valley (as observed in a shallow gravel pit I mile to the north).

POLLEN RESULTS

The silt-sized sediments in the lacustrine units and the peat from the lower till were sampled for pollen analyses (Figure 11). The pre-Woodfordian spruce zones (Sirkin and Stuckenrath, 1980) are dominated by arboreal pollen of pine, spruce, and oak with *Ericaceae*, birch, willow, poplar, and minor hickory, hemlock, and larch (Appendix). The nonarboreal pollen contains pondweed, grasses, composites, and *Thalictrum*. The basic significance of the overall results is that they support the evidence that the deposits record environments dissimilar from typical Late Wisconsin sections. The finest-grained portion of the uppermost lacustrine beds (from which a single anomalous age of 26,600± BP was first obtained) produced only a single oak pollen grain. The presence of such hardwood flora could be evidence of far-traveled contamination from the ancestral Genesee River headwaters to the south. Subsequently, silts from this unit indicated a pre-Woodfordian spruce zone pollen assemblage (Appendix, Diagram 1), and provided ages more consistent with the bulk of the sediments above and below the peat-bearing till (Figure 10).

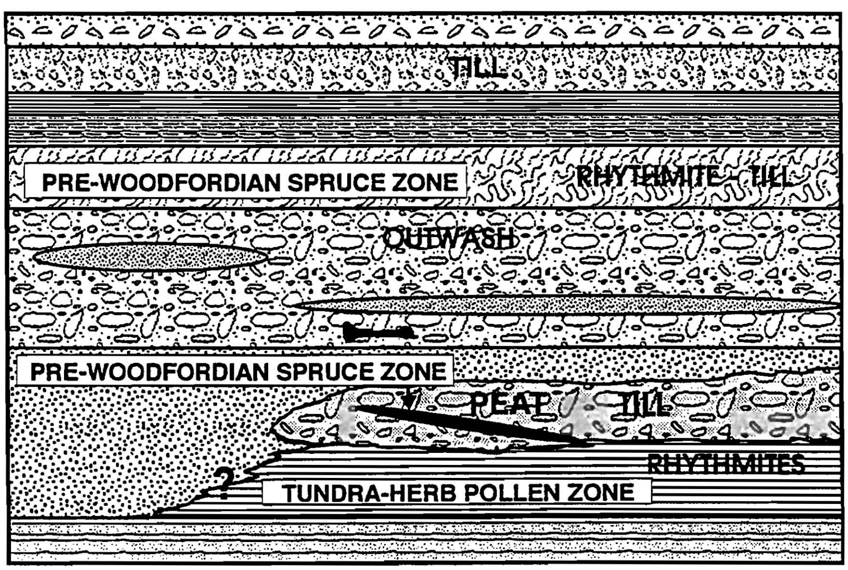

Figure 11. General pollen associations determined by Les Sirkin (See Appendix).

The lower lacustrine unit contains an apparent pre-Woodfordian tundra assemblage (herb pollen zone; Appendix, Diagram 3) with willow, rose, and herb pollen (Sirkin and Stuckenrath, 1980). This implies an important vegetation change between the lower lacustrine unit and the environment associated with the earliest ice advance and recession in the section.

AGE DETERMINATIONS

At first glance radiocarbon ages obtained from the section (Figure 10) are somewhat contradictory or indeterminate. The ages are more consistent and coherent than may be apparent for a number of reasons. Unfortunately, the older peat ages are apparently very near the normal limit of routine AMS C^{14} dating (±48,000 years BP). In addition, some of the sediments sampled come from proglacial lakes (all dated proglacial lakes represent advances, not recessions) fed by streams that must have drained "established" forests in the upper reaches of the deglaciated Genesee Basin. Such forests, supplying detritus to a proglacial lake during readvance of Middle Wisconsin ice up the Genesee River basin, may have been supplying fine-grained organic debris from plants representing a significant time interval preceding Middle Wisconsin glaciation. Thus organic detritus with a range of radiocarbon ages may have been entering the proglacial lake from the upstream (southern) reaches of the basin and accumulating in fine-grained sediments. Finally, the bone date from the glacial outwash has the usual age uncertainties typically associated with bone material (more specific amino acid analyses are pending at INSTAAR, University of Colorado by T.W. Stafford Jr.).

Despite these problems, compelling evidence exists in the section (Figure 10) to suggest that the ages (Table 1), taken as a whole, support a Middle Wisconsin ice advance into west central New York at approximately 35,000 years BP for the following reasons:

1) Most of the finite ages from both the upper and lower lacustrine units cluster consistently in the interval between 33,000 and 36,000 years BP, well within the range of reliable carbon 14 age determinations. This interval straddles the late Port Talbot, early Plum Point interstadials and the intervening ice advance documented in Ontario, Canada. The ages fit reasonably well with an ice advance correlative with units such as the Titusville till, Meadowcliffe Till, or Seminary Till from regions north and west of western New York State. The dates are close to ages of 41,900± and 39,900± years BP from the Cayuga Lake trough by Bloom (1967, see reference comment), which imply Middle Wisconsin ice damming of the Cayuga Lake outlet around the same time (also see Schmidt, 1947).

2) The significantly greater range of ages obtained from the peat incorporated in the lower till are consistent with an ice advance over terrain where older peat bogs of pre-Middle Wisconsin age would have been established (Figure 12).

Thus the apparent reversal of ages between the basal rhythmites and overlying till can be logically attributed to the advance of ice that incorporated the peat in the lower till and deposited it over "younger" (Middle Wisconsin) proglacial lacustrine sediments, which contained organic residues from contemporaneous forest cover growing near the advancing ice margin. The relatively high organic content of the basal lacustrine beds indicates an ice advance into a lake receiving drainage from well-vegetated (tundra?) areas, rather than proglacial lakes of an ice recession, where vegetation cover might be sparce (as is generally inferred from late Wisconsin recessional lake sediments in the Genesee Valley). The advance of the ice across an older peat stratum with pre-Middle Wisconsin ages, in itself, suggests a Middle Wisconsin advance, following development of significant vegetation cover.

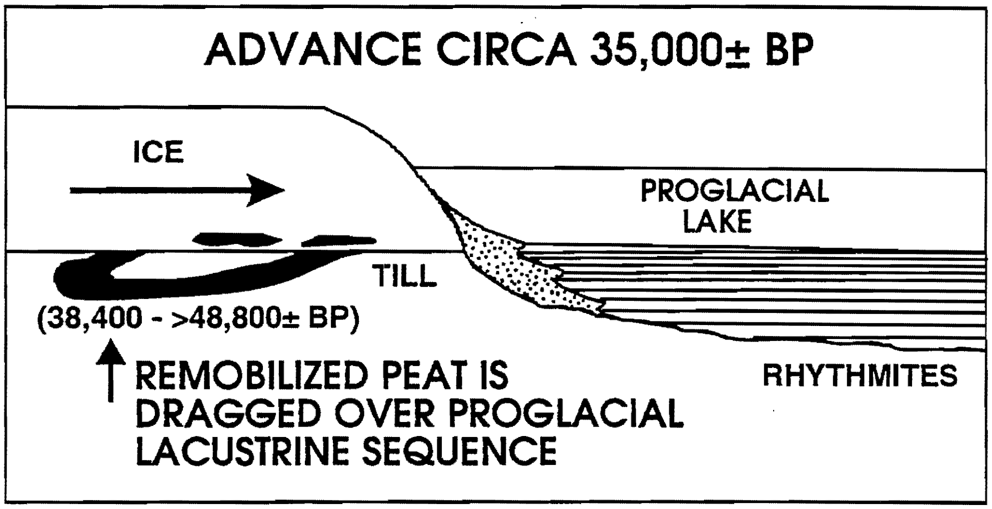

Figure 12. Inversion of radiocarbon age relations by remobilization of peat and redeposition onto younger proglacial lake sequence during advance.

The wider range of ages in the peat samples may represent the extended time interval of peat development, now scrambled and compressed by overriding ice. In the field it was impossible to reconstruct any coherent peat stratigraphy from scattered peat fragments in the till, or to determine if the peat beds were upside down from the scattered inclusions. Thus any small peat sample chosen for an AMS C^{14} date could have come from the top, middle, or base of a highly compressed peat sequence, affected by the passage of both Middle Wisconsin and Late Wisconsin ice sheets. A range of ages representing a period in excess of 10,000 years does not seem unreasonable for the peat, if it represents a considerable portion of the likely time interval between Early and Middle Wisconsin ice advances (~40 kyr to ~60 kyr BP?).

