LATE PLEISTOCENE HISTORY OF NORTHWESTERN NEW YORK

by Parker E. Calkin

Introduction

A number of excellent surficial geological studies have been made over northwestern New York. The most comprehensive were done by Leverett (1902), and by Kindle and Taylor (1913). These authors and others, including: J. W. W. Spencer (1881-1915), G. K. Gilbert (1887-1908), A. W. Grabau (1901-1920), H. L. Fairchild (1902-1932), F. G. Taylor (1895-1939) have roughly delineated moraines, glaciolacustrine features (plate 2), and have discussed chronologies of glacial retreat, lake formation, and the Niagara Gorge formation. However, with the exception of two recent areal studies (Blackmon, 1956 and D'Agostino, 1957), and local studies by E. Muller, these works were based largely on topographic features. Insufficient detail has been uncovered to clearly define the end moraines and drift sheets in most of the northwestern New York area. Furthermore, there was never any agreement among authors as to the number or importance in the glaciolacustrine chronology of moraines that were delineated.

Although in recent months, the writer and others have begun an attempt to reevaluate and add to the knowledge of this area, this paper is largely a compilation of the published literature. Emphasis is placed on features observed in following the road guide accompanying this report.

Evidence from striae and drumlins of the latest ice advance and interpretation of heavy mineral provenance (Dreimanis, 1961; Connally, 1964) suggests that Western New York was strongly affected by the Erie Glacial Lobe which was fed by an ice flow northeast or north-northeast of Lake Ontario. During its general southwesterly advances, ice also spilled out southeasterly from the Erie Basin into western New York.

Pre-Valley Heads (Port Huron?) Time

Preliminary studies suggest that most of the morphologic glacial features of northwestern New York may be attributed to the Port Huron Stade (Late Cary) and the following interstade; however, evidence of prior glaciation is found in adjacent area (see fig. 1).

In southwestern New York, the existence of pre-Illinoian glaciation may be inferred in the Salamanca re-entrant (Allegany Park area) from relationships of subsequent deposits (Muller, 1965). In the same area, terrace remnants containing possible morainal material above the Allegany River may be Illinoian in age (MacClintock and Apfel, 1944). Isolated erratics and spotty occurrence of till mapped by Muller (1963) southeast of the Wisconsin terminal moraine in Chautaugua County may belong to the outer-phase Illinoian of Shepps et. al. (1959).

At the well-studied Don and Scarborough beds locality near Toronto, the clayey York Till resting on Ordovician bedrock is considered by most to be of Illinoian age. Here also, much of the Sangamon interglacial episode is represented by the overlying Don beds which contain more than 70 species of trees and herbs, 20 diatoms, and 6 mammals. These fossils suggest that the maximum mean annual temperature was at least 6°F warmer than at present (Terasmae, 1960; Goldthwait et. al., 1965).

Correlations of many fossil organic sites in Quebec (Terasmae, 1958) Southern Ontario (Terasmae, 1960; Goldthwait et. al., 1965), Ohio (Goldthwait et. al., 1965) and in Cattaragus and southwestern Erie Counties, New York (Muller, 1964; 1965) indicate a very complex history of glacial fluctuation during the Wisconsin Glaciation. In the London to Toronto area of southern Ontario, four distinct Wisconsin advances prior to the Port Huron are represented by the Sunnybrook, Southwold, and Port Stanley tills which in turn help define the St. Pierre, Port Talbot, Plum Point and Lake Erie interstades. Finite dates of 52,000 B.P. and 63,000 B.P. by the Groningen (Netherlands) Laboratory on peat beneath gravel, till, and thin lacustrine beds at Otto, Cattaragus County (Muller, 1964) may suggest correlation respectively to Scarborough (St. Pierre interstade) beds and Sunnybrook till of the Toronto area.

Elsewhere in southern New York, the Olean Moraine (MacClintock and Apfel, 1944) and the Kent (Binghamton), Lavery, and Hiram tills of Chautauqua County (Muller, 1963) may be correlated with the drifts of southern Ontario. Till and interstade beds exposed in banks of Clear Creek near Gowanda in southernmost Erie County have been dated at greater than 38,000 years and may correlate with the Otto or St. Pierre interstade beds (Muller, 1960; 1965).

Valley Heads (Port Huron?) and Later Time

End Moraines

The last significant glacial readvance in western New York is probably marked by the outstanding, Valley Heads/Lake Escarpment end moraine ridges of southwestern New York (plate 2). This morainal system, the most continuous in New York State, is characterized by strong topographic expression with conspicuous kame knobs and kettle depressions. It is presently considered to be of Late Cary age and probably in part equivalent to the Port Huron Moraine of Michigan. Analysis of spruce from marly silt overlying gravel near a mastodon site in southwestern Erie County, yields a minimum date of 12,020 B.P. for recession from this terminal moraine (Muller, 1963). Muller (1963, p. 48) notes "that this date may be measureably younger than the Valley Heads maximum is afforded by --" an average of 12,370 B.P. for two dates taken on spruce from proglacial Lake Iroquois deposits at Lewiston, New York. Evidence from western New York, of recessional history from Cary (Kent) terminal moraine provides no adequate demonstration of the magnitude of the readvance to the Port Huron

(Valley Heads Moraine) [Muller, personal communication]. However, the probable correlatives of the Valley Heads Moraine near Hamilton, Ontario the Paris and Galt Moraines (Goldthwait et. al., 1965; Muller, 1963) may have followed a pronounced retreat of the Erie Lobe (Karrow, 1963).

