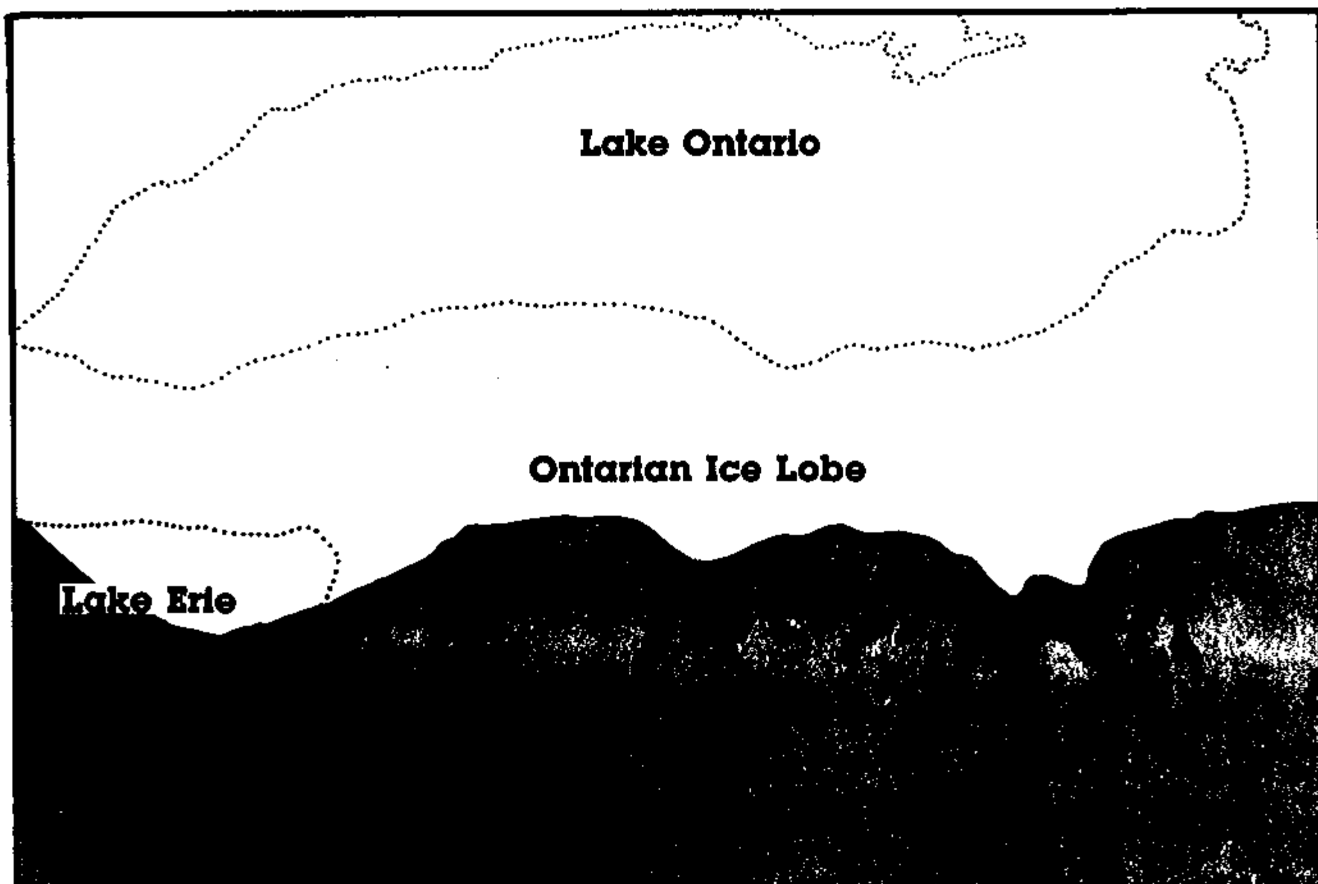
# 51st Annual Meeting
# Friends of the PLEISTOCENE

## Late Wisconsinan Deglaciation of the Genesee Valley
## May 27-29, 1988




Department of Geological Sciences
State University College
Geneseo, New York

**GUIDEBOOK**

**51st Annual Meeting**

**FRIENDS OF THE PLEISTOCENE**

# LATE WISCONSINAN DEGLACIATION OF THE GENESEE VALLEY

**State University College**
**Geneseo, New York**

**May 27-29, 1988**


Editor: William J. Brennan

Contributors: Duane D. Braun
William J. Brennan
Ernest H. Muller
Richard A. Young



Guidebook distributed by:

William J. Brennan
Department of Geological Sciences
State University College
Geneseo, New York 14454

Price: $15 (if available)

## TABLE OF CONTENTS

# INTRODUCTION AND OVERVIEW

**Ernest H. Muller**

## Welcome to Genesee country!

The Iroquois word which means "beautiful valley" is fittingly applied to the basin occupied by the only river that flows all the way across New York State. In so doing, the Genesee River drains a basin of 6425 sq. km., of which 97% is in New York (Fig. 1). From its headwaters in Pennsylvania, the Genesee descends 2000 feet in 163 miles to enter Lake Ontario near Rochester at Ontario Beach.

## Field Trip Plan

We have planned this itinerary with a view both to developing a framework of regional relationships and to providing opportunity for field discussion of evidence bearing on the Late Wisconsinan and Holocene history of the Genesee Valley from Belmont to Lake Ontario (Fig. 2).

## Proposed field program

Saturday, We travel south from Geneseo to Mount Morris across an actively evolving portion of the Genesee floodplain developed on the floor of former "Lake Geneseo". At Mount Morris we enter Letchworth Park, with stops to consider the character and origins of the steeply incised middle Genesee, including the Mount Morris and Portageville High Banks with the St. Helena Reach between them.

Before noon we enter the upper Genesee Valley at Portageville, crossing the river at Whiskey Bridge to examine proglacial sedimentary facies relationships of the Valley Heads Moraines and recent changes in the Genesee River channel. We will climb the slumping valley wall, with lunch at an overlook midway up, to examine the stratigraphic section below the Valley Heads Moraine. We continue south by bus, stopping to examine the relationships of till and lake sediments that predated the Valley Heads Moraine. From Black Creek near Belfast we return by way of the Canisteo Valley.

Sunday. We drive north, along the open valley of the lower Genesee with stops to consider meltwater channels, moraines and proglacial lake features developed during post-Valley Heads glacial recession. After lunch northeast of Rochester near the shores of Irondequoit Bay, several stops deal with evidence of the valley-filling drift sequence, the Irondequoit aquifer and Holocene history of the embayment. We plan to be back in Geneseo before 4:30 P.M.

Two days is a short time in which to hammer out a friendly consensus on problems in an area as large and complex as the Genesee Valley. To accomplish our objectives in the short time at hand will require full cooperation of all participants in assembling promptly at the end of each stop.

## Introduction to the Field Area

The Genesee River carries runoff from two structurally related but topographically contrasting physiographic units -- the Ontario Lake Plain of

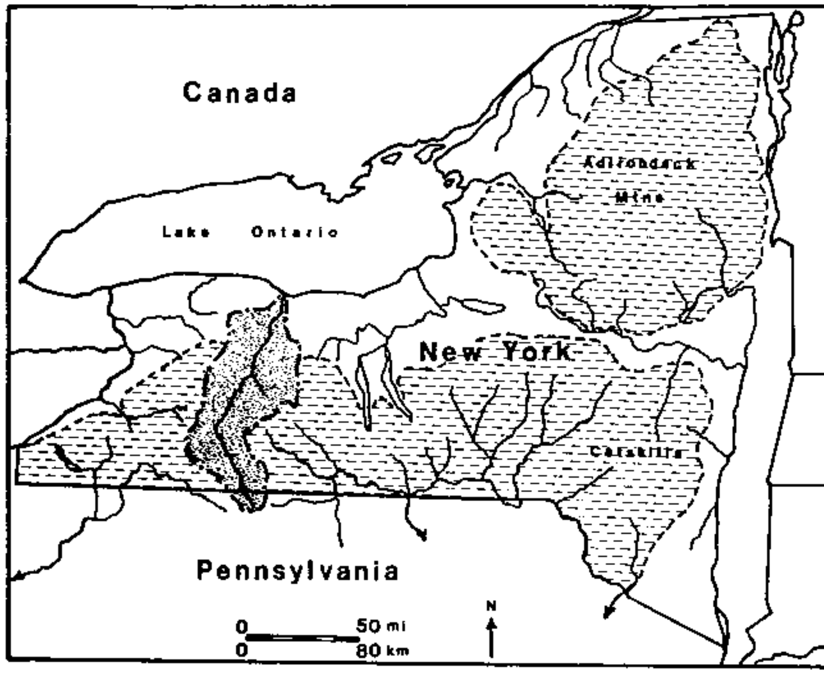



Figure 1. New York State, showing uplands by dashed pattern and the Genesee Valley by lined pattern.

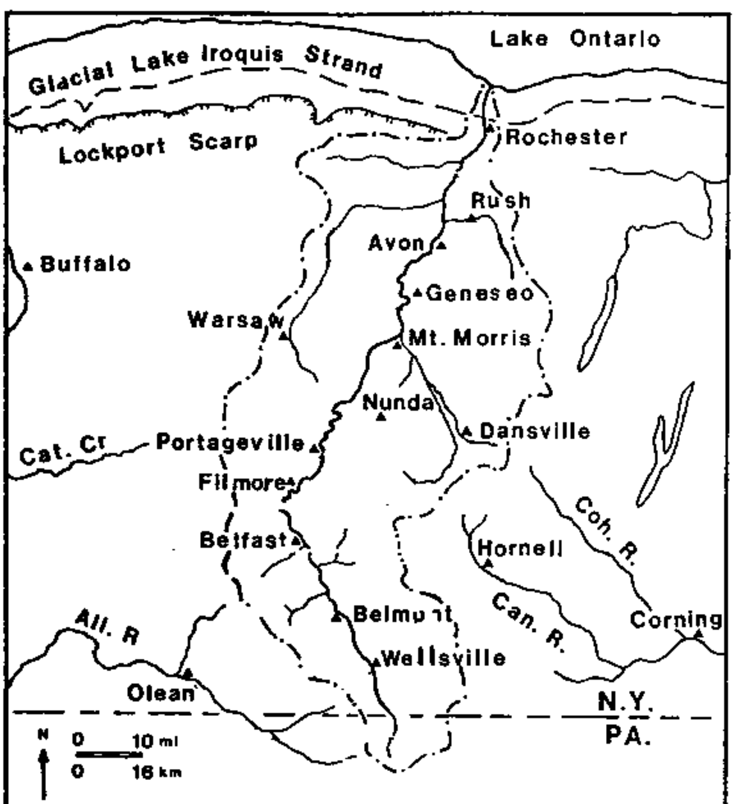



Figure 2. Central New York showing the Genesee Basin and locations mentioned in the text.

the Interior Lowlands and the Glaciated Allegheny Plateau subprovince of the Appalachian Plateau (Fig. 1). Both areas are underlain by relatively undeformed Upper Silurian and Devonian strata. Located on the southeast limb of the Appalachian geosyncline, the strata are progressively younger southward as a function of regional dip at a fraction of a degree.

The lake plain is a drift-mantled, extensively drumlinized belt of low relief. The plateau is a heavily dissected upland with thick valley fill and moderate relief. Summit accordance is increasingly well preserved southward, indicating decreasing glacial reduction of the upland. Physiographic distinctions between the two subprovinces relate primarily to differences in rock resistance and to the greater duration and intensity of glacial erosion in the lake plain as compared to the plateau.

Far from being a completely integrated drainage basin, the Genesee Valley involves several physiographically distinctive components. The sharply incised middle Genesee gorges are in marked contrast to the more open upper and lower reaches. The upper Genesee extends for some 60 miles from the southern headwaters north to Portageville. The river rises in Potter County, Pennsylvania on plateau remnants with summits about 2500 feet above sea level. Cols 250 to 450 ft. below this summit accordance separate the Genesee headwaters from tributaries of the Susquehanna and Allegheny Rivers. From this drainage divide, the East, Middle and Western Branches of the Genesee River flow 20-25 miles north in sharply defined, youthful valleys, v-shaped in transverse profile. North of the confluence of these branches near the State Line, kame and stream terrace remnants flank a gradually widening valley. Stratigraphic relationships south of the Valley Heads Moraine between Portageville and Belmont will be our focus on the afternoon of the first day.

The middle Genesee extends from Portageville through Letchworth State Park to Mount Morris, a distance of almost 15 miles. The Genesee River flows in a confined and deeply incised canyon that contrasts with the open valley reaches both up- and downstream. At both south and north ends of the park, cliffed sandstone walls enclose a meandering valley floor barely wider than the stream channel. The Portageville High Banks at the south end of the park include a reach that has been proudly termed the "Grand Canyon of the East". The Mount Morris High Banks at the north end of the park contain the Mount Morris flood control dam and reservoir. In the St. Helena reach between these two high-bank areas, the Genesee is mildly sinuous on a valley floor a half-mile wide. The origins of Letchworth landscapes will be considered Saturday morning.

The lower Genesee begins at Mount Morris where the river enters the broad glacial trough drained by Canaseraga Creek. With actively developing meanders, the river swings freely across the flat floor of postglacial "Lake Geneseo", thence back northwest to cut through the "Fowlerville Plug" of glacial and fluvioglacial drift. North past Avon and Rush, the Genesee again meanders freely on the broad, flat bed of proglacial lakes Avon and Scottsville. Once a granary of the Northeast, this is still an area of productive farms and handsome country estates. We will travel through this portion of the valley Sunday morning en route to Irondequoit Bay. In Rochester, the flood plain again narrows where the Genesee cuts through the Pinnacle Hills Moraine. A short distance to the north, the river enters the postglacial Rochester Gorge cut across the Niagara Cuesta. North of Rochester, the valley floor, little wider than the stream channel, is incised 100 to 150 feet below the floor of glacial Lake Iroquois.

Several miles east of the Rochester Gorge is the broad, drift-filled partly inundated valley of tiny Irondequoit Creek. The contrast between the youthful Rochester Gorge and the overfit Irondequoit Valley were early recognized as evidence of drainage derangement. Indeed, the importance of the Irondequoit aquifer has stimulated studies of this drift-filled reach of the Ancestral Genesee River which will receive attention on the afternoon of the second day of this field conference.

**Previous Work**

Hypotheses relative to pre-glacial development of the Genesee drainage basin were early discussed by Hall (1943), Grabau (1894; 1908), Whitbeck (1902), Fairchild (1908; 1925; 1928) and von Engeln (1961).

Regional relationships were described or mapped in reconnaissance scale by Leverett (1902), Fairchild (1932), MacClintock and Apfel (1944), Muller (1975) and Cadwell (in press). The framework of proglacial lake history in the Genesee Valley was well established by Fairchild (1896; 1908; 1926; 1928) supplemented by Chadwick (1917; 1923), Chadwick and Dunbar (1924) and recently by Muller and Prest (1986) and Muller and others (in press).

Aspects of Genesee Valley geology have been treated in detail in guidebooks (Univ. of Rochester Geology Department, 1956; Young and Kreidler, 1957; Hewitt, 1973) for New York State Geological Association fieldtrips (available from the Executive Secretary, M.P.Wolff, Geology Department, Hofstra University, Hempstead, NY, 11550) and Liebe (1981) for the National Association of Geology Teachers (available from Secretary-Treasurer William Brice, Dept. of Geology and Planetary Science, Univ. of Pittsburgh-Johnstown, Johnstown. PA, 15904). Theses and dissertations by Street (1963), Connally (1964), Terlecky (1969; 1974) and Grossman (1973) deal with Quaternary geology of areas in the Genesee Valley.

Sediment transport patterns of the Genesee were studied by the U.S.Geological Survey (Mansue and others, 1981). The drift-filled course of the preglacial Genesee in Monroe County received particular attention as a groundwater reservoir (Waller and others, 1982).

The age of the Wisconsinan terminal moraine in the Genesee headwaters has been recently debated (Sevon and others, 1975; Crowl and Sevon, 1980; Cotter, 1983; Cotter and others 1984, 1985).

**Where do we stand in 1988?**

Studies that have led directly to our extending an invitation to the Friends of the Pleistocene to meet in the Genesee Valley began in 1975 under sponsorship of the New York State Geological Survey as part of an International Joint Commission investigation of the impact of land use on water quality of the Great Lakes (Muller and others 1981). In addition to the IJC investigation, Young has participated in subsequent engineering, environmental and geohydrologic studies in Monroe and Livingston Counties. The detailed stratigraphy of the upper Genesee has been mapped by Braun (Braun and others, 1984, 1985). Brennan has applied geophysical methods to problems of drift stratigraphy (Brennan and others, 1984).

Problems of interregional correlation rest far more heavily than we would wish on the tracing of moraines and shorelines, and on intepretation of their relationships in the Erie, Genesee, Finger Lakes and Mohawk regions. Materials suitable for radiocarbon dating are rare. Efforts toward paleomagnetic correlation give promise, yet to be realized, of an objective and independent basis for testing the existing chronologic framework.

Although we have been brash enough to publish our current views on the morphogenic history of the Genesee Valley (Muller and others, in press), we are under no illusion that the final word has been written. We may state our preferred interpretations with conviction, but few facts are firmly established, as yet.

Problems of long distance correlation are knotty and the clues are tantalizingly ambiguous. We trust that you, too, will find them challenging. If so, clearer interpretation of relationships and a better understanding of the problems involved will develop from this, the 51st Annual Meeting of the Friends of the Pleistocene.

**End Moraines**

In the absence of a firmly established basis for dating, the Laurentide Ice Sheet is presently considered to have built its terminal moraine (Fig. 3) at the southern headwaters of the Genesee Basin about 20,000 years ago. A single piece of wood collected from till exposed in Rush Creek, east of Fillmore, yielded a date of 25,450 (+6657, -3600) (QC-238). This is strong evidence that the Upper Genesee, at least, was free of ice immediately before it was overrun in Late Wisconsinan time.

Kame deposits marking recessional ice border positions during the Nissouri Stadial are discontinuously distributed along upper Genesee valley margins and in tributary valleys. Two such retreatal positions of the waning ice tongue are mapped south of Shongo near the State Line, and at Stannards (Stage Bs, Fig. 3), 4 miles south of Wellsville. Ice marginal positions are similarly indicated in the right-bank tributary valleys of Scio and Phillips Creeks (Stage Bp, Fig. 3). Oscillatory ice retreat is indicated, but no marked readvance is recognized as having interrupted general withdrawal from the Upper Genesee at least as far north as Angelica.

Previous investigators have construed features in the vicinity of Angelica as evidence of glacial readvance (Stage C, Fig. 3). Low arcuate mounds of clay till constrict the Genesee Valley southwest of Angelica and the Black Creek Valley southwest of Belfast. A morainal ridge all but closes the valley of Angelica Creek at its mouth and kame delta gravels choke its right-bank tributary. MacClintock and Apfel (1944) mapped this as Binghamton Moraine on the basis of content of exotic clasts and depth of leaching. Muller (1975) related this position to the Kent Moraine in northwestern Pennsylvania. Morphostratigraphic tracing and long distance correlation remain inconclusive, but recent mapping (Braun and others, 1984) supports the view that these features, now termed the Angelica Moraine, mark a readvance of at least 20 miles. On the basis of limited radiocarbon data, this advance postdates a lake

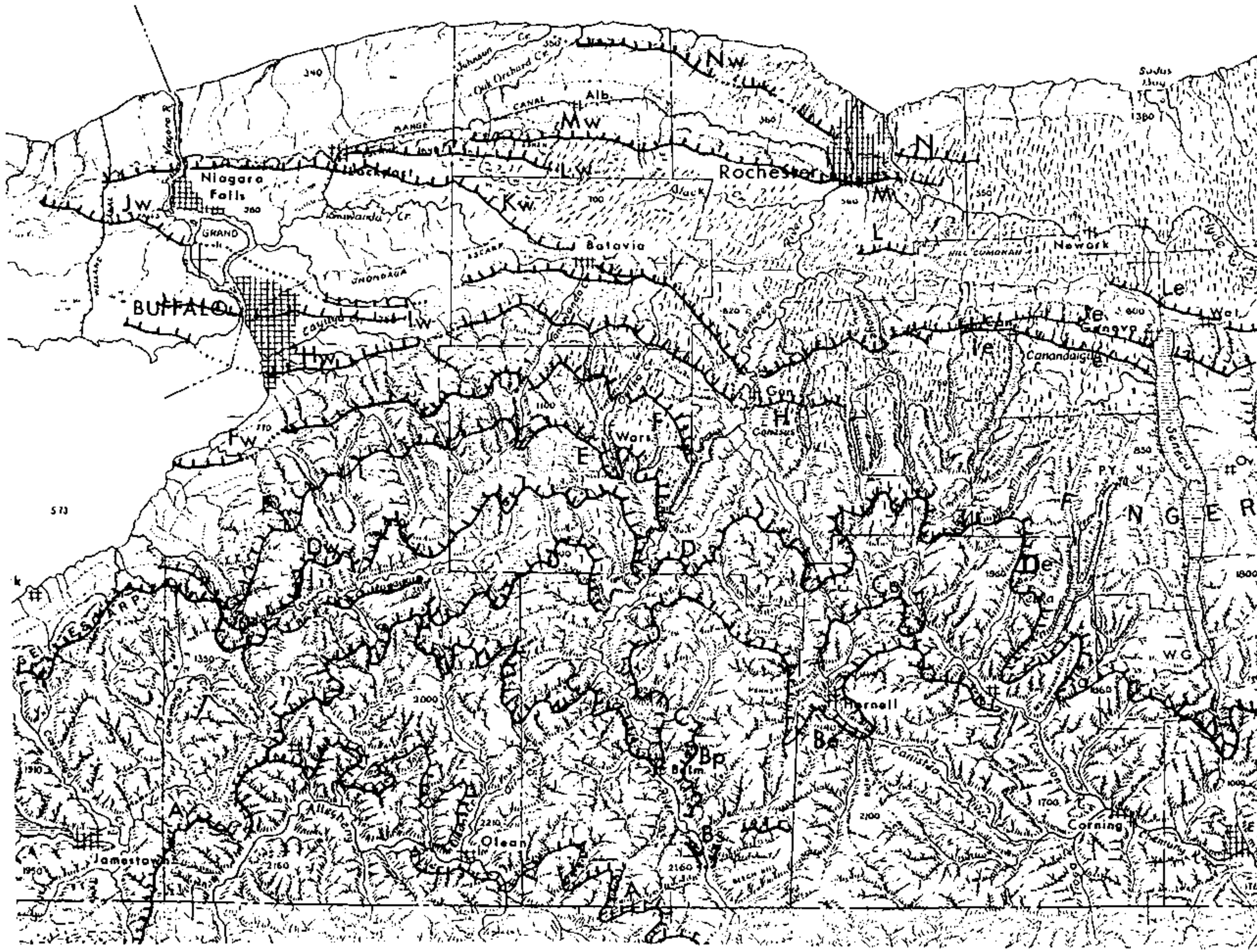



Figure 3. Moraines and glacier marginal positions in the Genesee Valley. See Table I for identification and tentative correlation.

TABLE I. TENTATIVE CORRELATION OF MORAINES

| Western New York | Genesee Valley | Finger Lakes |
|---|---|---|
| Carlton Moraine (Nw) in Lake Iroquois | Carlton Moraine (N) in Lake Iroquois | Oswego Moraine in Lake Iroquois |
| Albion Moraine (Mw) | Pinnacle Hills Mor. (M) dammed Lake Dawson | ? |
| Barre Moraine (Lw) built in Lake Avon | Mendon Kames (L) dammed Lake Avon | Junius Kames/Waterloo Moraine (Le) dammed Lake Warren III |
| Batavia Moraine (Kw) built in Lake Warren III | Hopper Hills (K) kame delta into Warren III | |
| Niagara Falls Mor. (Jw), dammed Lake Warren I | "Lima Moraine" (J) dammed Lake Hall and falling waters | Geneva/Union Springs Moraine (Je) dammed Warren I |
| Buffalo Moraine (Iw) dammed Lake Warren I | Canandaigua Moraine (I) ("Fowlerville plug") dammed Lake Hall | Mapleton Moraine (Ie) dammed Lake Newberry |
| Alden Moraine (Hw) dammed Lake Warren I | "Geneseo Moraine" (H) dammed Lake Hall | "Long Point Moraine" dammed Lake Newberry |
| Marilla Moraine (Gw) dammed Lake Whittlesey | "LaGrange Moraine" (G) dammed Lake Dansville | Not identified. probably dammed Lake Newberry |
| Hamburg Moraine (Fw) dammed Lake Whittlesey | "Perry Moraine" (F) dammed Lake Dansville | Not identified. probably dammed Lake Newberry |
| Gowanda Moraine (Ew) fronted on Lake Maumee | "Castile Moraine" (E) south of Silver Lake dammed Lake Dansville | Not identified. Dammed Lakes Ithaca and Watkins |
| Lake Escarpment Moraines (Dw), in Lake Maumee | Valley Heads Moraines (D), dammed Lake Belfast-Fillmore | Valley Heads Moraines (De) comprise the drainage divide. |
| | "Kent Moraine" (C) built in Lake Belfast-Fillmore | Arkport Moraine (Ce) |
| | Angelica Moraine (B) reinstated Lake Wellsville (?) | Almond Moraine (Be) |
| Wisconsinan Terminal Moraine (A) | | |

stage that began about 16,000 years ago, and may be tentatively correlated with the Erie Interstadial in Ontario.

The Valley Heads Moraine (Stage D, Fig. 3), built during the Port Bruce Stadial, is a complex of morainal ridges and associated stratified drift. Elsewhere, this moraine comprises the divide separating the St. Lawrence drainage from that of the Allegheny and Susquehanna Rivers. For several miles south of Portageville, the Genesee River breaches multiple ridges of this moraine complex. Displaced from its ancestral channel, the Genesee incised the canyon that begins at Portageville. North of Portageville, it cuts through a fifth moraine tentatively correlated with the Gowanda Moraine (Stage E, Fig. 3) in the Erie Basin.

The extent of recession during the Mackinaw Interstadial of western Ontario is not independently documented in the Genesee Valley. On the basis of morphostratigraphic tracing of uncertain reliability, moraines which trend northwest from the St. Helena Reach are tentatively correlated, with the Hamburg and Marilla Moraines (Stages F and G, Fig. 3) of the Erie Basin. The Hamburg and Marilla Moraines were built about 13,000 years ago during the Port Huron readvance which impounded glacial Lake Whittlesey in the Erie Basin.

A subdued morainal ridge extending eastward from Geneseo is matched by a similar ridge that trends northwest from the Genesee Valley through Peoria. This moraine, hereafter referred to as the Geneseo Moraine (Stage H, Fig. 3), is tentatively correlated with the Alden Moraine in the Erie Basin and therefore postdates the early high stage of Lake Warren. The Canandaigua and Geneva Moraines of the western Finger Lakes, enter the Genesee Basin north of Conesus Lake and cross the Genesee Valley as the "Fowlerville plug" (Stages I and J, Fig. 3) correlated with the Niagara Falls and Buffalo Moraines in the Erie Basin. Like the Geneseo Moraine, they record recessional episodes during the Port Huron Interstadial.

Southeast of Oakfield, the Batavia Moraine (Stage K, Fig. 3) is a distinct morainal ridge, where it rises onto the Onondaga Scarp at Deep Pond. In thin drift on the Onondaga Bench the moraine is weakly developed and its eastward continuation is lost. The Waterloo Moraine (Stage Le, Fig. 3), north of the Finger Lakes, is in similar sequential position, but cannot be traced westward to the Genesee Valley. Several authors (Fairchild, 1909; Shumaker, 1957, Muller 1983; Fullerton (1980) correlated the Batavia and the Waterloo Moraines. The 45-mile gap between mapped portions of the Batavia and Waterloo Moraines is all the more frustrating because, abutting against the Onondaga Scarp, both played critical roles in ponding proglacial meltwater. Considerations presented on later pages suggest correlation of the Waterloo Moraine with one of the Barre Moraines (Stage L, Fig. 3), rather than the Batavia Moraine.

The Pinnacle Hills (Stage M, Fig. 3) in the southern outskirts of Rochester form a distinct ridge rising above the level of proglacial Lake Dawson into which they were built. They correspond to the Albion Moraine (Stage Mw, Fig. 3) which slopes westward obliquely across the Lockport Scarp. Indirect evidence indicates another ice marginal position north of Rochester, represented by the Carlton Moraine (Stage N, Fig. 3).

**Proglacial Lake Sequence**

Because the Genesee drains northward, a continuous sequence of lakes was impounded at the ice margin from the beginning of its recession until it withdrew from New York State. Successively lower outlets, uncovered by the retreating ice, controlled the levels of these lakes. The relationships of ice marginal positions to lake levels afford primary morphostratigraphic criteria for long distance correlation in western New York.

The proglacial lake sequence began with the impounding of primary lakes (# 1, Fig. 4) in the Eastern, Middle and Western Branches of the Genesee River in Pennsylvania. As the ice sheet receded across the State Line, these lakes coalesced to form glacial Pennsylvania Lake (# 2, Fig. 4). Several short-lived lake stages followed while outflow gained access westward by the canyon of Honeoye Creek along the State Line to the Allegheny River (# 3, Fig. 4) In time, the Stone Dam Col at the head of Honeoye Canyon became the control for glacial Lake Wellsville (# 4, Fig. 4). Glacial Lake Belfast-Fillmore (# 6, Fig. 4) was initiated by uncovering of the Cuba Outlet, west via Oil Creek to the Allegheny River. Glacial readvance restored Lake Wellsville which again gave way to Lake Belfast-Fillmore after glacial retreat from the Angelica Moraine.

Recession from the Valley Heads Moraine opened outlets to the Canisteo River in the Susquehanna drainage basin. First among the southeastward draining lake stages was glacial Lake Nunda, (# 6, Fig. 4) controlled by the Swains Outlet to Canaseraga Creek. Subsequently expanding into Canaseraga Trough, glacial Lake Dansville (# 8, Fig. 4) drained via the Burns Outlet to the Canisteo River. Meanwhile, northward outflow from moraine-dammed Lake Portageville (# 7, Fig. 4) began to incise the Letchworth canyons.

Glacier recession to the Geneseo Moraine uncovered the Pearl Creek Channel. With threshold at 1000 ft., this channel controlled westward-draining Lake Hall (# 9, Fig. 4). At first, sediment at the mouth of the channel was deposited in a shallow lake in Oatka Trough controlled by shifting and ephemeral channels across stagnant and decaying ice ultimately to reach early Lake Warren. The Alden and Geneseo Moraines were built in glacial Lakes Warren and Hall, respectively. Both were modified by shore processes at a later, lower stage of Lake Warren which spread from the Erie Basin into the Genesee watershed (# 10, Fig. 4). The Pearl Creek outlet was abandoned as westward drainage developed across stagnant, debris-laden ice, near Lynwood, 7 miles southwest of LeRoy.

Withdrawal of the ice margin from the Onondaga Scarp initiated a fast-moving sequence of events as outflow from the Erie-Huron Basin shifted rapidly from the Mississippi to the Mohawk drainage system in an act of avulsion that has seldom been matched. Drift cover was swept from bedding surfaces northwest and northeast of LeRoy as the outflow shifted obliquely across the Onondaga Bench. The Taylor Channel may have provided the first stable outlet control, building a delta at 700 feet into the lake impounded in the Genesee Valley. Anastamotic channels east of Caledonia record catastrophic outflow into glacial Lake Avon (# 12, Fig. 4) in the Genesee Valley which in turn was controlled by the Rush-Victor channels to the east.