3) Although the ^{14}C ages extend over a substantial range, several finite ages within the upper and lower lacustrine units are internally consistent, averaging about 35,000 BP. This interval (30,000 to 38,400± BP) would include the formation of proglacial lakes associated with ice damming of the valley both

during advance and retreat of Middle Wisconsin ice. Thus the ages represent the lacustrine "events" bracketing the time that ice occupied the valley, not simply the age of the advance. This analysis indicates that the major advance appears to be constrained to a time near 35,000±BP (average of consistent, finite lacustrine dates in closest agreement).

This average age is very close to the revised oceanic iceberg detritus horizon (Heinrich event H-4) dated at 35.5 kyr BP (Bond and others, 1993). The iceberg horizons are correlated with the Dansgaard-Oeschger warm-cold oscillations recorded in the Greenland ice cores (Bond and others, 1993). Current climatic modeling associates the large-scale iceberg releases in the North Atlantic with episodes of continental ice sheet expansion. The northern hemisphere cooling event near 35.5 kyr BP is well constrained by AMS dates on both ice core and ocean core samples. The finite Genesee Valley dates are in reasonably good agreement with this oceanic and ice core data. The time interval indicated is consistent with ages that have been obtained in association with the Titusville Till event in northwestern Pennsylvania, where radiocarbon ages indicate that an advance probably occurred between 40,000 and 33,120 years BP (Muller and Calkin, 1993).

4) The time interval involved is a portion of the radiocarbon time scale for which precise calendar-year corrections are not available. Thus it is likely that some of the apparent inconsistencies would be less marked if atmospheric or other corrections were available for these older dates, as they are for the interval younger than 20,000 years BP.

MECHANISMS AND IMPLICATIONS FOR ICE SHEET TRANSPORT AND EROSION

The Finger Lakes region and adjacent portions of west-central New York are often cited as examples of the effects of deep glacial scour by continental ice sheets and the formation of closed glacial troughs eroded below sea level. Bedrock subcrop data show that the structure and stratigraphy of the local rocks have focused erosion in some major valleys along the south-dipping ramp of the resistant Onondaga Formation, eroding the less resistant shales above. The buried Genesee Valley is equivalent to one of the larger Finger Lakes in overall shape and dimensions, and its N-S bedrock profile also follows the Onondaga subcrop ramp south of Avon, NY. It differs from a true Finger Lake because of the large, through-flowing Genesee River, which was capable of effectively eroding an outlet to Lake Ontario and of filling its proglacial lake basins with sediment more effectively than its smaller counterparts . Given the magnitude of the deep ice scouring of the buried bedrock landscape, the location and shallow depth of the Middle Wisconsin section on the west margin of the Genesee Valley seem anomalous.

The southernmost extent of Late Wisconsin ice in western New York was at least 80 miles (130 km) further south than the Middle Wisconsin site in northern Livingston County. This means that a significant thickness of Late Wisconsin ice overrode the site without severely eroding this shallow section of weakly consolidated sediments. This evidence appears to contradict the common scenario associated with ice advances through this region. There is an obvious need to explain why an ice advance of a magnitude similar to those that eroded the deep Finger Lake troughs appears to have slid over the surface of the drift near Avon with so little effect on this occasion. The duration of the advance, the basal ice velocity, the ice thickness, and the basal ice thermal regime all influence the degree of erosion by an ice sheet. Several factors may partially explain the lack of erosion.

An ice advance into a proglacial lake is one of the ways to potentially reduce the weight of the ice mass and to reduce friction at the glacier base. At this latitude the ice front was continuously advancing into a gradually rising lake, controlled by the outlet cols along the valley divides (Muller and others, 1988). After the ice overrode the region for several miles the effect of buoyancy in the lake at the latitude of this site would no longer be a factor. This would seem to leave reduced friction at the contact of the ice with fine-grained, water-saturated sediments as the major contributor to reduced ice erosion. This suggests that the deeper, more dramatic scour of the Finger Lakes troughs occurred either during times of prolonged ice flow under significantly different thickness and thermal conditions and/or at significantly different ice velocities.

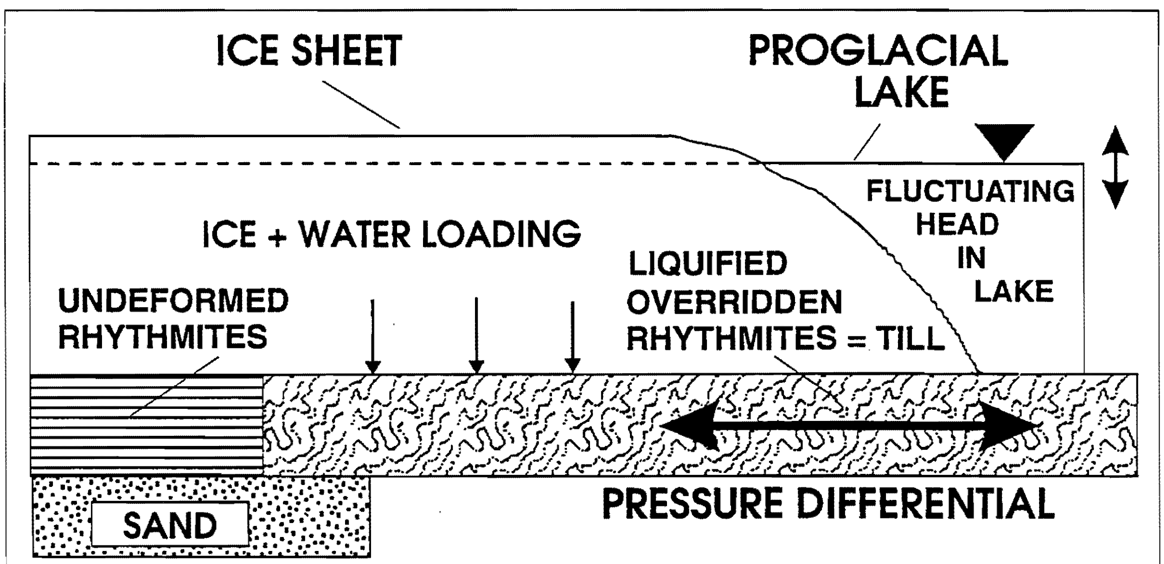

Figure 13: Diagram of hypothetical ice conditions leading to spontaneous liquefaction of rhythmite sediments by ice advance into proglacial lake.

RHYTHMITE LIQUEFACTION TO PSEUDO "TILL" TEXTURE

Massive clay beds at the site contain evidence of spontaneous liquefaction of rhythmically bedded silts and clays, and provide evidence of the importance of processes involving hydrostatic head fluctuations across and through saturated sediments at the ice margin. The textural modifications produced by this process can provide insight into the interpretation of some commonly observed till textures.

Figure 13 illustrates the conditions hypothesized to have created the variable textures observed in the "rhythmite till" (Figure 10). The abrupt lateral transition from undisturbed to deformed rhythmite bedding and, suddenly, to massive silty clays within a single bed (Figures 13, 14) demonstrates that the structures within the continuous unit were modified from an initially uniformly bedded unit. It is likely that marked or sudden head differentials could develop between the glacial meltwater on (or within) the glacier and the waters in the adjacent proglacial lake. Such head variations could be caused by fluctuations in lake outlet elevations or by changing seasonal meltwater conditions near the ice front and could fluctuate markedly. These potentially rapid fluctuations could create instantaneous pressure changes and strong hydraulic gradients within the water-saturated sediments near the glacier-lake interface. These steep gradients or potentially sudden pressure changes could cause spontaneous liquefaction of the fine-bedded rhythmite textures, converting them to the more massive textures observed at the northern Livingston County site (Figure 14).

