GEOLOGY OF THE NORTHERN APPALACHIAN BASIN WESTERN NEW YORK

Field Trips Guidebook for New York State Geological Association 54th Annual Meeting

October 8 — 10, 1982
Amherst, New York

Department of Geological Sciences
State University of New York
at Buffalo
Edward J. Buehler and Parker E. Calkin
Editors

In Conjunction With
11th Annual Meetings Eastern Section
American Association of
Petroleum Geologists

State University of New York at Buffalo

DEPARTMENT OF GEOLOGICAL SCIENCES

FACULTY OF NATURAL SCIENCES AND MATHEMATICS

19 October 1982

Dear Guidebook Authors/ Field Trip Leaders,

Thank you again for your fine contributions to the 54th Ann. NYSGA Field Conference. I have heard good things about all the trips and I think the trips all went well in the eyes of participants. There is no real way that you can be compensated for all your work but I did have some extra guidebooks which I am distributing to each of you. In addition, I enclose some extra reprints for the first-listed guidebook authors.

Yours sincerely,

Parker

Parker E. Calkin
General Chairman 1982

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NEW YORK STATE GEOLOGICAL ASSOCIATION

54th ANNUAL MEETING

October 8-10, 1982

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GUIDEBOOK FOR FIELD TRIPS IN WESTERN NEW YORK, NORTHERN PENNSYLVANIA AND ADJACENT, SOUTHERN ONTARIO

Edward J. Buehler

and

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Editors

Department of Geological Sciences
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Published by the New York State Geological Association. Guidebook available from the executive secretary: M.P. Wolf, Geology Department, Gittleson Hall, Hofstra University, Hempstead, New York 11550.

HISTORY OF THE NEW YORK STATE GEOLOGICAL ASSOCIATION (NYSGA)

The first intercollegiate state field meeting was organized at Hamilton College in Clinton, N.Y., by Nelson C. Dale in 1925 (we think we are the oldest state intercollegiate geological association in the nation). The meetings became an annual event and consisted of an evening banquet and a Saturday trip to local outcrops near the host institution. A few hand-out sheets were prepared by the field director for this occasion. These early trips emphasized the stratigraphy, geomorphology, structural, glacial, and economic geology of the Pre-Cambrian, Lower Paleozoic and Pleistocene strata of the state - this pattern continues. The excursions were popular with 150-200 students, faculty and "working" geologists attending; personnel from the New York State Museum were also active contributors.