Retreat from the Valley Heads border is marked in western New York by a series of recessional moraines, from south to north: Gowanda, Hamburg, Marilla, and Alden Moraines of Leverett (1902); the Buffalo Moraine of Kindle and Taylor (1913); and the Barre and Albion Moraines of Leverett (1902). In most cases, the tracing of these moraines is difficult. In the Erie and Ontario Lowlands, they were laid down in proglacial lakes resulting in subdued topographic expression and poor continuity (see Fairchild, 1932). Because of the lacustrine deposition and heavy proglacial drainage, much of the morainal material is very sandy and kame moraine ridges are common.

The Gowanda Moraine is probably related closely in age to the Valley Heads and has been reduced in most areas to an erosional boulder remnant of Lake Whittlesey action. The Alden and Marilla Moraines may be distinguished from the Hamburg Moraine largely through glacial marginal drainage channels (Fairchild, 1932). The latter extends from Hamburg 25 miles into the northwestern part of the adjoining Wyoming County where it connects with the interlobate (Erie/Ontario Lobes) moraine of the Valley Heads system near Batavia. With sharp knolls of 20 to 50 feet relief, together with inclosed basins and sloughs; and also with its great width near Batavia (plate 2), the Hamburg Moraine is by far the most conspicuous ridge in northwestern New York. The other moraines for the most part show relief of much less than 50 feet or are identified by slight boulder concentrations only.

The extent to which these moraines represent advances or significant halts in recession of ice from the Valley Heads moraine in western New York is not known. However, the glacier had retreated north of the Niagara Escarpment by 12,370 B.P. and probably north of the Ontario Basin, never to return into western New York, by 12,500 to 10,500 B.P. according to bracketing dates of Lake Iroquois (Karrow et. al., 1961).

Glacial Great Lakes

The correlation of events in the Erie - Ontario Lake Basin with "Late" Pleistocene chronology of the upper Great Lakes has posed a serious problem to some (See Hough, 1963; Bretz, 1964; Wayne and Zumberge, 1965; Hough, 1966), but most workers now agree that proglacial Lakes Whittlesey, Warren, Wayne, Grassmere, Lundy and/or possibly Dana and an Early Algonquin existed prior to the Two Creeks interstade and formation of Lake Iroquois.

The highest Great Lakes strandline recognized in Erie County and western New York rises from 778 feet at the Pennsylvania State Line to 910 feet, 73 miles northeast at East Aurora (plate 2). Here it dies out at the position of the former ice border. This tilted strand is dated as 12,660 B.P. in southern Ontario (Goldthwait et. al., 1965; Hough,

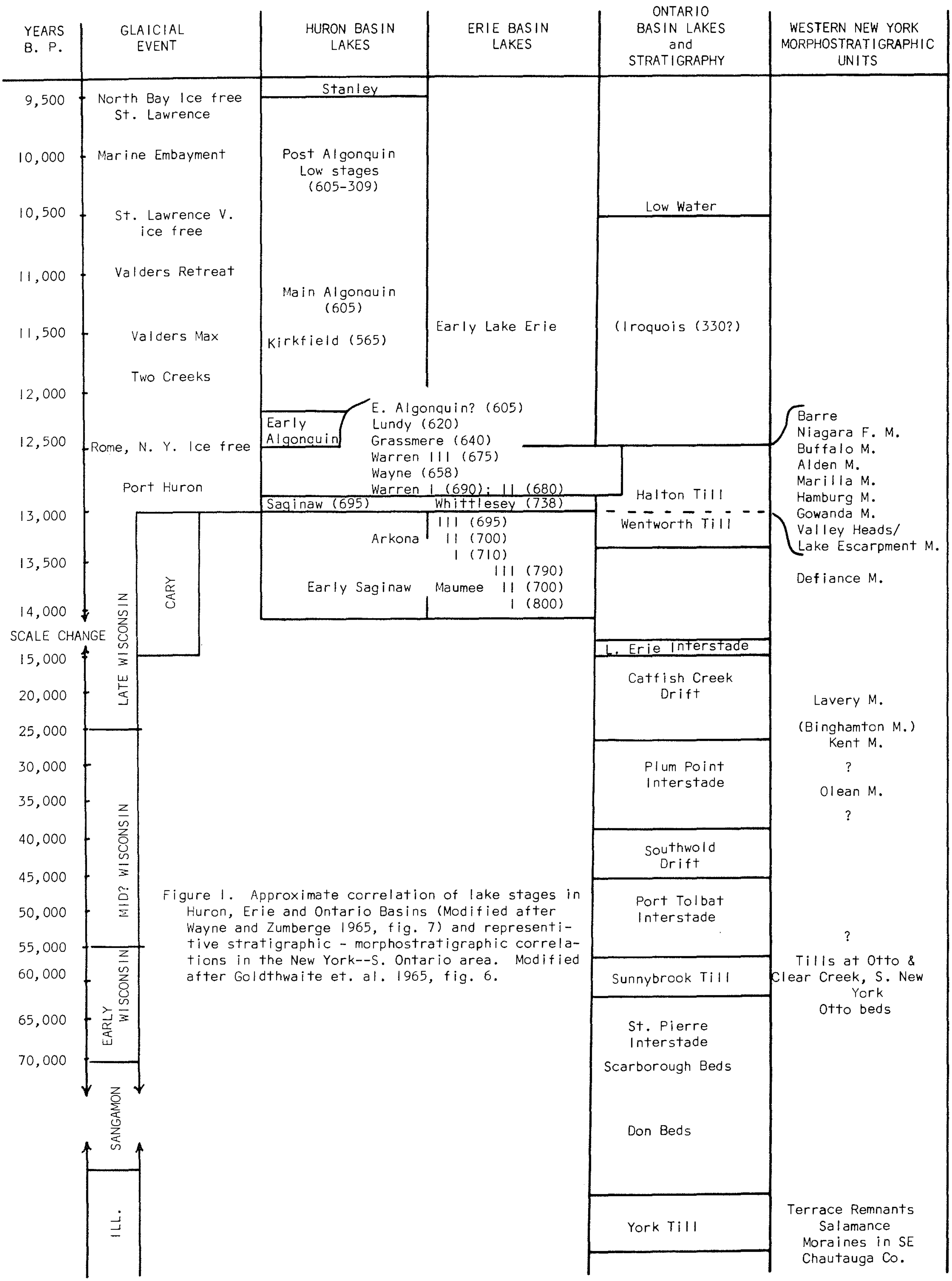
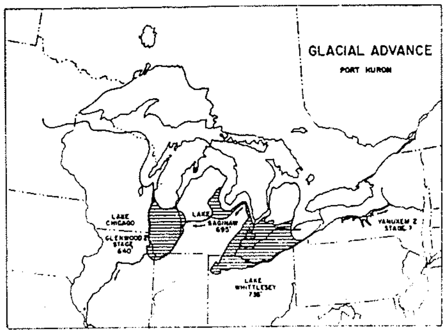