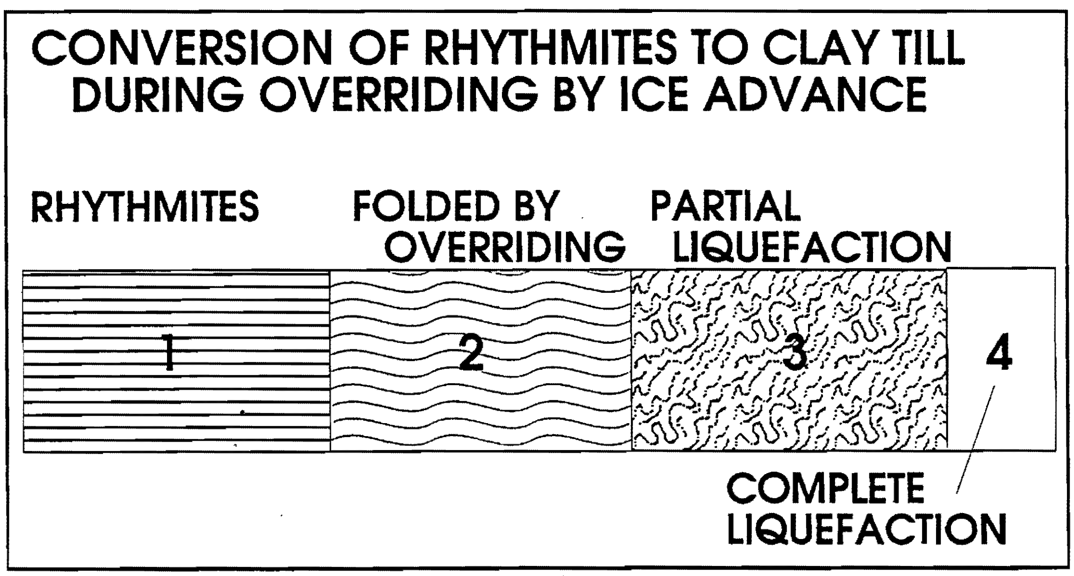

Figure 14. Diagrammatic lateral transition of ryhthmite sequence (1) to massive clay "till" (4) seen in pit exposures

The lateral transition from rhythmites to structureless silty clays creates a massive texture similar to a clast-poor till. Because proglacial lacustrine sequences commonly contain numerous ice-rafted drop stones, the massive beds also contain occasional clasts. In a case where the liquefaction of a rhythmite sequence has been complete, or where the exposures are limited, it is possible to mistake such a massive silt-clay unit for a true till. The implications for interpretation of the associated events are obvious.

There are numerous descriptions of clast-poor, silt-clay tills in engineering boring logs, in test pit logs, and from natural exposures that may imply a similar origin. An awareness of their possible significance is important to stratigraphic studies. One potential way to identify such "liquefied rhythmite tills" is by grain-size analyses. Many rhythmite couplets (varves) have a bimodal particle-size distribution that readily shows up in pipette size analyses when plotted as Phi size versus cumulative weight using a probability scale. The two size populations, usually interpreted as "summer" and "winter" varve couplets, should be diagnostic of the bimodal populations associated with rhythmite bedding.

In this part of central New York, given the common condition of ice advances into proglacial lakes along north-draining valley axes, the existence of "liquefied rhythmite tills" may be more common than has been recognized. It is not unusual in glacial drift sections for tills of sharply contrasting textures to be noted, some being very stony, whereas other tills in the same section are notably clast-poor. The mechanism described above is a reasonable alternative to explain some of these marked textural differences. Simple mechanical "mixing" or shearing of a rhythmite sequence by overriding ice is unlikely to destroy the laminated texture completely, as is obvious from observed basal till exposures that contain inclusions consisting of large masses of deformed rhythmites.

DISCUSSION AND SUMMARY

The occurrence of such a shallow, Middle Wisconsin glacial section, covered by relatively thin Late Wisconsin sediments, has important implications for the existence of a more widespread, fragmentary record of Early and/or Middle Wisconsin sequences in central and western New York. Although the major north-draining glacial valleys are the most obvious locations where such sections might be preserved in the subsurface, the ability of shallow, unconsolidated glacial sediments to survive overriding by ice of 80 miles (130 km) suggests that Early or Middle Wisconsin sequences may have survived ice scour in other locations.

Assumptions concerning the lateral continuity of Late Wisconsin tills and lacustrine units, even over short distances, are obviously subject to potential errors. The complex section described in northern Livingston County contains several Middle Wisconsin units located well above the modern floodplain elevation. The individual exposed beds are relatively indistinguishable from younger tills, outwash gravels, and rhythmites seen elsewhere in restricted outcrops along the Genesee Valley. In the absence of continuous exposures or other compelling field evidence, it would be appropriate to be cautious when making lateral or temporal correlations of similar units in the absence of datable materials. The same cautions apply to correlations based on engineering or geophysical parameters, including subsurface boring logs, paleomagnetic data, and seismic images, where there may be lateral discontinuities, unrecognized unconformities, or significant gaps in data sets.

It is clear from the Livingston County site that thick, rhythmically-bedded sequences of lacustrine sediments cannot always be attributed to Late Wisconsin recessional proglacial lake stages. Glacial advances apparently can override and preserve lacustrine sequences that appear to have been formed contemporaneously with ice advancing southward up the north-draining Genesee Valley. Without adequate chronological data and exposure of both upper and lower contacts of lacustrine sediments with older and younger units, it would be difficult to adequately access the geologic context of such isolated lacustrine exposures.

Finally, the existence of clast-poor, clay-rich tills in locations where the physiography was favorable for the creation of proglacial lakes during major ice advances may have resulted from the spontaneous liquefaction of lacustrine sediments by hydraulic head differences along ice margins. Comparison of grain-size distributions of lacustrine rhythmites and clast-poor "clay" tills might provide a means discerning till sheets that were potentially derived from ice advances overridding proglacial lacustrine sequences. The possible presence of organic residues, pollen, or scattered wood fragments in such lacustrine-derived tills might provide additional means by which to study glacial chronology and pre-Late Wisconsin events. AMS radiocarbon dating of small samples, while subject to obvious contamination errors, can provide additional information to improve stratigraphic knowledge of Middle to Late Wisconsin chronology.

ACKNOWLEDGEMENTS

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Reviews by Ernest H. Muller and Paul F. Karrow helped the writers to improve the text and the discussion of unresolved issues.

TABLE 1. RADIOCARBON AGES KEYED TO FIGURES 3,4 & 10

IB = Irondequoit Bay bar; Figures 3,4 (Boring No., feet below lake level)
LC = Livingston Co. gravel pit; Figure 10 (horizon)
GV =Genesee Valley south of Mt. Morris near Pioneer Road (PR) or Keshequa Creek (K) intersections with railroad (depth).

LOCATION; DEPTH or HORIZON[Ψ]	AGE in YEARS BP	LAB. NO.*	MATERIAL
IB (B-2,129 ft)	11,790±80	AA-8632	Organic silt
IB (B-3,132.5 ft)	11,340±90	AA-8636	Organic silt.
IB (B-2, 257 ft)	32,000±550	AA-8633	Organic silt.
IB (B-2, 350.5 ft)	21,320±170	AA-8634	Organic silt.
IB (B-3, 137 ft)	9,300±65	AA-8637	Thin peat.
LC (top basal rhythmites)	33,950±650	AA-10790	Gray/black rhythmites, fine organics.
LC (top basal rhythmites)	35,350±770	AA-10791	Gray/black rhythmites, fine organics.
LC (peat in lower till)	38,400±1200	AA-8638	Compressed peat bed.
LC (peat in lower till)	>44,350	AA-10789	Compressed peat bed.
LC (peat in lower till)	>48,800	AA-10789R2	<u>Repeat of above</u>; Rigorous acid-base cleaning.
LC (lower outwash)	30,285±480	AA-8639	Mastodon(?) rib bone (δ^{13} C = -20.3).
LC ("rhythmite till")	26,680±300	AA-8640	Very thin black lamina from rhythmite couplet. (Exposed, contaminated?)
LC ("rhythmite till")	35,000±740	AA-10792	Tiny black organic blebs in rhythmite couplets.
LC ("rhythmite till")	43,700±2100	AA-12126	Abundant, fibrous organic debris in silt-clay matrix.
GV (PR; 31-35 ft)	10,730±150	I-9952	Organic residue seived from lacustrine silts, clays.
GV (K; ~32 ft)	11,160±160	I-9972	Wood with bark, 1 in dia. branching specimen within peat and 0.8 ft above peat -lacustrine transition

[*Ages from University of Arizona, NSF-Arizona AMS Facility (AA) & Teledyne (I)]
[ΨBelow <u>lake level</u> for bar sites; bar projects ~5 feet above lake to elev. 250± ft]

<u>Note added in proof:</u> AA-8639 bone was redated (amino acid extraction) at **45,800±2800 (CAMS-14611)** T. Stafford (INSTAAR). <u>Wood</u> from lower till gave finite age, **46,337±2982 (AA-14584).** These new age determinations suggest Port Talbot I age for reworked peat and bone included in lower till-outwash sequence. The ice-overridden bone fragment (crushed and impregnated with clay) must have been reworked into outwash associated with (directly above) lower till and probably was derived from same older interstadial deposits as peat bed <u>in</u> lower till.