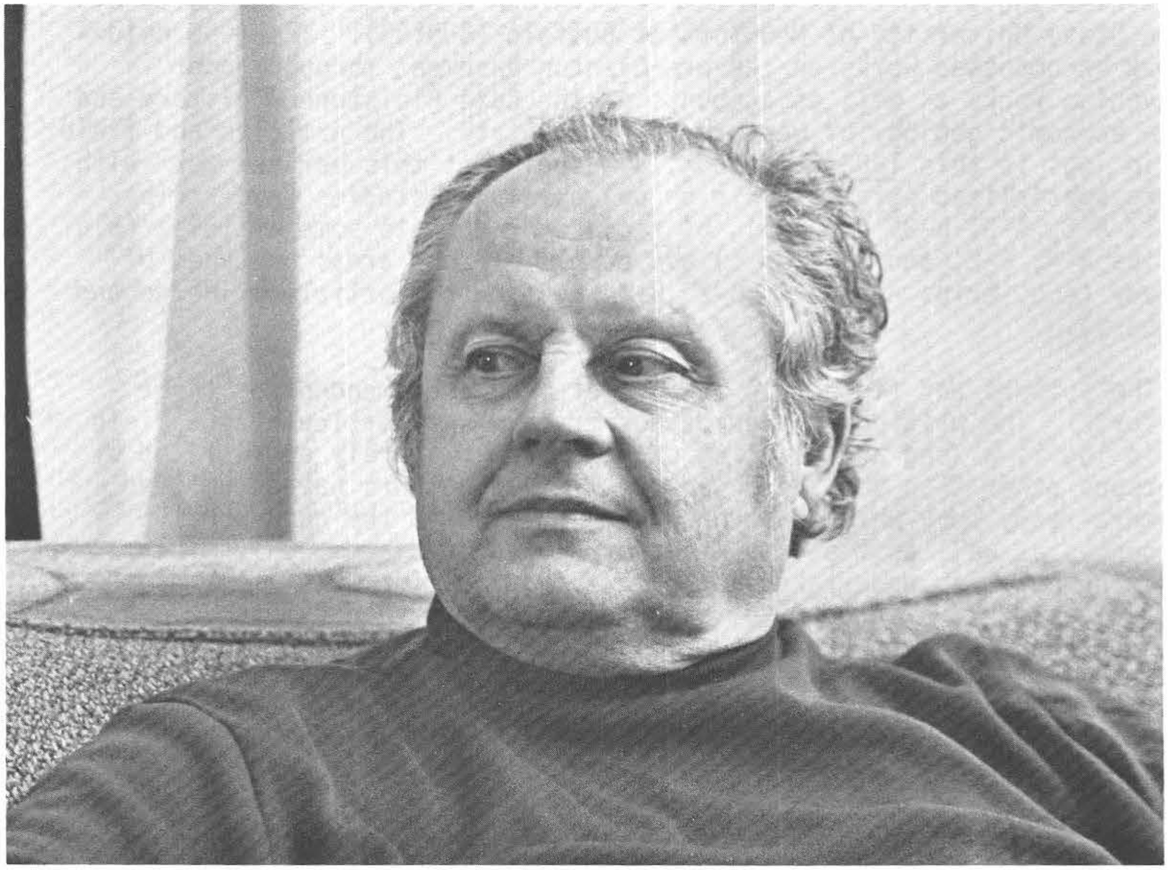
After the war years, 1942-45, the meetings were resumed and extended to cover two field trips (hard rock and soft rock) for the weekend; a small "guidebook" was also prepared. It remained a loose-knit, self-perpetuating, but unmanaged association until 1953 when Kurt E. Lowe (as "permanent secretary") undertook the responsibility of providing some organization. More host institutions were added, and in 1956 the first bound guidebook with a choice of seven trips were offered by the University of Rochester. All the guidebooks have been reprinted and sold since that time (see last page of guidebook for list and prices).

Through Kurt Lowe's efforts, more of a "routine" arrangement became established at each meeting, and the number of participants increased from 150 in 1954 to over 400 in 1965. The field guidebooks became more sophisticated and detailed, and now also included petrology, paleoecology, sedimentology, and engineering-environmental geology. They grew from 100 pages (1963) to over 400 pages (1977), but, as with the attendance, they are now decreasing to more manageable levels (see reverse side).

A constitution was developed by Kurt Lowe in 1969, when Phil Hewitt became the first Executive Secretary. Daniel F. Merriam took over the responsibility in 1973, and through the society's emphasis on stratigraphy, structure and paleontology, sought the establishment of a national affiliation with the American Association of Petroleum Geologists (A.A.P.G.).

A few years ago the organization celebrated its 50th annual meeting, and looks forward toward the further development and understanding of N.Y. State geology, and its resource potential and environmental problems during the next 50 years.

Presently under the direction of Fred Wolff, the Association has become incorporated, continues its affiliation with the national A.A.P.G , and still remains a loose-knit organization. It continues to host annual field trips throughout N.Y. State for its students and faculty, and promotes the sale of guidebooks to anyone interested in the regional geolgoy of this classic, diverse area.



DEDICATION TO

EDWARD J. BUEHLER
(1916 - 1982)

Professor Edward Buehler died on July 12th while much of this publication was in preparation. He had been instrumental in the organization of these joint meetings and of this guidebook. He also had organized the 1966 NYSGA meeting and guidebook. Ed loved western New York; he received his Bachelor's and Master's degrees from the University of Buffalo. In 1948 after receiving his Doctor of Philosophy degree in geology from Yale University, he returned to teach at his Alma Mater with a major emphasis on paleontology. In deep appreciation of his devotion and contributions to the understanding of western New York geology, this volume is affectionately dedicated.