Figure I. Approximate correlation of lake stages in Huron, Erie and Ontario Basins (Modified after Wayne and Zumberge 1965, fig. 7) and representitive stratigraphic - morphostratigraphic correlations in the New York--S. Ontario area. Modified after Goldthwaite et. al. 1965, fig. 6.



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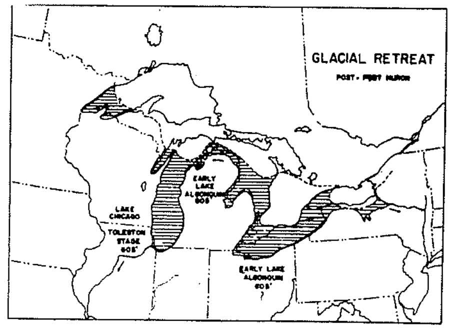

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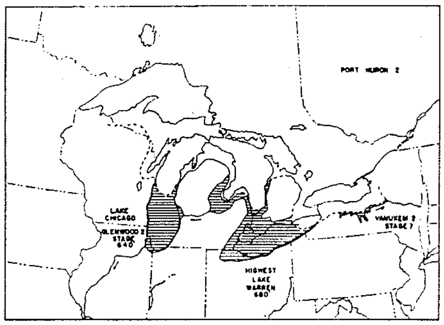

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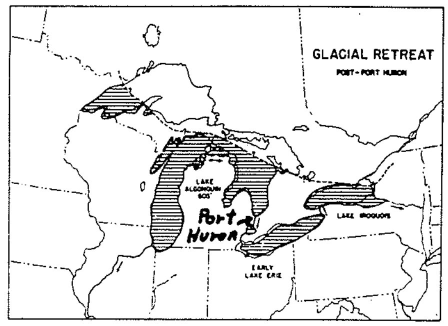

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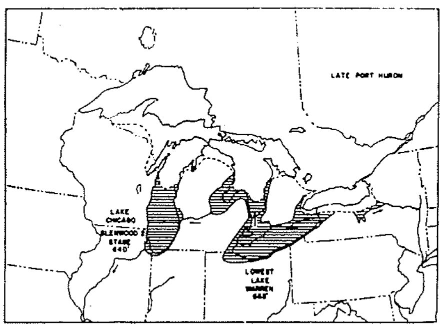


2C

Figure 2. Some stages of the Great Lakes. After Hough (1963).