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APPENDIX
Pollen Histograms

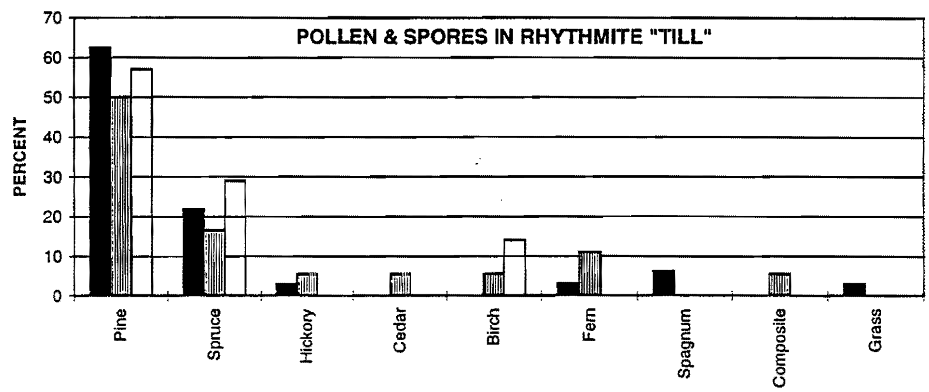


Diagram 1. Pollen and spores from three random samples near top of overridden rhythmite section (Figure 10) or "rhythmite till". Histogram bars are precentages of total sample, based on number of grains counted. Samples: Solid bars (32 grains); Open bars (7 grains); Hatchured bars (18 grains).

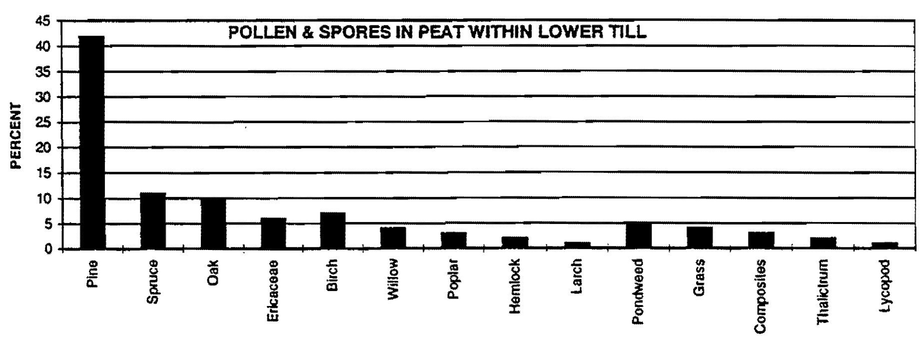


Diagram 2. Pollen and spores from single sample of overridden peat incorporated in lower till (Figure 10). Histogram bars are percentages of total sample, based on numbers of grains counted. Number of grains in sample = 102. Abundant Spagnum spores not included.

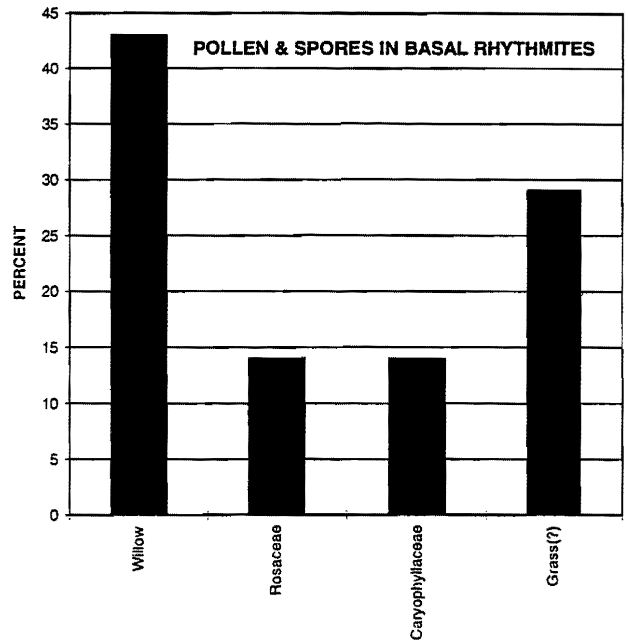


Diagram 3. Pollen and spores from single sample of lower rhythmites below peat-bearing till (Figure 10). Recovery was only seven grains (converted to percent for histogram) plus a few reworked, oxidixed pre-Cenozoic spores (not included in graph).

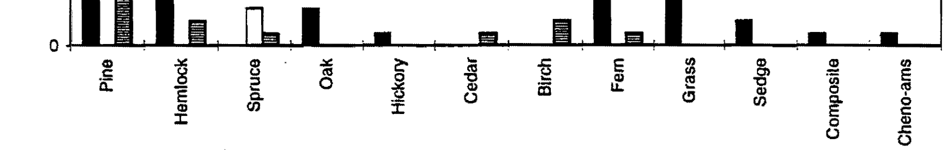


Diagram 4. Pollen and spores from three samples in borings for I-390 in Genesee Valley. Histogram bars are pecentages of total sample, based on number of grains counted. Samples: Solid bars = Boring at Pioneer Rd.and railroad crossing; depth 32 ft., contact of peat over lacustrine clay (44 grains). Hatchured bars = boring at Keshequa Creek and railroad; depth 31.5 ft., in peaty clay section (23 grains). Open bars (17 grains) are sample from same boring as hatchured bars, but at 32 ft. depth, just above wood dated at 11,160 years BP. Peat - lacustrine clay transition is at 32.8 ft in same boring. Peat over lacustrine silts and clays contact is assumed to be glacial-postglacial transition in local section. Date on organic residue seived from lacustrine sediments below peat at depths from 31 to 35 feet gave age of 10,730 years BP. Borings are separated by ~1 mile along valley axis. Compare low amounts of spruce pollen with Diagram 5 for sediments of similar age at Irondequoit Bay (spruce counts were 3 and 1 grains in open and hatchured bar plots respectively, Diagram 4). (Sites 2 mi. south of Mt. Morris; trip route map.)

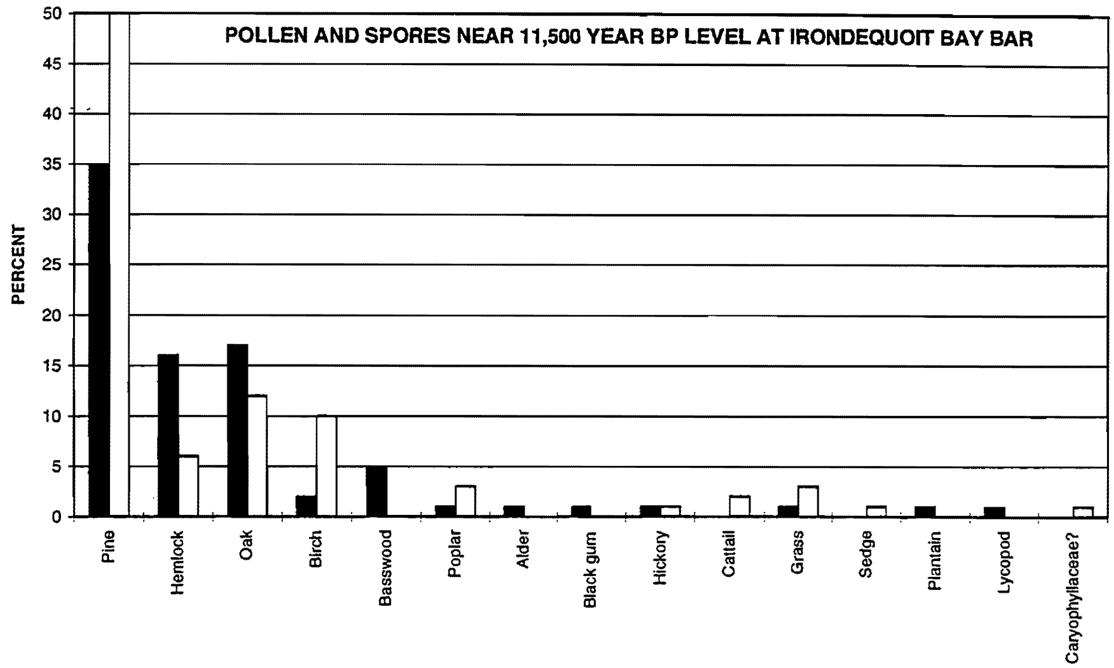


Diagram 5. Pollen and spores from 2 samples near base of postglacial lacustrine sequence at Irondequoit Bay bar. Histogram bars are percent of total sample, based on numbers of grains counted. Samples: Solid bars = depth of 133.5 feet below lake level, boring B-2 (Figures 3, 4). Plot based on 82 grains from organic clay and silt horizon. Open bars = depth of 137 feet below lake level, boring B-3 (Figures 3,4). Plot based on 88 grains from organic clay and silt horizon. Note relative absence of spruce pollen; compare with Diagram 4 horizon of similar age.