FOREWORD AND ACKNOWLEDGEMENTS

In 1966 Dr. Edward Buehler assembled a comprehensive group of geologic articles and field trips for the first NYSGA meetings to be hosted by the State University of New York at Buffalo (SUNYAB). The 1966 guidebook encompassed works on the pre-Clinton basement through Upper Devonian rocks as well as accounts of the Late Pleistocene history and the economic geology of western New York State. The articles and field trip logs in this 1982 guidebook, a) supplement this earlier work with more specialized stratigraphic or paleontologic studies (e.g. trips A-1, A-2, A-3, B-3, and B-5), b) replace or up-date some articles (e.g. trips A-4, B-2, and B-3), or c) add completely new areas or types of study (e.g. trips A-5, A-6, B-2, B-4, B-5 and the article by Hodge and others).

The authors and field trip leaders are to be commended for their efforts. The writing of a guidebook article and particularly a field trip log is a long and exacting process that is rarely appreciated by persons other than the field geologist. The authors have submitted camera-ready copy which has considerably lessened the guidebook expenses as well as the time necessary for assemblage. Only minor editing has been undertaken. The fact that the format of some of the articles is at variance with the specific format requested should not hinder the articles' great usefulness.

The faculty and staff members of the Department of Geological Sciences have all contributed to the joint meetings with which this guidebook is associated; their help and cooperation is gratefully acknowledged. The guidebook has been printed and bound by the printing facility of the State University of New York at Buffalo. Funds for most of the printing have been generously provided by the Conferences in the Disciplines Committee of SUNYAB. This has allowed the Organizing Committee to maintain the tradition of low registration costs.

Parker E. Calkin
General Conferences Chairman and
President of NYSGA for 1982
Meeting