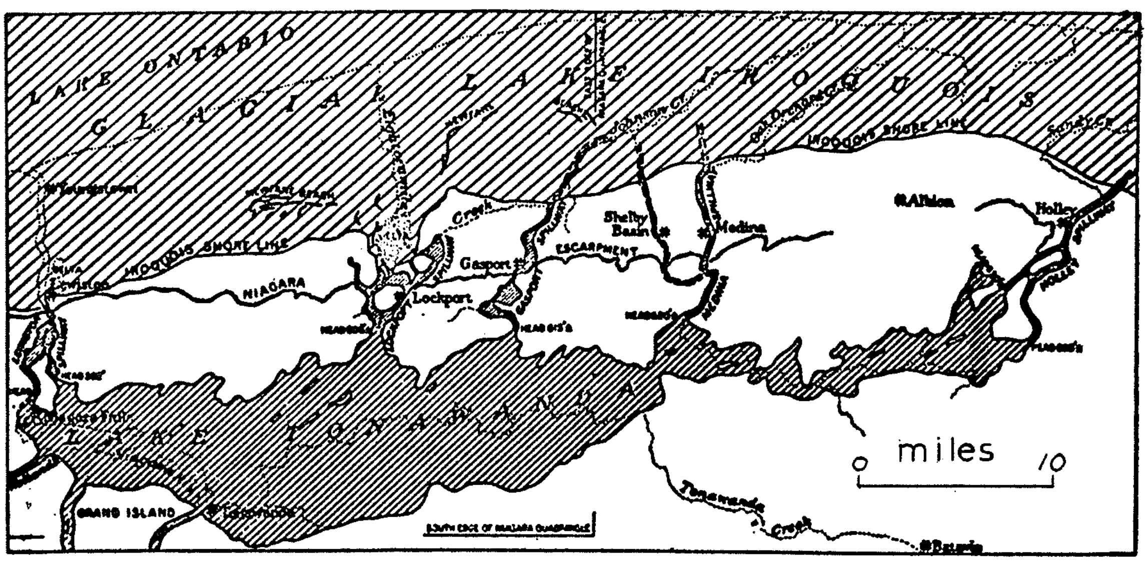


Figure 3. Sketch map of Lake Tonawanda and spillways. Shoreline of Lake Tonawanda approximate. After Kindle and Taylor (1913).

1963) and is correlated with the Lake Whittlesey stage (fig. 2A). This lake may have been initiated at Port Huron (Valley Heads) time and discharged westward across the thumb of Michigan. Lake Whittlesey, at its peak twice the size of Lake Erie, was slowly extinquished with retreat of ice from the Hamburg Moraine to the position marked by the Alden Moraine (Leverett, 1895). Its waters escaped westward through the Grand River channel to Lake Chicago (fig. 2A).

A lower outlet allowed formation of Lake Warren (fig. 2B), which according to Hough (1963) had three phases, the last (fig. 2C) following an intervening lower water stage of Lake Wayne. The lowest Lake Warren strand may be dated about 12,000 B.P. or 11,860 by organic material from a Tupperville, Ontario sight (Dreimanis, 1964). In the area of East Aurora (plate 2), the first Lake Warren beach occurs as a single ridge at about 840 feet and "lower" Warren beaches occur as short, multiple ridges from 810 to 840 feet (Blackmon, 1956).

Weekly developed strandlines from 810 to 770 feet in the East Aurora area may relate to, respectively, Lake Wayne, Grassmere, and Early Lake Lundy. Occurrences of beaches assigned to these Great Lake stages are not plentiful in northwestern New York and levels are of uncertain correlation (see Kindle and Taylor, 1913).

The assignment of a number of beach ridges at about 700 feet and lower is uncertain. Fairchild's (1906) and Hough's (1958) Lake Dana stage, an assumed intermittent stage between Warren and Iroquois, may well be Kindle and Taylor's (1913) low stage of Lake Lundy or Hough's (1963) Early Lake Algonquin (fig. 2D [See Wayne and Zumberge, 1965]. The outlets for these lakes is not well defined (Hough, 1963), but may have been eastward to the Marcellus Channels near Syracuse (Fairchild, 1906).

As the ice margin retreated from the Niagara Falls Moraine and subsequently from the Barre Moraine positions in western New York (plate 2), and from the divide between the Oneida and Mohawk basins to the east (Muller, 1965), the Mohawk outlet to the Hudson River was opened. The resulting gradual lowering of the lake level (Early Lake Algonquin--fig. 2D) to below the Niagara escarpment initiated formation of Early Lake Erie and Lake Iroquois and necessarily, the Niagara River, Niagara Gorge, and Falls at Lewiston, New York (fig. 2E). Dates for this event from New York and Ontario average about 12,370 B.P. (Goldthwait et. al., 1965). Glacial ice still blocked drainage east toward the St. Lawrence Valley (Karrow et. al., 1961).

Ridge Road (U. S. Route 104) which follows the Iroquois beach ridge gives clear demonstration of postglacial isostatic tilt as it rises eastward from about 377 feet at the Niagara River to about 450 feet above sea level 180 miles away near Rome, New York (Muller, 1965). Further ice retreat uncovering the northern end of the Adirondacks (MacClintock and Terasmae, 1960) and opening of the St. Lawrence

Valley caused draining of Lake Iroquois about 11,000 years ago. Because of isostatic depression by ice in the St. Lawrence Valley, the water level in Ontario Basin first dropped to 230 feet or more below the present level of Lake Ontario, an event dated at about 10,150 B.P. (Goldthwait et. al., 1965). Postglacial uplift of the outlet near Kingston, Ontario, has brought the lake to its present level of about 246 feet.