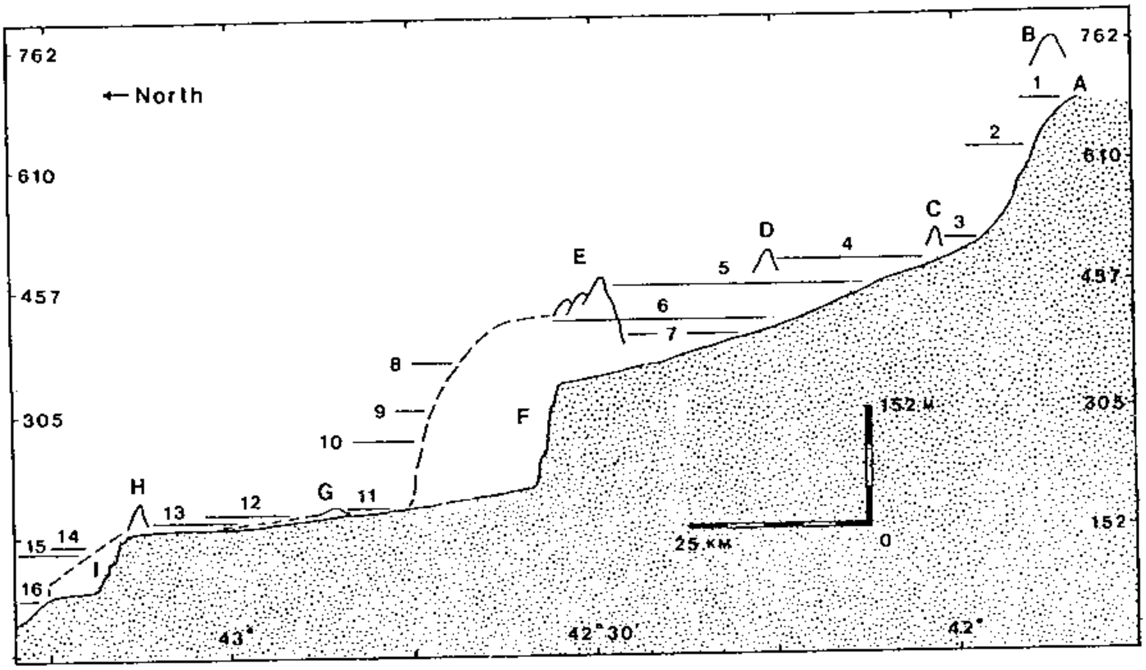



Figure 4. Genesee proglacial lake succession as related to the present long profile of the Genesee River from its southern watershed (right) to Lake Ontario (left). Lake stages identified below.

| | Name of lake | Elevation | Outlet and drainage basin |
|---|---|---|---|
| 1. | Primary lakes | >2100 ft. | Allegheny and Susquehanna |
| 2. | Lake Pennsylvania | 2060 ft. | Rose Lake Col to Allegheny |
| 3. | Lake Shongo | 1680 ft. | Stone Dam Col to Allegheny |
| 4. | Lake Wellsville | 1580 ft. | Stone Dam Col to Allegheny |
| 5. | Lake Belfast-Fillmore | 1500 ft. | Cuba Outlet to Allegheny |
| 6. | Lake Nunda | 1320 ft. | Swains Outlet to Susquehanna |
| 7. | Lake Portageville* | <1320 ft. | N. thru Valley Heads Moraine |
| 8. | Lake Dansville | 1220 ft. | Burns Outlet to Susquehanna |
| 9. | Lake Hall | 1000 ft. | Pearl Cr. Outlet to Erie Basin |
| 10. | Lake Warren | 900 ft. | Grand R. to Michigan Basin |
| 11. | Lake Geneseo* | <650 ft. | N. across "Fowlerville Plug" |
| 12. | Lake Avon | 700 to 580 ft. | Rush channels to Mohawk |
| 13. | Lake Scottsville* | <540 ft. | N. across Pinnacle Hills Moraine |
| 14. | Lake Dawson | 462 ft. | Fairport Channels to Mohawk |
| 15. | Lake Iroquois | 425 ft. | Rome outlet to Mohawk |
| 16. | Lake Ontario | 246 ft. | St. Lawrence |

* Lakes indicated by an asterisk were moraine-dammed and persisted after withdrawal of the ice sheet.

A. Genesee-Allegheny watershed.
B. Late Wisconsinan Terminal Mor.
C. Ice margin at Stannards.
D. Angelica Moraine
E. Valley Heads Moraine
F. Letchworth Gorge
G. Fowlerville morainal plug
H. Pinnacle Hills Moraine
I. Rochester Gorge

During subsequent glacial readvance, the ice margin may have again mounted the Onondaga scarp at the north edge of the plateau on its salients north of Batavia and west of Syracuse. Apparently the ice dam yielded earlier in the west than near Syracuse, for shore features of the late, low stage of Lake Warren (Warren III) (# 10, Fig. 4) are traceable into the Genesee Valley.

Following abandonment of the Syracuse Channels, the ponded water in the Genesee Valley was again isolated from the Finger Lakes area. Lake Avon ended with the opening of lower eastward outlets near Fairport. A moraine-dammed remnant, here named Lake Geneseo, (# 11, Fig. 4) probably persisted for some time south of the "Fowlerville plug". Drainage control by the Fairport Channels defines glacial Lake Dawson (# 14, Fig. 4) which was impounded north of the Pinnacle Hills Moraine. Moraine-dammed Lake Scottsville (# 13, Fig. 4) south of the Pinnacle Hills Moraine existed only until its outlet cut the present Genesee River channel northward through this barrier.

**Acknowledgement:**

Figures 1 and 4 are adapted from Muller and others, in press, and are included here with permission of Northeastern Science Foundation, publishers of "Northeastern Geology".

## REFERENCES

Braun, D.D., Helfrick, E.W., Jr., Olenick, G.F. and Brennan, W.J., 1984, Using secular variation of geomagnetic declination to test the age of the Kent Moraine in the Genesee Valley, New York: a progress report; Geological Society of America Abstract with Programs, Vol. 16, p. 5.

Braun, D.D., Daniel, T.A., Smith, L.B. and Brennan, W.J., 1985 Quaternary geologic and geomagnetic record of the upper Genesee Valley, New York; Geological Society of America Abstracts with Programs, Vol. 17., p. 7.

Brennan, W.J., Hamilton, M., Kilbury, R., Reeves, R.L., and Covert, 1984, Holocene and late Pleistocene secular variation of the horizontal component of the geomagnetic field in western New York; Earth and Planetary Science Letters, Vol. 70, p. 363-372

Cadwell, in press, Surficial Geologic Map of New York State: Niagara Sheet. New York State Geological Survey Map and Chart Series

Calkin, P.E. and B.H.Feenstra, 1985, Evolution of the Erie Basin Great Lakes; p. 149-170 in Karrow, P.F and Calkin, P.E., eds, Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30.

Chadwick, G.H., 1917, The lake deposits and evolution of the lower Irondequoit Valley; Rochester Academy of Science Proceedings, Vol. 5, p. 123-160.

Chadwick, G.H., 1923, Glacial lake problems; Geological Society of America Bulletin, Vol. 34, p. 499-506.

Chadwick, G.H. and Dunbar, E.H., 1924, Genesee glacial lakes; Geological Society of America Bulletin, Vol. 35, p. 669-676.

Connally, G.G., 1964, The Almond Moraine of the western Finger Lakes Region, New York. Ph.D. dissertation, Geology Department, Michigan State University, 102 p.

Cotter, James F.P., 1983, The minimum age of the Woodfordian deglaciation of northeastern Pennsylvania and northwestern New Jersey; Ph.D. dissertation, Lehigh University, 151p.

Cotter, J.F.P., Evenson, E.B., Sirkin, L.E., and Stuckenrath, R. 1984, The interpretation of 'Bog-bottom' radiocarbon dates in glacial chronologies;

p. 299-316 in Mahaney, W.C., ed., Correlation of Quaternary Chronologies; GeoBooks, Norwich, U.K.
Cotter, J.F.P., Ridge, J.C., Evenson, E.B., Sevon, W.D., Sirkin, L.A and Stuckenrath, Robert, 1985 The Wisconsinan history of the Great Valley, Pennsylvania and New Jersey, and the Age of the "Terminal Moraine"; p. 1-59 in Evenson, E.B., ed., 48th Friends of the Pleistocene Field Conference, Lehigh University Bethlehem, PA.
Crowl, G.H. and Sevon,W.D., 1980, Glacial border deposits of Late Wisconsinan age in northeastern Pennsylvania; Pennsylvania Topographic and Geologic Survey, General Geology Report 71, 68p.
Engeln, O.D. von, 1961, The Finger Lakes Region. Cornell University Press, Ithaca, NY, 147p.
Fairchild, H.L., 1896, Glacial Genesee Lakes; Geological Society of America Bulletin Vol. 7, p.423-452.
Fairchild, H.L. 1897, Lake Warren shorelines in western New York, and the Geneva Beach; Geological Society of America Bulletin, Vol. 8, p 269-286.
Fairchild, H.L., 1908, Pleistocene history of the Genesee Valley; New York State Museum Bulletin, Vol. 118, p. 70-84.
Fairchild, H.L., 1909, Glacial waters in central New York; New York State Museum Bulletin 127, 66p.
Fairchild, H.L., 1919, The Rochester Canyon and the Genesee River baselevels; Rochester Academy of Science Proceedings, Vol. 6, p 1-55.
Fairchild, H.L., 1923, The Pinnacle Hills or the Rochester Kame Moraine; Rochester Academy of Science Proceedings, Vol. 6, p. 141-194.
Fairchild, H.L., 1926, The Dansville Valley and drainage history of western New York; Rochester Academy of Science Proceedings Vol. 6. p. 217-242.
Fairchild, H.L., 1928, Geologic story of the Genesee Valley and western New York. Pub. by author, printed by Henderson-Mosher, Inc., Rochester, NY, 216p.
Fairchild, H.L., 1932a, New York physiography and glaciology west of the Genesee Valley; Rochester Academy of Science Proceedings, Vol. 7, p. 97-135
Fullerton, D.H., 1980, Preliminary correlation of Post-Erie interstadial events (16,000-10,000 radiocarbon years before Present), Central and Eastern Great Lakes Region, and Hudson, Champlain and St. Lawrence Lowlands, United States and Canada. U.S.Geological Survey Professional Paper 1089, 52p.
Grabau, A.W., 1894. The preglacial channel of the Genesee River; Boston Society of Natural History Proceedings Vol. 26, p.359-369. Grabau, A.W., 1908, Preglacial drainage in central-western New York; Science ns Vol. 28, p 527-534.
Grossman, W.L., 1973, Bedrock geology of the Caledonia Quadrangle. Unpub M.S. thesis, Department of Geology, University of Rochester
Hall, James. 1843, Geology of New York, 4th District. Albany,
Hewitt, P.C. (ed.) 1973, Guidebook to Field Trips, 45th Annual Meeting, New York State Geological Association, p. A1-113. Available from Exec. Secretary M.P.Wolff, Dept. of Geology, Hofstra University, Hempstead, NY 11550
Leverett, Frank, 1902, Glacial formations and drainage features of the Erie and Ohio Basins. United States Geological Survey Monograph 41, 802p.
Liebe, Richard (ed.), 1981, Field Guidebook for The Geology of the Genesee Valley Area of Western New York. National Association of Geology Teachers - Eastern Section. Available from William Brice, Dept. of Geophysics and Planetary Science, University of Pittsburgh - Johnston, Johnstown, PA 15904.

MacClintock, Paul and Apfel. E.T., 1944, Correlation of the drifts of the Salamanca Re-entrant, New York; Geological Society of America Bulletin Vol. 55, p. 1143-1164.
Mansue, L.J., Young, R.A. and Soren, J., 1981, Hydrologic influences on sediment-transport patterns in the Genesee River Basin, New York. U.S.Environmental Protection Agency, Open File Report, 33p.
Muller, E.H., 1977a, Quaternary Geology of New York, Niagara Sheet, 1:250,000. New York State Museum Map and Folio Series, No 25. Albany, NY.
Muller, E.H., 1977b, Late glacial and early postglacial environments in western New York; New York Academy of Sciences Annals Vol. 288, p. 223-233.
Muller, E.H., Young, R.A., Rhodes, D.D., Willette, P.D., Wilson, M.P. and Fakundiny, R.H., 1981, Surficial geology of the Genesee Valley. New York State Geological Survey Open File Report No. 2102.056, 174p.
Muller, E.H. and Prest, V.K., 1986, Glacial lakes in the Ontario Basin; p. 213-228 in Karrow, P.F. amd Calkin, P.E. (eds.) Quaternary evolution of the Great Lakes. Geological Association of Canada Special Paper 30.
Muller, E.H., Braun, D.D,., Young, R.A., and Wilson, M.P., in press, Morphogenesis of the Genesee Valley; Northeastern Geology.
Sevon, W.D.,Crowl, G.H. and Berg, T.M., 1975, The Late Wisconsinan drift border in northeastern Pennsylvania. Guidebook 40th Annual Field Conference of Pennsylvania Geologists, Pennsylvania Geological Survey, Harrisburg, PA, 108p.
Shumaker, R.C., 1957, Till texture variation and Pleistocene deposits of the Union Springs and Scipio Quadrangles, Cayuga County, New York, 44p.
Terlecky, P.M., 1970, The origin, stratigraphy and post-depositional history of a late Pleistocene marl deposit near Caledonia, NY. Ph.D. dissertation, Geology Department, University of Rochester, 161p.
University of Rochester Geology Department Staff, Guidebook for Fieldtrips in Western New York. New York State Geological Association, 121p. Available from Exec. Secretary M.P.Wolff, Dept. of Geology, Hofstra University, Hempstead, NY 11550.
Waller, R.M. and others, 1982, Geohydrology of the preglacial Genesee Valley in Monroe County, New York. U.S.Geological Survey Open File Report 82-552, microfiche, Ithaca, NY
Whitbeck, R.H., 1902, The pre-glacial course of the middle portion of the Genesee River; American Geographical Society Bulletin, Vol. 34, p. 32-33.
Wilson, M.P. 1981, Catastrophic discharge of Lake Warren in the Batavia-Genesee region. Ph.D.dissertation, Syracuse University, 107p.
Young, R.A. and W.D.Rhodes, 1973, Late glacial and postglacial geology of the Genesee Valley in Livingston County, New York; p. E1-E21 in Hewitt, P.C. (ed.), New York State Geological Association Guidebook. Available from Exec. Secretary, M.P. Wolff, Dept. of Geology, Hofstra University, Hempstead, NY 11550.
Young, W.H. and W.L.Kreidler, (eds) Guidebook, 29th Annual Meeting, New York State Geological Association, 66p. Available from Exec. Secretary M.P.Wolff, Dept. of Geology, Hofstra University, Hempstead, NY 11550

## GEOPHYSICAL INVESTIGATIONS OF GLACIAL DRIFT

**William J. Brennan**

## GRAVIMETRIC INVESTIGATIONS

### Introduction

The Genesee Valley in Livingston County, New York is a northerly-trending glaciated trough filled with till, glaciolacustrine silt and clay, and lesser amounts of sand and gravel. Similar drift-filled bedrock valleys have been investigated using gravimetric methods for the purpose of locating ground water resources (Hall and Hajnal, 1962; McGinnis and others, 1963; Rankin and Lavin, 1970), to reconstruct preglacial drainage relationships (Foose and Cunningham, 1968; Calkin and others, 1974; Roberts, 1975; Wilson and others, 1983), or to determine depth of fill (Gieschen, 1974). In many previous studies the gravimetric method has been used primarily as a reconnaissance tool. The principal objective in the present study has been to determine the shape of the bedrock floor of the Genesee Valley. The density contrast between the unconsolidated fill and the enclosing marine sedimentary bedrock is such that the acceleration of gravity over the deepest parts of the valley, where the maximum thickness of the fill is approximately 625 feet, is diminished by up to 5.7 milligals. Although some wells have penetrated the entire thickness of the unconsolidated fill, they are too sparsely distributed to provide a clear picture of the shape of the bedrock floor of the valley.

### Methods

In order to better delineate the bedrock floor of the valley, differences in the acceleration of gravity were measured with a gravimeter, along the Jones Bridge, Pioneer and Everman roads (see road log map for day 1). The measurements were corrected for tidal and instrument drift, elevation, Bouguer effect, latitude and terrain (zone J along Jones Bridge and Pioneer roads; zone M along Everman Road) to give the Bouguer gravity. The regional gradient of the gravitational field along each survey line was obtained by extending the measurements for three to five miles to the east and west of the valley margin and fitting, by graphical means, smooth profiles to the Bouguer gravity. The regional gradient profiles were confirmed by comparison with profiles taken from the Bouguer Gravity Map of Western New York (Revetta and Diment, 1971). The residual anomaly along each survey line was then obtained by subtracting the regional gradient from the Bouguer gravity (Fig. 1).

Interpretation of each residual anomaly was done by iterative construction of a series of two-dimensional models using depth to bedrock information from a few wells along each survey line and an initial assumed density contrast. In all instances it was possible to construct a model for which the calculated gravitational effect was within 0.2 milligal of the residual anomaly observed at each station. The density contrast from the best-fit model of each series, along with bedrock density obtained from gamma ray density logs, was used to determine the density of the valley fill along each survey line.

### Regional Gravity

It is readily apparent from the Bouguer gravity map that most identifiable

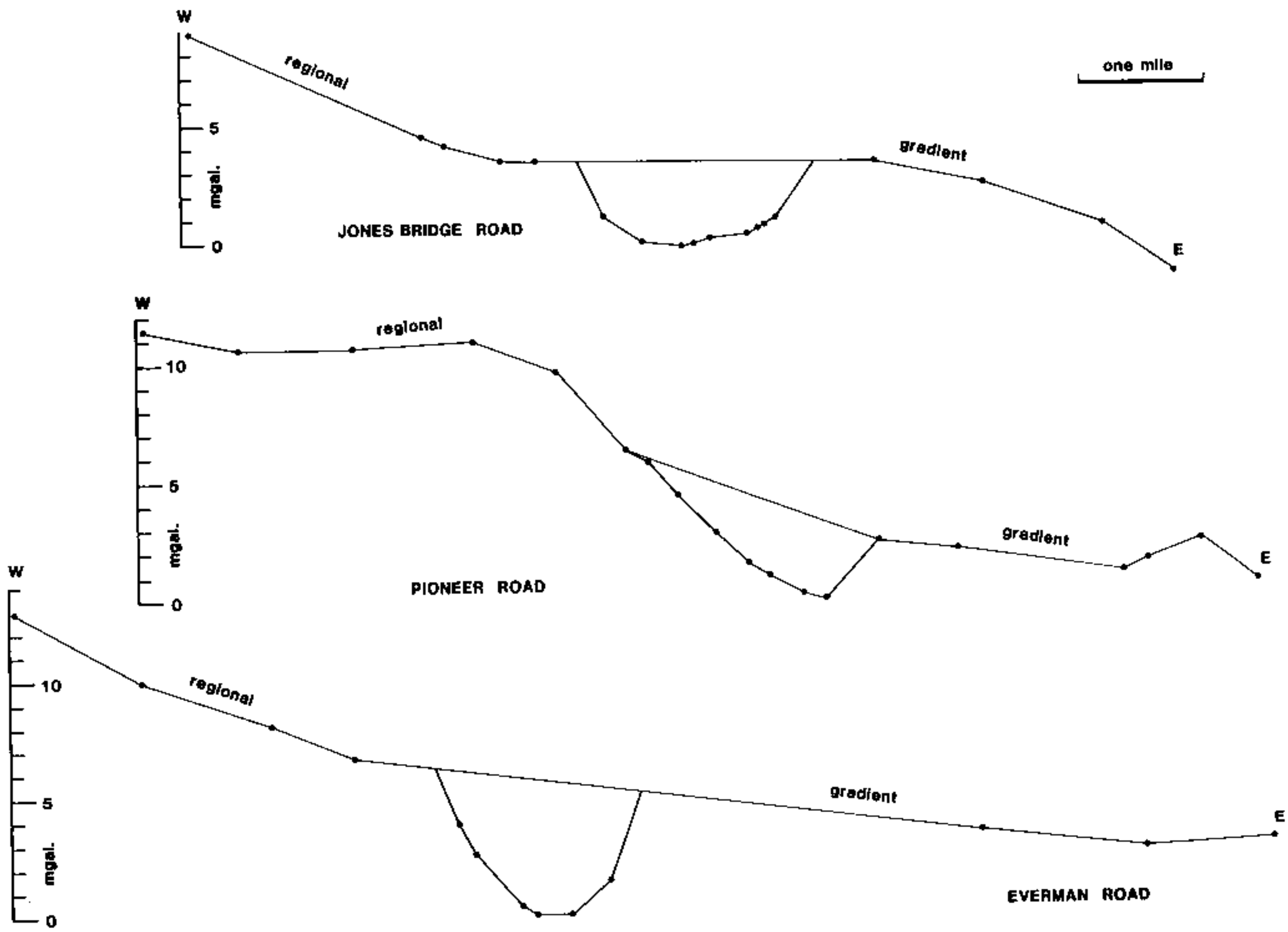



Figure 1. Bouguer gravity profiles across the Genesee Valley. Valley margins are located at points where the bouguer gravity and regional gradient intersect. Locations of profiles are given on Field Trip Route Map for day 1 (p. 40).

anomalies result from sources that are of regional size. The horizontal dimensions of all individual anomalies indicate that they result from anomalous density distributions located at depths below the top of the basement complex. This is to be expected given that the stations at which measurements were made were spaced approximately one mile apart. In addition, the simple structure and uniformity of the Paleozoic stratigraphic section, which are demonstrated by many geophysical logs of wells which are available and the extensive literature which is much too voluminous to be cited here, suggest that few sources should be expected to exist there. In Livingston County the principal feature of the Bouguer gravity map consists of a smooth decline of approximately 30 milligals from west to east. It is not surprising, therefore, that most anomalous features of the gravitational field with horizontal dimensions of one mile or less result from sources located in the glacial drift.

**Results**

Gravity profiles along the three lines surveyed are illustrated in Figure 1. The maxima of the residual anomalies are -3.6 mgals along Jones Bridge Road, -3.4 mgals. along Pioneer Road and -5.7 mgals. along Everman Road. The maximum thickness of fill and density contrast of the best-fit model for each survey are 500 feet (density contrast=-0.6 gm./cc.) for Jones Bridge Road, 500 feet (density contrast=-0.6 gm./cc.) for Pioneer Road and 625 feet (density contrast=-0.85 gm./cc.) for Everman Road (Fig. 2).

**Discussion**

In order to obtain average bulk density estimates for the fill, the density of the Paleozoic bedrock was determined from gamma ray density logs obtained from gas wells in the area. Healy (1970) has compared in situ density measurements made by borehole gravimeters and gamma ray density logs run in boreholes in the Basin and Range and found that densities obtained by the gamma ray density logging compared favorably with those obtained using a borehole gravimeter. His results lend support to the contention that the gamma ray density logs in the Genesee Valley give an accurate measure of bedrock density. Along both Jones Bridge and Pioneer roads bedrock consists of the shales of the Hamilton Group with a mean density of 2.6 gm./cc. Along Everman Road the bedrock consists of the shales of the Genesee and Sonyea Formations, for which the average density is 2.7 gm./cc. These densities are in general agreement with the results of Wilson and others (1983), who report an average density of 2.61 gm./cc. for shales and siltstones of the Arkwright and Conewango Groups in Chautauqua County, and also Roberts (1978), who reported an average density of 2.59 gm./cc. for shales of the Vernon Formation in Onondaga and Madison Counties. In both studies densities were determined by laboratory measurement on samples collected from outcrop.

The best-fit density contrast of -0.6 gm./cc. obtained for the Jones Bridge Road and Pioneer Road surveys indicates that the average bulk density of the fill in both locations is approximately 2.00 gm./cc. Along Everman Road the best-fit density contrast of -0.85 gm./cc. indicates that the density of the fill there is approximately 1.85 gm./cc. Both values fall within the range of average densities of alluvium in large valleys in the Basin and Range (1.69-2.22 gm./cc.) reported by Healy (1970), drift-filled valleys in Saskatchewan (1.80-2.15 gm./cc.) reported by Hall and Hajnal (1962), and

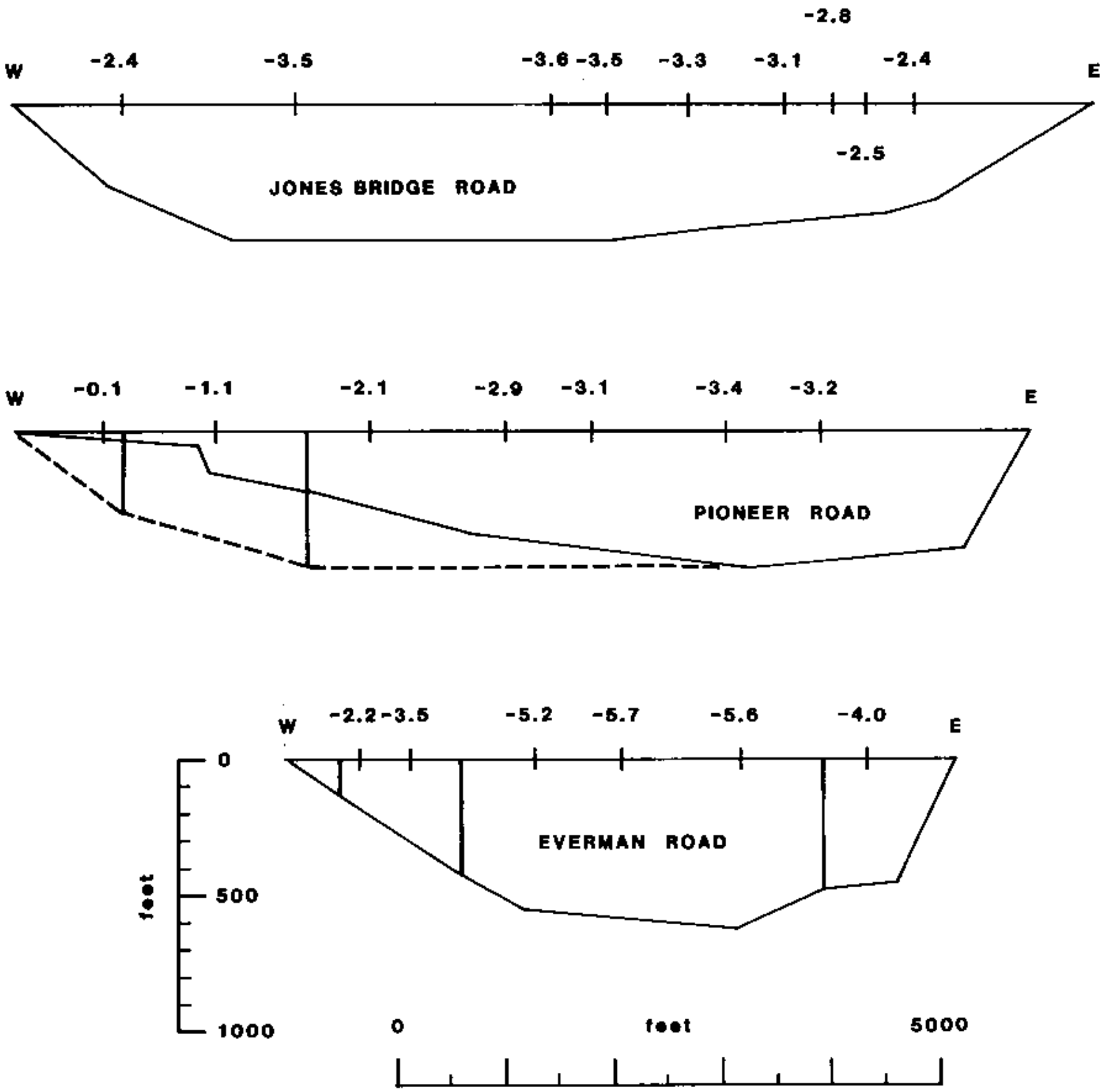



Figure 2. Cross-valley sections. Solid lines delineate configurations of valley floors obtained from best-fit gravity models. Dashed line delineates valley floor configuration at west end of Pioneer Road obtained from gas wells. Residual gravity anomaly values are given in milligals. Vertical lines are gas wells.

drift-filled valleys in central New York (1.84-2.04 gm./cc.) reported by Roberts (1978). In contrast, McGinnis and others (1963) found bulk density of fill to range between 2.06 and 2.54 gm./cc. in bedrock valleys in northern Illinois, and Wilson and others (1983) have reported a density range of 1.77 gm./cc. (sand) to 2.47 gm./cc. (till) for the fill in shallow upland valleys in southwestern New York. All density values discussed here are for a condition of 100 percent saturation.

Because there is little information on lithology of the fill from those few wells which extend to bedrock the cause of the lower density along Everman Road is unknown. McGinnis and others (1963) found that the density of drift in a bedrock valley in Illinois increased as a function of percent sand and gravel, but not as a function of depth. In contrast, Wilson and others (1983) found that the density of the fill in shallow upland valleys in southwestern New York was highest for till and lowest for sand. The measurements reported by Hall and Hajnal (1962) do not exhibit a simple correlation between density and lithology or depth. THey found that silt units (1.80 gm./cc.) and some sand units (1.90-2.15 gm./cc.) have the lowest densities, and clay units (2.00 gm./cc.) and till units (2.10-2.20 gm./cc.) have the highest densities. Roberts (1978) found highest density in fill composed of gravel and sand/clay mixtures (2.04 gm./cc.) and lowest density in fill composed of sand/silt (1.89 gm./cc.) and silt/clay mixtures (1.93 gm./cc.).

The inconsistent relationship between lithology and density is almost certainly due, in part, to the incomplete information on texture contained in generic terms like till, sand, and silty clay. Sorting and packing of framework grains exert strong influence on density, but such information is not adequately conveyed by brief lithologic labels. It also seems likely that some of the differences in density reported for similar lithologies may be the result of conditions during and following deposition. Lowland valleys in which the fill consists primarily of lacustrine sediment which has not been over ridden by ice, rather than till and/or other sediment which has been over ridden by ice, are more likely to have fill of relatively low density. Because the Genesee Valley has been occupied by a series of proglacial lakes the fill consists in large part of glaciolacustrine silt and clay, and it seems reasonable to speculate that the lower density fill along Everman Road is due to the presence of a greater thickness of lacustrine silt and/or clay. This interpretation is supported by the predominance of silt and clay reported in the upper 20 to 40 feet of engineering borings located along the present course of I-390.

Although the models calculated for the Jones Bridge Road and Everman Road surveys indicate the typical "U" shape of a glaciated trough (Fig.2), the cross-sectional shape of the valley along Pioneer Road is quite asymmetrical. However, depth to bedrock determinations obtained from two gas wells along the road have revealed that the assymmetry is due to a diminished density contrast along the west side of the valley rather than assymmetry in the bedrock floor. A deposit of ice-contact stratified drift is located at the west end of Pioneer Road. The gravels which comprise the deposit are exposed in several gravel pits in which the presence of abundant calcite cement can be observed. In much of the deposit the cement appears to fill all of the pore space, and it is clear that the gravity model delineates the lateral extent and aggregate thickness of the cemented gravel.

Depth to bedrock data from scattered wells indicate that from Jones Bridge Road the bedrock floor rises progressively northward. At the airport at Geneseo the fill is approximately 490 feet thick, and at Fowlerville Road, seven miles northward along the valley, it is approximately 100 feet thick. This shallowing of the floor is due to the presence of the Onondaga Formation at the bedrock surface. The Onondaga is much more resistant to erosion than the shales above and below, and the floor of the northern portion of the valley simply follows the Onondaga Formation up the regional dip. A similar shallowing near the southern termination of the valley at Dansville causes the longitudinal profile of the valley to resemble that of a bath tub.

The thalweg of the valley is at an elevation of approximately 465 feet at Fowlerville Road, and it descends to approximately 70 feet in elevation at the airport at Geneseo. The elevation is 70 feet at Jones Bridge Road, which is near the point where the bedrock floor crosses the top of the Onondaga Formation into the softer shales above. The thalweg is at an elevation of 70 feet at Pioneer Road and then continues its descent down to -40 feet at Everman Road.