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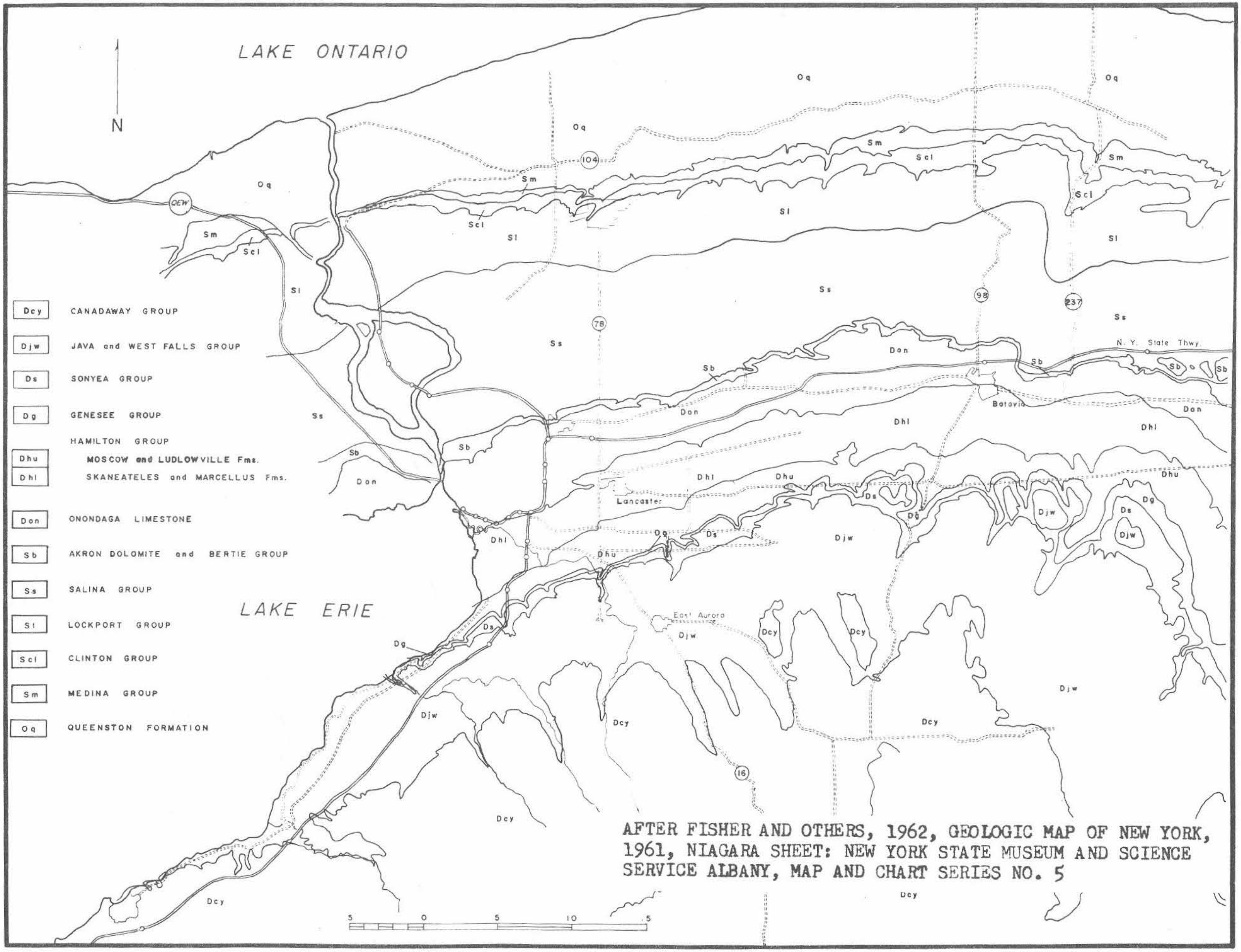
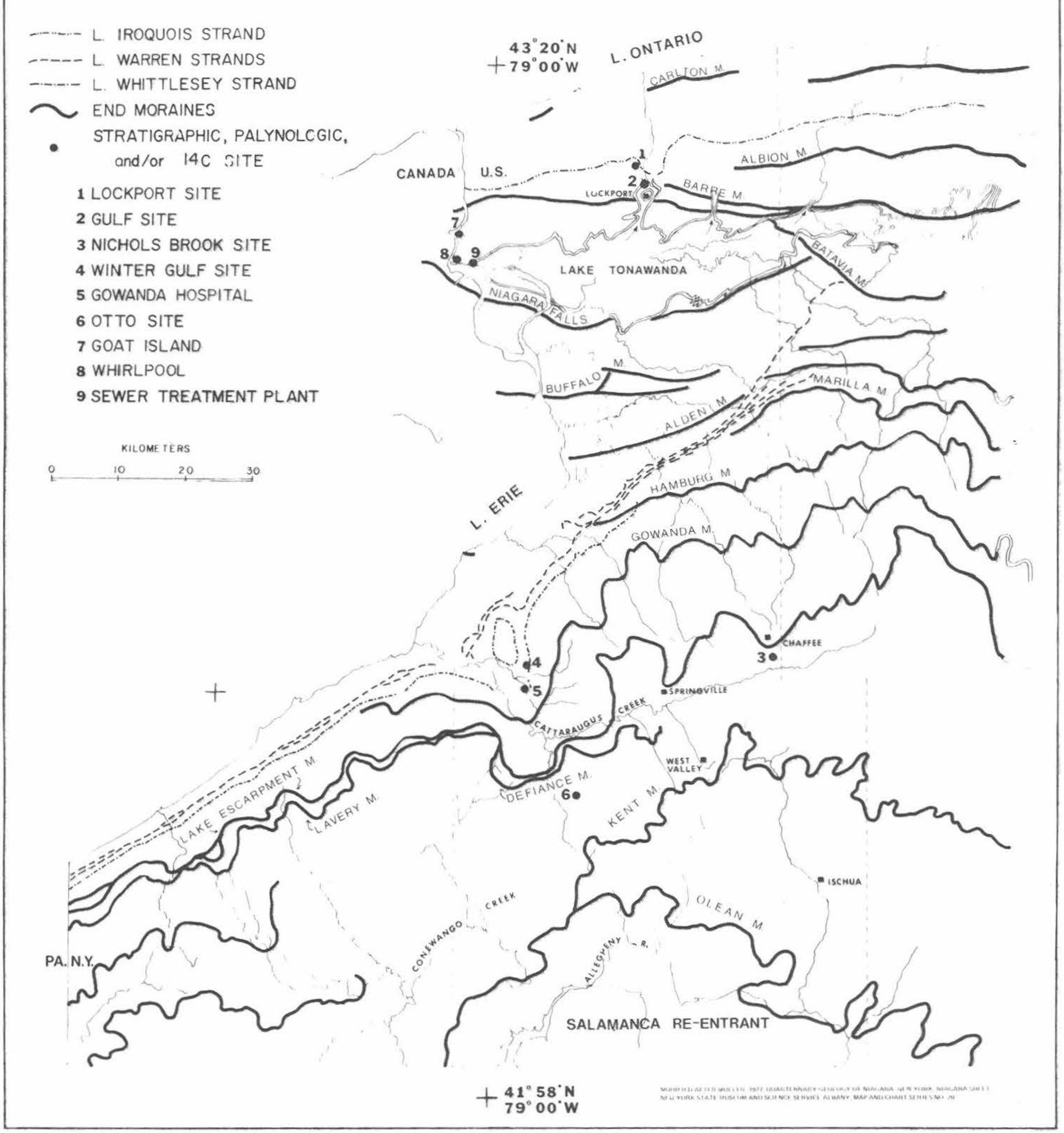