NOTE: These pollen histograms are based on small grab samples acquired from split spoon cores at Irondequoit Bay and from the Genesee Valley near Pioneer Road and Keshequa Creek, or from exposed horizons sampled for dating at the gravel pit in Livingston Count (Figure 10). The samples from engineering borings completed by the New York State Department of Transportation for Interstate Route I-390 and for the sand bar at Irondequoit Bay were all limited by the fact that the projects were designed to obtain undisturbed engineering samples for highway or bridge designs, projects which were not under the control of the authors. Therefore, larger or more continuous vertical samples for pollen profiles were not available. The apparent Middle Wisconsin ages were not originally anticipated when the opportunistic samples were first collected. The pollen histograms for the drill hole test samples were completed on remaining core samples only after the ages and locations indicated the unusual chronologic potential of the sites.

Limited access to the gravel pit site of Figure 10 was also a problem. The site was being excavated at an accelerated rate for borrow materials, as well as for planned construction of an "industrial park". Work on the best and deepest parts of the site was severely limited by ongoing heavy equipment activity, construction work schedules, water levels in portions of the pit, and liability concerns of the owners. The collecting was limited to a few occasions and the emphasis was placed on documenting stratigraphic relationships and a rapid search for datable material before the best exposures were lost. Collections for pollen profiles were not contemplated until the unanticipated age results were obtained, by which time the largest peat bed exposure and most of the organic-rich, upper "rhythmite till" had been removed. The earliest organic materials collected were mostly sacrificed for the radiocarbon dating of samples. The finer lacustrine clays yielded little pollen, which appeared to be concentrated in the siltier units. An attempt was made obtain pollen data from as close as possible to each of the dated horizons, or from correlative horizons where the original sample locations were no longer accessible.

ROAD LOG FOR LOWER GENESEE RIVER VALLEY
MIDDLE WISCONSIN SITES AND GLACIAL GEOMORPHOLOGY
IRONDEQUOIT BAY OUTLET
TO
CANASERAGA VALLEY

Objective: The purposes of this trip are: 1) to visit the two sites in the Genesee Valley that have recently produced the evidence for a Middle Wisconsin glacial advance (~35,000 BP) to near the latitude of Avon, N.Y., and 2) to provide an overview of the glacial geomorphology created by Late Wisconsin ice withdrawal from the lower basin. The events that shaped the surficial geomorphology of the modern valley are clearly related to the major proglacial lake stages, lake outlets, and prominent moraines found from the Valley Heads Moraine near Dansville to the Pinnacle Hills Moraine in Rochester. The geologic setting of the valley can best be appreciated by referring to the review articles by Muller and Calkin (1993), and Muller and others (1988). The existence of shallow, Middle Wisconsin tills, outwash gravels, and lacustrine sequences close to the elevation of the modern flood plain within the broad, open valley pose obvious questions about mechanisms of Late Wisconsin ice erosion and the possibility that similar sections might be preserved elsewhere in central New York.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Trip mileage record begins at Irondequoit Bay outlet, west side, end of Sea Breeze Expwy. (From U. of Roch. take I-590 north past exit for Rt.104 East, which becomes Lake Rd. or Sea Breeze Expressway and ends 2.7 miles north at boat ramp parking lot).

STOP 1. SITE OF BAYMOUTH SAND BAR BUILT DURING RISE FROM EARLY LAKE ONTARIO (Text, Figure 2).

This location provides one of the few good overviews of the relief and topography of Irondequoit Bay, the baymouth bar, and their relationships to Lake Ontario. An orientation stop will be made here to view cross sections and large out-of-print and unpublished maps of geology and engineering projects used to interpret sections cored through the sand bar to depths of 380 ft (116 m). The geologic relations in the accompanying text Figures (2-8) will serve to focus the discussion. The Rt. 104 bridge can be seen spanning the bay 1.5 miles to the south. Eighteen deep exploration boring logs provide a detailed cross section view at the bridge (Kappel and Young, 1989) to compare with the subsurface stratigraphy cored by water test wells and bridge foundation borings at the sand bar.

The long bluffs to the southeast (nearly parallel to the bar), which dominate the anomalous northeast trend of the adjacent Bay shore appear to represent the north edge of an unnamed moraine potentially correlative with the Carleton Moraine to the west. An ice-contact feature in this position would reasonably account for the unusual and abrupt change in the shoreline trend near the mouth of the Bay. Gravel pits formerly present on the south edge of this northeast-trending ridge exposed coarse gravel and sand sequences with a southerly dip (now under Stony Pt. development), presumably the outer margin of a kame moraine complex. Older topographic maps (1935) show the top of the bluffs to have been a low ridge bordered by a large irregular depression at its western limit.

A moraine of this age would have been closely associated with the withdrawal of ice near the time of Lake Iroquois, whose shoreline (elev. 435± ft) lies about a mile south of the moraine at the location of Ridge Road (Rt. 104).

5.4	**5.4**	Return to Sea Breeze Expwy. and follow north to Browncroft Blvd. Exit, turn left on Rt. 286 toward Penfield.
6.3	**0.9**	Turn right on Landing Rd. at bottom of hill.
6.7	**0.4**	Park on left in lot for Ellison Pk.

Stop 2. KAME MORAINE FAN AND LACUSTRINE SEQUENCE ON SOUTH EDGE OF BURIED PINNACLE HILLS MORAINE

Walk down slope to Irondequoit Creek, proceed upstream to Bridge and cross to exposure of Kame moraine fan with lacustrine sequence at top.

The Pinnacle Hills Moraine (most prominent pre-Lake-Iroquois moraine) extends from the University of Rochester to near this location where it is largely buried within the older Irondogenesee Valley, covered by lacustrine sediments from contemporaneous and younger glacial Lakes Dana(?), Dawson, and Iroquois (Fairchild, 1928; Muller and others, 1988). The till portion of the moraine seems to divide the groundwater regimes of the upper and lower Irondequoit Basin. The subsurface extent of the moraine was studied in a series of borings in an attempt to define its effect as a groundwater barrier and is described in Kappel and Young (1989). Groundwater upstream of the moraine appears to be locally more saline (trapped road salt) than groundwater north of Browncroft Blvd. This stop will utilize the cross section Plates from Kappel and

Young (1989) and Figures from Young (1988) to examine what is currently known of the moraine and the size of the buried valley.

8.7	2.0	Return to Sea Breeze Expwy. (I-590) and continue to Blossom Rd. I-490 junction. Continue south on I-590 (center lanes).
12.7	4.0	I-590 curves west and Pinnacle Hills moraine can be seen clearly on right side past Winton Rd. Exit ~ 2 miles to north (radio towers).
13.5	0.8	Exit to I-390 S.
22.6	9.1	Take Rush Exit 11 to Rt. 251W.
23.5	0.9	Follow signs through complex intersections.
27.2	3.7	Drive west on Rt 251 across Genesee R. to left turn on River Rd. (south)
28.8	1.6	Stop at gravel pit on left opposite old Valley Sand & Gravel sign.

STOP 3. VIEW AREA OF MIDDLE WISCONSIN SECTION SHOWN IN FIGURES 9, 10, 11 OF ACCOMPANYING ARTICLE.

The geology of this stop is the focus of the second half of the accompanying article. It is the only complex Middle Wisconsin exposure in New York with the range of different depositional units, organic horizons, and multiple ice advances recorded both in this pit and the DeWitt gravel pit immediately to the south. The site has changed markedly over the past 4 years due to rapid extraction since the peat horizon in the lower till was discovered by R.A. Young in November 1990. The major units and dated horizons have largely been removed, but most of the section from the "bone horizon" (Figure 10) upward is still visible (June 1994) and the only good wood specimen found to date was collected from the lower till during a summer 1994 visit (submitted to Arizona AMS Lab). The local deposits have been used to supply "clay" to the new Monroe County (Rochester) Mill Seat Landfill in Riga, thus accounting for the rapid excavation of the lower rhythmite section, largely below the present pit water table. The success of this visit will depend entirely on how much excavation or construction occurs during the 1994 summer season and how much remains of the section shown in Figure 10 between this writing (June 1994) and October of 1994.