Lake Tonawanda

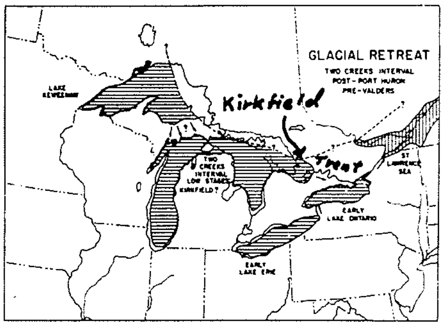
With glacial recession and lowering of the water level to Lake Iroquois, the Onondaga and Niagara escarpments were uncovered and waters pouring from the newly formed Detroit, St. Clair and Niagara Rivers respectively (fig. 2E), were impounded between the two cuestaform ridges forming Lake Tonawanda (fig. 3). During its early history, Lake Tonawanda extended nearly 58 miles from near Rochester to beyond Niagara Falls, Ontario and averaged 6 miles in width. The best developed beaches are at about 629 feet.

Lake Tonawanda had five outlets (fig. 3), each forming a falls and gorge where it drained over the Niagara escarpment to Lake Iroquois. Because of isostatic tilt, illustrated by Ridge Road, and because of its proximity to Lake Erie, outlets at Lewiston and Lockport carried most of the discharge. At Lockport, the largest spillways were the Gulf and the northeast trending gorge now utilized by the Erie and New York State Barge Canals. Subsequently, because of more rapid incision of the Lewiston outlet and because of sedimentation, Lake Tonawanda decreased in size. Oak Orchard swamp east of Lockport, and other swamps now surrounding Tonawanda Creek are successors of Lake Tonawanda.

Formation of Niagara Falls and Gorge

Niagara River and Falls formed, and cutting of the Gorge into the escarpment began at Lewiston with the opening of the Mohawk Valley and subsequent draining of Early Lake Algonquin? water to form Early Lake Erie and Lake Iroquois. The initiation of gorge-cutting, previously dated by Lake Iroquois at 12,080 to 12,600 B.P., occurred at Lewiston, N. Y. rather than at the mouth of the buried St. Davids' Gorge, Ontario a few miles to the west (fig. 5). The reason for not using the old channel appears to be that its drift fill and the over-topping Barre Moraine (Plate 2) put its effective floor some 60 feet above the escarpment level at Lewiston (Taylor, 1933).

Thereafter, through more than 12,000 years the notch of the Falls has retreated approximately seven miles to its present position. The height of the Falls has been maintained through this period by the tough, Middle Silurian, Lockport Dolostone caprock, while undermining through the weaker Rochester Shales and other, less competent formations



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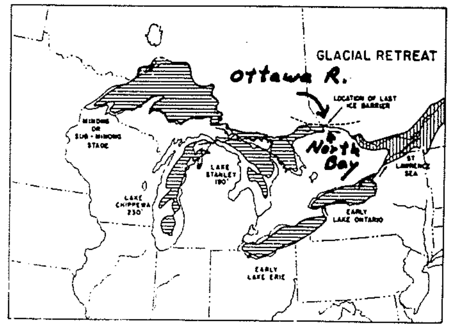

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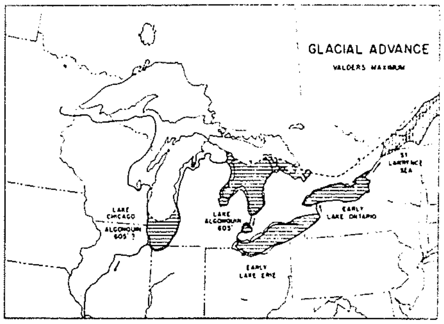

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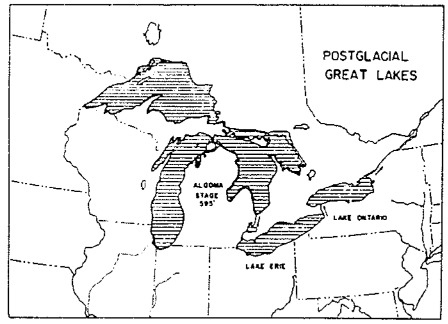

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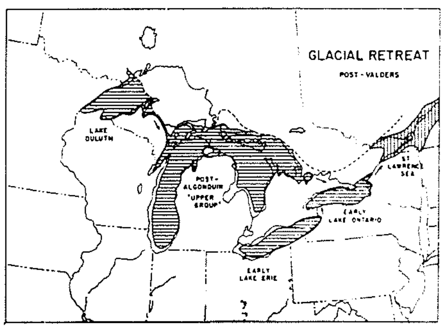