AFTER FISHER AND OTHERS, 1962, GEOLOGIC MAP OF NEW YORK, 1961, NIAGARA SHEET: NEW YORK STATE MUSEUM AND SCIENCE SERVICE ALBANY, MAP AND CHART SERIES NO. 5


L. IROQUOIS STRAND
L. WARREN STRANDS
L. WHITTLESEY STRAND
END MORAINES
STRATIGRAPHIC, PALYNOLCGIC, and/or 14C SITE
1 LOCKPORT SITE
2 GULF SITE
3 NICHOLS BROOK SITE
4 WINTER GULF SITE
5 GOWANDA HOSPITAL
6 OTTO SITE
7 GOAT ISLAND
8 WHIRLPOOL
9 SEWER TREATMENT PLANT
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CARLTON M.
ALBION M.
CANADA
U.S.
LOCKPORT
BARRE M.
LAKE TONAWANDA
BATAVIA M.
NIAGARA FALLS
BUFFALO M.
MARILLA M.
ALDEN M.
L. ERIE
HAMBURG M.
GOWANDA M.
CHAFFEE
SPRINGVILLE
CATTARAUGUS CREEK
WEST VALLEY
DEFIANCE M.
KENT M.
LAKE ESCARPMENT M.
LAVERY M.
ISCHUA
OLEAN M.
CONEWANGO CREEK
ALLEGHENY R.
PA. N.Y.
SALAMANCA RE-ENTRANT
41°58'N
79°00'W

GEOPHYSICAL SIGNATURE OF CENTRAL AND WESTERN NEW YORK

D.S. HODGE AND R. ECKERT - Department of Geological Sciences, State University of New York at Buffalo, 4240 Ridge Lea Road, Amherst, New York 14226

F. REVETTA - Department of Geological Sciences, State University of New York at Potsdam, Potsdam, New York 13676

INTRODUCTION

Central and western New York encompasses part of the northwest portion of the Appalachian basin. The structure of this region is simple with Paleozoic rocks that overlie Precambrian crystalline basement dipping gently to the south. The interpretation of the rocks in the basement depends principally on analysis of the gravity, magnetic and heat flow anomalies. Locally, seismic surveys have detailed the Paleozoic structure. This paper includes maps of the Bouguer gravity, aeromagnetic, and temperature gradients of central and western New York with a brief interpretation. Description of regional heat flow, and seismicity along with a cross-section of seismic stratigraphy of Paleozoic rocks provides additional geophysical data for geologic interpretation.