### Conclusion

In the Genesee Valley, the gravimetric method has proven to be useful as a means of delineating, locating and disproving the existence of buried channels. Given the density contrasts that have been observed in the area, a buried channel with 100 feet of sedimentary fill should have an associated gravity anomaly in the range 0.5 to 1.0 mgals.

## PALEOMAGNETIC INVESTIGATIONS

### Introduction

Fine grained detrital sediments,including glacial varves are known to possess a stable detrital remanent magnetism (DRM) which records the geomagnetic field at the time and location of deposition (Johnson and others, 1948). However, the fidelity of records in glacial varves has been the subject of considerable debate. Anomalously low remanent inclinations, given the latitude of sampling sites have been attributed to an initial inclination error (Johnson and others, 1948), bedding error (Griffiths, 1955), compaction (Blow and Hamilton, 1978) and post-depositional remagnetization (Barton and McElhinny, 1981). While it has been demonstrated in laboratory experiments that water currents may affect the direction of remanent declination (Rees, 1961,1964), unequivocal field evidence of such current rotation has not been reported. Barton and McElhinny (1981) have attributed some changes in remanent declination in wet-sediment piston cores to long-term post-depositional remagnetization, however, most studies of glacial varves and other densely compacted sediments indicate that the "lock-in" time for acquisition of remanence is at most a few varve years (Verosub, 1975). A more complete discussion of these factors is presented by Brennan and others(1984) and Ridge (1985). It has been demonstrated, in applications where independent means of correlation are available, that the remanent declination recorded in glacial varves at sites known to be of the same age are identical (Brennan and others, 1985; Ridge, 1985). Thus, there is strong evidence that glacial varves faithfully record the declination of the geomagnetic field but not its inclination.

## Deglacial History of Western New York

Retreat of the ice sheet from the southern limit of glaciation began approximately 20,000 years ago (Muller, 1977), and by 12,300 years B.P. the ice had receded to the southern shoreline of Lake Ontario (Calkin and Brett, (1978). During deglaciation, a series of proglacial lakes (Fairchild, 1908) were the sites of deposition of varved clays. Since the areas inundated by lake waters shifted northward with the retreat of the ice front,the age of the varved clays decreases in the same direction. Recessional moraines provide a means of correlating between widely separated exposures.

## Sampling of Glacial Varves

Because sections of glacial varves deposited in proglacial lakes in the Genesee Valley do not span more than a few hundred varve years, it was not possible to sample over a long interval of time by simply sampling up a stratigraphic section. As an alternative, sampling of the basal portion of the varves at many sites along the path of ice recession was conducted. This procedure is based on the stratigraphic onlap relationship which exists between the basal portion of any section of varves and the underlying till. As the ice front retreated northward past a site a change from deposition of till beneath the ice to deposition of varves in the adjacent proglacial lake occurred. Thus, the basal varves become progressively younger to the north along the path of ice retreat, and the sampling sites are arranged in proper chronologic order in the same direction.

Sampling was restricted to those sites where the varves were horizontally stratified, undeformed and resting directly on till. At each site the bottom half meter (approximately 20 varve years) was sampled with two to three samples taken from an average of three to five horizons. This procedure was used to average out any noise of sedimentological origin in the record (Verosub, 1979). Each sample consists of a 1"x1" right circular cylinder core which spans from one to four varve years. The mean declination obtained at each sampling site represents the geomagnetic declination during the interval of deposition of the varves at the site.

## Laboratory Procedures

The remanent magnetism of the cores was measured on a PAR SM-2 spinner magnetometer at the Department of Geological Sciences, State University College at Geneseo. All cores were subjected to step-wise alternating field demagnetization in fields of up to 400 oersteds, in order to remove viscous remanence acquired after deposition and to verify the stability of the remanence. More detailed descriptions of the laboratory procedures are given by Brennan and others (1984) and Ridge (1985). In general, the glacio-lacustrine varves were found have magnetic properties which suggest that they faithfully record the geomagnetic declination at the time and place of deposition.

## Results

Secular variation records of geomagnetic declination have now been obtained from glacial varves in the Genesee Valley and the the eastern portion of the

Finger Lakes region (Brennan and others, 1984; Braun and others,1985) and the western Mohawk Valley (Ridge, 1985). The composite record obtained from these studies is is nearly continuous over the interval from approximately 16,000 B.P. to 12,300 B.P. (Brennan and others, 1985). A number of aspects of the composite record (Fig. 3) are noteworthy. First, the declination was 45-50 degrees west of north at the time that the ice front reached the southern shoreline of Lake Ontario. This very distinctive feature may prove to be an easily identifiable aspect of other secular variation records obtained in the eastern Great Lakes region and, therefore, an easily identifiable time interval. Second, a number of discontinuities, which appear to represent short intervals of time, are present. The discontinuity in the Genesee Valley record between 14,000 B.P. and 13,000 B.P. has been interpreted as the result of the Port Huron readvance (Brennan and others, 1984). The Shed Brook Discontinuity reported by Ridge (1985) is represented in both the magnetic and stratigraphic records. Several discontinuities exist in the Genesee Valley record during the Nissouri Stadial. They result in part from oscillation of the ice front as indicated by inter-stratification of varves and till. In some instances the remanence of the varves appears to have been altered as a result of mechanical loading as the varves were over ridden by grounded ice. There is a larger discontinuity in the Genesee Valley record which spans the time interval during which the Valley Heads Moraine was deposited. In general, the effect of a hesitation or readvance of the ice front on a secular variation record obtained by sampling the base of a stratigraphic onlap is the same as that produced by an interval of non-deposition and/or erosion on a conventional vertical section in that an abrupt discontinuity in the record often results. However, the secular variation of the geomagnetic field is periodic, and it is possible that some unrecorded intervals might be masked by the periodicity.

**Lake Ontario Cores**

During the summer of 1979, seven Alpine piston cores, up to 7.3 meters in length, of lacustrine sediments from Lake Ontario were obtained through the cooperation of Dr. Deborah Hutchinson of the U.S. Geological Survey. Two of the cores were unusable because of excessive mechanical deformation. All but one (7021) of the remaining five were collected from sites where the present water depth is in excess of 142 meters, which is consistent with their location, at the time of deposition, within that portion of the Lake Ontario basin occupied by Lake Admiralty (Coleman, 1922). Since Coleman has estimated the elevation of the Lake Admiralty Shoreline (now submerged and located approximately 40 km. east of Hamilton, Ontario) at approximately 67 meters below present lake level, the water depths at the coring sites were probably in excess of 70 meters during the period of low lake levels which immediately followed deglaciation of the region. Thus, unconformities, and therefore discontinuities in the magnetic records, should not be expected in the cores. One of the cores (7024) penetrated the varve-till contact at a depth of 4.8 meters in the core.

The remanent magnetism of each of the five cores was measured on a Digico long-core magnetometer made available by Dr. Donald Woodrow of Hobart and William Smith College, Geneva, New York. Since the instrument is capable of measuring only the horizontal component of magnetization, and since piston cores are not oriented with respect to azimuth, the records contain only relative declination (Fig. 4). The cores were not demagnetized,but the

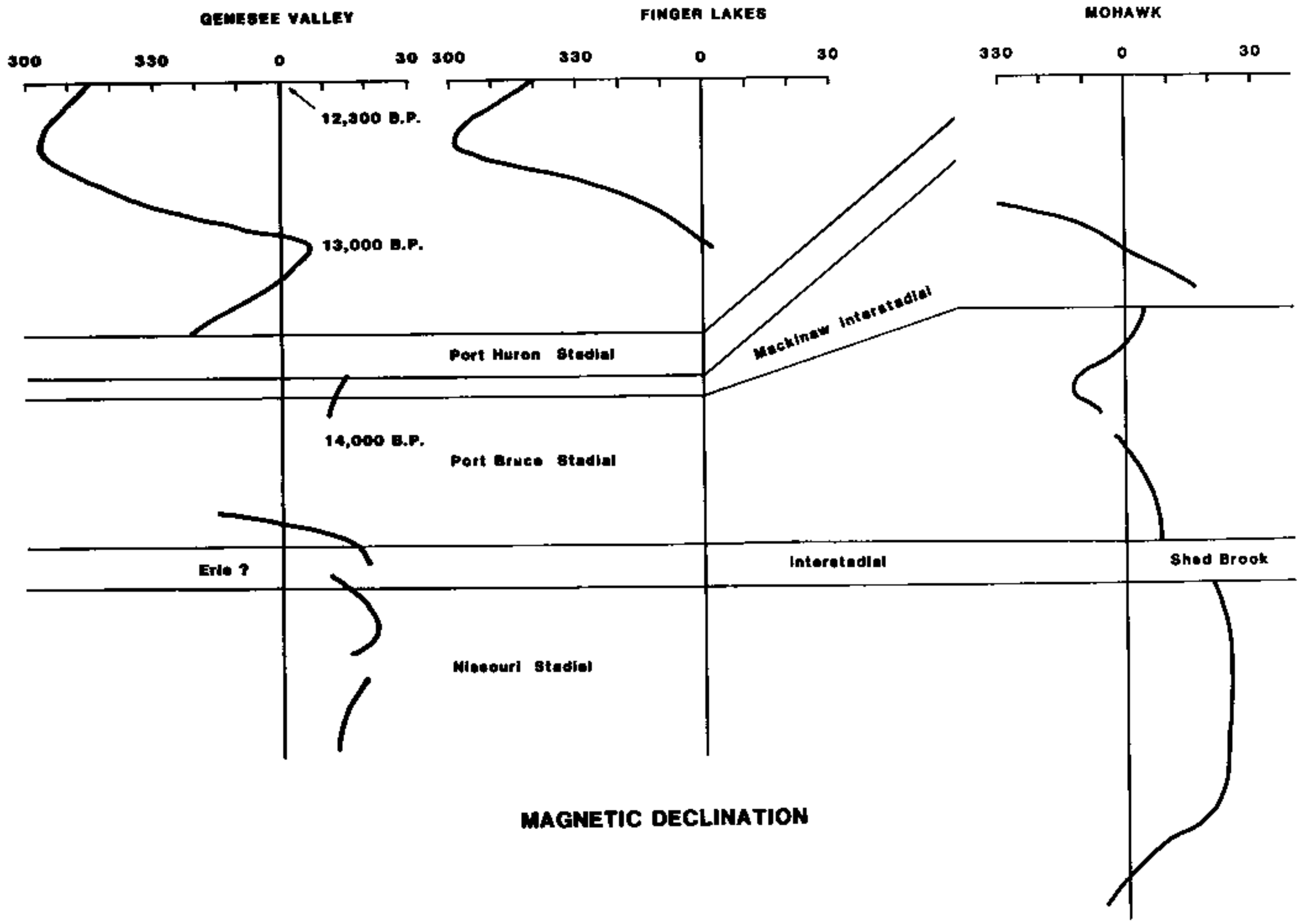



Figure 3. Combined declination records from the Genesee Valley, eastern Finger Lakes and western Mohawk regions.

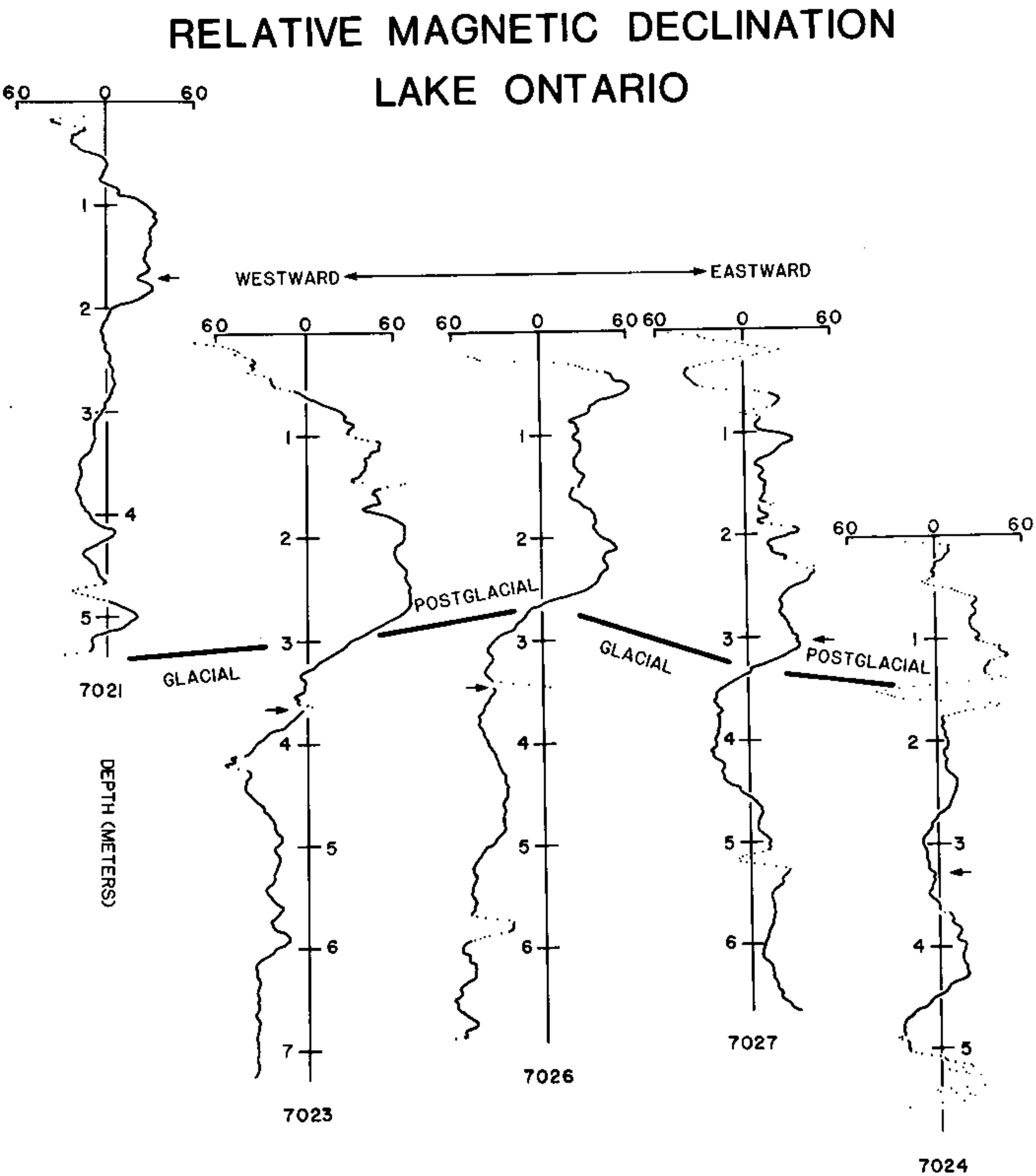



Figure 4. Relative magnetic declination records of piston cores from Lake Ontario. Small arrows indicate joints in the plastic liner tubes.

stability of the remanence was verified by alternating field demagnetization of samples taken from two (7024, 7026) of the cores. In addition, the cores were examined by X-radiography and later opened and examined visually to identify stratigraphic contacts and the presence of deformed intervals.

## Discussion

Some effects of mechanical deformation of the piston cores are apparent in their magnetic records (Fig. 4). The most conspicuous examples are the inclined trend of the entire length of core 7026, the inclined offset between 2.7 and 4.2 meters in core 7023, the more than 100 degree range in declination of core 7023, and the uppermost meter of cores 7023, 7026 and 7027. The irregularities in the uppermost portions appear to be the result of the slurry-like consistency of the sediment, and the inclined trends and unrealistic range are more likely the result of rotation of the core barrel. In addition, the peak-to-peak amplitude of maxima of equivalent declination features are not of consistent magnitude. It is well known that many magnetic records obtained from piston cores are of questionable fidelity (Brennan and others, 1984). Creer and Tucholka (1982) have enumerated the difficulties which are encountered in any attempt to correlate magnetic records of lake cores when some of the cores are not well dated or record the geomagnetic field with questionable fidelity.

The piston core records reported here are typical in that they are not of sufficiently high fidelity that they can be used as a basis for correlating among the other basins of the Great Lakes. Despite this limitation, they do share a general similarity of gross features which may be of value in local correlations. In particular, the prominant westward swing in declination just below the varve-till contact in core 7024 is most significant. The location of core 7024 is such that the ice front retreated past it shortly after passing the sites sampled in outcrop along and immediately south of the shoreline of Lake Ontario. Therefore the declination recorded in the basal varves just south of the shoreline should be identical with that recorded in the till in core 7024, because till was being deposited beneath the ice at the location of core 7024 while varves were being deposited in the proglacial lake located a short distance south. Thus, the secular variation record obtained from outcrop sampling of basal varves appears to be continuous with that obtained from core 7024. It should be noted that most till lithologies in the region are not well suited to record the geomagnetic field. Only a few exposures of fine-grained, clay-rich till of limited areal extent have been found to have a stable remanence.

## Conclusion

The secular variation record of geomagnetic declination which has been obtained from glacial drift in central and western New York is internally consistent and has proven to be useful as an independent means of correlation. As the interval covered by the record is increased in length, and additional partial records are obtained from more localities in the region, this technique will become more useful as a correlation tool.

## REFERENCES

Barton, C.E. and M.W. McElhinney, 1981, A 10,000 yr. geomagnetic secular

variation record from three Australian maars, Geophysical Journal, Roy. Astron. Soc., v. 67, pp. 465-485.
Blow, R.A. and N. Hamilton, 1978, Effect of compaction on the acquisition of a detrital remanent magnetization in fine-grained sediments, Geophys. Jour., Roy. Astron. Soc., v. 52, pp. 13-23.
Braun, D.D., E.W. Helfrick, G.F. Olenick and W.J. Brennan, 1984, Using secular variation of geomagnetic declination to test the age of the Kent Moraine in the Genesee Valley, New York: a progress report: Geol. Soc. Amer. Abstr. with Programs, v. 16, no. 1, p. 5.
Braun, D.D., T.A. Daniel, L.B. Smith and W.J. Brennan, 1985, Quaternary geologic and geomagnetic record of the upper Genesee Valley, New York: Geol. Soc. Amer. Abstr. with Programs, v. 17, no. 1, p. 7.
Brennan, W.J., M. Hamilton, R. Kilbury and R.L. Reeves, 1984, Late Quaternary secular variation of geomagnetic declination in western New York, Earth and Planet. Sci. Lett., v. 70, pp. 363-372.
Brennan, W.J., J.C. Ridge and D.D. Braun, 1985, Stratigraphic interpretation of remanent declination in Pleistocene glaciolacustrine clays of New York, 5th General Assembly, Int. Assoc. of Geomag. and Aeronomy, Prague, Czechoslovakia, p. 179.
Calkin,P.E. and C.E. Brett, 1978, Ancestral Niagara River drainage: stratigraphic and paleontologic setting, Geol. Soc. A. Bull., v. 89, pp. 1140-1154.
Calkin, P.E., D.S. Hodge, D.E. Champion, E.T. Oakesford and E.C. Palmer, 1974, Gravity delineation of the preglacial Cazenovia River Valley, Western New York State, U.S.A., Zeitschrift fur Geomorphologie N. F., v. 18 no. 3, pp. 247-259.
Coleman, A.P., 1922, Glacial and post-glacial lakes in Ontario, Ont. Fish Res. Lab., Univ. Toronto 10, 76 p.
Creer, K.M. and P. Tucholka, 1982, Construction of type curves of geomagnetic secular variation for dating lake sediments from east central North America, Can. Jour. Earth Sci., v. 19, pp. 1106-15.
Fairchild, H.L., 1908, Glacial waters in central New York, New York State Mus. Bull. 127, 66p.
Foose, R.M. and C.G. Cunningham, 1968, Preglacial Connecticut River course near Amherst, Massachusetts, Geol. Soc. Amer. Bull., v. 79, pp. 1415-1418.
Gieschen, Paul A., 1974, Gravimetrically determined depths of fill in the upper Susquehanna River system: procedures and interpretations, Masters dissertation, SUNY College at Oneonta, Oneonta, New York, 90 p.
Griffiths, D.H., 1955, Remanent magnetism of varved clays from Sweden, Mon. Not. R. Astron. Soc., Geophys. Suppl. 7, pp. 103-114.
Hall, D.H. and Z. Hajnal, 1962, The gravimeter in studies of buried valleys, Geophysics, v. 27, pp. 939-951.
Healy, D.L., 1964, Gravity survey of northern Hot Creek Valley, Nye County, Nevada, U.S. Geol. Surv. Technical Letter: Central Nevada-18, 22p.
Healy, D.L., 1970, Calculated in situ bulk densities from subsurface gravity observations and density logs, Nevada Test Site and Hot Creek Valley, Nye County, Nevada, U.S. Geol. Survey Prof. Paper 700-B, pp. B52-B62.
Johnson, E.A., T. Murphy and O.W. Torreson, 1948, Pre-history of the Earth's magnetic field, Terr. Magn. Atmos. Electr., v. 53, pp. 349-372.
McGinnis, L. J. Kempton and P. Heigold, 1963, Relationship of gravity anomalies to a drift-filled bedrock valley system in northern Illinois, Illinois State Geol. Surv. Circular 354, 24 p.
Muller, E.H., 1977, Quaternary geology of Niagara, New York, Niagara Sheet, New York State Mus. Sci. Serv., Map and Chart Series 28.

Rankin, W.W.E and P.E. Lavin, 1970, Analysis of a reconnaissance gravity survey for drift-filled valleys in Mercer Quadrangle, Pennsylvania, Journal of Hydrology, v. 10, pp. 418-435.

Rees, A., 1961, The effect of water currents on the magnetic remanence and anisotropy of some sediments, Geophys. Jour. Roy. Astron. Soc., v. 5, pp. 235-51.

Rees, A., 1964, Measurements of the natural remanent magnetism and anisotropy of magnetic susceptibility of some Swedish glacial silts, Geophys. Jour. Roy. Astron. soc., v. 8, pp. 356-69.

Revetta, F.A. and W.H. Diment, 1971, Simple bouguer garvity anomaly map of Western New York, New York State Mus. and Sci. Service, Map and Chart Ser. no. 17.

Ridge, J.C., 1985, The Quaternary glacial and paleomagnetic record of the West Canada Creek and western Mohawk Valleys of central New York: Ph.D. dissertation, Syracuse University, Syracuse, New York, 471 p.

Roberts, Richard B., 1978, Gravity studies of buried valleys north of Green Lakes State Park, central New York, Masters dissertation, Syracuse University, Syracuse, New York, 100 p.

Verosub, K.L., 1975, Paleomagnetic excursions as magneto-stratigraphic horizons: a cautionary note, Science, v. 190, pp. 48-50.

Verosub, K.L., 1979, Paleomagnetism of varved sediments from western New England: variability of the paleomagnetic recorder, Geophys. Res. Lett., v. 6, pp. 241-244.

Wilson, M.P., D.N. Peterson and T.F. Ostrye, 1983, Gravity exploration of a buried valley in the Applachian Plateau, Ground Water, v. 21, no. 5, pp. 589-596.

# WISCONSINAN DEGLACIAL HISTORY OF THE GENESEE VALLEY FROM THE TERMINAL MORAINE TO THE VALLEY HEADS MORAINE

**Braun, D.D., Geog. & Earth Sci., Bloomsburg University, Bloomsburg, Pennsylvania, 17815**

## Introduction

The upper Genesee Valley extends 84 km from the Wisconsinan terminal moraine at the drainage divide between the Genesee basin and the Allegheny and Susquehanna basins to the Valley Heads moraine at the south (upstream) end of the Letchworth bedrock gorge. The valley has a continuous northward trend, N $30^0$ W for 56 km to Caneadea and then N$30^0$E to Letchworth gorge and beyond (Fig. 2, Muller's intro.). Wisconsinan deglaciation was marked by a sequence of proglacial lakes held in by retreating ice, whose levels were controlled by a succession of 3 major outlets that become lower northward in the direction of recession (Fig. 5, Muller's intro.). Recession was interupted at times by readvance that restored earlier lake levels. That all 3 outlets are westward to the Allegheny implies that ice was thicker to the east, blocking eastward outlets, and that the ice margin orientation during recession from the Terminal to the Valley Heads position was basically northwest to southeast across the region. The lake sequence as presently known (Fig. 5, Muller's intro.), has few differences from Fairchild's (1896) original sequence.

Essentially all glacially derived sediment on the valley sides and under the present valley floor were deposited in a lacustrine environment. These sediments are primarily clay-silt to silt-sand rhythmites, sand, sand and gravel, and diamicton. An ice-marginal bench or shoulder, in places a true kame terrace, is composed of sand and gravel, rhythmites, diamicton, or bedrock and marks the erosional and depositional base level afforded by the different lake levels. The river, from its headwaters to Letchworth gorge, has progressively incised into the sediments, accentuating the height of the ice marginal shoulder. From Caneadea to Letchworth, the height of the shoulder is mostly erosional with the present floodplain incised well below the original proglacial lake floor.

## Proglacial lake sequence

Ice withdrawal from the Genesee-Susquehanna divide resulted in impoundment of "primary" lakes (Fairchild, 1896) in the three tributary valleys of the Genesee in northern Pennsylvania (Fig. 1). The primary lakes initially drained across cols 1,2, and 3 (Fig. 1) into the Allegheny or Susquehanna basins. Glacial retreat opened cols 4 and 5 between the tributary valleys and further ice recession to the New York - Pennsylvania border permitted formation of a single glacial Lake Pennsylvania (Fairchild, 1896) draining through col 3 at about 2050 ft (625 m).

Renewed recession initiated local ice-marginal drainage across col 7 at 2110 ft (643 m) and out through partially drift-filled Stone Dam col at 1770 ft (539 m) (Fig. 2). Two more km of ice recession opened a col at 1710 ft (521m) (Fig. 2), now floored by boulder-lag coverd bedrock, that led to rapid draining of glacial Lake Pennsylvania and partial removal of the Stone Dam col drift-plug. The Stone Dam channel was the deepest and longest-lived sluiceway and controlled proglacial lake level until ice receded 35 km north to open the

Cuba outlet (Fig. 2). Wisconsinan glacial drift in the bottom of the 700 ft (212 m) deep bedrock sluice shows that the channel had been cut to its present elevation in pre-Wisconsinan time.

Continued ice recession opened the main entrance to Stone Dam Col and then a significant stillstand of the ice occurred. This stillstand developed a head of outwash within the Genesee Valley at Stannards (Fig. 2) with flow southward and a second head of outwash at the confluence of the Genesee and Cryder valleys with flow to the northwest. Between these 2 kame deltas at 1700 ft (518 m), the Genesee Valley was occupied by glacial Lake Shongo (Muller and others, in press)(Fig. 2). Lake Shongo coexisted with the next proglacial lake to the north in the sequence until drained by further incision of the drift plug in the Stone Dam sluice.

Renewed ice recession left the 1700 ft (518 m) Stannards head of outwash as the initial threshold of glacial Lake Wellsville (Fairchild,1896) (Fig. 2). During the first 5 km of recession, only 20 ft (6 m) of threshold incision occurred as shown by the 1680 ft (515 m) topset-foreset contact on a delta built into the Genesee Valley by Growner Brook near Wellsville (Fig. 2). In the next 5 km of recession, 80 ft (24 m) of threshold incision occurred as shown by a 1600 ft (488 m) head of outwash at Scio (Fig. 2). For the remaining 18 km of recession to the next lower outlet there would be no further lowering of Lake Wellsville's level. The lack of lake lowering is attributed to final removal of the drift plug and cessation of incision with development of a low 3.7 ft/mi (.6 m/km) gradient bedrock floor in the Stone Dam sluice.

At maximum extent, Lake Wellsville extended 30 km from Stannards to near Belfast, providing a base level for terrace development at elevations from 1660 to 1620 ft (506 to 494 m). Nearly all benches cited by Fairchild (1896) as wave-cut terraces are actually ice marginal deposits. Ice recession during the Lake Wellsville stage was continuous but with temporary stillstands marked by groupings of ice marginal deposits and channels. Minor ice margin oscillations and/or debris flow events are marked by intercalation of rhythmites and diamictons. First Lake Wellsville drained when the ice retreat opened the 1500 ft (457 m) elevation Cuba outlet (Fig. 2).

Lake Belfast-Fillmore (Fairchild, 1896), controlled by the Cuba sluice, came into existence temporarily as ice recession continued to near Fillmore (Fig. 3). Then a short lived 24 km readvance to the Angelica area overran Lake Belfast-Fillmore and very temporarily reestablished Lake Wellsville. Recession again took place with the ice retreating an unknown distance northward of the present Valley Heads position (Erie interstade?) and again temporarily impounding Lake Belfast-Fillmore. At Caneadea, organic matter near the base of rhythmites from this recession have yielded a radiocarbon age of 16,380, +660,-710 (DIC 1884)( W.J. Brennan, S.U.C. at Genseo, oral communication).

A second readvance of at least 30 km constructed the bulk of the Angelica-Belmont moraine and restored Lake Wellsville for the last time. Upon final recession from the Angelica-Belmont moraine, Lake Belfast-Fillmore again succeeded Lake Wellsville. A prominent kame delta complex at Rushford marks a significant stillstand of this final recession and, upon further ice withdrawal, it may have divided Lake Belfast-Fillmore into two lakes, one at

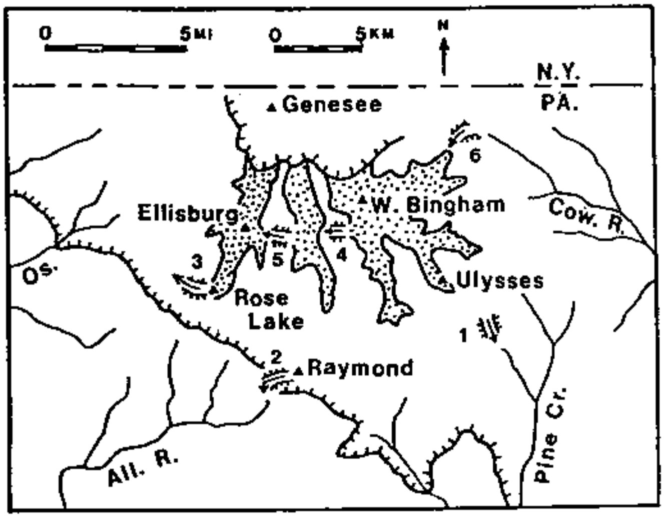



Fig. 1 Map showing terminal Wisconsinan border and primary lakes in the Genesee headwaters. Numbered cols are discussed in the text.

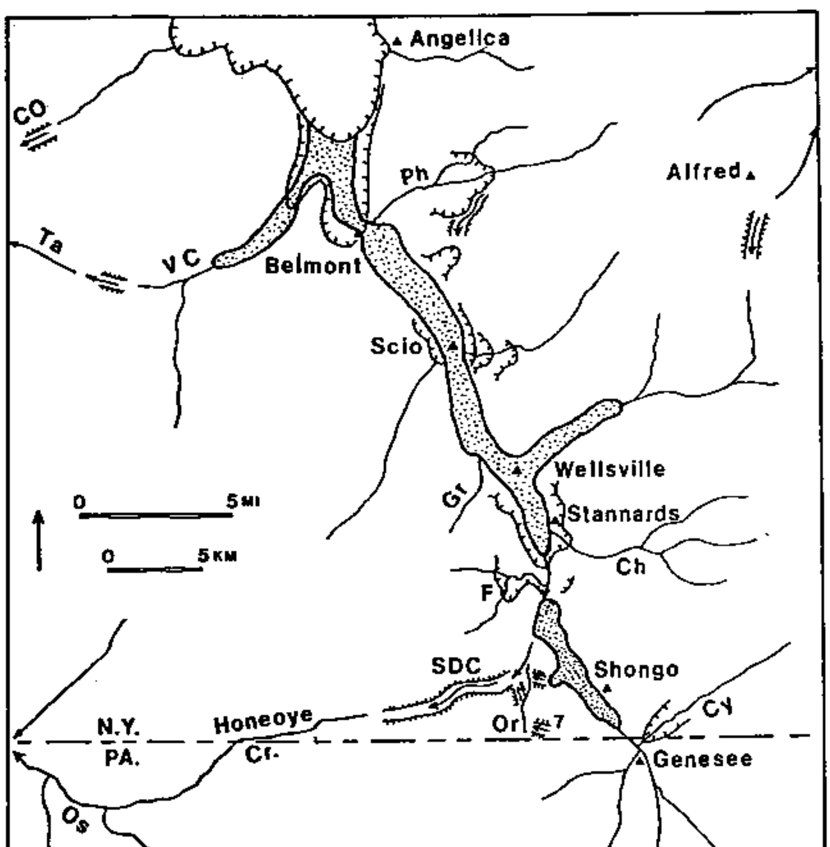



Fig. 2 Map of glacial Lake Shongo and glacial Lake Wellsville. Lake Wellsville is shown at its maximum extent during deposition of the Angelica - Belmont moraine. Both lakes drained through Stone Dam col (SDC) to the Allegheney River by way of Honeoye and Oswayo (Os) creeks. CO = Cuba Outlet, Cy = Cryder Creek, Gr = Growner Brook, Or = Orebed Creek, VC = Van Campen Cr.