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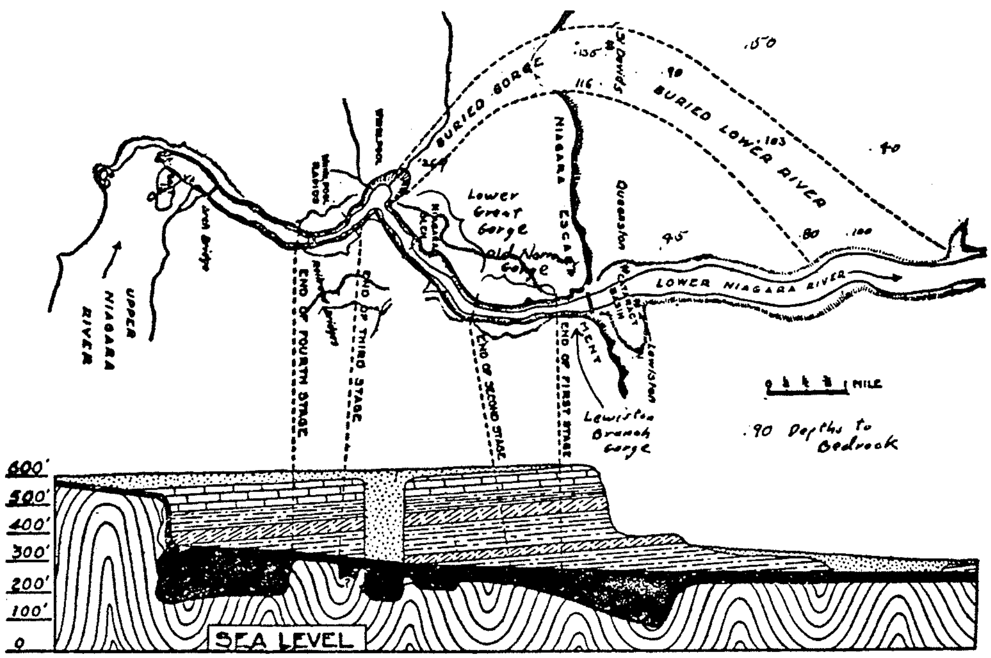
Figure 4. Some stages of the Great Lakes. After Hough (1963).



Figure·5. Sketch map of Niagara Gorge showing the varying width and depth. Vertical section is along Gorge and River with Vertical scale greatly exaggerated. Depth of water shown in black. Modified after Forrester (1926 & 1928).

below caused a joint block by joint block recession (see associated papers for stratigraphic column).

Correlation of Niagara Gorge Enlargement With Great Lakes History

The time involved in the cutting of the Niagara Gorge was at one time considered a reasonably accurate measure of the duration of time since deglaciation. Estimates of postglacial time obtained in this manner vary by a factor of ten, ranging from as little as 3,000 years to more than 35,000 years (Kindle and Taylor, 1913). The principle complicating factor comes with the realization that the volume of water discharging through the Niagara River has varied considerably since its initiation. Such changes resulted from the varying size and number of outlets, other than through Lake Erie, afforded the upper Great Lakes waterbodies by action of the former ice margin. The various sections of the Niagara Gorge shown in figure 5, were correlated with the Great Lakes History by Kindle and Taylor in 1913 and by Taylor again in 1933. Since this time, radiocarbon dating and more detailed research elsewhere in the Great Lakes Basin makes this correlation out of date and a new interpretation, based on the present data available, is given below. Although such correlation is highly conjectual and based on meager proof, it may serve as a very temporary base of reference and help to illustrate how variation of discharge controlled the rate of Niagara Falls, recession. The interpretation below is based on the Great Lakes History according to Hough (1963) and others as outlined below and is partially illustrated in figures 2 through 5.

1. Cataract Basin and Lewiston Branch Gorge Sections

Moderate discharge required is correlated with formation of Lake Tonawanda outlets and Chicago outlets for Lake Algonquin and Early Lake Erie (12,000 B.P. to 11,850 B.P.) See figures 2E and 5.

2. Old Narrow Gorge Section

Decrease in volume of discharge is correlated with ice recession, the Two Creeks Interstade, and consequent opening of the Trent River outlet at Kirkfield, Ontario. The upper lakes (Algonquin/Kirkfield) then drained directly into Lake Iroquois. Carrying only discharge of Lake Erie, a narrow and shallow gorge section was cut. See figures 4A and 5.

3. Lower Great Gorge Section

Increase in volume is correlated with: a) Valders ice advance and subsequent blocking of outlet to the Trent Valley (Hough, 1963); or with b) isostatic rebound of the outlet at Kirkfield. Either event subsequently returned the Main Algonquin discharge through Lake Erie. At Niagara University (head of Old Narrow Gorge) the gorge widens perceptably and deepens slightly, the greater width continuing southward nearly to the Whirlpool. At the Whirlpool, the southwestward retreating Falls intersected the previously cut, but drift filled St. Davids' Gorge at right angles thereupon turning sharply southeastward to reexcavate the head of the older gorge. See figures 4B and 5.

4. Whirlpool Rapids Section

The decreased volume at this stage is correlated with recession of the ice front from the Valders terminal moraine and opeining successively of outlets at Kirkfield to the Trent Valley and at North Bay, (9,500 B.P. after Terasmae and Hughs, 1960) to Lake Ontario and to the St. Lawrence River respectively (see Chapman, 1954). See figures 4C, 4D and 5. It is possible that as the ice front reached North Bay, the Trent River outlet at Kirkfield was closed by isostatic recovery. A narrow gorge may have been cut south of the area shown in figure 5; however, this section would have been enveloped by cutting of the Upper Great Gorge.

5. Upper Great Gorge Section

The large increase in discharge through the Falls necessary to initiate this section is correlated with abandonment of the North Bay outlet to the Ottawa River due to isostatic rebound. For the first time, the entire discharge of the three upper Great Lakes flowed through two southern outlets (Hough, 1963). The Chicago outlet, resting on bedrock, could not cut deeper; however, the Port Huron outlet to Lake Erie, resting in till, was cut down and further concentrated the flow through Lake Erie and over the Falls (4,200 B.P. after Dreimanis, 1958; Farrand, 1962). Separation of the Horseshoe and American Falls is estimated to have taken place from 600 to 700 years ago (Taylor, 1933).