REGIONAL GEOLOGY

Within the central and western portion of New York State the geologic structure is relatively simple with Cambrian through Devonian shales and limestones dipping gently to the south. The thickness of this sedimentary sequence is about 900 meters at the shore of Lake Ontario and thickens southward to over 3050 meters. Precambrian crystalline basement rocks underlie these Paleozoic sediments and a thin veneer of glacial debris covers most of the area; glacial material may reach a thickness of 300 meters in some valleys. The Precambrian basement of New York State is exposed in two areas, the Adirondack Mountains and Hudson Highlands. Precambrian rocks are also exposed directly to the north of Lake Ontario in Ontario.

The Precambrian rocks from Ontario and the Adirondacks were deformed and metamorphosed about 1,000 Ma B.P. during the Grenville Orogeny. The rock types that are found in the Grenville Group include metagabbros, meta-anorthosites, granitic, charnockitic, and syenitic gneisses; amphibolite; and meta-sedimentary marbles, calc-silicates, quartzite, and para-gneisses.

In central and western New York, Lower and Middle Cambrian rocks are absent with Late Cambrian or Early Ordovician rocks lying uncomformably on the Precambrian basement. Lower Ordovician rocks are absent. The Late Ordovician Taconic Orogeny resulted in uplift of eastern New York while the rest of the State was inundated by the sea which deposited limestones that interfinger with shales and impure sandstones. The Taconic Orogeny reached its climax in the late Ordovician with the Taconic Mountains in the east eroding and producing the sediment that forms the Queenston Delta, a delta that prograded west to midcontinent and south into Virginia. The Queenston Formation in central and western New York consists of sandstone and siltstones.

Deposition of the wind-blown Whirlpool Sandstone marks the beginning of the Silurian Period. During this time, the Taconic Mountains were uplifted and the Queenston Formation was eroded and redeposited to the west as the Medina Group. A white quartz sand (Thorold-Kodak) which terminates in the Oneida Conglomerate near Oneida, New York was deposited above the Medina Group. Subsequently, shales and anydydrites were deposited in the Clinton, Salina, and the Cayugan Series.

Early Devonian rocks (Helderberg Limestone) crop out between the Hudson River and Cayuga Lake; the Helderberg west of Cayuga Lake has been locally removed or is variable in thickness due to erosion. Subsequently deposited were the Oriskany and Glenerie Formations which are overlain by the shales and siltstones of the Esopus and Carlisle Center Formations.

During Middle Devonian the Onondaga Limestone formed with deposition being terminated by the Acadian Orogeny which was centered in New England and the Canadian Maritime Provinces. Large amounts of material were eroded from the New Acadian Mountains and deposited as shale in a westward prograding delta called the Catskill Delta. A brief interruption in delta deposition occurred and the Tully Limestone was formed during this time. The delta resumed deposition and has been divided into eight groups which are principally composed of shale and are listed here from oldest to youngest: Hamilton, Genesee, Sonyea, West Falls, Java, Canadaway, Conneaut, and Conewango. The four oldest groups extend across the entire state, whereas the younger groups either were never deposited in the Catskills or were subsequently eroded after deposition.

Early Mississippian rocks resemble the underlying Devonian sequences with deposition ending in the Middle Mississippian. Only one brief period of deposition occurred in Early Pennsylvanian, the Olean Conglomerate.

In the Mesozoic, vertical uplift and erosion predominated and during the Cenozoic Era the physiographic provinces of New York were shaped. The landscape has been modified by at least four major advances of ice during the Pleistocene.

BOUGUER GRAVITY ANOMALY MAP

The Bouguer gravity anomaly map of western New York (Fig. 1) indicates a series of gravity highs and lows extending north-northeast across New York State and Lake Ontario into southern Ontario, Canada. The magnitudes, gradients and wavelengths of these anomalies indicate that they are caused by density contrasts within the Precambrian basement. The overlying Paleozoic rocks appear to have little effect on the regional gravity contours. The gravity contours trend north-south while the contacts of the Paleozoic formations trend east-west.