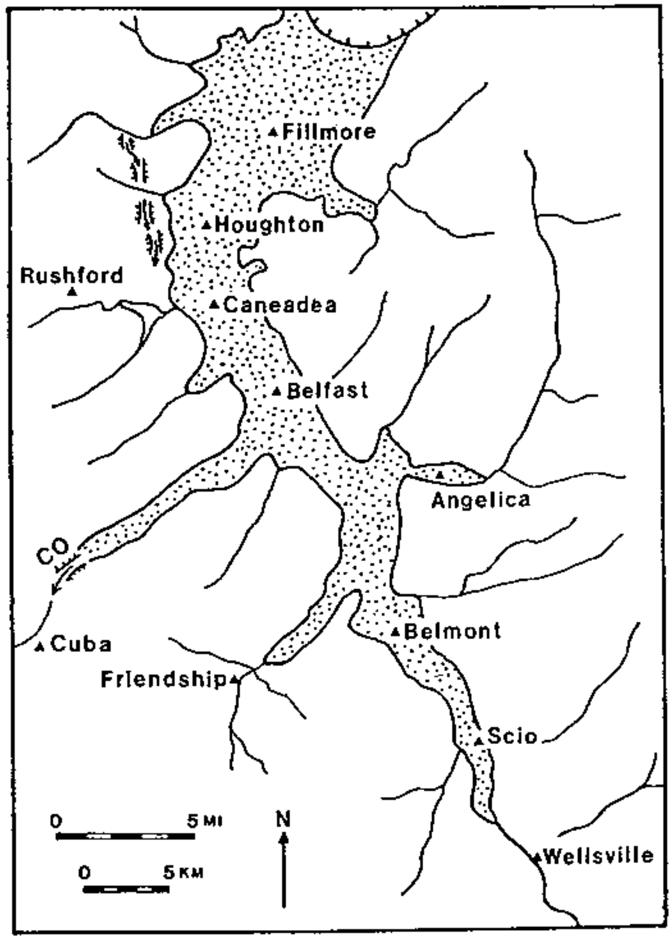



Fig. 3 Map of glacial Lake Belfast - Fillmore near its maximum extent during building of the Valley Heads moraine. The lake drained through Black Creek valley or the Cuba Outlet (CO) via Oil Creek to the Allegheney River.

Belfast and one at Fillmore (Fig. 3). Between Rushford and the Valley Heads margin, a series of ice marginal channels have been cut in rock along the west flank of the Genesee Valley (Fig. 3). Along both flanks of the Genesee valley, prominent ice marginal deposits form morainic loops across tributary valleys and temporarily impounded lakes in those valleys.

Glacial recession an unknown distance northward was followed by readvance that built the Valley Heads triplet of morainic loops across the Genesee Valley south of the Letchworth gorge and Portageville. No interruption of proglacial lake sedimentation is observable underneath the Valley Heads diamictons, implying that ice did not recede far enough during the Erie Interstadial (about 15,500 yrs. ago) to interrupt glacial ponding (Braun et al, 1985). The readvance could have removed all evidence for the Erie Interstade underneath the Valley Heads moraines but there is also a lack of evidence for an Erie Interstade hiatus south of the moraines. If continued deposition during the Erie Interstade occurred in a moraine dammed nonglacial lake, then part of the rhythmite section under the Valley Heads diamictons is nonglacial (Fig. 5, Lake unit between Dl & Du; Day 1, Stop 5, Fig. 2, unit 4).

About 14,000 BP, recession from the Valley Heads Moraine exposed for the first time outlets from the Genesee Valley southeast to the Susquehanna via Canaseraga Creek and Canisteo River. Southwest draining 1500 ft (457 m) elevation Lake Belfast-Fillmore was succeeded by southeast draining 1360 to 1320 ft (415 to 402 m) Lake Nunda (Fairchild, 1896). The Valley Heads formed a drift barrier across the Genesee Valley at a minimum elevation of 1400 ft (427 m) and continued to dam the upper Genesee, impounding Lake Portageville (Fairchild, 1896) even after ice withdrawal (Fig. 3). North-draining Lake Portageville shrank progressively as a result of outlet incision into the drift threshold and sediment infilling until its extinction well before 7500 BP, the radiocarbon age of terrace deposits 40 to 60 ft (12 - 18 m) above the present channel. Lake Portageville was a shallow short-lived lake, at most 50 ft (15 m) deep at Fillmore where the lake floor was as low as 1350 ft (411 m) and should more appropriately be named post-glacial Lake Fillmore.

**Position of the Late Wisconsinan terminus**

The key question since the time of Chamberlain (1883) and Lewis (1884) has been the relative age of the drift on either side of the Salamanca re-entrant along and within Lewis' terminal moraine. Consensus through the 1970's (Muller 1977) has been that the terminus is older (early or mid Wisconsinan Olean Drift) on the east side of the re-entrant than on the west side (late Wisconsinan Kent drift). In the Genesee valley, the Late Wisconsinan terminus has been placed at the Angelica-Belmont margin (MacClintock and Apfel 1945, Connally 1964; Muller 1965,1977) and correlated with the Kent moraine in Ohio (Muller 1963,1965,1975,1977; Connally 1964). This reasoning is based on differences in landforms and deposits observed to either side the Angelica-Belmont margin (Table 1). While the differences observed are real, the age significance of the differences is uncertain. Muller (1960,1965,1975,1977) has also traced the Kent moraine around the Salamanca re-entrant to the Angelica-Belmont margin.

| Feature Observed | Difference observed (north vs. south side) |
|---|---|
| clast lithology and heavy minerals in the drift | (bright vs. drab) |
| amount of glacial modification of the landscape | (much vs. little) |
| continuity of glacial cover | (continuous vs. discontinuous) |
| depth of weathering and leaching of the drift | (shallow vs. deep) |
| degree of periglacial modification | (slight vs. significant) |

Table 1 Differences in landforms and deposits to either side of the Angelica-Belmont moraine.

Craft (1976), working on the west side of the Salamanca re-entrant, and Lafleur (1979), working at the apex of the re-entrant, have questioned the mid-Wisconsian age assignment (White 1969; Muller 1975,1977) for the Titusville-Olean drift to the south of the Kent margin. They have suggested that the Titusville-Olean drift is earliest late Wisconsinan. Working on the southeast side of the Salamanca re-entrant in northeastern Pennsylvania, Crowl and Sevon (1980) argue that the Olean drift of Lewis' terminal moraine is Late-Wisconsinan in age. This age assignment is supported by radiocarbon dates and palynology work in the Great Valley area of Pennsylvania and New Jersey (Cotter 1983; Cotter and others,1985). In the Genesee Valley , this age assignment would make Lewis' Terminal Moraine at the Genesee divide the late Wisconsinan terminus and make the Angelica-Belmont margin a younger readvance position within the late Wisconsinan.

**Testing the age of the Angelica-Belmont moraine**

To test whether the Angelica-Belmont moraine is the late Wisconsinan terminus or a readvance within the late Wisconsinan (Fig. 4), the glacial geology of Genesee Valley both north and south of the moraine was mapped to identify:

1. Presence or absence of mid and/or early Wisconsinan paleosols under and south of the Angelica-Belmont margin.
2. Sampling sites in clayey rhythmites to either side of the Angelica-Belmont margin to develop a magnetostratigrahy using secular declination changes .
3. Lithostratigraphic relationship of the Angelica-Belmont margin to the Valley Heads margin to the north and to the unnamed margins to the south.

No paleosols have yet been identified under or in front of the Anglica - Belmont margin. Magnetic declination values are essentially the same in rhythmites both above and below the diamicton that makes up the bulk of the Anglica-Belmont moraine. This geomagnetic evidence leaves two choices (Fig. 4):

1. The Angelica-Belmont margin was occupied by ice for 100 yrs. or so, too short an interval to show a significant change in declination.
2. The Angelica-Belmont margin was occupied by ice for about 1000 yrs. or some multiple of 1000 yrs., the time it would take the sinusoidal declination change record to repeat a particular value.

The lithostratigraphy developed from the mapping provides the strongest case for Angelica-Belmont margin being a readvance position north of the late Wisconsinan terminus rather than the late Wisconsinan terminus itself (Fig. 5). The key observation is that the diamicton that makes up the body of the Angelica-Belmont moraine is directly traceable northward to the Valley Heads moraine (Fig. 5, unit $D_u$). Lake sediments separate this upper diamicton from a lower diamicton throughout the area. This lower diamicton, at or near river level in this part of the Genesee Valley, passes underneath the Valley Heads and Angelica-Belmont positions and continues southward towards the terminal moraine in Pennsylvania (Fig. 5,unit $D_l$). This diamicton is taken to represent the late Wisconsinan Nissouri stade recession from the terminal moraine in Pennsylvania. This implies that the 100 to 300 ft (30 to 90 m) thick section of glacial deposits underneath river level throughout this area is from the Nissouri stade advance and/or older advances.

**Did early or mid Wisconsinan ice extend south of the late Wisconsinan ice ? Evidence from outside the Genesee Valley**

High resolution oxygen isotope records from deep sea cores show a continuous trend in the oscillatory record from minimum ice volume in the early Wisconsinan to maximum ice volume in the late Wisconsinan (Shackleton and others,1983; Martinson and others,1987). Sea level changes derived from the oxygen isotope records (Shackleton 1987) and raised coral reef studies (Bloom and others, 1974) show sea level lowering to be greatest in the late Wisconsinan.

Sedimentology and paleontology of the classic Scarborough Bluffs section strongly suggests no ice advanced south of Lake Ontario until the late Wisconsinan (Eyles and Eyles 1983; Eyles 1987; Westgate and others, 1987). Many sites in Illinois and Ohio with paleosols on supposed mid or early Wisconsinan deposits have recently been reclassified as Illinoian (Kempton, 1985; Totten, 1987). Ongoing work of my own in northeastern Pennsylvania shows supposed early Wisconsinan deposits to be Illinoian. At diverse sites in the western United States, extending from the Yellowstone region to the Sierra Nevada, recent work indicates little or no ice extent in the early to mid Wisconsinan (Atwater and others 1986; Baker 1986, Dorn and others, 1987; McCoy 1987; Oviatt and others,1987).

These studies, taken together, strongly support the contention that glaciers were most extensive in the late Wisconsinan and, by inference, support the view that the terminal moraine at the Genesee divide is late Wisconsinan rather than early or mid Wisconsinan in age.

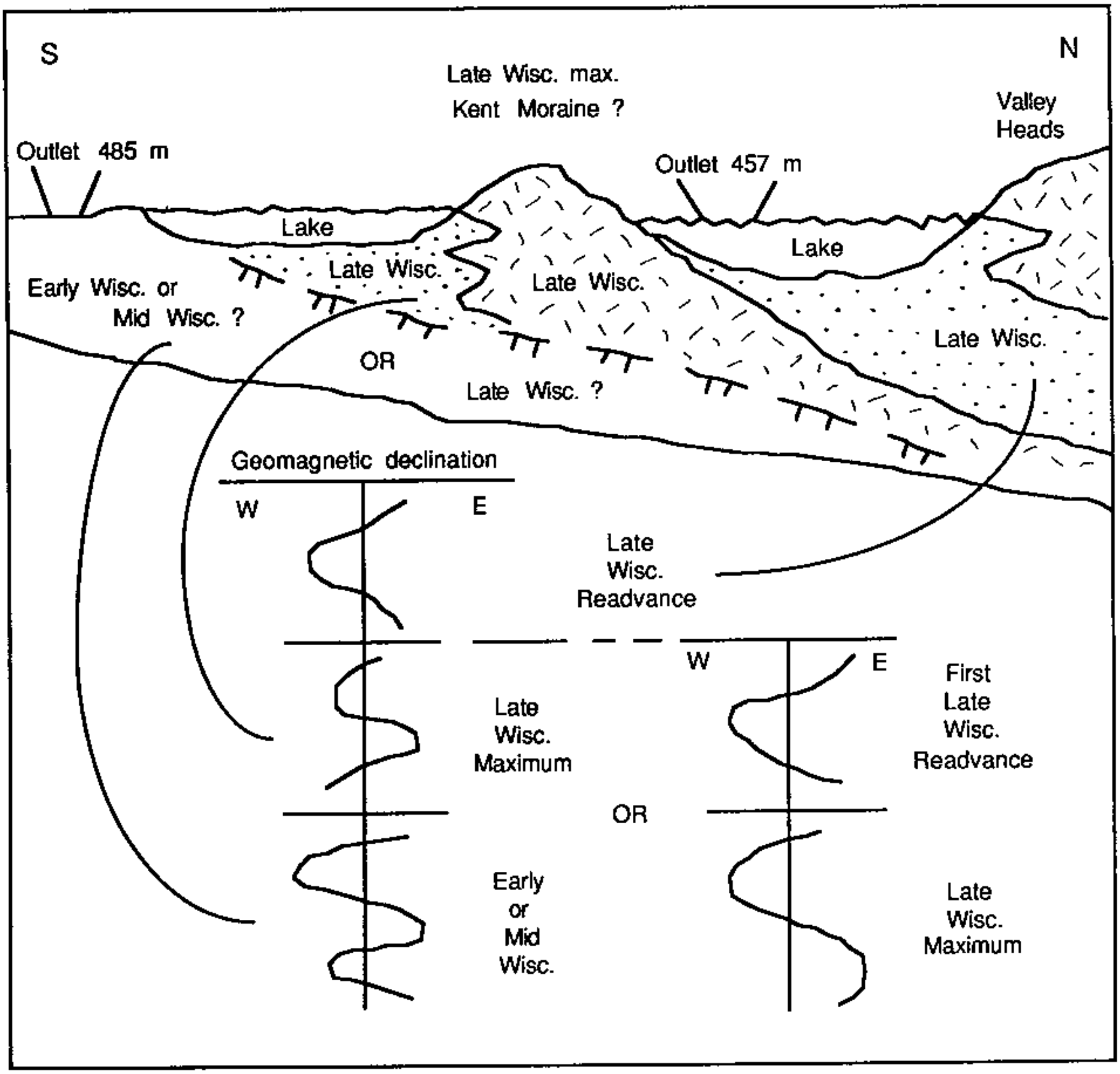



Fig. 4 Schematic cross-section of the Genesee Valley showing two alternative hypotheses concerning the age and geomagnetic declination record of the Wisconsinan glacial deposits.

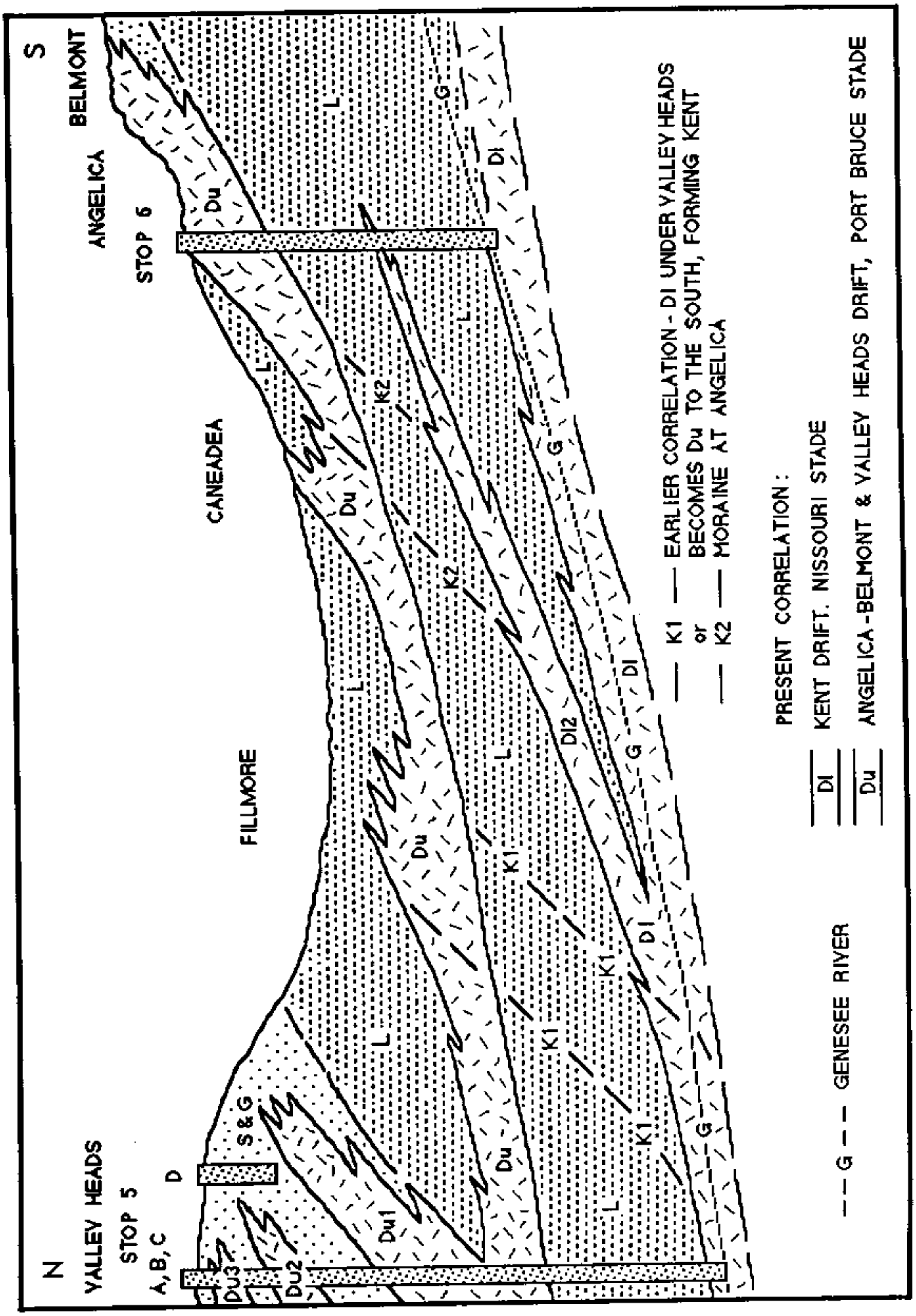



Fig. 5 Schematic cross-section of the Genesee Valley from the Valley Heads to the Angelica-Belmont moraines.

**Tentative correlation of Genesee Valley ice margin positions**

| West of Genesee | Genesee | East of Genesee |
|---|---|---|
| Lake Escarpment or Ashtabula | Valley Heads near Portageville | Valley Heads near Dansville |
| Findley Lake | Rushford | Arkport |
| Clymer | Angelica-Belmont | Almond-Alfred |
| Kent | Terminal Moraine | Terminal Moraine |

**Acknowledgements**

This research has been supported by the Petroleum Research Fund, administered by the American Chemical Society. Willian Brennan of the Geosciences Dept., S.U.C. at Geneseo, assisted the project by providing access to the Spinner Magnetometer and by running many of the geomagnetic samples. Ernest Muller provided the original impetus for the project and has generously provided information from his earlier work in the area. Todd Daniel, Edward Helfrick, Gerald Olenick, and Larry Smith assisted with the field mapping and sampling. Thanks are extended to Robert Graham for permission to examine the extensive exposures on his property at stop 5.

**REFERENCES**

Atwater, B.F., Adam, D.P., Bradbury, J.P., Forester, R.M., Mark, R.K., Lettis, W.R., Fisher, G.R., Gobalet, K.W., and Robinson, S.W.,1986, Afan dam for Tulare Lake, California, and implication for the Wisconsin glacial history of the Sierra Nevada: Geol. Soc. Am. Bull., v. 97, p.97-109.

Baker, R.G., 1986, Sangamonian(?) and Wisconian paleoenvironments in Yellowstone National Park: Geol. Soc. Am. Bull., v. 97, p.717-736.

Bloom, A.L., Broecker, W.S., Chappell, J.M.A., Matthews, R.K., and Mesolella, K.J., 1974, Quaternary sea level fluctuations on a tectonic coast: New $^{230}Th/^{234}U$ dates from the Huon Peninsula, New Guinea: Quat. Res. v.4, p.185-205.

Braun, D.D., Daniel,T.A., Smith, L.B., and Brennan, W.J., 1985, Quaternary geologic and geomagnetic record of the upper Genesee Valley, New York: Geol. Soc. Am. Abs. with Prog., v. 17, p.7.

Chamberlain, T.C., 1883, Preliminary Paper on the Terminal Moraine of the Second Glacial Epoch: U.S. Geol. Sur., 3rd Annual Report, p.291-402.

Connally, G.G., 1964, The Almond Moraine of the western Finger Lakes Region, New York: Ph.D. diss., Geo. Dept., Michigan State Univ., 102p.

Cotter, J.F.P., 1983. The minimum age of the Woodfordian deglaciation of northeastern PA and northwestern N.J.: Ph.D. diss., Lehigh Univ., 151p.

Cotter, J.F.P., Ridge, J.H., Evenson, E.B., Sevon, W.D., Sirkin, L., and Stuckenrath, S., 1985, The Wisconsinan history of the Great Valley, Pennsylvania and New Jersey, and the age of the "Terminal Moraine": in Evenson, E.B. (Ed.), 48th Friends of the Pleistocene Field Conf., p.1-59.

Craft, J.L., 1976, White City sand and gravel pit stop discussion, in Ward, A.N., Jr. (ed.), Bedrock and Glacial Geology of Northwestern Pennsylvania in Crawford, Forest, and Venango Counties, 41st Annual Field Conf. of PA Geologists, p.28-32.

Crowl, G.H. and Sevon, W.D., 1980, Glacial border deposits of late Wisconsinan age in northeastern Pennsylvania: PA Geol. Sur Gen. Geo. Report 71, 68p.

Dorn, R.I., Turrin, B.D., Jull A.J.T., Linick, T.W., and Donahue, D.J., 1987, Radiocarbon and Cation-ratio ages for Rock Varnish on Tioga and TAhoe Morainal Boulders of Pine Creek, Eastern Sierra Nevada, California, and their Paleoclimatic implications: Quat. Res., v.28, p.38-49.

Eyles,C.H. and Eyles,N., 1983, Sedimentation in a large lake: A Reinterpretation of the late Pleistocene stratigraphy at Scarborough Bluffs, Ontario, Canada: Geology, v. 11, p.146-152.

Eyles, N. and Westgate, J.A., 1987, Restricted regional extent of the Laurentide Ice Sheet in the Great Lakes basins during early Wisconsin glaciation: Geology, v. 15, p. 537-540.

Fairchild, H.L., 1896, Glacial Genesee Lakes: Geol. Soc. Am. Bull., v.7, p.423-452.

Kempton, J.P., Berg, R.C., and Follmer, L.R., 1985, Revision of the stratigraphy and nomenclature of glacial deposits in central northern Illinois, in Illinoian and Wisconsinan stratigraphy and environments in northern Illinois: The Altonian revised: Ill. Dept. Energy and Nat. Resources Guidebook 19, p.1-20.

LaFleur, R.G., 1979, Glacial geology and stratigraphy of western New York Nuclear Service Center and vicinity, Cattaraugus and Erie Counties, New York: U.S. Geol. Sur. Open File Report 79-989, 17p.

Lewis, H.C., 1884, Report on the terminal moraine in Pennsylvania and western New York: Pa Geol. Sur., 2nd Ser. Report Z, 299p.

MacClintock, Paul and Apfel, E.T., 1944, Correlation of the drifts of the Salamanca Re-entrant, New York: Geol. Soc. Am. Bull., v.55, p.1143-1164.

Martinson, D.G., Pisias, N.G., Hay, J.D., Imbrie, J., Moore, T.C., and Shackleton, N.J., 1987, Age dating and the orbital theory of the ice aged: Development of a high resolution 0 to 300,000-year vhronostratigraphy: Quat. Res., v.27, p.1-27.

McCoy, W.D., 1987, Quaternary aminostratigraphy of the Bonneville Basin, western United States: Geol. Soc. Am. Bull., v. 98, p.99-112.

Muller, E.H., 1960, Glacial geology of Cattaraugus County, New York: 23rd Friends of the Pleistocene Field Conf., Guidebook.

Muller, E.H., 1965, Quaternary geology of New York: in Wright, H.E. and Frey, D.G. (eds.), Quaternary of the United States: Princeton Univ. Press, 922p.

Muller, E.H., 1975, Pleistocene Geology, in Tesmer, I.H., Geology of Cattaraugus Co.: Buffalo Mus. Nat. Sci. Bull. 27, p.12-20.

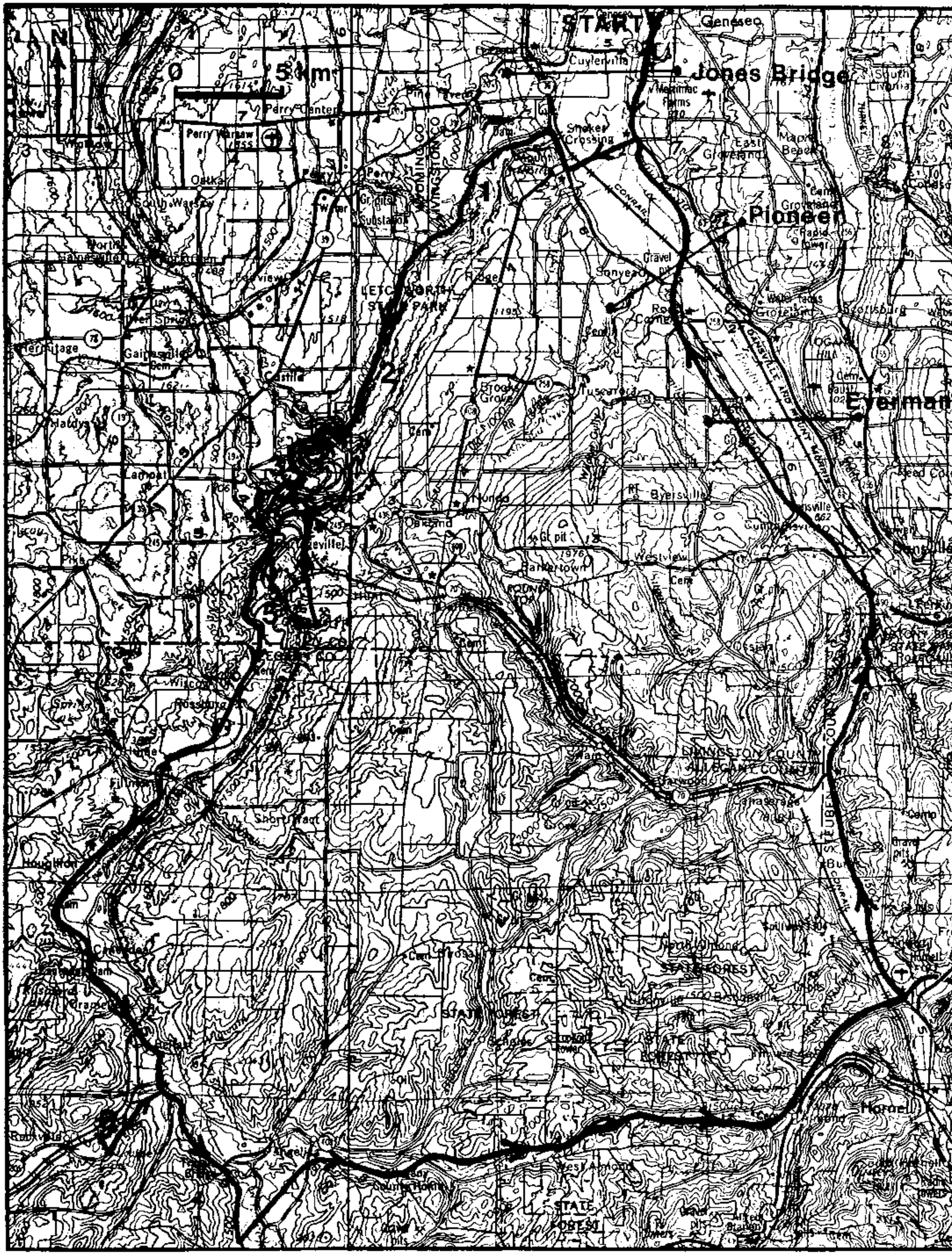
Muller, E.H., 1977, Quaternary Geology of New York, Niagara Sheet: N.Y.S. Museum Map and Folio Ser. 25.

Oviatt, C.G., McCoy, W.D., and Reider,R.G., 1987, Evidence for a shallow early or middle Wisconsin-age lake in the Bonneville basin, Utah: Quat. Res. v.27, p.248-262.

Shackleton, N.J., Oxygen isotopes, ice volume and sea level: Quat. Sci. Rev. v.5, p.183-190.

Totten, S.M., 1987, Stratigraphy of Tills in northern Ohio, in Pre-Woodfordian stratigraphy of north-central Ohio: Midwest Friends of the Pleistocene, 34th Field Conference, p.1-25.

White, G.W., Totten, S.M., and Gross, D.M., 1969, Pleistocene stratigraphy of northwestern PA: PA Geol. Sur., 4th Ser., Gen. Geo. Report 55, 88p.




FIELD TRIP ROUTE MAP - DAY 1

## ROAD LOG - DAY 1 - Saturday, May 28, 1988
## Letchworth Gorge; Valley Heads and Angelic-Belmont moraines

Duane D. Braun, Geog. & Earth Sci.,
Bloomsburg Univ., Bloomsburg, PA 17815

| Cum. Mil. | |
|---|---|
| 0.0 | **START.** Leave parking lot of Onondaga Hall. |
| 0.1 | **TURN RIGHT** onto US 20A (south). Genesee valley floor to right (west) |
| 0.7 | **CONTINUE STRAIGHT AHEAD** onto Rt. 63 (south). US 20A turns to right. |
| 1.9 | Cross over I-390 |
| 3.6 | **BEAR RIGHT** onto Rt. 408 (southwest). Cross under I-390 and continue on Rt. 408 across Genesee Valley floor towards Mt. Morris. Portage Escarpment directly ahead and to left. |
| 5.4 | Enter Mt. Morris |
| 5.8 | **TURN RIGHT** onto Rt. 36 (north). |
| 6.7 | Cross Genesee River. To the left is the downstream end of the Mt. Morris reach of the Genesee post-glacial bedrock gorge. |
| 7.0 | **TURN LEFT** (south) onto entrance road to Letchworth State Park. Start climb up Portage escarpment. |
| 7.4 | Entrance gate to Letchworth State Park. |
| 7.7 | On the right are sand and gravel pits in a sequence of hanging deltas built into glacial Lake Warren. |
| 9.9 | **TURN LEFT** into parking lot beside Genesee gorge. |

**STOP 1. VIEWPOINT OF INCISED MEANDER BEND OR GOOSENECK IN THE POSTGLACIAL BEDROCK GORGE OF THE MT. MORRIS REACH.**
**Leader: Dick Young**

Photos 1-4

Bedrock is exposed nearly to the top of the bluffs at a higher elevation than the next stop upstream at the St. Helena drift-filled reach. The upper part of the bedrock wall and the top of the gooseneck are composed of the Rhinestreet black shale while the lower part of the gorge wall is composed of the Cashaqua shale. The area below is within the flood storage pool of the Mt. Morris dam and may be partly submerged depending upon the amount of rainfall that has fallen prior to the trip.