Origin of St. Davids' Buried Gorge

Another problem of the Niagara Falls area is the origin of the drift-filled St. Davids' Gorge (fig. 5). Apparently as deep and nearly as wide as the Upper Great Gorge (Kindle and Taylor, 1913; Forrester, 1926; International Joint Commission, 1953), it extends from the Whirlpool to the town of St. Davids, Ontario at the Niagara escarpment. As evidenced by deep drilling in the Iroquois Plain, it extends northward from St. Davids across the Iroquois Plain to the Lower Niagara River. Its length and apparent cross section would suggest that it was carved by a discharge approaching that of the present Niagara River and may have taken more than 4,000 to 8,000 years to form. Recent palynological study by Dr. Paul Karrow may suggest an "Early" Wisconsin or Sangomon origin for some units of the drift fill (J. Terasmae, personal communication). Geophysical survey and stratigraphic drilling are being undertaken currently by Dr. Terasmae of the Geological Survey of Canada.

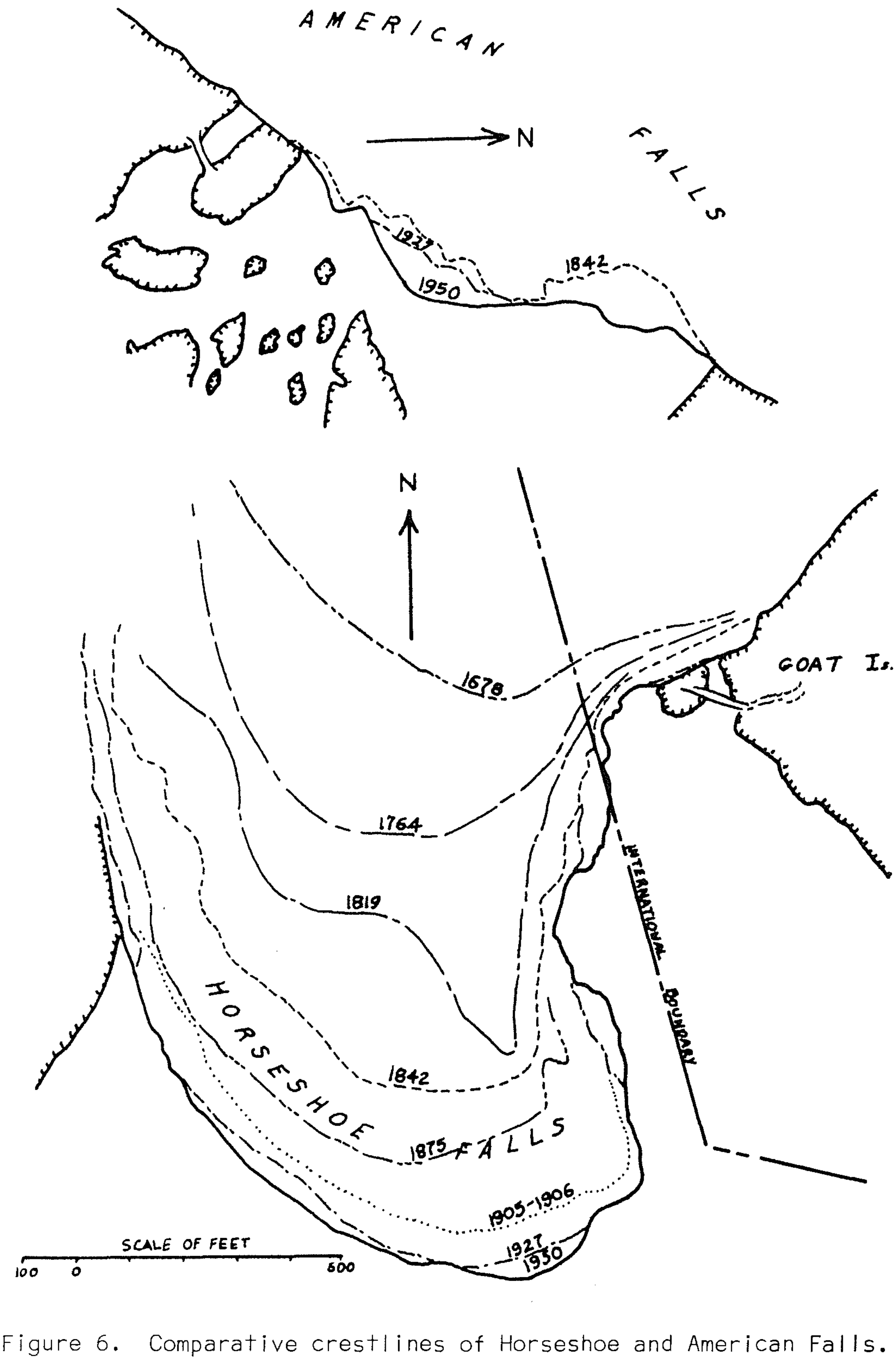
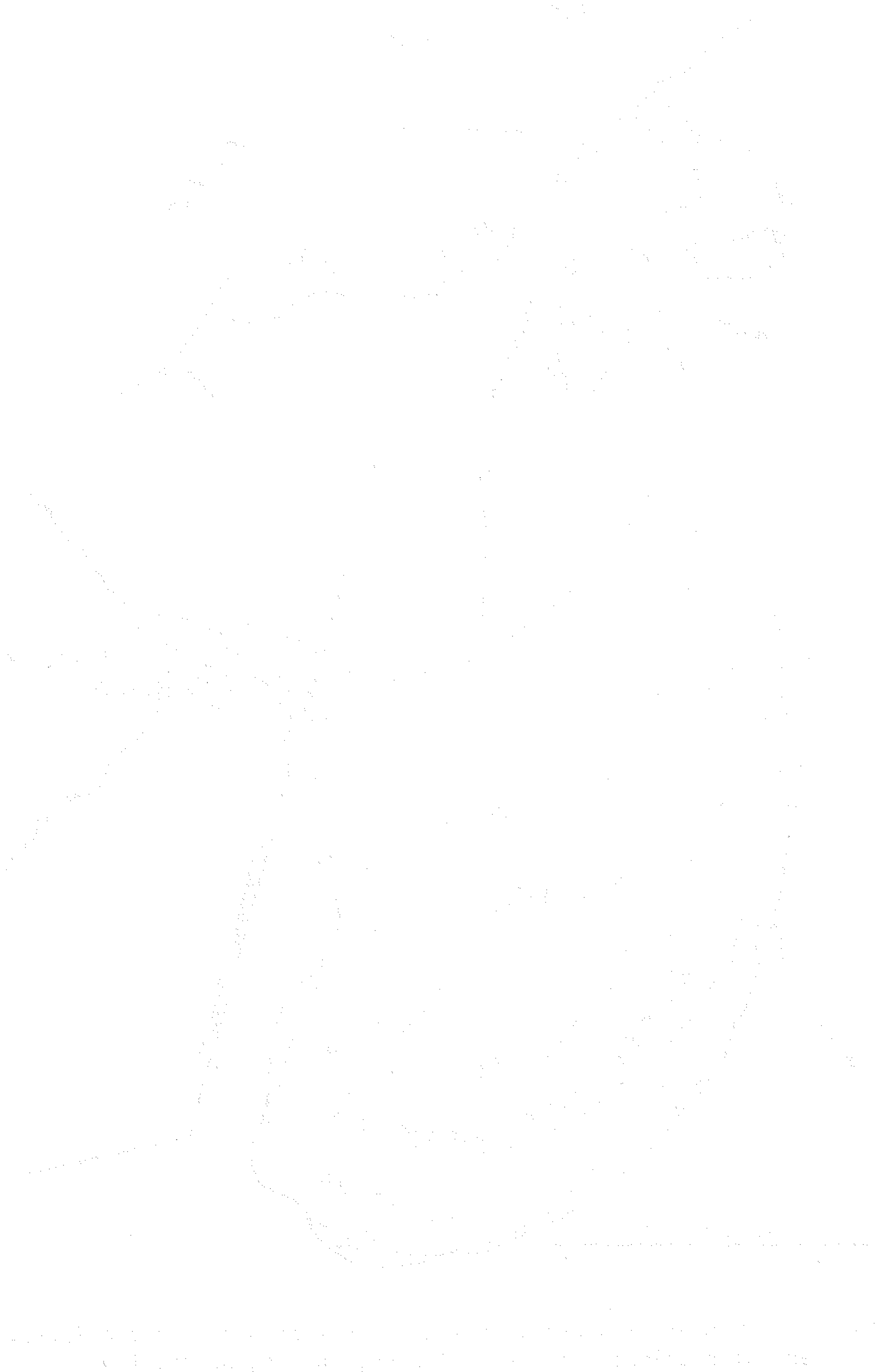