The DeWitt pit immediately adjoining this site on the south will be included in the stop, either by hiking across both pits (water conditions permitting) or rejoining the bus for a short ride. Participants are cautioned to stay clear of all vertical bluffs where very large (and dangerous) segments of the clay and gravel units slump unexpectedly.

If conditions are favorable time spent at this stop may be extended and the remainder of the trip shortened to allow interested participants to adequately explore this important, but rapidly disappearing site.

32.8	**4.0**	Continue south on River Road to Rt. 5. Left at intersection.
34.7	**1.9**	Cross Genesee River and turn right on River St.
35.6	**0.9**	South to Rt. 39., Right (south).
36.15	**0.55**	Cross Conesus Lake outlet.
36.5	**0.35**	Fowlerville Rd. Delta on right built into glacial Lake Scottsville.
36.9	**0.4**	Moraine (hill) on left is site of mastodon find in 1989-90.

STOP 4. INFORMAL STOP ALONG ROUTE 39 OVERLOOKING VALLEY

A broad moraine complex arcs westward into valley (road follows crest for ~3 miles) marking late Wisconsin ice readvance. The moraine is undated but is thought to be approximately correlative with Alden, Buffalo, and Niagara Falls moraines (Muller and others, 1988). The moraine is difficult to see from any given point but fills the Genesee Valley between Geneseo and Avon (1:24,000 topo.), eliminating the broad flood plain characteristic of the Genesee Valley to the north and south of this reach.

40.1	**3.2**	Roots Tavern Rd. on right marks end of obvious moraine ridges on topographic map.
43.6	**3.5**	Enter Geneseo, turn right on South St. at Courthouse.
44.2	**0.6**	West to Route 63 (down hill).
52.3	**8.1**	Route 63 northwest to Peoria and Alden moraine.

STOP 5. VIEW OF ALDEN MORAINE NEAR ICE POSITION THAT CREATED LAKE HALL (~1000 ft elev.)

Genesee Valley south of this latitude was filled by Lake Hall as the ice retreated from the Alden moraine. The drainage outlet was to the west through the prominent Pearl Creek outlet where a large delta was built into the Wyoming Valley. The next leg of the trip will descend into the Pearl Creek outlet channel and climb out to the south, descending to the shoreline of glacial Lake Warren near elevation 845 ft.

53.2	**0.9**	Peoria Rd. south to Old State Rd.
54.2	**1.0**	Right on Old State Rd.
55.2	**1.0**	Left on Moag Rd. (south).
55.8	**0.6**	Left on Morrow.
56.4	**0.6**	Right on Simmons Rd. (south).
58.0	**1.6**	Left on Kendall-Barber Rd.
58.8	**0.8**	Right on Covington Rd. (SE)
60.0	**1.2**	Left on New Road to STOP 6

STOP 6. STOP AT LAKE WARREN BEACH WITH VIEW EAST INTO GENESEE VALLEY

Glacial Lake Warren was the last large glacial lake embayment extending southward into the Genesee Valley from an ice front position near the N Y State Thruway between 13,000 to 12,600 BP (Muller and Calkin, 1993). The shoreline presently indicates a relative postglacial uplift to the north of approximately two feet per mile between Geneseo and Victor, New York. Lake Warren shoreline features can clearly be seen along both sides of the Genesee Valley on aerial photographs southward to the latitude of Letchworth Park, whereas no other glacial Lake shorelines are still as obvious or continuous.

64.4	**4.4**	Descend eastward into valley to Cuylerville (right on Rt 36, left on Rt 39 & 20A).

END OF FORMAL TRIP

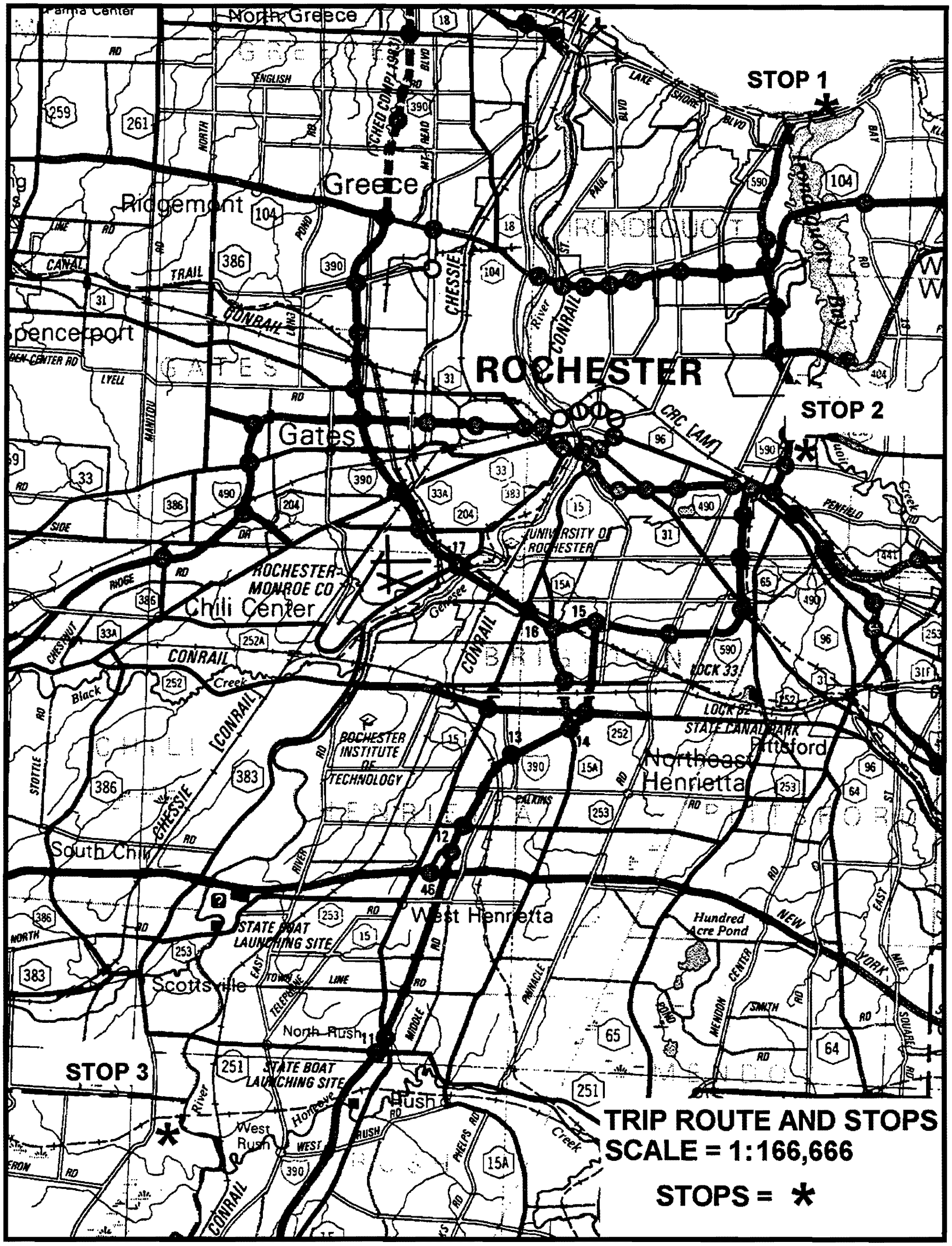
At this point in the trip, depending on the hour, the time spent at the Middle Wisconsin site (STOP 3), and the ongoing developments at the Akzo-Nobel salt mine collapse at Cuylerville (March 1994 to present), the trip will either focus on an informal tour of the salt mine collapse area (with maps and cross sections of the mine), or continue to the Valley Heads Moraine near Dansville.

Access to the Akzo-Nobel collapse areas is restricted and more subsidence is anticipated in the near future. There may be little to see of the areas actually subject to the most severe collapse due to their location in wooded areas, due to tight security, and due to restricted road access. An alternative possibility is a stop at the Geological Sciences Department at SUNY Geneseo to view the maps, slides, and aerial video footage assimilated since the beginning of the collapse problem in March 1994.