Leave parking lot by **TURNING LEFT** onto the Park access road to continue upstream (south).

14.5 **TURN LEFT** into pullover for Gardeau overlook.

**WE WILL REMAIN IN THE BUSES**

## STOP 2. ST. HELENA OPEN OR DRIFT-FILLED REACH

Viewpoint of broad valley eroded out of the drift-fill. Extensive exposure of diamictons, sands, rhythmites, and gravels occur in the center of the valley. The Gardeau shale is exposed on the valley side walls. The key question is whether this is a preglacial Genesee course, an interglacial Genesee course, or an "over deepened" closed trough eroded as ice climbed the escarpment and flow was concentrated in a shallow pre-existing tributary valley.

Leave pullover, **TURNING LEFT** to continue upstream (south) on the Park access road.

21.1 **TURN LEFT** into overlook parking lot.

## STOP 3. "GRAND CANYON" OF NEW YORK STATE
**Leader: Dick Young**

Photos 5-8

Deepest portion of the Letchworth bedrock gorge, 500 ft. high vertical wall exposing Gardeau shale (lower 200 ft.), West Hill shale (middle 100 ft.), and Nunda sandstone and shale (upper 200 ft.).

Leave parking lot, **TURNING LEFT** onto Park access road to continue upstream (south).

22.8 Pass Glen Iris Inn on the left.

23.1 **TURN LEFT** into road to Middle Falls parking lot. Upper Falls is to the right, upstream, and is capped by Nunda sandstone.

23.6 Park along right side of the road at Middle Falls.

## STOP 4. MIDDLE FALLS OF THE GENESEE.
**Leader: Dick Young**

Photos 8-12

The 50 ft. high falls is capped by a sandstone unit within the West Hill shale. Across the valley (east) and above the bedrock gorge are slumps in a 200 ft. thick section of proglacial lake sands and rhythmites. James Hall described the section in the 1840's, noting wood fragments in the rhythmites. Recent attempts to find wood fragments have not been sucessfull. In the 1840,s, an attempt was made to bring the Genesee Valley canal across the slumping slope, an effort that ended in failure for the whole canal project.

Return to Park access road.

23.8 **TURN LEFT** to continue upstream (south). Go past Glen Iris Inn again and road to Middle Falls.

24.6 Cross under Conrail bridge over the gorge. The upper falls is downstream to the left.

25.2 Exit from Letchworth State Park. **TURN LEFT** onto Rt. 19A (south). View to the left is the upstream end of the Portageville reach of the Letchworth bedrock gorge. From this point southward, the Genesee has cut a broad floodplain in diamictons, sand & gravel, sand, and rhythmites that partly fill the preglacial Genesee valley. All materials within the valley have been deposited in a lacustrine environment.

25.5 **BEAR RIGHT,** continuing on Rt. 19A through Portageville (south).

26.2 **TURN LEFT** and cross bridge over Genesee River.

26.3 **TURN RIGHT** (south) at end of bridge onto dirt road.

26.6 **TURN RIGHT AT T** onto narrow dirt road, and immediately **BEAR LEFT** passing a hog barn on the right.

27.1 **TURN LEFT AT T,** driving uphill and then bearing right across a high terrace. To left is an abandoned meander scar with a wooded meander core or spur in the center of it.

27.3 **Dead end road** overlooking Genesee River. Road has been cut off by slumping triggered by undercutting of the slope by the Genesee River.

**STOP 5: Cross-section through the Valley Heads Moraines - Portageville or Graham Farm section.**
**Leaders: Duane Braun and Ernie Muller.**

**TWO OPTIONS:**

1. **HIKE - 100 m rise on as much as a 30$^0$ slope & 2.75 km length -** Examine 5A and then hike upslope across 5B; lunch at overlook at top of 5B. During lunch there will be a short discussion of the Genesee Valley morphology and recent channel changes of the Genesee River that are visible from the overlook (weather permitting). Then hike to 5C where there will be another discussion and on to 5D where the buses will be waiting.

Photos 12-17

2. **BUS -** Examine 5A and lower part of 5B. Return to buses for lunch. Then take buses to 5D and rejoin the "hikers".

**Basic questions:**

1. Does the river level diamicton (D1, Fig. 1; Unit 1, Fig. 2) represent the late Wisconsinan Nissouri stade recession or a younger event ?

2. Where is the Erie interstade hiatus in the section ?

3. All sediments have been deposited in a proglacial lake. What is the sedimentologic evidence for lacustrine deposition, especially within the diamicton and the sand & gravel units ?

4. Are the diamictons resedimented "till" (for the most part sediment flows from nearby ice) or are they true till deposited directly from the ice and if so, are they lodgement or melt-out till ?

5. Has slumping repeated or confused the stratigraphy on the face of the bluff (middle part of 5B) ?

**SITE 5A description:**

To reach site 5A, walk downstream along the edge of the terrace following the white blaze and/or pink ribbon trail down to river level. Return to where buses are by retracing the trail down or follow a rough pink ribbon trail that slants upward across the face of the outcrop.

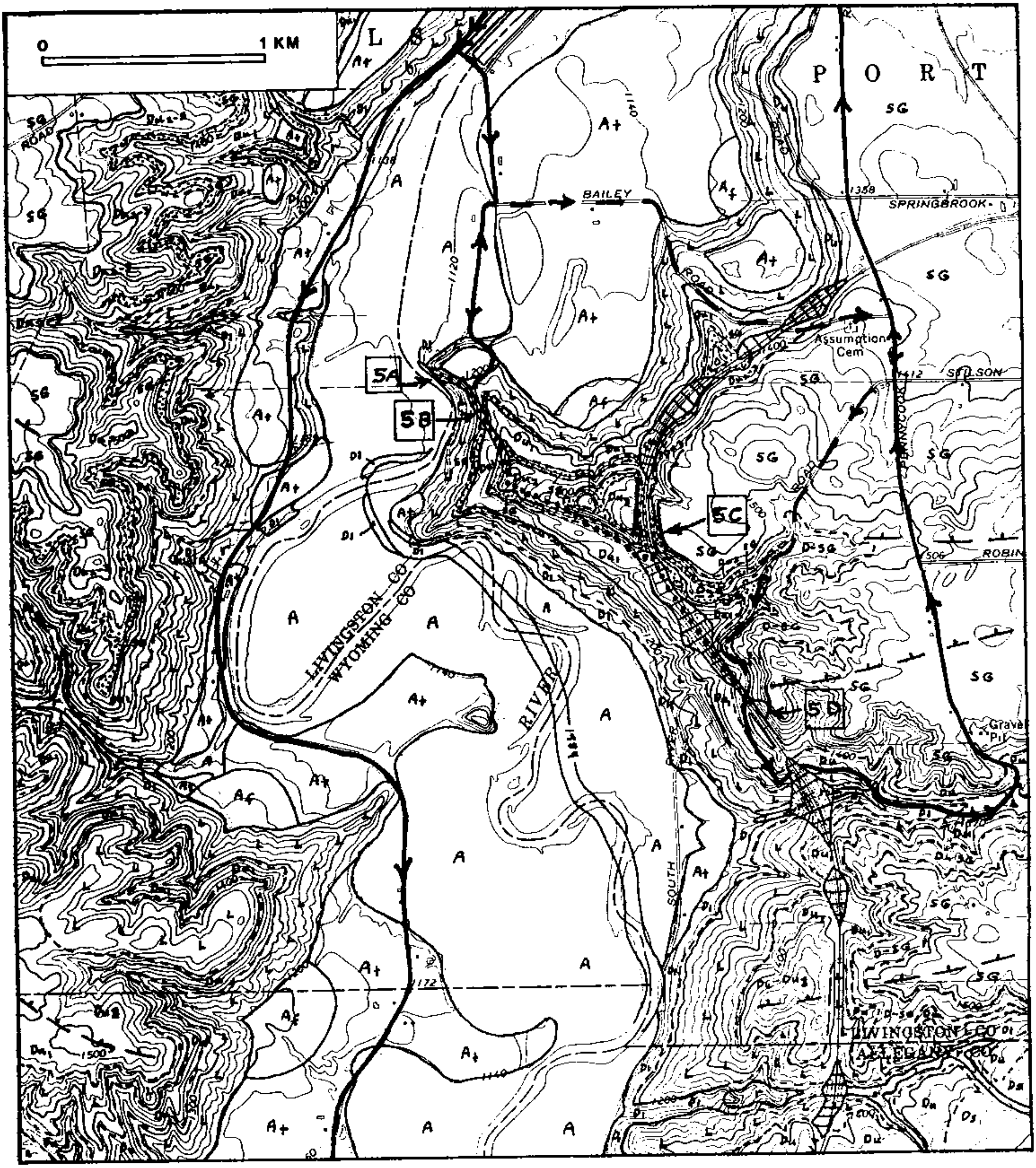
At river level a diamicton is exposed (Fig. 1, D1; Fig. 2, unit 1)(river stage permitting) that has been traced southward in the Genesee Valley (Fig. 6, Terminal Moraine to Valley Heads Moraine text section) underneath and beyond the Angelica-Belmont moraine. This diamicton has also been traced up tributary valleys where it becomes the basal diamicton in contact with bedrock (Fig. 1) and is enriched in local angular bedrock clasts (changes from "bright" to "drab" drift). Along the middle of the Genesee Valley, the diamicton is underlain below river level by 30 to 100 m of interbedded rhythmites, sand, sand & gravel, and diamicton of unknown age. The river level diamicton is interpreted to represent the late Wisconsinan Nissouri stade recession.

The diamicton is overlain by silt-clay rhythmites (Fig. 2, unit 2) that are draped over irregularities on the diamicton surface. There is a transitional zone between the two units where thin diamicton layers (debris bands) interbed with silt-clay layers containing abundant dropstones. Soft sediment deformation structures are common. Above the transition zone, dropstones become rare and animal tracks and burrows are common on some bedding planes in the rhythmites. A massive to planar bedded to rippled sand overlies the rhythmites (Fig.2, unit 3). This in turn is overlain, on the eastern (upstream) end of site 5A, by silt-clay rhythmites (Fig. 2, unit 4) that form the base of site 5B. At the sand-rhythmite contact, the base of the rhythmites is slickensided and shows movement of the rhythmites towards the center of the valley. On the western (downstream) end and middle of site 5A, the sand of unit 3 is overlain by sand & gravel of a Genesee River terrace. Within the terrace sediments are blocks, up to several meters across, of tilted silt-clay rhythmites that slid into the channel when it was forming the terrace and undercutting the rhythmites of unit 4.

**SITE 5B description:**

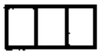
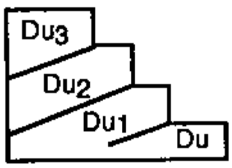
Site 5B starts at an abandoned township road that was cut off by slumping and extends upslope along the headwall scarp of the large slump that involves the whole 85 m high and 500 m long face of the bluff that rises above the river (Fig. 1). Follow the white blaze and/or pink ribbon trail upslope.

The section starts in silt-clay rhythmites (Fig. 2, unit 4) that exhibit exceptionally thick clay bands, up to several cm thick, that are often thicker then adjacent silt bands. A short covered interval separates the rhythmites from the overlying interbedded sands, gravels, and occasional silts of unit 5

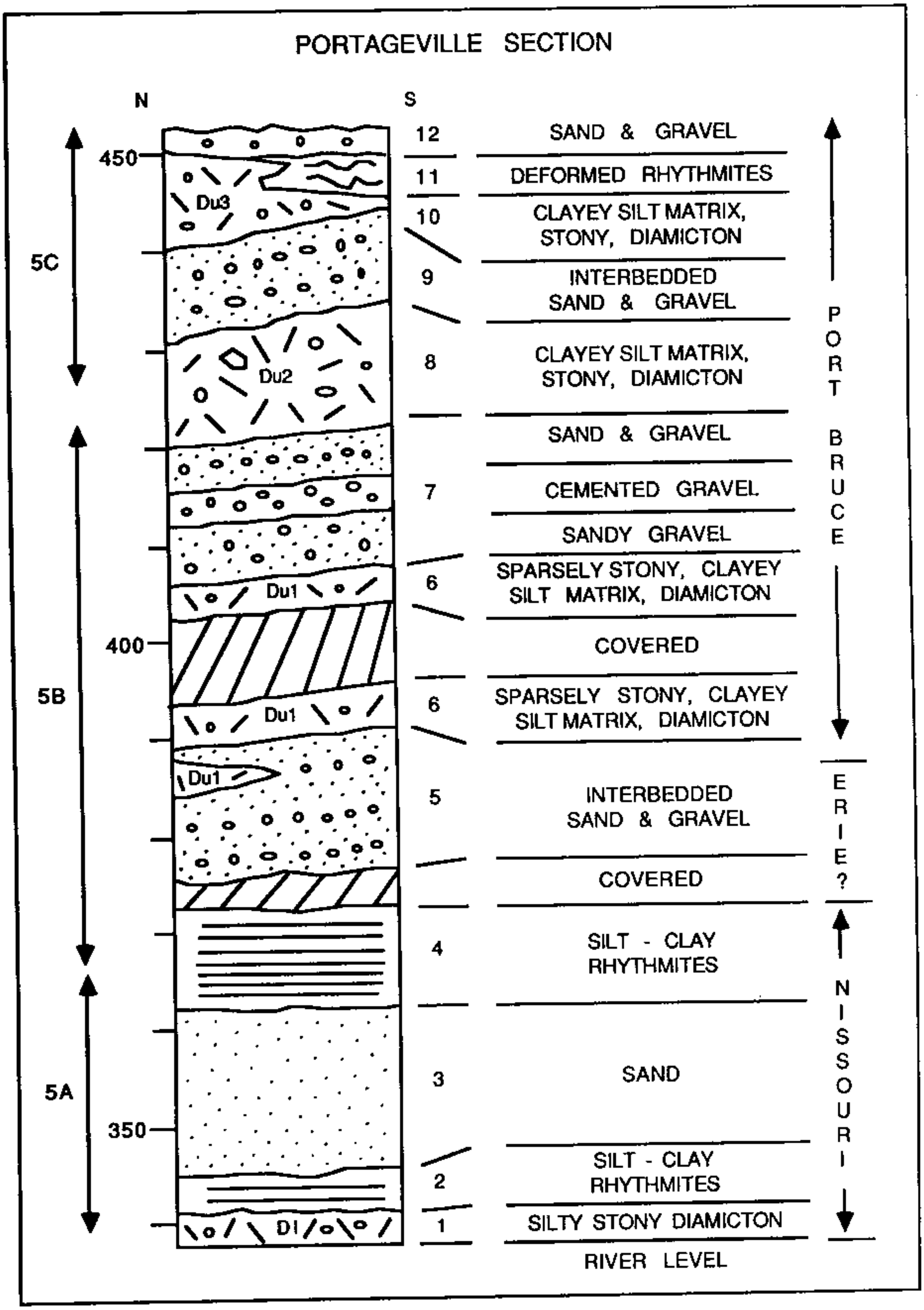



STOP 5, Fig. 1 Glacial deposit and field trip route map of the area around stop 5. Dotted line marks hiking route. (Portageville 7 1/2' quad.)

## LEGEND

| | |
|---|---|
|  | RAILROAD EMBANKMENTS |
| A | ALLUVIUM OF ACTIVE FLOODPLAIN |
| $A_t$ | ALLUVIAL TERRACES -- 6+m ABOVE RIVER |
| SG | SAND & GRAVEL -- CAPS VALLEY HEADS; SEPARATES UPPER DIAMICTON UNITS Du1, Du2, & Du3 |
| D-SG | INTERTONGUED DIAMICTON AND SAND & GRAVEL; TRANSITIONAL UNIT AT THE HEAD OF OUTWASH OF EACH OF THE VALLEY HEADS MARGINS |
|   | UPPER DIAMICTON SEQUENCE, THREE SEPARATE TONGUES (Du1, Du2, & Du3) UNDER THE VALLEY HEADS MARGINS; SINGLE UNIT (Du) CONTINUES SOUTH |
| L | LACUSTRINE SEQUENCE; SAND, SILT-CLAY RHYTHMITES, AND SANDY GRAVEL |
| Dl | LOWER DIAMICTON, STONY, CLAYEY - SILT MATRIX |

**STOP 5,** Fig. 1A Map legend for Stop 5, Fig. 1.




**STOP 5,** Fig. 2 Stratigraphic section through the Valley Heads moraines. Also shown on Fig. 5 in the Terminal to Valley Heads moraine text section.

(Fig. 2). Unit 5 or the covered zone under it is the most likely interval to place the Erie Interstade hiatus, but no evidence for such a hiatus has yet been observed.

The upper part of the sand & gravel of unit 5 is interbedded with the massive, sparsely stony, clayey silt to silty clay matrix diamicton of unit 6 (Fig. 2). This diamicton is relatively clay rich, probably from incorporation of underlying rhythmites, and is texturally similar to what are considered Port Bruce readvance diamictons in the Erie basin. The key reason to assign this diamicton to the Port Bruce readvance is that it forms the core of and intertongues with the head of outwash sand & gravel of the outermost Valley Heads Moraine (Fig. 1, Du1 area in southeast corner of map). This diamicton is traceable southward along the flanks of the Genesee Valley to the Angelica-Belmont moraine and thereby also indicates a Port Bruce age for the Angelica-Belmont moraine (Fig. 6, Terminal Moraine to Valley Heads Moraine text section).

This massive clayey diamicton intertongues with the overlying sand & gravel of unit 7 (Fig. 2). A pebble to cobble gravel interval within unit 7 is cemented by calcium carbonate. The sands & gravels are overlain by diamicton (Fig. 2, unit 8; Fig. 1, Du2) that caps the top of the bluff at 5B. This diamicton slopes upward to the east to become the basal diamicton at 5C. The diamicton also slopes upward to the south until just north of 5D (Fig. 6, Terminal Moraine to Valley Heads Moraine text section) where it forms the core of the middle of the three Valley Heads Moraine loops (Fig. 1).

**LUNCH at top of bluff overlooking the Genesee Valley**

Discussion over lunch of:

1. Stratigraphic sequence observed at 5A & 5B.

2. Overall Genesee Valley morphology. The view southward from the overlook shows a prominant shoulder on the flank of the Genesee Valley that marks the lake floor elevation at cessation of glaciation. The nearest part of the shoulder is the outermost Valley Heads Moraine. The Genesee River has incised the glacial sediments to form the present "inner" Genesee Valley.

3. The Genesee River during the 1972 hurricane Agnes flooding cutoff a sequence of four meander loops south of the overlook. The most prominent one is immediately southwest of the overlook (Fig. 1). Since 1972, the river has developed a new series of bends that are rapidly migrating today.

Follow white blaze and/or pink ribbon marked trail to 5C and gather there temporarily for short orientation talk.

**SITE 5C description:**

An abandoned 30 m deep and 400 m long railroad cut provides a complete exposure of the upper part of the sequence (Fig. 2, units 8 - 12). All units slant upward from north to south across the face of the cut. Diamicton unit Du2 (Fig. 1; Fig. 2, unit 8) starts below roadbed level at the north end of the cut and rises well above roadbed level at the south end of the cut (where we will first enter the cut). Diamicton Du2 is overlain by sand & gravel (Fig. 2, unit 9) that is in turn overlain by the uppermost diamicton in the sequence (Fig. 1, Du3; Fig.2, unit 10). This diamicton forms the core of the inner most Valley Heads Moraine, the crest of which crosses the northern end of the railroad cut (Fig. 1). Deformed rhythmites of unit 11 record the oscillatory nature of the ice front during deposition of the upper most diamicton. The section is capped by sand & gravel (Fig. 2, unit 12) deposited during final recession of the ice from the Valley Heads position.

Follow railroad bed .75 km to the south (upstream) to where a township road embankment crosses the railroad bed. Site 5D is the sand & gravel quarry on the left (east) beyond and above the road embankment.

**SITE 5D description:**

The sand & gravel deposits are the head of outwash of the middle of the three Valley Heads Moraine positions (Fig. 1). The lower part of these deposits are contemporaneous with the deposition of diamicton Du2 (Fig. 1; Fig 2; Fig. 2, unit 8). The upper deposits may be contemporaneous with units 10 - 12 at site 5C. South dipping foreset beds dominate the exposure. The Lake Belfast-Fillmore level of about 457 m (1500 ft) would place the foreset-topset contact near the top of the 463 m (1520 ft) slope that rises above the pit.

- - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - -

**Road Route from STOP 5A to STOP 5D.**

27.3 **TURN AROUND.**

27.5 **TURN RIGHT AT T**

27.9 **BEAR RIGHT** (actually straight ahead) (east) onto Bailey Road.

29.2 **TURN RIGHT AT T** (south) onto Pennycook Road.

29.4 **TURN RIGHT** (east) onto Stilson Road.

30.5 **STOP 5D** Stilson Road crosses old railroad bed. Park along the side of the road. Stop 5D is the sand and gravel pit immediately to the east and above the railroad grade.

- - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - - -

Continue ahead (south) on Stilson Road.

30.7 **TURN LEFT AT T** onto Pennycook Road.
Pass through tunnel under railroad embankment. Deep tributary valley to right is bedrock floored. Stony diamicton overlying the bedrock has been traced downstream to the elevation of similar exposures of diamicton in the Genesee River bank, and is correlated with the river level till at STOP 5A. The valley is between the outermost (most southerly) Valley Head margin and next one to the north. The road curves to left uphill passing successively across sand, diamicton, and the sand and gravel cap of the second Valley Head margin. To the right are abandoned gravel pits.

32.2 **STRAIGHT AHEAD** on Pennycook Road, passing Robinson Road on the right and crest of third or inner-most Valley Heads margin, the one capping the railroad cut at STOP 5C.

32.7 **CONTINUE STRAIGHT AHEAD** AT Stilson Road intersection, remaining on Pennycook Road.

32.9 Pass Bailey Road on the left.

33.0 Cross old railroad grade. Now on 1360 ft. (415 m) outwash surface cut by ice marginal and Genesee drainage. As one continues north, the surface steps down to 1340 ft. (409 m) and 1320 ft. (402 m). These outwash surfaces slope eastward and are graded to the Swains Col. The surfaces are part of the connecting channel between moraine dammed Lake Portageville in front (south) of the Valley Heads Margin and proglacial Lake Nunda in back (north) of the same margin. Upon ice recession from the Valley Heads margin, the Genesee drained east through Swains Col then southeast into the Canisteo River, part of the Susquehanna drainage system.

33.2 **CONTINUE STRAIGHT AHEAD** at intersection. Now crossing 1340 ft. surface. This is one of the more likely areas to place a preglacial course of the Genesee, going from left to right (east) towards the town of Nunda.

34.5 Cross Conrail Tracks.

34.8 **TURN LEFT AT T** (west) onto Rt. 436. For next 1/2 mi (.7 km) Rt 436 runs on a narrow ridgeline of glacial material that filled the preglacial Genesee Valley, preserved as a spur within a large incised loop in Letchworth gorge. Test drilling at the turn of the century for a proposed hydroelectric project showed no rock to be present to a depth of 300+ ft. (100 m). The top of this ridgeline also represents the south side of the Gowanda moraine.

36.0 Cross bridge over Genesee River. To left (east) is the broad partly excavated preglacial Genesee Valley. The bedrock ledge crossing the river bed marks the boundary between the glacial fill and the bedrock valley wall of the preglacial Genesee Valley. To the right (west) bedrock rises progressively higher along the sides of the river as it enters the head of the Letchworth bedrock gorge.

36.1 **BEAR RIGHT** at intersection in Portageville, continuing on Rt. 436.

36.4 **BEAR LEFT** at intersection onto Rt. 19A (south)

37.1 Pass earlier turnoff to STOP 5, continue on Rt.19A.

38.3 To left is a view across the Genesee Valley to STOP 5A and 5B.

38.7 On the right is an old Bluestone or flagstone quarry with extensive slumps in lacustrine sediments upslope of the quarry. On the left is the large abandoned meander loop seen from STOP 5B.

40.1 On the left across the valley is STOP 5D at the top of the bluff. We are now passing outside (south) of the outermost Valley Heads margin.

40.3 to 40.9 On both sides of the road are a series of unpaired terraces, often containing oxbow lakes, from progressive downcutting of the river. To the right, the bluffs rising above the terraces are composed predominately of lacustrine sands and rhythmites.

41.9 Cross Wiscoy Creek.

42.3 To the right and behind us (northwest), a large multiple diamicton exposure is visible.

43.7 More unpaired terraces. To the right (west) are badlands cut in lacustrine sediments. Forest cover obscures how bad it is until one tries to walk the area.

45.2 Enter Fillmore. To the left but not visible from the road, Rush Creek enters the Genesee valley from the east. It is the site of 25,000 yr. wood date from basal diamicton overlying bedrock.

45.4 **CONTINUE STRAIGHT AHEAD** at intersection in Fillmore. Rt. 19 comes in from the right. Rt. 19A becomes Rt. 19 south.

45.5 Cross Cold Creek. Several multiple diamicton exposures are upstream but are difficult to get to.

48.5 More unpaired terraces.

48.9 Enter and continue through Houghton. Wells at Houghton penetrate 150-200 ft. of glacial sediments below river level, producing 300 gpm from gravels at depth.

51.7 To the left across the valley is a large exposure. It is called Crystal Bluffs due to gypsum rosettes that have crystallized in the lacustrine and diamicton units and are now weathering out of the exposure.

52.1 Enter Caneadea. To the left across valley is the large Caneadea bluff exposure but it is difficult to see from road and to reach on foot.

52.3 **CONTINUE STRAIGHT** on Rt 19. Intersection with Rt. 243 on right. West one and one half miles on Rt 243 is Rushford Lake, originally built to supply water to the Genesee Valley Canal and now a recreational lake. The nearly 100 ft (30 m) high dam crosses a narrow bedrock gorge cut into the south side of the preglacial Caneadea Creek Valley. The Rushford ice margin forms a leaky natural dam across the rest of the valley. Spring discharge of several cubic feet per second issues from a gulley in the downstream side of the natural dam which, below lake level, is composed of lacustrine sands. While the water is clear, it will be interesting to see if discharge through the dam increases and becomes sediment laden in the future, indicating incipient piping failure.

52.7 Cross Caneadea Creek. This is where the Rushford margin crosses the Genesee valley. It is probably equivalent to the Arkport margin to the east in the Canisteo Valley.

53.9 Enter and continue through Oramel on Rt. 19.

56.0 Enter Belfast.

57.2 Cross Black Creek.

57.4 **TURN RIGHT** onto Rt. 305 (west). Entrance to Black Creek Valley, outlet for glacial lake Belfast-Fillmore at 1500 ft (457 m).

60.3 **TURN RIGHT** onto Lake Road. To left (west) the Angelica-Belmont margin crosses Black Creek Valley.

61.1 After passing a lake on the right, **TURN RIGHT** into a dead end road.

61.2 **DEAD END** at Black Creek. Park and walk down right bank of Black Creek to exposure a short distance downstream.

**STOP 6: Cross-section through the Angelica-Belmont Moraine - Black Creek section.**
**Leader: Duane Braun**

This is one of the more accessible and complete sections through the Angelica-Belmont Moraine. The site is about one km inside the outer edge of the moraine.

**Basic questions:**

Photos 18-21

1. If the Angelica-Belmont Moraine is equivalent to the Kent Moraine and marks the maximum extent of late Wisconsinan ice, where is the mid Wisconsinan hiatus in the section underneath the upper most diamicton that makes up the moraine ?

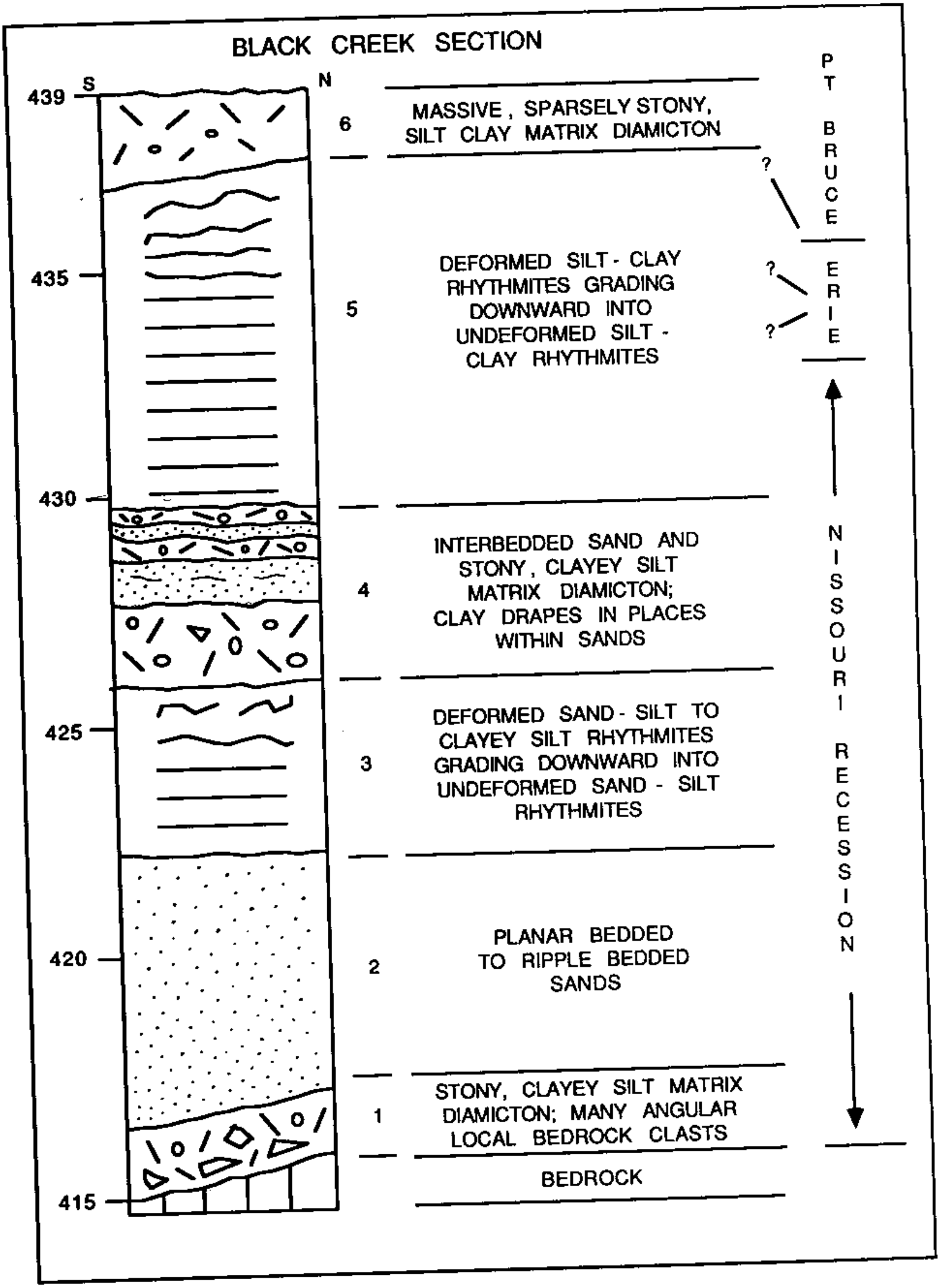
2. Are the diamictons resedimented "till", especially the upper most diamicton, or are they true till deposited directly from the ice, especially the basal diamicton in contact with bedrock ?

3. What parts of the rhythmite units are true seasonal varves and what parts are shorter term sediment influx events ?

**Site 6 description:**

Upstream of the road, bedrock is exposed in the bed of Black Creek. A very stony diamicton (Fig. 1, unit 1), dominated by angular local bedrock clasts, overlies the bedrock. The diamicton slopes downstream, passing below stream level and the main part of the exposure on the downstream side of the bridge.