Figure 6. Comparative crestlines of Horseshoe and American Falls. After International Niagara Falls Engineering Board, 1953.



Niagara Falls in Recent Time

The Niagara River and hence the Niagara Falls continues to carry the surplus water of the upper Great Lakes seaward from Lake Erie to Lake Ontario. The mean flow of the river is about 200,000 ft^3/sec. and because of the immense stroage capacity of the upper lakes, the flow is remarkably steady. The normal flow is measured by a few thousand cubic feet per second which are diverted into Lake Superior from the Albany River watershed in Canada, and it is reduced by somewhat similar amounts diverted by the Chicago Sanitary and Ship Canal from Lake Michigan into the Mississippi River, and by the Welland Canal and the New York State Barge Canal directly into Lake Ontario (International Joint Commission, 1953).

A considerable body of information regarding recent rates of recession of the Horseshoe and American Falls is summarized in figure 6. Average recession for the Horseshoe Falls has apparently decreased from an average of 4.2 ft/yr between 1842 and 1906, to 3.2 ft/yr between 1906 to 1927, to 2.2 ft/yr from 1927 to the last survey in 1950. Several factors contributing to this reduction of cutting include: 1) the regional dip of the capping Lockport Dolostone, diminishing the height of the Falls by about 20 ft/mile; 2) a southward thickening of the capping Lockport Dolostone; 3) diminishing discharge of the Niagara River as a result of increased diversion for power purposes (10,000 ft^3/sec. in 1906, to greater than 100,000 ft^3/sec. at present).

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