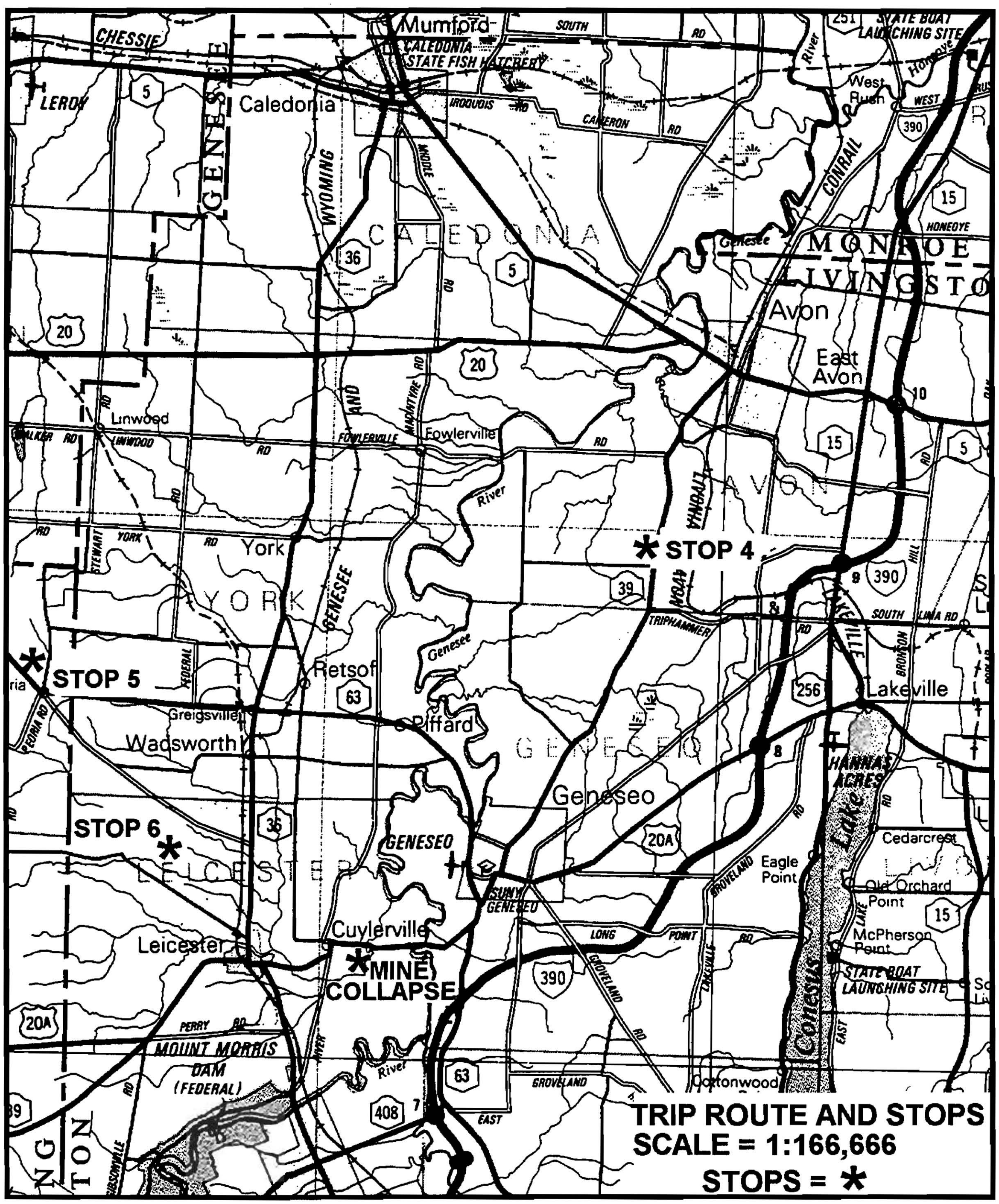
The Valley Heads Moraine complex at the south end of the Canaseraga Valley is a unique, multi-lobate feature, which extended south and west from Dansville in a series of semicircular arcs (not unlike some piedmont lobes in plan view). The lobate extensions of the moraine actually turned northward up a small valley east of Dansville, creating a unique, north-facing moraine with an impounded lake on its northern edge. The unique morainal ridge morphology and associated spillway features make this moraine complex a worthwhile study of the local influence of topography on ice flow. The moraine segments can be readily viewed from roads or regional overlooks, and can provide an alternative ending to this general overview of Genesee Valley glacial geomorphology and stratigraphy if the Middle Wisconsin gravel pit is largely mined out by October 1994, or if the ongoing events at the Akzo-Nobel mine prevent productive viewing of this serious environmental problem. The trip from Cuylerville to Dansville to view the moraine is about 15 miles on I-390. Alternatively, if the hour is late, the formal trip will return to Rochester via I-390, a distance of about 30 miles.

REFERENCES CITED

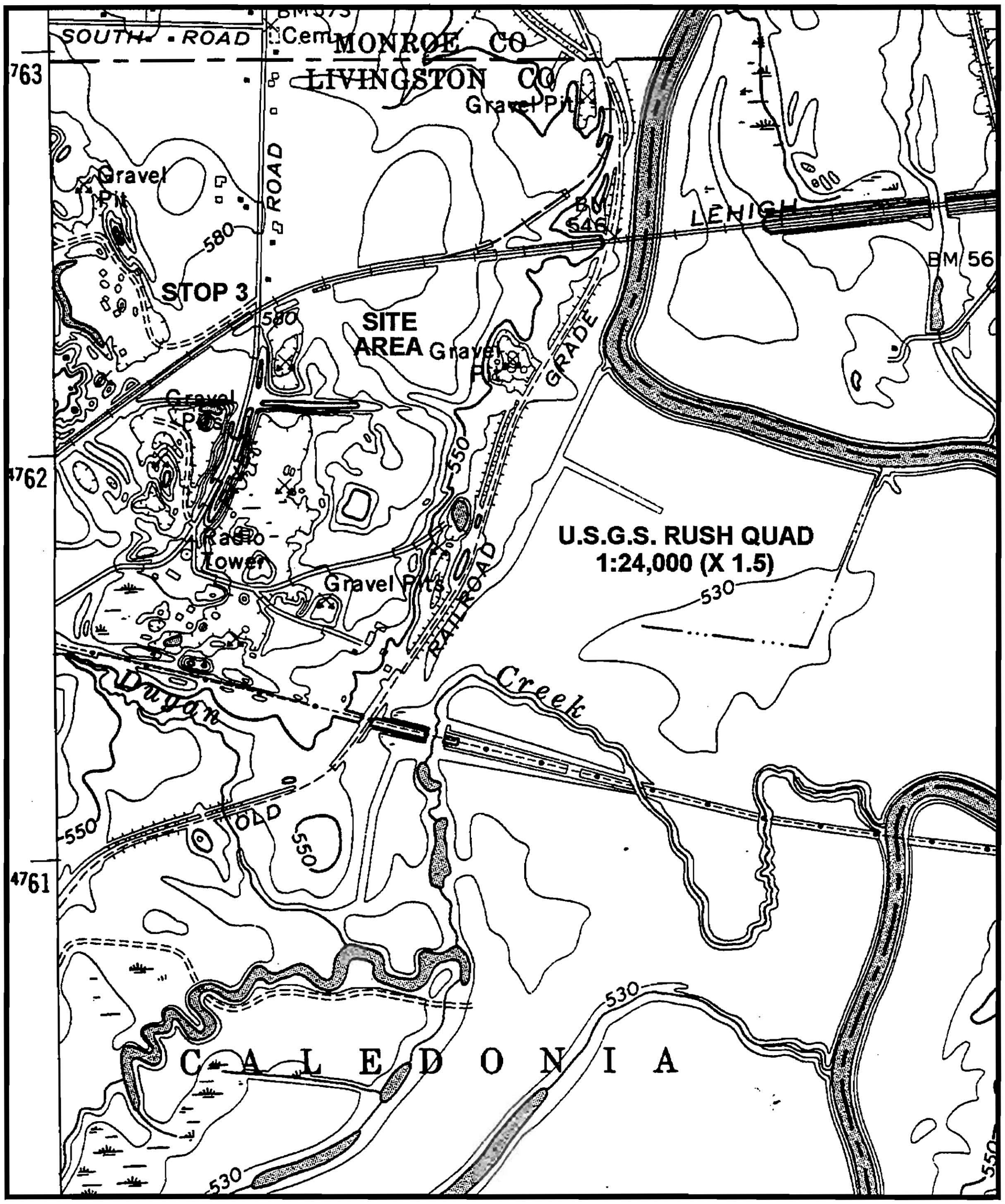
(See accompanying article)