The diamicton is overlain by sands (Fig.1, unit 2) that are in turn overlain by sand-silt rhythmites (Fig. 1, unit 3) that become more clayey up-section. These rhythmites are deformed immediately underneath the next overlying unit




**STOP 6,** Fig. 1 Stratigraphic section through the Angelica-Belmont moraine. Also shown on Fig. 5 in the Terminal to Valley Heads moraine text section.

(Fig.1, unit 4) composed of stony diamicton layers interbedded with sands containing clay drapes. Silt-clay rhythmite layers (Fig. 1,unit 5) overlie and are draped across the uppermost diamicton layer of unit four. Debris bands and dropstones are common in the lower meter or so of the rhythmites. In places, the rhythmites exhibit laminated bedding with as many as 5 silt-clay laminae per cm.

Up-section, the rhythmites of unit five become increasingly deformed as the base of the upper most massive clayey diamicton (Fig. 1, unit 6) is approached. Its texture is like that of the main diamicton mass that forms the morainic loop across the Genesee Valley near Angelica. (We will cross that feature on our return to Geneseo.)

If the upper diamicton represents the maximum extent of late Wisconsinan ice, a mid Wisconsinan weathering profile should be present somewhere in the section underneath the diamicton. I can't find it, maybe you can. It is doubtful that the weathering profile has been eroded away by the advance that deposited the upper diamicton since we are only one km inside the edge of the advance. The ice would have been thin and would have been sliding across saturated clay rich lake sediments. The present age interpretation (Fig. 1, right side; Fig. 6, Terminal Moraine to Valley Heads text section) is that the basal diamicton (unit 1) and the middle diamicton (unit 4) represent late Wisconsinan Nissouri stade recessional events. The upper diamicton represents the maximum extent of the Port Bruce stade readvance.

61.2 **TURN AROUND** and retrace route to Rt. 305.

61.9 **TURN LEFT** (east) onto Rt. 305

64.8 **TURN RIGHT** (south) onto Rt. 19.

66.1 To the left (east) across the Genesee Valley, kames mark the Angelica-Belmont margin.

67.9 Passing bridge over Genesee on the left, shale outcrops appear on the right. The Genesee is presently against the west valley wall of the preglacial valley.

70.5 Approaching Rt. 17 intersection. The broad Van Campen Creek valley on the right (west) lacks evidence of ever functioning as a lake outlet. On left (east) the Angelica moraine loops across Genesee Valley.

70.6 Pass over 4-lane divided highway, Rt. 17.

70.7 **TURN LEFT** onto East-bound entrance ramp for Rt. 17.

71.1 Cross Genesee River. Angelica moraine directly ahead. To the right, a distinct bench along the east wall of the Genesee Valley is an ice margin position that marks the extension of the Angelica margin southward 3 mi (4 km) to Belmont. The Belmont margin marks the maximum readvance of the ice that reformed glacial Lake Wellsville while the Angelica margin marks a slightly recessional position.

72.4 Crossing the surface of the Angelica moraine. The surface is dissected by several small streams that start on the crest of the moraine to the right (south) of Rt. 17. The moraine is composed of 20 to 80 ft (6 - 24 m) of massive, sparsely stony diamicton overlying deformed clayey rhythmites.

73.8 Rt. 17 curves to the right (southeast), passing under a side road and outside of the Angelica margin. To the left (north), Angelica Creek enters a bedrock gorge where the mass of the Angelica moraine blocks the preglacial Angelica Valley. Town of Angelica on the left(north). Rt. 17 will now go eastward up the Angelica Creek Valley.

78.6 To the right, a notch in ridgeline is the 1830 ft (558 m) sluice from this valley to the next valley to the south.

82.8 Crossing drainage divide between Genesee and Canisteo systems (St. Lawrence - Susquehanna divide). Now traveling east in the valley of Karr Valley Creek.

84.8 Directly ahead the Almond moraine (Connally, 1964) blocks the mouth of Karr Valley Creek.

87.8 On the left (north) is a large sand and gravel pit in Almond moraine.

88.4 On the right (south) is the bedrock gorge of Karr Valley Creek where it has been diverted by the mass of the Almond margin (locally called Sand Hill).

88.8 Curve left and enter the Canacadea Valley. To the right (south) and upstream 1.7 mi (2 km) is the Alfred Station margin. The close juxtaposition of these two margins suggests a correlation with the closely paired Belmont-Angelica margins.

90.6 To the right (south), is a flood control reservoir (Almond Lake). It is placed at entrance to the Canacadea Creek bedrock gorge. The Canacadea Creek has been diverted through the ridge by ice margin blockage where the creek originally entered the main Canisteo Valley (straight ahead to east). The large bedrock knob to the left (east) of the dam is an umlafburg. The size and depth of the bedrock channel and the low elevation of the ice marginal deposits in the entrance to the channel suggest that the bedrock gorge was cut to near its present depth by pre-Wisconsinan ice marginal drainage.

92.0 To the right (south) is knob and kettle landscape par excellence.

93.4 Enter main through valley of the Canisteo River. Arkport margin kames on either side of the valley on left (north). The valley floor outwash plain is graded to the Valley Heads margin 7 miles (10 km) to the north (left).

94.2 **TURN RIGHT** onto the exit ramp for Rt. 36 North.

94.8 **TURN LEFT** onto Rt. 36. Head up valley (north). As with other through valleys to the east of here in the Finger Lakes district, the present St. Lawrence-Susquehanna divide is the Valley Heads margin. In this through valley the original divide was 20 (32 km) south of the Valley Heads margin and is now marked by a 600 ft (183 m) deep sluice-way cut into rock.

101.8 Rt. 70 enters Rt. 36 from the left (west). Immediately north of the intersection is the outermost Valley Heads margin (head of outwash). This is also the outlet for Glacial Lake Dansville, the proglacial lake that developed as the ice retreated from the Valley Heads position. For the next 6 miles (10 km) Rt. 36 follows along side and then descends across the Valley Heads margin into the broad glacial trough of Canaseraga Creek.

106.4 On the right are large exposures of glacial deposits within Stony Brook State Park. Out of sight upstream is a deep bedrock gorge.

107.7 Cross under I-390

108.2 **TURN LEFT** onto the entrance ramp for I-390 north. I-390 runs along the west side of the Canaseraga trough. The trough held, at a water surface elevation 1220, the 650 ft. (200 m) deep glacial Lake Dansville as the ice receded northward.

114.2 To the right (east), Everman Road crosses the floor of the Canaseraga trough. A gravity profile and well data indicate 625 ft. (190 m) of glacial fill lies under the floor of the valley at this site.

118.6 I-390 bears right (northeast) and crosses the Canaseraga trough. When the ice front was at this point, glacial Lake Dansville was still in existence and discharging water across the Valley Heads margin. On the left (southwest) is a gravel pit in ice contact stratified drift that must have been deposited in the deep waters of Lake Dansville. On the right Pioneer road crosses the Canaseraga trough. A gravity profile along the road indicates 500 ft. (152 m) of glacial fill underneath the valley floor.

123.7 **TURN RIGHT** onto exit ramp for Rt. 63.

124.2 **TURN RIGHT** onto Rt. 63. To the left (west) Canaseraga Creek is joined by the Genesee River.

126.1 Cross over I-390.

126.2 To the left (west), Jones Bridge Road crosses the Genesee Valley. A gravity profile along the road indicates 500 ft. (152 m) of glacial fill underneath the valley floor.

127.3 **CONTINUE STRAIGHT AHEAD,** Rt 20A joins Rt 63 from the left (west).

128.0 **TURN LEFT** into entrance road to Onondaga Hall.

## GENESEO TO THE PINNACLE HILLS
Ernest H. Muller

### In the vicinity of Geneseo

Geneseo (Fig. 1) is situated on the break in slope where a bedrock bench with discontinous veneer of till and lake clay descends westward more than 200 feet to the flat floor of the partly drift-filled Genesee Valley. The valley floor, at 550 to 570 feet above sea level and extending south of the "Fowlerville plug" (see Canandaigua Moraine, below) for 15 miles up Canaseraga Trough, is the basin of moraine-dammed Lake Geneseo, a former finger lake. Though it has all the appearance and serves the functions of a floodplain, this valley floor is only partly a product of the fluvial processes by which it is now actively evolving.

South of the "Fowlerville plug", the Genesee River meanders freely over an open flat-floored valley, 560 ft. above sea level and 1.5 miles wide. Fan-tailing meander loops with amplitude of .5 mile are particularly well developed for several miles north and south of Geneseo. Several meander necks have been cut off within the past few decades. About a mile SW of the SUNY College campus, an oxbow lake formed by 1975 where a fan-tailed meander had been mapped in 1950. Abutments of the bridge by which NY-39 crosses the Genesee River south of Geneseo were in imminent danger of direct attack if cutoff had not been prevented by the timely placement of rip-rap to protect a narrowing meander neck.

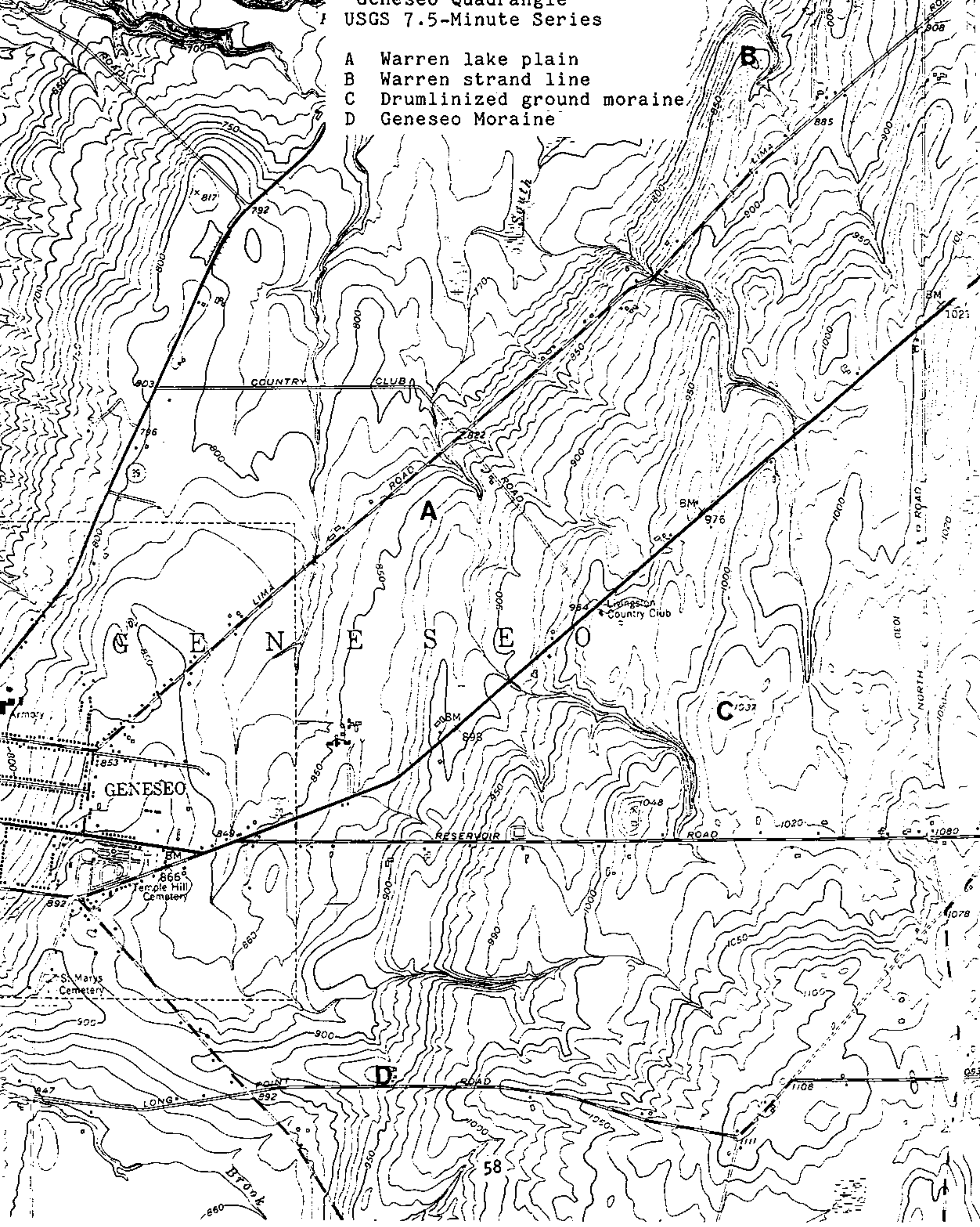
A north-south belt (A, Fig. 1) paralleling the valley floor, and extending 1-2 miles east of Geneseo, ranges in altitude from 800 to 900 feet. Except for the gorges of several small creeks that have cut steeply into the western margin of this belt, slopes are gentle. Relief is low to moderate, and topographic grain is weakly developed parallel to former ice flow. This, too is former lake floor, though lake sediments are too thin and discontinuous to mask the mildly modified glacial topography. Morainal topography is modified and diffuse, reflecting its having been deposited in a lacustrine environment marginal to a grounded ice cliff. With a bit of effort, one can trace the shoreline of Lake Warren at about 875 feet on the slopes a few miles northeast of Geneseo (B, Fig. 1).

Above the Warren strand, the glacially modified plateau northeast of Geneseo, rises only to 1000 - 1100 feet, and is strongly drumlinized with local relief of 100 ft. (C, Fig. 1). In this zone, ice marginal deposition was subaerial and several morainal ridges can be mapped with confidence.

### Geneseo Moraine and Glacial Lake Hall

A low, rounded east-west ridge (D, Fig. 1), followed for some distance by Long Point Road, lies athwart the grain of drumlinized topography on the southern edge of the Village of Geneseo. This is the Geneseo Moraine which dammed glacial Lake Hall in the Canaseraga/Genesee Trough.

Near the east edge of the Geneseo map sheet, the ridge displays characteristic morainal topography. Westward, passing below 1000 feet above sea level, however, the ridge is so smoothed and subtle as to readily escape recognition even though it rises 20 to 50 feet above the topography on either side. The


Geneseo Quadrangle
USGS 7.5-Minute Series
A Warren lake plain
B Warren strand line
C Drumlinized ground moraine
D Geneseo Moraine
A
B
C
D
G E N E S E O
GENESEO
Armory
Temple Hill Cemetery
St. Marys Cemetery
Livingston Country Club
COUNTRY CLUB
RESERVOIR ROAD
LONG POINT ROAD
South
Brook

contrast reflects differences between moraine deposited subaerially and in a proglacial lake. No trace of the moraine is recognized on the valley floor. Lake Hall, ponded in the Genesee Valley south of the Geneseo Moraine, drained west through the Pearl Creek Channel at 1000 feet into Wyoming Valley. Stream load at the west end of Pearl Creek Channel was initially deposited in shallow glacial Lake Oatka with its level controlled by shifting channels across the Oatka-Tonawanda Creek divide south of East Bethany. Beginning above 1000 feet, these channels supplanted one another in quick succession; Lake Oatka was drained and a subaerial fan was built at the mouth of Pearl Creek.

**Canandaigua Moraine and the "Fowlerville Plug"**

Conesus Creek flows north from Lakeville for 2 miles breaching the outer ridge of a compound moraine before being sharply deflected westward by the low but continuous and well-defined Canandaigua Moraine.

Westward the moraine belt separates into two or three broad, low, smoothed swells, arcuate in plan view. The outermost swell curves southwest to form the divide between Conesus Creek and Jaycox Run. At one point the two streams are within 500 feet of each other before diverging to reach the Genesee 5 miles apart.

Below 650 feet, the smoothly rounded swells flatten further as broad fans sloping toward the Genesee River, narrowing the valley floor in places to less than a quarter of a mile. This is the "Fowlerville plug", the expression of ice-marginal deposition in 400-ft. deep water of Lake Hall. West of the river, the changing character of the morainal belt is mirrored in reverse with increasing elevation above the river.

In April, 1973, a large slide mass slumped into the Genesee where Oxbow Lane approaches the river (Young and Rhodes, 1973). Other partially healed slide scars tell of similar occurrences. In these scars are exposed one or more diamict layers intercalated with varved lake sediments. Some diamict units appear to be subaqueous debris flows, but lodgment till comprises most of the exposure in the Oxbow Lane landslide scar. The Fowlerville plug appears to record an unstable ice margin which we have tentatively correlated with the Buffalo and Niagara Falls Moraines.

**Batavia Moraine**

Evidence as to relationships of Lake Warren to the Batavia Moraine and the penetration of Lake Warren into the Genesee Valley have long been troublesome and ambiguous.

In brief, as mapped, both the southern shoreline of Lake Warren III, and the Caledonia meltwater channel system postdate the Batavia Moraine, and neither shows evidence of subsequent glacial overriding west of the Genesee Basin, yet, to account for them as produced during a single episode of ice recession demands an improbable see-sawing of the ice margin. Alternatively, either the mapping of strandline relationships or the southeastward continuation of the Batavia Moraine may be called into question.

In the Erie Basin it has been accepted that low-level eastward-draining Lake Wayne intervened between Lake Warren's high and low levels (Warren I and

Warren III)(Calkin and Feenstra, 1985). The demise of both early and late Warren stages involved withdrawal of the ice sheet from northward salients of the Onondaga Escarpment, either north of Batavia, or west of Syracuse. While diverse lines of evidence confirm this interpretation, the manner in which it occurred remains clouded in uncertainty about ice marginal positions during the reversals of outflow direction.

As early as 1897, Fairchild wrote:

> "The working hypothesis held by glacial geologists favored the opening of the Mohawk outlet before the Ontario ice lobe retreated from the Lockport-Batavia highland. In such event, the Warren waters, imprisoned in the Erie-Huron basin, would have been at once lowered toward the Iroquois (Mohawk-Hudson outlet) by the withdrawal of the ice-dam from the Helderberg escarpment north of Batavia and the Warren shoreline would not extend east of that highland." (p. 271).
>
> "The Lockport" [now the Batavia] "moraine was, undoubtedly the eastern limit of the Warren water for a considerable time ... but the withdrawal of the ice-front from that portion did not produce immediate lowering of the water or terminate the beachmaking process." (p. 271).

By 1909, Fairchild inferred an episode of "free drainage" eastward between two episodes of ponding:

> "Certainly the ice barrier had to recede or back away and open low passage through Syracuse and eastward in order to allow the river flow which cut the channels. It is equally certain that when Lake Warren subsequently occupied central New York, at about 880 feet altitude, the Syracuse passes were closed. But while the ice front was readvanced at Syracuse, so as to hold back the Warren waters, it was necessary in order that the waters could enter central New York at all that the ice barrier should recede in the Oakfield district." (p. 50)

Fairchild hypothesized that, during the earlier recession, the ice border withdrew in the Syracuse area while it was still pressed against the Onondaga Scarp north of Batavia (Fig. 2A). As a result, withdrawal from this western fulcrum released Warren waters in catastrophic flooding that cut the Caledonia channels (Fig. 3) as well as the associated Rush-Victor channels east of the Genesee Valley.

The western pivot point, Fairchild assumed to be where the Batavia Moraine (initially named Lockport Moraine by Leverett, 1895) mounts the Onondaga Scarp at Phelps Pond (Fig. 2a). Recent palynologic study of sediment cores collected in Phelps Pond have yielded several radiocarbon dates of which the oldest, on basal clayey gyttja, is 11,900 + 100 years (informal communication, Norton S. Miller, N.Y. Biological Survey).

The Batavia Moraine is readily traceable eastward along the crest of the Onondaga cuesta for about 6 miles, but question arises where the moraine crosses NY-98. East of Route 98, the drift cover over Onondaga Limestone is thin. Morainal topography is lacking and no other basis has been recognized

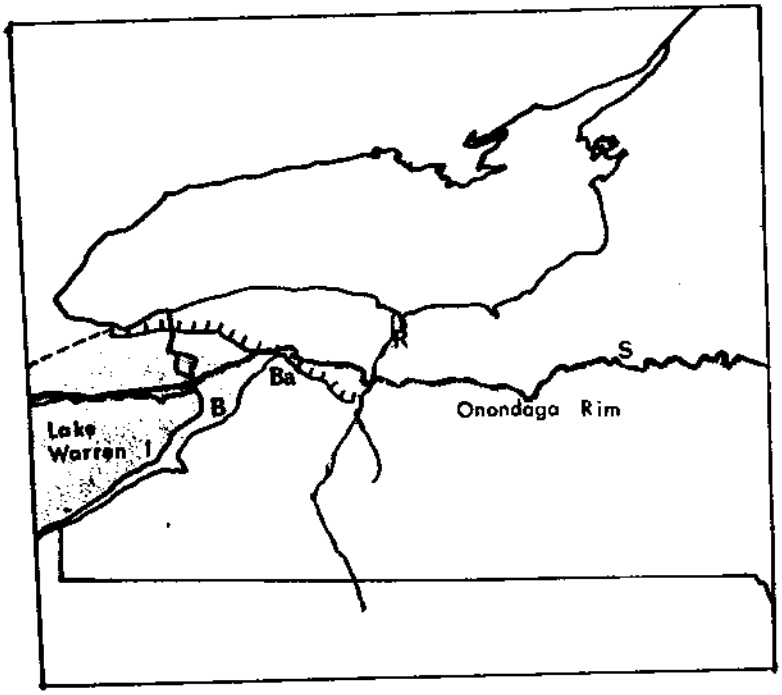



Figure 2A. Schematic representation of moraines and strandline of glacial Lake Warren, impounded on the east by the Batavia Moraine abutting on the Onondaga Escarpment north of Batavia.

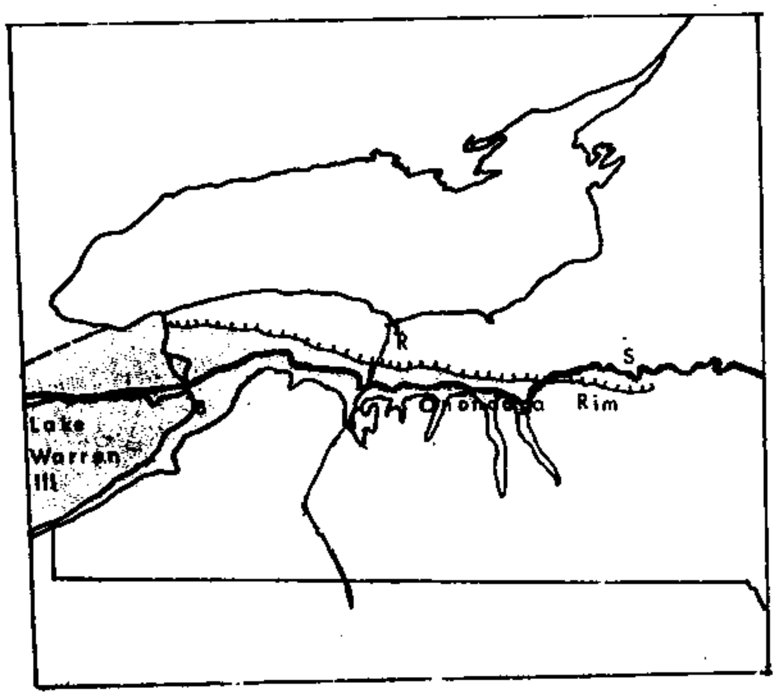



Figure 2B. Schematic representation of moraines and strandline of glacial Lake Warren III, impounded on the east by the Waterloo Moraine abutting on the Onondaga Escarpment north of Skaneateles.

for mapping an ice-border position along this cuesta rim.

On the other hand, only a half-mile gap separates the Batavia Moraine at NY-98 from a discrete southeastward trending morainal ridge, 15 to 25 ft. high. This ridge converges with morainal topography of the Niagara Falls Moraine that can be traced southeastward for 15 miles, passing north of Stafford and crossing the Oatka Valley 1.5 miles south of LeRoy. Leverett (1902) mapped this alignment as Batavia Moraine, placing it well south of the Caledonia channels (Fig. 3). Muller (1977) showed this trend as a dotted boundary, inferring possible but unverified relationship to the Batavia Moraine.

Whether this moraine alignment is part of the Niagara Falls Moraine or the younger Batavia Moraine, it must predate the Caledonia meltwater channels (Fig. 3) to the north, for they show no evidence of glacial modification.

**Eastward-draining meltwater channels**

Glacial drift is thin on the crest of the Onondaga Cuesta northeast of Batavia and northwest of LeRoy, whether because of non-deposition or subsequent erosion. Conceivably the sole of the glacier was stripped of debris by basal shear as it rose over the scarp. Alternatively the rock surface may have been swept clean by meltwater, at first beneath and then at the edge of the ice.

Unmistakeable evidence of meltwater escape eastward begins north of LeRoy and in channel erosion at Limerock (midway between LeRoy and Caledonia (Fig. 3)). The nature of stream bedload in places during this stage is illustrated by boulder gravel dumped on uneroded till as a gravel terrace now exposed in an abandoned pit of the Dolomite Products Company, 2 miles northeast of LeRoy. Meltwater probably occupied ice-walled channels, shifting progressively northward as the ice margin retreated across the scarp face of the Onondaga cuesta. Channel continuity is obscured in places by postglacial deformation which Fairchild ascribed to subsidence due to dissolution of underlying soluble rock. More striking than evidence of subsidence, however, are "pop-ups" with vertical relief of several feet, which appear to have resulted from release of residual stress since deglaciation.

In this scenario of rapidly changing events, Taylor Channel (Figs. 3 and 6) provided the first stable outlet threshold. Terrace gravel remnants suggest that inflow into the channel began above 770 feet above sea level, and cut down to Onondaga bedding surfaces about 50 feet lower. Given the ephemeral conditions of ice retreat from the scarp, it is unlikely that the Taylor Channel was long in use. Entering the ponded Genesee waters, the Taylor torrent built a prominent delta at about 715 feet, across which it shortly incised a 20-ft. deep channel. This channel may simply have resulted from incision as bedload diminished, but may also indicate lowering of the level of ponding in the Genesee Valley while Taylor Channel was in use.

The evidence argues for intensified flooding and rapidly changing marginal meltwater drainage (Fairchild, 1909; Wilson, 1981). A slight weakening of the ice barrier between the head of Taylor Channel and Caledonia was enough to release floods that eroded to bare rock south of the Fairgrounds and aggraded a gravel terrace only slightly lower than the level of incision at the mouth of Taylor Channel. Further slight recession of the ice margin allowed flooding waters to aggrade a gravel plain some 20 feet below the earlier terrace. The

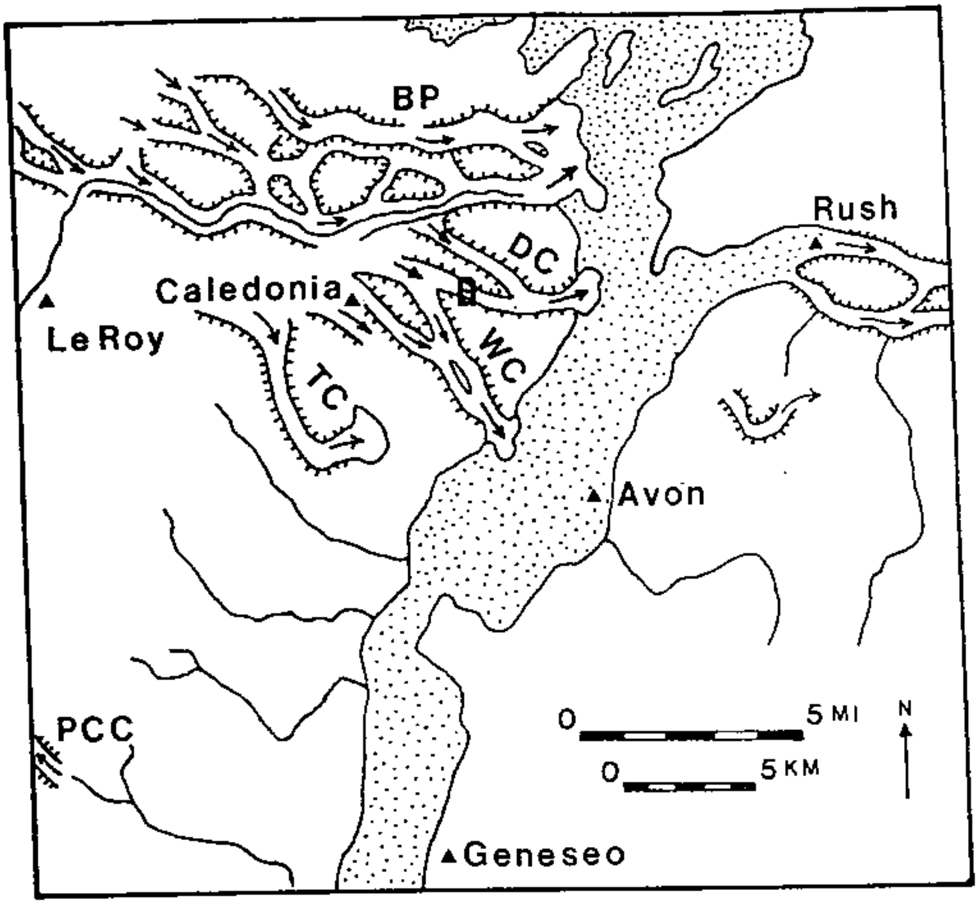



Figure 3. Glacial meltwater channels near Caledonia. BP = Blue Pond Channel; DC = Dugan Creek Channel; WC = White Creek Channel; TC = Taylor Creek Channel. D is the site of test drilling summarized in Figure 4.

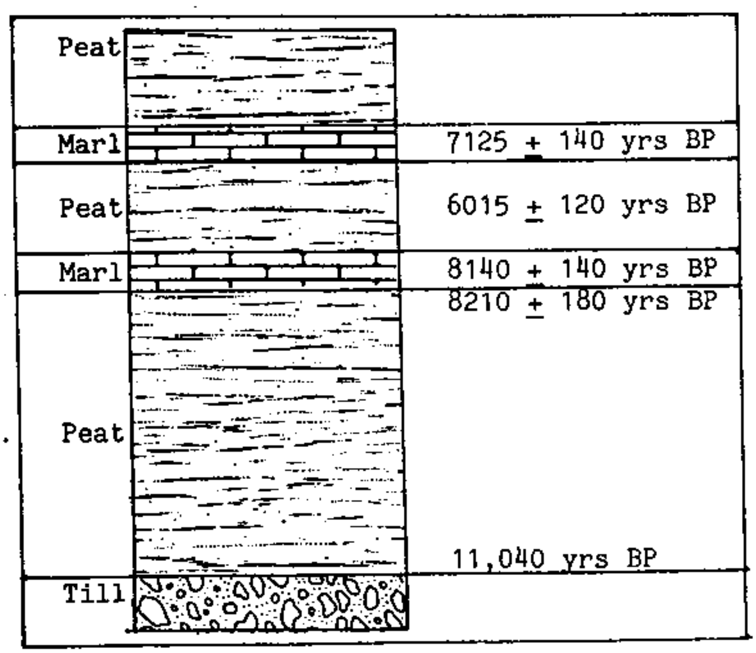



Fig. 4. Stratigraphic relations and radiocarbon data for test hole at Baker, 3.5 mi. east of Caledonia. (R.A.Young, personal communication).

eastward outlet controlling the level of ponding in the Genesee Basin during building of the Taylor Delta is by no means clear. It may have been on the Onondaga Limestone draining via the North Avon umlaufberg (M, Fig. 5), or a bit east on the Honeoye Falls Quadrangle. The evidence cited above indicates that abandonment of the Taylor Channel was closely followed by erosion and associated progradation at successive levels that suggest lowering outflow controls eastward from the Genesee Valley.

At this stage, subglacial drainage may have taken effect in the drumlinized lowland north of the Onondaga Cuesta. The anastamotic drainage pattern (Figs. 3 and 7) involves overfit channels occupied now by White Creek, Dugan Creek, Oatka Creek and Blue Pond. They head, not along the base of the Onondaga scarp, but as far as 4 miles to the north. In general they begin as broad, shallow depressions at elevations ranging from 700 to 650 feet, yet it is probable that for a time at least, they operated simultaneously. Evidence such as divide-crossing channels, low channel sinuosity, flood-exposed rock surfaces, and cross-channels hanging at both ends suggests unusual discharge conditions. Terrace gravels indicative of bar and overbank deposition are extensive, but disintegration features (glacial karst) are surprisingly limited.

Where these channels enter the Genesee Valley, each has aggraded deltaic or terrace gravels at approximately 600 feet, the level of glacial Lake Avon controlled by outflow through the Rush-Victor channel system, draining east toward the north end of Seneca Lake basin. Glacial Lake Avon extended south up the Genesee Valley beyond Avon. Alluviation at the southern end of Lake Avon apparently provided the source for a small sand dune field 1.7 miles north of Avon. Moraine-dammed Lake Geneseo may have come into existence as a separate body of water at this time.

**Lake Warren III**

Following an interval of "free drainage" eastward, Lake Warren (Warren III) was restored in the Erie Basin at a level some 20 feet lower than Warren I. This required glacial readvance onto the Onondaga Scarp west of Syracuse or north of Batavia. The scarp north of Batavia apparently was not covered, or was uncovered before retreat at Syracuse, for Lake Warren III spread into the Genesee Valley (Figure 2B). Fairchild (1897) showed that this was, indeed, the case by tracing a Warren strand from the Erie Basin into the Genesee Valley and across it to the east. Identification of specific strandline features may be questioned, particularly across the salient of the Onondaga Escarpment north of Batavia, but there can be little doubt as to the sequence of events which they indicate.

As mapped, the Warren shoreline clearly cuts across and therefore postdates the Batavia Moraine. Briefly, the level of impoundment in the Genesee Valley was controlled by Lake Warren's threshold at the head of the Grand River in Michigan. This would have been the case while the ice margin stood at the Waterloo Moraine against the plateau west of Syracuse (Fig. 2B). Evidence in the channel system east of the Genesee, for example, near Victor southeast of Rochester, and near Phelps northwest of Seneca Lake, suggests that the glacier readvanced to the Waterloo Moraine, after meltwater flow had ceased in this channel system. If this was, indeed, the alignment that controlled the spread of Warren III into the Genesee Valley, the Waterloo Moraine cannot equate to the Batavia Moraine but may instead be correlable with a later moraine position in western New York represented perhaps by the Barre Moraines. The demise of Warren III occurred as The Gulf channel opened north of Skaneateles, west of Syracuse.

## Postglacial record in the Caledonia meltwater channels

A test hole drilled .3 mile west of the T-intersection with Cox Road (Fig. 4) (R.A.Young, informal communication) passed through 27 ft. of sand and gravel, 15 ft. of till and 12 more ft. of sand and gravel to shale at 54 ft. (584 ft. above sea level). The bedrock surface slopes northward, being 20 feet lower here than at the Onondaga riffle a mile south in White Creek Channel.

Radiocarbon-dating of a 9-ft section in marl and peat on Lacy Road at Baker (C, Fig. 7) (about 3 miles east of Caledonia) yielded age relationships illustrated in Figure 4 (R.A.Young, informal communication). The length of the interval before the beginning of basal peat accumulation is undetermined, so the basal peat date does not define the end of glaciation. The lack of lake sediment between peat and till is unexplained. The inversion of dates may reflect incorporation of "old carbon" in marl.

## REFERENCES


Calkin, P.E. and B.H.Feenstra, 1985, Evolution of the Erie Basin Great Lakes; p. 149-170 in Karrow, P.F and Calkin, P.E., eds, Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30.

Fairchild, H.L. 1897, Lake Warren shorelines in western New York, and the Geneva Beach; Geological Society of America Bulletin, Vol. 8, p 269-286.

Fairchild, H.L., 1909, Glacial waters in central New York; New York State Museum Bulletin 127, 66p.

Leverett, Frank, 1895, On the correlation of New York moraines with raised beaches of Lake Erie; American Journal of Science, Ser. 3, Vol. 50, p. 1-20.

Leverett, Frank, 1902, Glacial formations and drainage features of the Erie and Ohio Basins. United States Geological Survey Monograph 41, 802p.

Muller, E.H., 1977, Quaternary Qeology of New York, Niagara Sheet, 1:250,000. New York State Museum Map and Folio Series, No 25. Albany, NY.

Wilson, M.P. 1981, Catastrophic discharge of Lake Warren in the Batavia-Genesee region. Ph.D.dissertation, Syracuse University, 107p.

## FROM LAKE HALL TO THE PINNACLE HILLS MORAINE

8:00 AM.Leave campus.

MILE Drive north through Geneseo on Main St. (NY-39) to North St. (called Court St. on west side of Main St.)

0.0 **TURN RIGHT** (east) on North St. for one very long block (Fig. 1)

0.5 **BEAR LEFT** on Lima Road. During the Civil War, Camp Union at this corner was a regimental mustering and training site for raw Union troops. **CONTINUE** across flat-floored lake bed drained by south branch Jaycox Creek. As the road rises gently NE, Cashequa Shale crops out at base of roadside ditch.

2.9 Gully cut by tributary of South Fork Jaycox Creek. Shale shingle in roadside ditch just south of gully. A weakly developed strand at approximately 870 ft above sea level which trends NNE obliquely across the road, is ascribed to Lake Warren III.

3.7 Drumlinoid ridge on left, 2000 ft. north, off of Elm Welch Road is wave truncated at about 875 ft. The old Briggs-Laidlaw gravel pit exposed a thin bed of sand and gravel, locally cross-bedded with southeasterly dip

6.7 **TURN LEFT** (north) onto Route 256; cross Conesus Outlet; in 500 ft joined from the right by NY-15 from Lakeville. **CONTINUE** north on NY-256 and NY-15.

7.7 Overpass over I-390.

8.1 **TURN RIGHT** on Marshall Road, recrossing I-390 and curving northeast along the crest of the Canandaigua Moraine.

9.6 **TURN LEFT** (north) on Bronson Hill Road

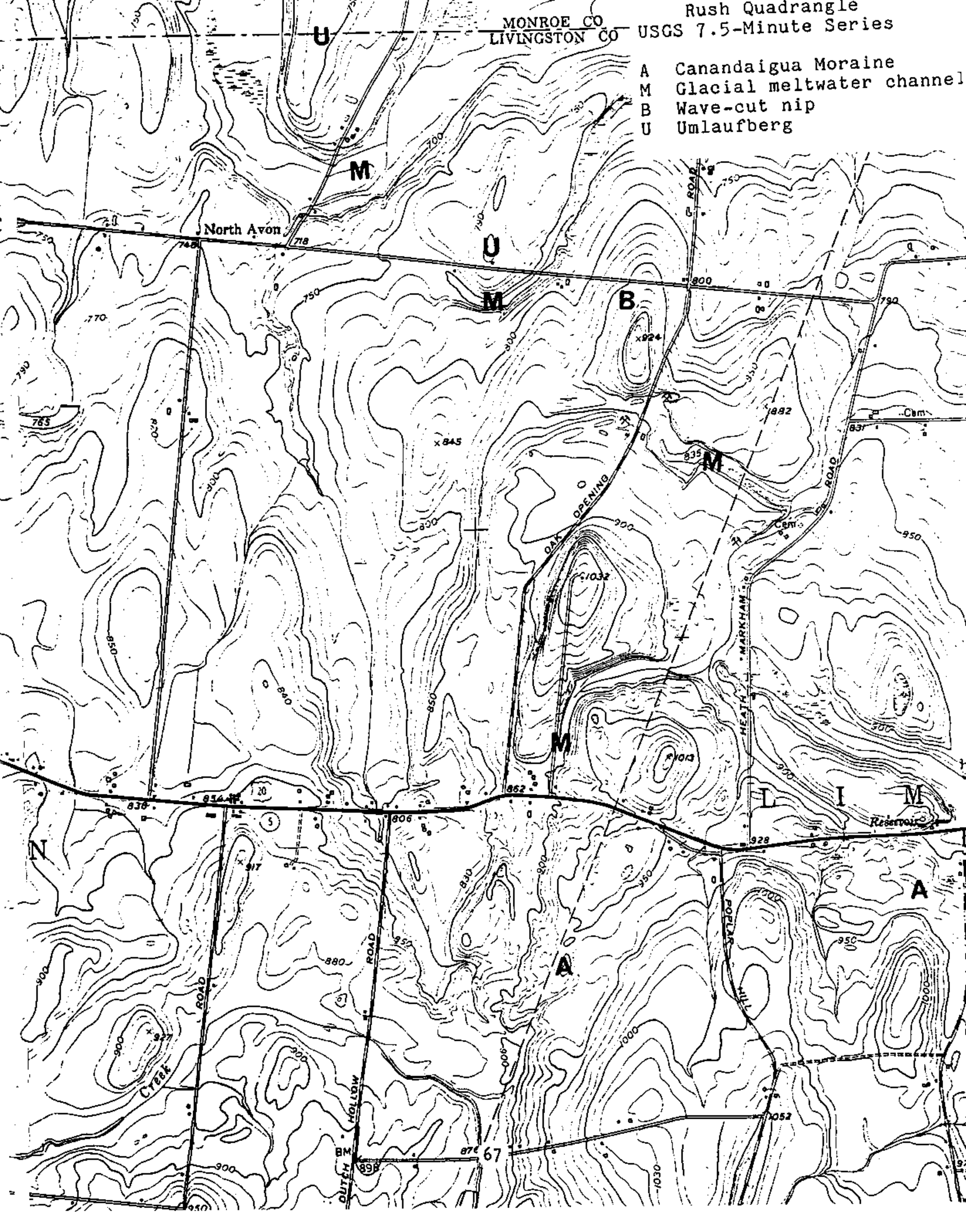
11.2 **TURN RIGHT** (east) US-20 and NY-5 toward Lima (C, Fig. 5). The north end of the drumlin on your right at this turn appears to have been wave-eroded. Pass NYSDEC regional office.

12.2 **TURN LEFT** (north) on Oak Openings Road. The Canandaigua Moraine (A, Fig. 5) is now behind you, south of US-20. A narrow meltwater channel, with its threshold at 905 ft (40 ft above the road) leads northeast from this corner (B, Fig. 5).

13.5 Cross minor meltwater channel with threshold at 845 ft. This is one of a number of well-defined meltwater channels which head below the Warren strandline. In .3 mi. turn west on North Avon Road

**STOP 1.** Relationships of meltwater channels and Warren shore features are observable in the northward reentrant of the Canandaigua Moraine midway between East Avon and Lima. A small channel with threshold at 840 ft. a.s.l. crosses Oak Opening Road, 1.3 miles north of US-20A (M, Fig. 5).

The Warren strandline cuts across and postdates the Canandaigua


Rush Quadrangle
USGS 7.5-Minute Series
A Canandaigua Moraine
M Glacial meltwater channel
B Wave-cut nip
U Umlaufberg
MONROE CO
LIVINGSTON CO
North Avon
OAK OPENING
HEATH MARKHAM ROAD
POPLAR HILL
DUTCH HOLLOW ROAD
Creek ROAD
Cem
Reservoir
L I M
N

Moraine; the presence of ice at the Canandaigua Moraine is implied in the derangement represented by these channels, yet the channels extend below the projected Warren shore.

A nip on the stoss end of "Drumlin 924" (SW of the corner of Oak Openings and North Avon Roads)(B, Fig. 5) was recognized by Fairchild as a wave-cut feature cut at the Warren shoreline.

**Basic questions:**

1. Is the nip on the stoss end of "Drumlin 924" indeed wave-cut?

2. Are the meltwater channels that cross Oak Openings Road subaerial? Or could they have been cut below the level of proglacial meltwater impondment?

3. If the channels are entirely subaerial, where, and how did the proglacially dammed meltwater drain?

4. If the channels were cut before Lake Warren III, would you expect them to show more evidence of wave modification?

**CONTINUE** west on North Avon Road (Fig. 5).

15.1 **TURN NORTH** in North Avon onto Gilbert Mills Road. Cross the North Avon incised meander and umlaufberg (U, Fig 5). Note channel dimensions.

16.3 **TURN WEST** at T-intersection. Cross arm of incised meander, floored on Onondaga Limestone. The channel begins at this point, without depositional evidence of the ice margin that for a time prevented meltwater taking a lower course to the across bare rock toward Five Points, 1 mile northeast.

17.7 **TURN WEST** on Honeoye Falls Road 19.1 Cross I-390, then NY-15. Continue west.

20.6 At Y-intersection with Avon-West Rush Road, **BEAR LEFT** onto the stem of the "Y", toward Avon. Eolian sand, blown from shoals near the south end of glacial Lake Avon, forms low dunes just north of the County Line.

23.4 In Avon, **TURN RIGHT** (west) onto US-20. Enter flood plain.

24.0 At bridge, U.S.G.S. gaging station on the right, near site of Berry's Tavern, built in 1784 and the first bridge across the Genesee, built in 1804..

24.5 **BEAR LEFT** onto Broadway, continuing west on US-20.

25.5 **TURN RIGHT** on River Road for Stop 2, or continue west on US-20.

**OPTIONAL STOP. Canawaugus Delta**

Extensive gravel operations have wreaked havoc with the large hanging delta built into glacial Lake Avon at 590 ft by meltwater escaping through the White Creek Channel. Present stream discharge seems inadequate to have cut the

**OPTIONAL STOP: Canawaugus Delta** (continued)

channel incised in the delta and graded essentially to the Genesee flood plain at about 540 ft above sea level.

The delta is composed of cobble gravel dominated by carbonate clasts, with almost no boulders. Minor carbonate cementation has caused little problem in the borrowing operation. Canawaugus was the name of the Iroquois village at this site.

26.7 **TURN RIGHT** onto Quarry Road. Onondaga Limestone nearly free of drift cover along Christie Creek and west of road.

27.6 **TURN LEFT** onto McCorkindale (Taylor Hollow) Road (Fig. 6), .5 mi to Taylor Delta

**STOP 2. Taylor Delta**

A compound hanging delta, built into proglacially ponded water in the Genesee Valley by meltwater discharge through the Taylor Channel. The Taylor Channel (B, Fig. 6) heads on the Onondaga Scarp west of Caledonia and was the first channel cut in this area as eastward outflow began to tap the Erie Basin. Initial discharge through the Taylor Channel (B, Fig. 6) aggraded toward a base level above 720 feet, spreading its gravel onto stagnant ice at the northedge of the delta (C, Fig. 6). The level of aggradation dropped about 10 ft while the major part of the delta was built. This level was in turn incised by a channel graded to a control at about 690 feet.

**Basic questions:**

1. Where was the ice margin when Taylor delta was built?

2. What controls determined the levels of ponding in the Genesee Valley while the Taylor delta was being built?

3. What flow duration and discharge are indicated?

4. Was the gravel which comprises this delta primarily derived from cutting of the Taylor Channel (Onondaga Limestone, black shale, etc.) or from glacial load (derived from north of the Onondaga Scarp?)

**CONTINUE** west on Taylor Hollow Road

29.1 **TURN RIGHT** on Sand Hill Road. Cross Taylor Channel. North of Taylor channel to east of the road is a delta remnant and to the west, a small area of stagnant ice deposits. Continuing north to Caledonia, Sand Hill Road trends obliquely across the flood-swept remnants of two minor ice marginal positions (E, Fig. 6) that were quickly abandoned as meltwater channeling shifted northward off the Onondaga Scarp at Caledonia. Enter Caledonia near the Fairgrounds.

32.1 **TURN RIGHT** on NY-5. In a block, **TURN LEFT**; in two blocks turn right on Iroquois Road

**CONTINUE** straight east on Iroquois Road. In the next two miles cross

Caledonia Quadrangle
USGS 7.5-Minute Series

A Taylor Delta
B Taylor Channel
C Ice-contact deposits
D Floodway channeling
E Morainal remnant

flood-swept surface (D, Fig. 6), channeled and terraced at elevations ranging from 660 down to 610 ft. (Fig. 2) Cement Plant Pond in the bed of White Creek Channel is one of several locations in the Caledonia channels where marl has been commercially exploited.

32.8 **TURN RIGHT** (south) on Feeley Road (Cox Rd.) (Fig. 7)

33.1 **TURN LEFT** on Cameron Road.

35.5 **TURN LEFT** on River Road. Cross Dugan Creek.

36.3 **TURN LEFT** into Finewood Gravel Pit

**STOP 3. Finewood Gravel or Cole Sand and Gravel Co. pits**

Extended gravel operations afford opportunity to consider the nature of deltaic deposition at the mouth of Dugan Creek Channel.

In the past, as much as 10 ft. of laminated lake sediments have been exposed overlying delta gravel near the abandoned railroad line at the east edge of the Cole Sand and Gravel Co. pit. Near the west edge of the Finewood Gravel Co. pit, the gravel thins on a sloping and essentially uneroded till surface.

Deltas at the mouths of White Creek (A, Fig. 7), Dugan Creek (B, Fig. 7), Oatka Creek and Blue Lake Channels are all approximately at 600 ft. and 10 to 20 ft. higher than the present elevation of the threshold in the Rush Channel which ultimately controlled outflow from glacial Lake Avon.

**Basic questions:**

1. Were the Caledonia channels occupied simultaneously, or in sequence?

2. What hydraulic conditions are suggested by the sedimentary structures observed?

3. Where was the ice margin when these channels were eroded?

**CONTINUE** north on River Road (Canawaugus Road in Monroe County)

37.8 **TURN RIGHT** on North Rush Road. Cross Genesee flood plain. Note Ground Hog Hill and nearby sand and gravel knobs rising above the flatland -- evidence again that the flood plain is lake bottom and not primarily a product of stream erosion.

41.2 Cross overpass and **TURN LEFT** to enter I-390, northbound.

At Junction with I-590, **BEAR RIGHT** toward Irondequoit Bay. Field Guide continued in section by R.A.Young.

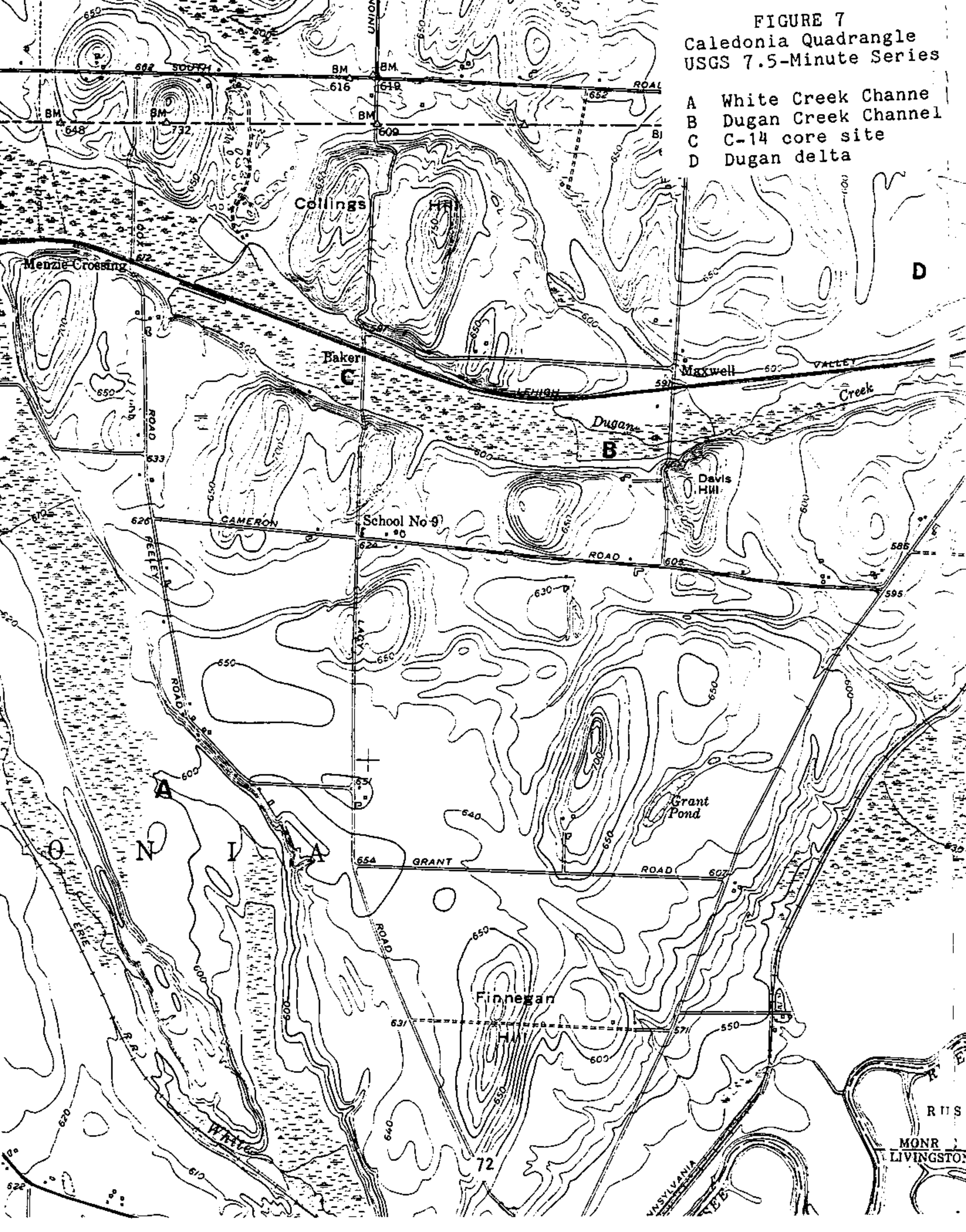



FIGURE 7
Caledonia Quadrangle
USGS 7.5-Minute Series

A White Creek Channe
B Dugan Creek Channel
C C-14 core site
D Dugan delta

# PLEISTOCENE GEOLOGY OF IRONDEQUOIT BAY

Richard A. Young

## Introduction

Irondequoit Bay occupies the northern end of a broad valley eroded below sea level by an unknown number of glacial advances and filled with a complex sequence of glacial and postglacial sediments. The northern portion of the valley beneath the modern bay preserves a diverse sequence of fluvial and lacustrine sediments resulting from fluvial entrenchment during the post-Iroquois low stage of Lake Ontario. The early erosional phase was followed by a complex sequence of fluviolacustrine infilling as the lake level rose in response to the southwestward tilting of the Ontario basin. The sequence of events associated with the post-Iroquois history can now be reasonably inferred from the lake level curves and history described by Anderson and Lewis (1985). Drilling in the valley for several engineering projects has produced a detailed, although incomplete, record of glacial and postglacial stratigraphy, which can be interpreted in light of the established late glacial history of Lake Ontario.

Significant differences can be demonstrated between the previously published glacial/postglacial history and the evidence provided by subsurface records. The differences fall into three major categories. First, the subsurface sedimentary sequence in the northern 6 miles of the bay area is separate and distinct from the largely till-filled reach to the south. The valley segments are divided by the buried Pinnacle Hills moraine within the valley near the latitude of Browncroft Boulevard. Second, the fine-grained bay sediments conceal a complex of fluvial gravel channels under the east side of the bay that were formed and covered as the lacustrine sequence prograded southward over the ancestral Irondequoit Creek delta. These geologic events have produced an unusual aquifer system, which is being utilized by two major well fields. Finally, the infilling of the bay by lacustrine sediments and ice-contact stratified drift (moraines?) produced an interesting glacial stratigraphy not well described or explained in the early published interpretations of the bay geology (Chadwick, 1917; Fairchild, 1928).

## Geologic Setting

This discussion will focus on the deglaciation of the valley from the latitude of the Pinnacle Hills moraine, at which time glacial Lake Dawson expanded westward (Muller and others, 1988) near a present elevation of 480 feet. The main portion of the Pinnacle Hills Moraine near Rochester (Figure 1) is a sandy kame moraine capped with a thin layer of till in places (Fairchild, 1923). This moraine can be traced to the west edge of the Irondequoit Valley near Browncroft Blvd., where numerous borings now demonstrate that the valley is filled by a thick mass of reddish till (Figures 2, 3). The floodplain elevation is approximately 250 feet, whereas the bay surface averages 246 feet. The rock floor of the buried valley slopes from 76 feet in elevation near Browncroft Blvd. to 131 feet below sea level at the sand bar.

A number of borings completed for sewer construction have been supplemented by

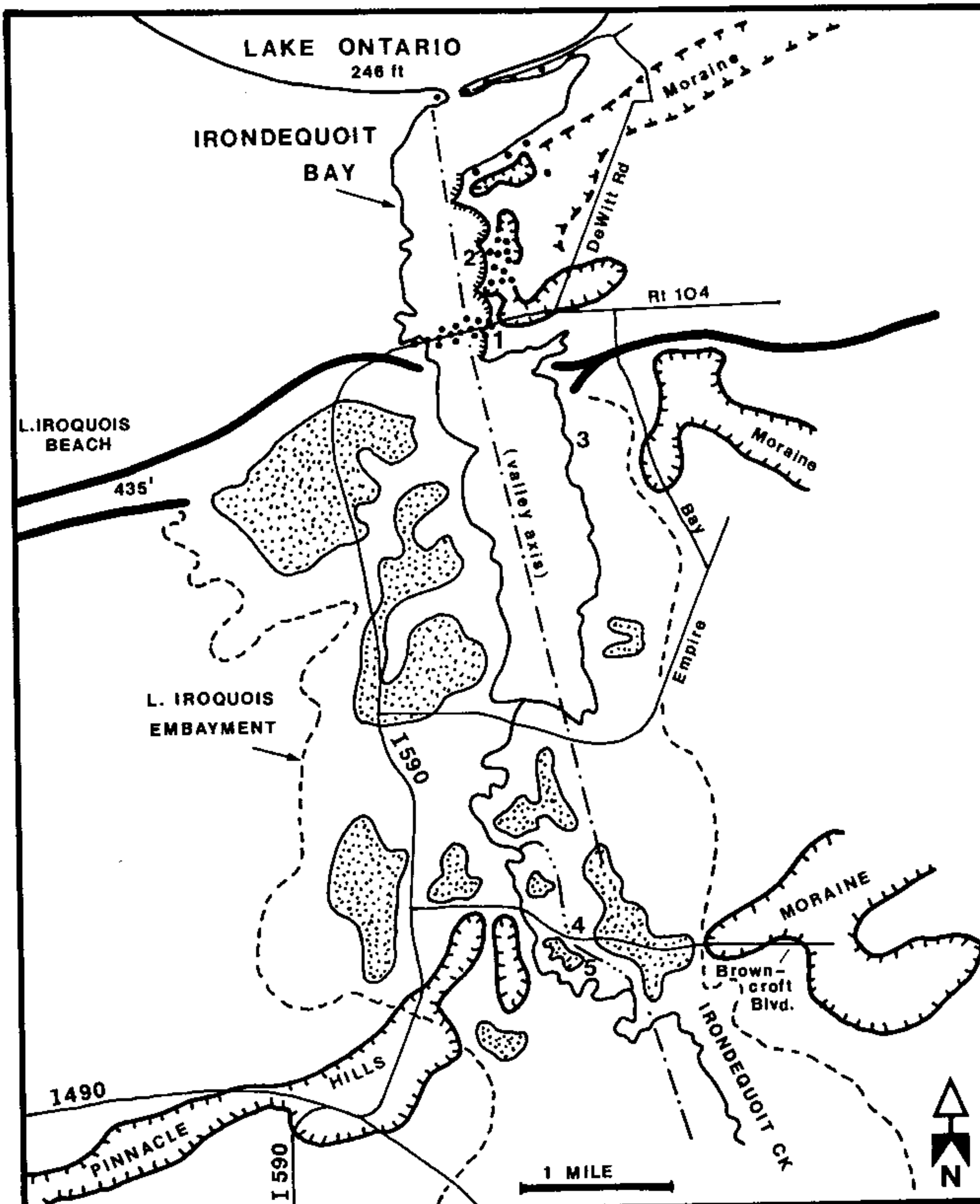



Figure 1. General location map of Irondequoit Bay. Fine dots are areas of least eroded Lake Iroquois sediments. Large dots are general locations of well sites. Stops: 1. Bluffs at Rt. 104 bridge; 2. Meander-scar bluffs (hachures) at DeWitt Rd. well field; 3. Glen Edith - - Iroquois sediments on till; 4. Browncroft Blvd. at Irondequoit Ck. (Figure 3); 5. Ellison Park section.

test holes completed by the United States Geological Survey (Ithaca, N. Y.) in connection with the National Urban Runoff Program (NURP) studies of the Irondequoit Creek basin. The drilling was undertaken to establish whether or not the buried valley constitutes a major, continuous aquifer.

Drill holes through the buried moraine have established that a 150-foot till plug fills the valley from near riverbed level down to the bedrock surface, which is currently at an elevation of 76 feet (Figure 2). In front (south) of the moraine an outwash fan extends along the axis of the valley. The area immediately behind the moraine (north) is filled with 117 feet of well-sorted, coarse sand and gravel resting on 16 feet of till above bedrock. These coarser sediments are buried beneath Lake Iroquois(?) sands and silts which, extend down to elevation 189 feet (57 feet below the modern floodplain of Irondequoit Creek).

The morainal till plug is reddish in color all the way to bedrock (Queenston Fm.), whereas the older till south of Browncroft Blvd. is a gray color. These relationships appear to establish that the Pinnacle Hills moraine involved a glacial surge or readvance across the red Ordovician strata, thus superimposing "red" till over an older "gray" till. The subsequent recession from this position formed a coarse gravel unit between the ice front and the moraine, much like the outwash apron that will be visited at the stop in Ellison Park.

**Ice Positions**

During general ice recession northward along the valley from the Pinnacle Hills position, the ice front appears to have stopped or readvanced at least twice south of the Lake Ontario shoreline. One moraine (Figure 1) was formed just south of the Iroquois strand (Ridge Road, Route 104), and the other where the northeastern perimeter of the bay changes to a more easterly trend (Carlton Moraine(?), Muller, 1977; Muller and Cadwell, 1986). These moraine positions are best substantiated by a comparison of both the older mapping (Fairchild, 1928; Chadwick, 1917) produced prior to urbanization of the bay perimeter and the new Finger Lakes Sheet of the Surficial Geologic Map of New York. When combined with the deep drilling data for the Town of Webster well fields, all of the maps, subsurface stratigraphy, and surface morphology indicate complex till, outwash(?), and kame moraine sequences at these latitudes (Figures 1 and 3). The presence of ice-contact deposition at these locations would also explain the obvious differences in the bay width, shoreline trends, and surface geomorphology.