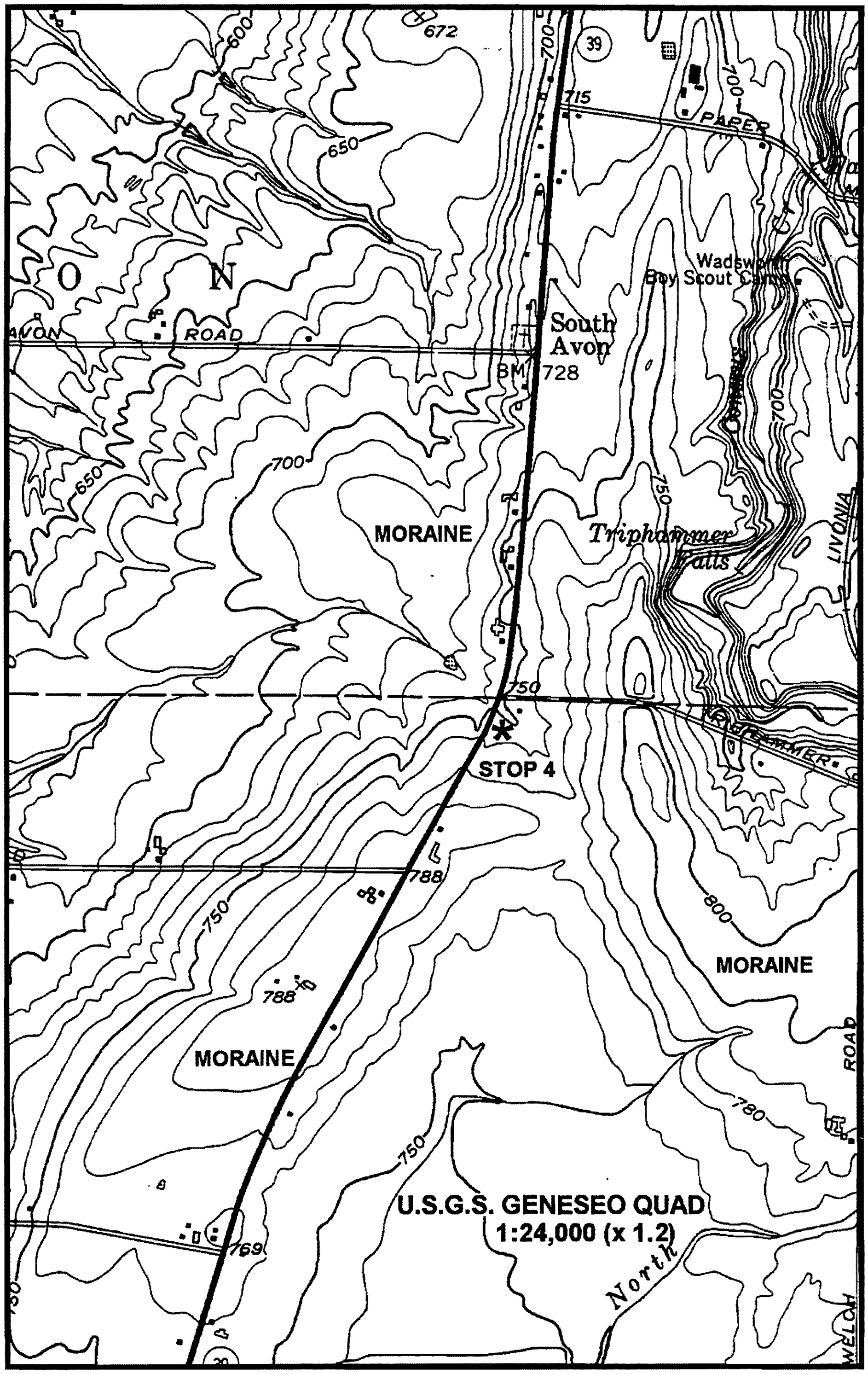
Location map of trip route and Stops 1-3. From N.Y.S.D.O.T. State Atlas.



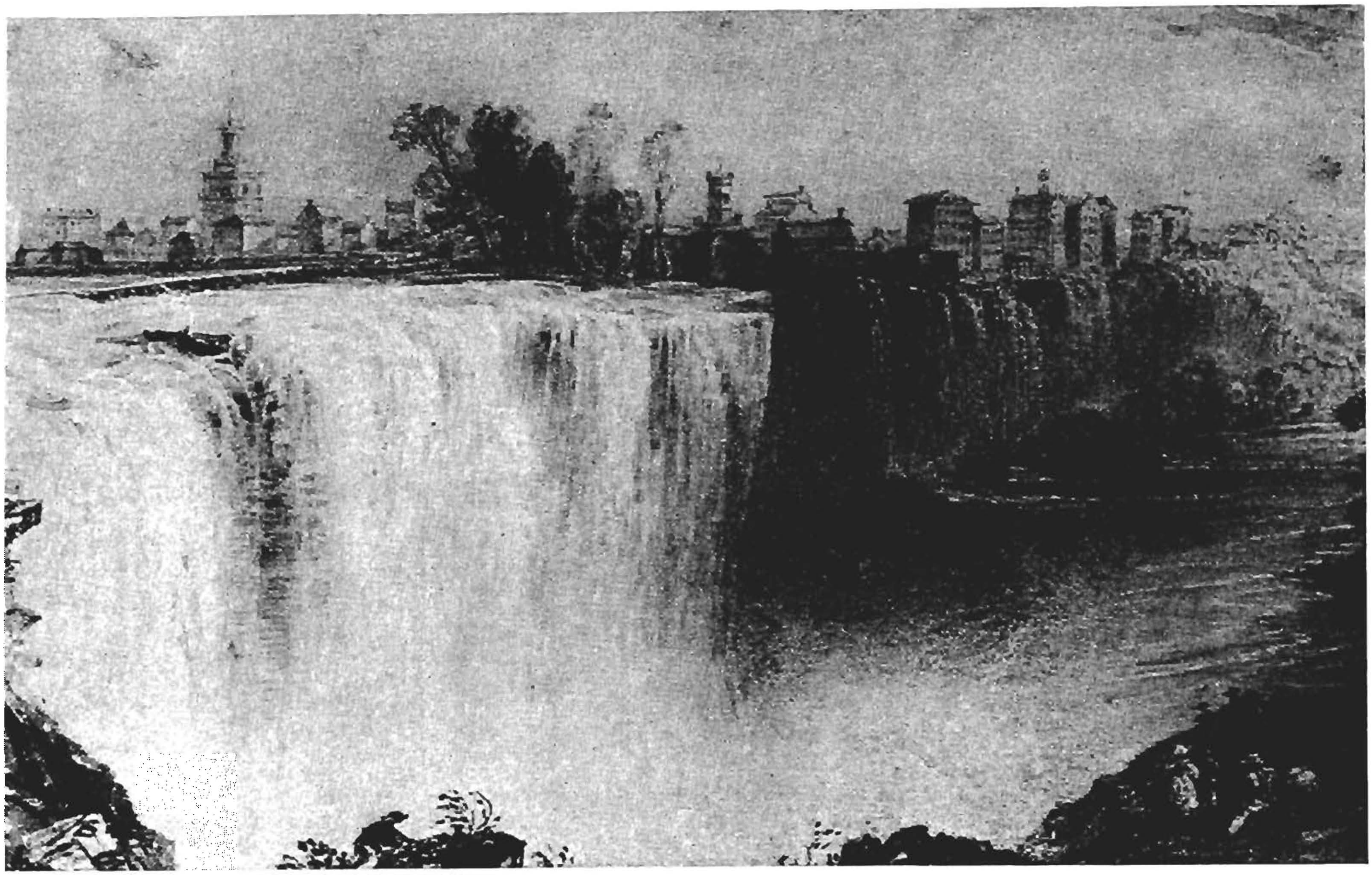
Location map of trip route and Stops 4-6. From N.Y.S.D.O.T. State Atlas.



Area of Stop 3. Gravel pit is much expanded from topography shown on this old map edition. Contour interval is 10 feet. Lehigh Valley Railroad has been removed.



Area of broad morainal topography near Stop 4. Contour interval is 10 feet.



Genesee Falls, Rochester, New York. ca. 1836
Painting by William H. Bartlett;
[From *The Course of Empire: The Erie Canal and the New York Landscape*, 1984, Memorial Art Gallery of the University of Rochester, Rochester, New York, p. 18]