**Post-Iroquois Low Stands of Lake Ontario**

The sequence of gradually rising lake levels following the significant drop from the Lake Iroquois strand (435 feet to -25 feet, present elevations) has been compiled and updated in detail by Anderson and Lewis (1985), including data developed by Young (1983) for Irondequoit Bay. The section preserved beneath the sand bar is shown in Figure 4. The basic geology recorded here involves a period of deep fluvial incision through the bay north of the Pinnacle Hills Moraine, which limited erosion upstream to a maximum of 35 feet below the modern floodplain. As Lake Ontario slowly rose to its present level, the bay was gradually filled in by the interfingering of fluviodeltaic and lacustrine sediments, including an axial, gravel-filled channel in the

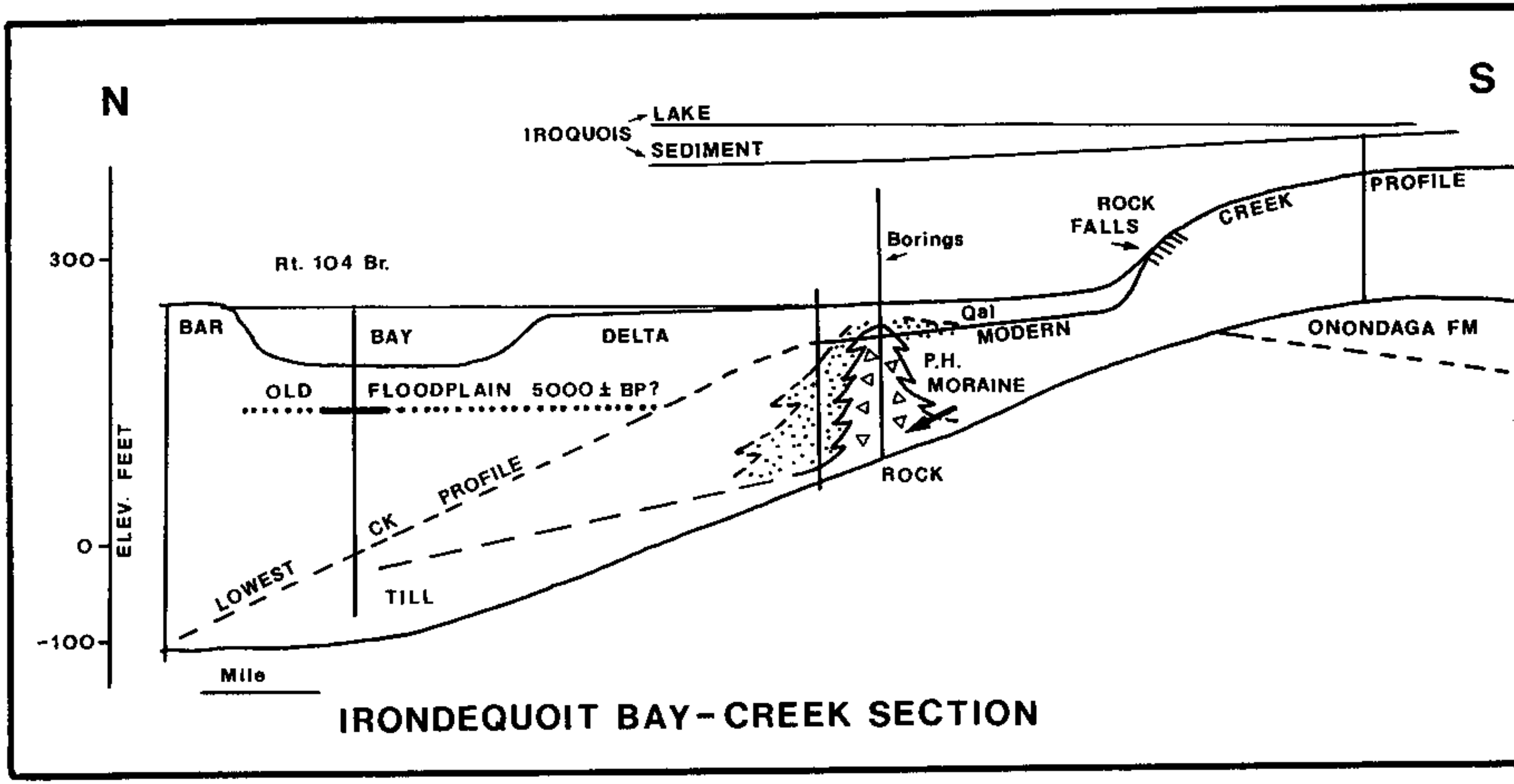
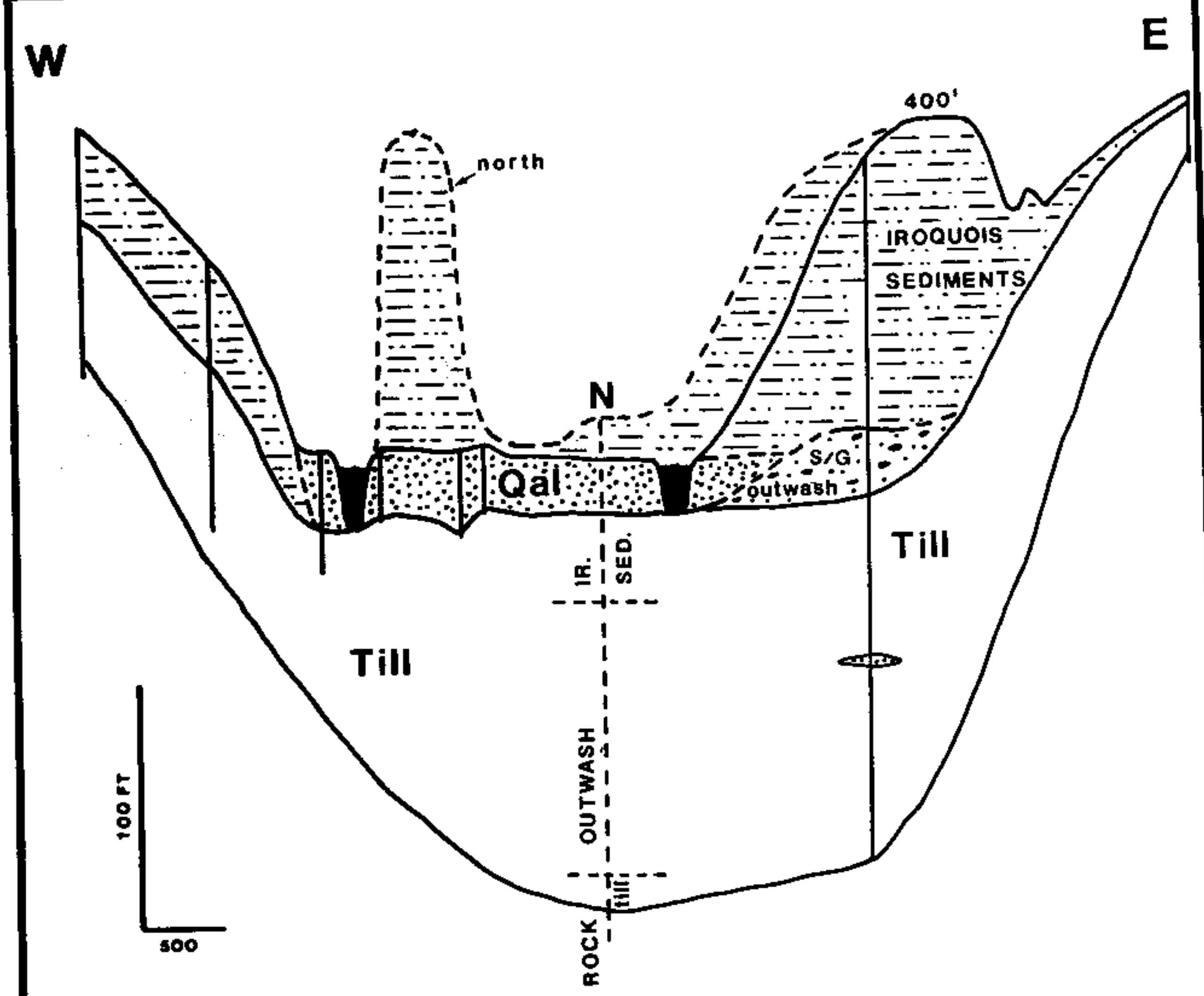



Figure 2. Generalized longitudinal section along Irondequoit Bay and Creek from Lake Ontario to East Rochester. Solid lines south of moraine indicate upper and lower limits of Creek incision, contrasted with inferred fluvial incision (dashed line profile) during post-Iroquois low stand of L. Ontario. Old floodplain (Figure 5) represents interval of slowed lake level rise indicated on Figure 6. Division of northern and southern valley segments by buried Pinnacle Hills (P.H.) Moraine indicates why groundwater aquifer analysis has been complicated. Note isolation of central bay depth by deposition of bar and delta. Borings through moraine also shown on Figure 3. Creek south of rock falls has responded to base level control resulting from superposition of Creek onto bedrock wall on east side of buried valley.




SECTION AT BROWNCROFT BOULEVARD STOP

Figure 3. Cross section near Browncroft Blvd. along Irondequoit Creek. Dashed profile and boring (N) located about 1000 ft. north of section line. Section is along approximate axis of buried Pinnacle Hills Moraine as defined by drill hole data from 25 wells (only key wells shown for clarity). Dashed boring shows thickness of glacial outwash deposited against back (north) side of moraine. Till is reddish and rock is Queenston Fm. in boring N. S/G (sand and gravel) outwash extends south to Ellison Park (Stop 5). Subsurface data show that Irondequoit Creek has not incised its bed through the moraine more than about 35 feet below the modern floodplain level, in spite of the magnitude of fluvial incision from the moraine north (see Figures 2, 5).

flowing currents. It also seems possible, given the evidence for local upward coarsening, that the ice front may have advanced slightly near the end of the lake deposition interval.

In portions of the bay, for example, near Glen Edith, the Lake Iroquois sediments thin abruptly toward the valley margins (east-west), forming a steep, narrow sedimentary wedge over reddish till. Within the lake sediment sequence north of the Route 104 Bridge a clay/silt bed from 2 to 6 feet thick occurs 10 to 20 feet above the modern lake level. A similar unit appears in some of the westernmost borings for the DeWitt Road well field. Occasionally, this unit and others like it show the effects of internal soft sediment deformation, whereas the beds above and below are horizontal. It appears that these major silt/clay horizons are potential time lines, which could be useful for the development of a better stratigraphic model for the depositional sequence in Lake Iroquois. A simple model would imply that the limited number of clay/silt units in the thick section of sands and silts represent those times when the ice front receded temporarily, resulting in short-term fining of the depositional sequence. However, the overall uniformity of the lacustrine section and the thin character of the clays suggest that these events did not represent either long intervals of time or major ice recessions.

**Relation Between Glacial and Postglacial Stratigraphy**

Figure 5 illustrates the inferred subsurface connection between the buried fluvial deposits near the eastern side of the bay and the permeable glacial sequence into which the valley was incised. The position of this fluvial section directly adjacent to the DeWitt well field for the Town of Webster is also supported by the geomorphic form of the eastern bay bluffs north of the Route 104 Bridge. The latest topographic maps and aerial photographs clearly indicate that the arcuate shape of these bluffs (Figure 1) has resulted from meanders impinging against the east side of the valley during some fluvial episode in the geologic evolution of the bay. The development of a floodplain at some lower elevation within the bay area could have been possible only when the level of Lake Ontario was at or slightly below the present level of the deepest portion of the bay floor (assuming some subsequent sediment deposition). The stratigraphic-geomorphic relationship implies that deltaic and fluvial sedimentation kept pace with the rise in lake level such that the valley was a floodplain/marsh throughout all but its northernmost end until the level of Lake Ontario was close to its present elevation. However, an alternative explanation is possible. A small bay may have existed during early postglacial time accompanying the relatively rapid (5 feet per century) lake level rise from the low post-Iroquois stand 11000 to 7000 years ago (Anderson and Lewis, 1985, Figure 9).

**Explanation For Fluvial Conditions (circa 7000 to 5000 B.P.)**

Following the early rapid rise of Lake Ontario from its lowest stand, the compilations of Anderson and Lewis (1985) indicate a decrease in the rate of rise to about 1.5 feet per century near Rochester between 7000 and 5000 years ago (Figure 6). During this period of relatively slow rise or near stability, the lake level was between 50 and 100 feet lower at Irondequoit Bay. This elevation difference corresponds closely with the greatest water depth of the modern central bay (75 to 80 feet). During this interval of slowing in the

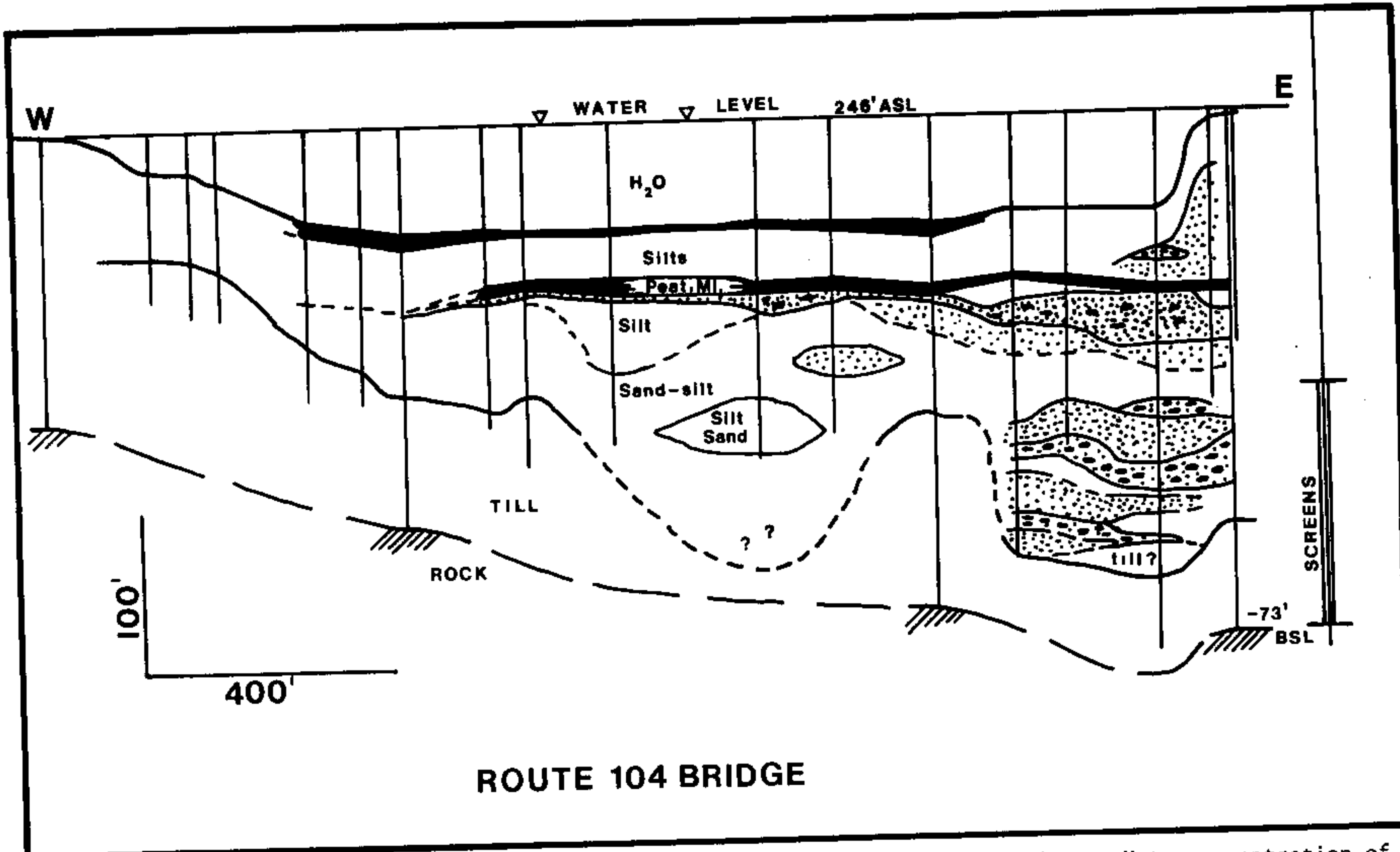



Figure 5. Cross section at Route 104 Bridge (See Figure 1) from NYSDOT boring data. Note concentration of fluvial gravel channel deposits on eastern side near bluffs. The only widespread gravel horizon is associated with the lower organic horizon inferred to be a floodplain surface about 90 to 100 feet below modern lake level. Schematic boring to east (well screen) indicates position of permeable glacial drift zone at Webster/DeWitt well field. Modern sedimentation (post 5000 B.P.) appears finer (more silty).

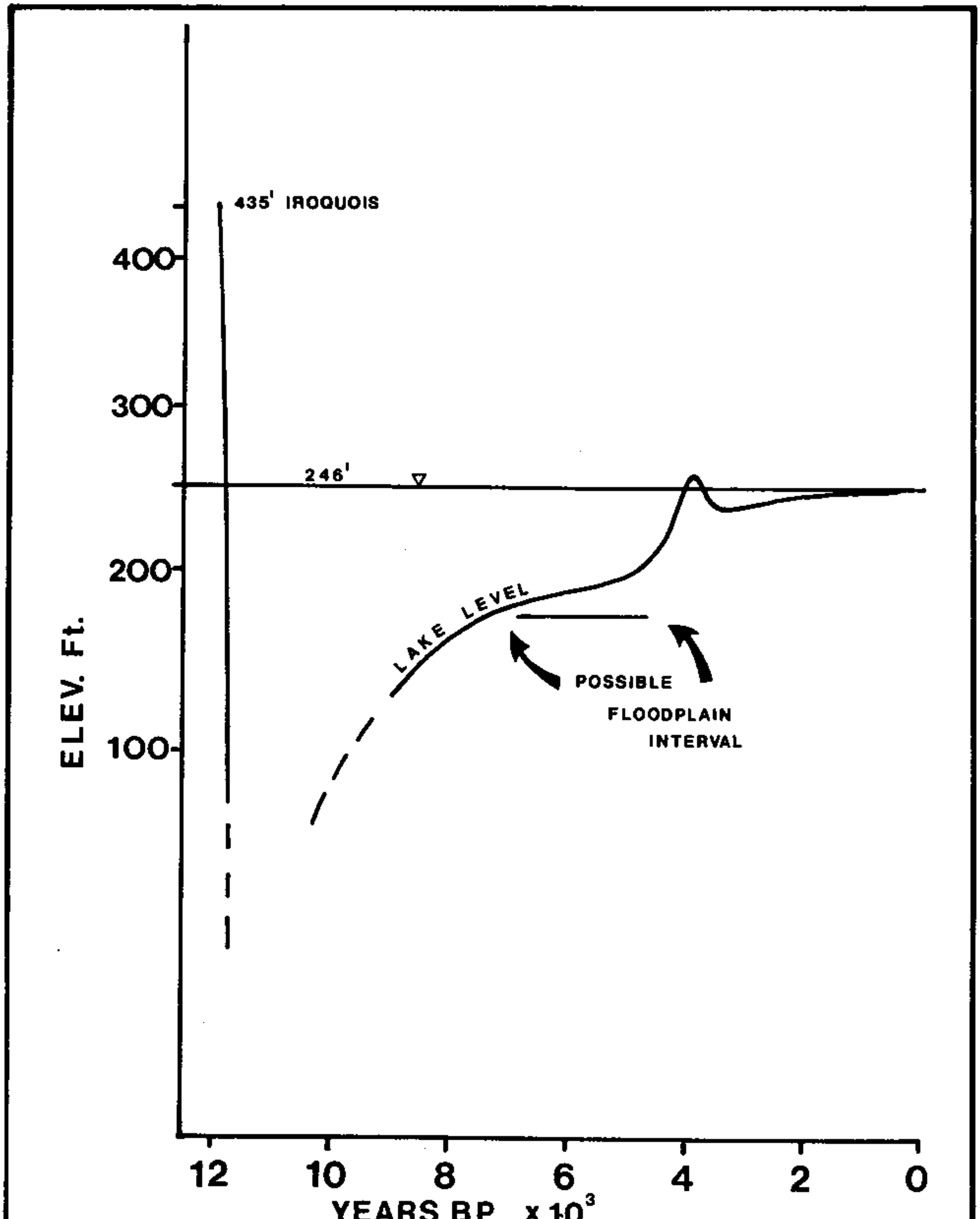



Figure 6. Approximate relative lake level curve from Anderson and Lewis (1985) modified for Irondequoit Bay area. Important feature is relative flattening of curve between 5000 and 7000 years ago, which might have allowed fluvial sedimentation to build a floodplain/marsh/delta into the central bay.

rate of lake level rise ancestral Irondequoit Creek could have prograded its delta and floodplain northward, potentially filling the valley nearly to the elevation of what is now the deepest part of the central basin. This would have created a temporary stream/delta/marsh and associated meandering stream in the valley north of the bridge at an elevation of 90 to 100 feet below the present lake. The subsequent increase in the rate of water level rise (about 6 feet per century between 4000 and 5000 years ago) would then have reflooded the valley (as at present), forcing deltaic sedimentation gradually back toward the head of the modern bay. This sequence of events would account for the irregular north-south bottom profile of the modern bay (Figure 2). Deltaic sedimentation to the south would be matched by barrier bar encroachment from the north, thus isolating the central bay floor as a deeper basin with a slower rate of sedimentation. This remnant of the deeper bay could represent the approximate level of the flooded marsh/delta surface (with a thin covering of younger sediments).

Stratigraphic evidence for this sequence of events, specifically the slowing down of lake level rise about 5000 to 7000 years ago, is contained in the Route 104 Bridge borings. At about 90 to 100 feet below modern lake level the bridge borings record a 5- to 10-foot-thick organic zone, which is continuous across the bay (about 35 feet below the present water-sediment interface). In most borings this peaty horizon is directly underlain by the only thin sand/gravel interval that is relatively continuous across the central portion of the bay. If the deepest portion of the bay (water) in early historic times was 80 feet (Chadwick, 1917), the organic horizon at 90- to 100-foot depths could represent such an early floodplain/marsh surface. It would be necessary to allow for variable amounts of sedimentation since that time (10 to 35 feet throughout the central bay), as a result of unequal areal sedimentation patterns.

The undercutting of the northeastern bluffs by stream meanders north of the present Route 104 Bridge could have begun during this period of relative lake level stabilization, and continued for an unknown interval, depending on the exact balance between lake level rise and fluvial/deltaic sedimentation rates. There is a 30-foot section of sands and gravels along the base of the northeastern bluffs directly below the 90-foot deep organic horizon (Figure 5). A thinner sand and gravel unit is present throughout much of the central bay at this same horizon. This fluvial deposit might correlate with the most recent episode of bluff undercutting. If this scenario is correct, the relationships on Figure 6 and the age of a similar horizon in Sodus Bay (Anderson and Lewis, 1985) dated at about 5000 years B.P. suggest that the meander scars date from this general period when fluvial aggradation temporarily became dominant over bay flooding.

If one considers the areal shape of the bay, it is obvious why the meanders only undercut the bluffs forming the protruding northeastern edge of the bay. If a straight line is drawn along the axis of the bay (connecting the points of greatest known bedrock depths) from the mouth of Irondequoit Creek to the sandbar inlet at Lake Ontario, the line is tangent to the protruding northeastern shore along the bluffs where the old meander scars are obvious. This section of the bay perimeter juts westward beyond the rest of the eastern shore because of the glacial geomorphology of moraine deposition (Figure 1). The subsurface and geomorphic evidence imply that the main course of ancestral Irondequoit Creek in post-Iroquois time followed the shortest axial distance through the valley (also the deepest bedrock profile). It undercut the eastern bluffs where they jutted furthest westward into the valley.

The excellent preservation of the relatively undissected bluff meander scars raises the issue of a reasonable age for their formation, given the unconsolidated nature of the glaciolacustrine section. Their state of preservation needs to be explained either as a result of the unusual erosional resistance of the underlying glacial sediments, or as a consequence of some potentially overlooked (undocumented) interval of fluvial erosion during a more recent low water stage. However, the documented physical constraints on the timing of permissible fluvial erosion at the elevation of the bay floor imply that a floodplain environment at this location could not have persisted much nearer to the present than about 4500 years ago. This is based on extrapolation of the radiocarbon ages constraining the generalized lake level curves of Anderson and Lewis (1985). The peat over sand interval in Irondequoit Bay (interpreted to represent the former floodplain) is matched by a similar stratigraphic sequence in coring from Sodus Bay at a depth near 80 feet below bay level. The radiocarbon constraints for the Sodus Bay peat horizon are between 4700 and 5500 years B.P. (Anderson and Lewis, 1985).

A younger, significant decrease in lake level, sufficient to uncover the deeper Irondequoit Bay floor (and permit fluvial meandering), does not seem supported by any line of evidence. The bluffs are protected from severe wave erosion by the geography of the bay and by a lack of drainage basin development along that particular section of bluff. It may be that the few clay horizons in the Lake Iroquois sediments have also helped keep the bluffs from eroding more severely and have prevented destruction of this unique record of relatively old meander undercutting.

**Practical Implications of Glacial/Postglacial History: Groundwater Aquifers**

The Town of Webster has developed two well fields along the bay, one at DeWitt Road near the meander-scarred bluffs and the other on the baymouth sandbar. The sandbar well field undoubtedly taps both the water of Lake Ontario (through the sand/gravel blanket that formed as the lake rose) and the fluvial channel to the south beneath the bay silts. Pumping of the DeWitt wells has been shown to affect the level of wells at the sandbar field, whose static levels are essentially at bay level. However, given the apparent postglacial history of the bay (even allowing for minor modifications in the events) the well field near DeWitt Road is clearly located within a glacial sequence, whereas the sandbar wells penetrate a younger postglacial bay fill.

One obvious way to explain the remarkable hydraulic connection of these two fields, located 6000 to 7000 feet apart, is through the buried fluvial channel beneath the eastern margin of the bay. The westernmost DeWitt wells penetrate a complex section of clay, sand, gravel, and till with a gravel sequence just above rock (Figure 5, screened interval). Some of the wells nearest the bay shore penetrate thick sections of sandy sediments very like the Lake Iroquois section exposed in the bay bluffs above lake level. Below lake level the sections logged in the DeWitt wells are diverse and contain thick sections of till and gravel. Four other test wells along the bluffs northeast of the meander scar section (Figure 1) between the main DeWitt Road field and the sandbar wells were not productive. Presumably these test wells are too far from the meander segment, and in addition, they lack permeable gravel horizons at comparable depths.

It appears that the DeWitt Road wells have fortuitously intersected a section of permeable glacial gravels which are at the same approximate elevation as the gravel intervals in the younger channels beneath the bay adjacent to the bluffs. As a result, there is a coincidental hydraulic connection between some of the permeable glacial materials just east of the bluffs near the bottom of the DeWitt Road wells and the postglacial channel fill under the bay (Figure 5). The unfortunate consequences of this juxtaposition of permeable strata have recently been demonstrated by a series of events involving gradual salt intrusion into the DeWitt Road well field.

Shortly after the Route 104 Bridge test borings were completed (1960-1965), DeWitt well numbers 9-12 were completed, and the pumping rate at the field was increased to a yearly average of about 4 million gallons per day. Within a few months wells 3 and 8 (southernmost wells closest to the bridge) showed a 4- to 6-fold rise in sodium levels (reaching 1000 to 1300 ppm). The salt has been attributed to connate bedrock brine contamination resulting from the increased pumping rates. However, the proximity of the well field to the bridge test borings in the bay bottom suggests that salt water, known to be present in the bottom of the bay (road salt runoff), could have migrated downward through these bridge foundation borings. The salt could have reached the lower gravels in the incised fluvial channels beneath the bay silts and clays, and then have been drawn into the well field.

If this is the correct geologic explanation for the salt intrusion, it is a good example of the usefulness of an improved geologic data base for dealing with environmental problems. A more adequate means of collecting, storing, and retrieving such geologic and environmental information would reduce the chances for such an occurrence in the future.

## ACKNOWLEDGEMENTS

I wish to acknowledge the cooperation of William Kappel, U.S. Geological Survey, Ithaca, for permitting me to participate in the aquisition of drilling data from the Irondequoit Basin. William Shearer, Town of Webster, and Wesley P. Moody, N.Y. State Department of Transportation, Albany, provided copies of relevant boring logs from the Irondequoit Bay region.

## REFERENCES

Anderson, T.W. and Lewis, C.F.M., 1985, Postglacial water-level history of the Lake Ontario Basin: in Karrow, P.F., and Calkin, P.E., eds., Quaternary Evolution of the Great Lakes: Geological Association of Canada Special Publication 30, pp. 231-253.

Chadwick, G.H., 1917, The lake deposits and evolution of the lower Irondequoit Valley: Proceedings of the Rochester Academy of Science, v. 5, pp. 123-160.

Fairchild, H.L., 1923, The Pinnacle Hills or Rochester Kame Moraine: Proceedings of the Rochester Academy of Science, v. 6, pp. 141-194.

Fairchild, H.L., 1928, Geologic Story of the Genesee Valley and Western New York: Published by the author, 215 p.

Muller, E.H., 1977, Quaternary Geology of New York, Niagara Sheet: N. Y. State Geological Survey Map and Chart Series No. 28, Scale 1:250,000.

Muller, E.H. and Cadwell, D.H., 1986, Surficial Geologic Map of New York,

Finger Lakes Sheet: N. Y. State Geological Survey Map and Chart Series No. 40, Scale 1: 250,000.
Muller, E.H., Braun, D.D., Young, R.A. and Wilson, M.P., 1988, Morphogenesis of the Genesee Valley: Northeastern Geology, v. 10, no. 2, pp. 112-133.
Young, R.A., 1983, Glacial Geology in west-central New York: A review of recent progress: Proceedings of the Rochester Academy of Science, Centennial Colloquium Issue, v. 15, no. 2, pp. 85-98.

## ROAD LOG

This section of the road log begins at the intersection of Interstate Routes I-390 and I-590 south of Rochester, about 9 miles from the Rush Exit (end of Muller road log).

00.0 **KEEP RIGHT** at I-390 / I-590 split; take I-590 north. I-590 trends east here and parallels the Pinnacle Hills Moraine, which can be seen as a conspicuous ridge about 1 and 1/2 miles to the north. **CONTINUE** on I-590 north through "can-of-worms" construction toward Lake Ontario.

09.0 **TAKE EXIT** for Route 104 East, Webster.

11.2 Cross Irondequoit Bay Bridge (view sand bar at north end of bay) and **EXIT** at Bay Road (first exit east of bridge).

12.3 **STOP 1.** Cross over Rt. 104 on Bay Road and return to westbound lane of Route 104, exiting at Rest Area near east end of bridge. View of bay and location of cross section in Figure 5. From the parking lot the group will walk north along the bay bluffs (Figure 1) through the DeWitt well field for the Town of Webster. The hike is about 2000 feet (one way) and includes a 150-foot climb down the lake bluffs to view the section of Lake Iroquois sediments. The first part of the walk is through a gravel pit, which appears to contain a transition from Lake Iroquois sediments upward into glacial outwash, or a marginal facies of the lake sequence where the ice was grounded near the shore.

12.7 **STOP 2.** Includes hike across morainal/eolian topography at crest of bluffs and inspection of bluff section. Composite of well logs in well field will be available.

**RETURN** to Rt. 104 (westbound) and connect with I-590 south to Empire Blvd. Exit.

16.1 **EXIT** onto Empire Blvd. heading east toward head of bay. Cross Irondequoit Creek at head of bay and observe exposures of Lake Iroquois sediments in bluffs near road (brief stop if time permits).

18.9 **LEFT** at intersection with Bay Road.

19.9 **TURN** off (left) to Glen Edith.

20.4 **STOP 3.** View of exposed section of Lake Iroquois sediments resting on red till in road cut (access restricted, private property). Good view from road apron.

20.9 **RETURN** to Bay Road, **TURN RIGHT** (south).

24.0 Bay Road crosses Empire, becomes Creek St.; **CONTINUE** to Browncroft Blvd. **TURN RIGHT** (east). Note that site of longest boring on Figure 3 is on hill descending toward Irondequoit Creek (no place to stop bus).

25.1 **STOP 4.** Valley View Party House turnoff (old Browncroft Blvd.). View drilling sites, sedimentary section, and location of Figure 3.

25.5 **RETURN** to Browncroft Blvd. (west) and continue 0.2 miles to Landing Road. **TURN LEFT** on Landing Rd. to Blossom and **LEFT** again toward Ellison Park entrance off Blossom Road.

27.1 Entrance to Ellison Park.

27.7 **STOP 5.** Enter park and stop at exposure of glacial outwash covered by Lake Iroquois sediments. Note location relative to Stop 4 on Figure 1. This section represents outwash associated with the Pinnacle Hills Moraine immediately to the north. Drilling between Stops 4 and 5 intersected fresh, gray till at depth, rather than the red tills seen to the north and at the sites along Browncroft Blvd.

Official end of formal stops. Time permitting, the bus will return to Geneseo via the southern Irondequoit Creek Valley through the large kame-esker, ice-contact drift complex that formed as the ice withdrew toward the Pinnacle Hills position (Powder Mill Park). Exposures and surface geomorphology in this poorly studied area suggest that geomorphic interpretations of "obvious" glacial landforms are not without pitfalls.