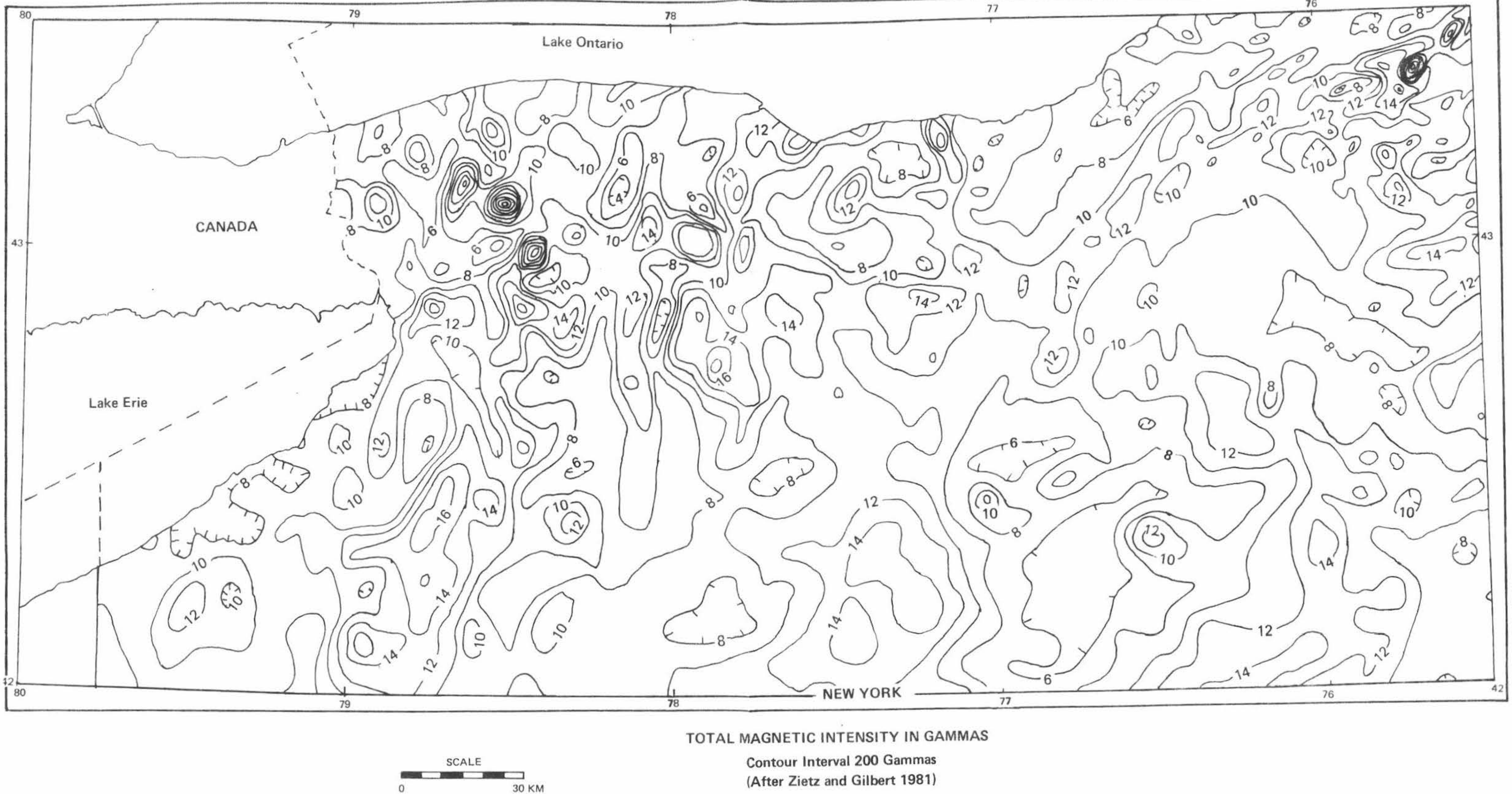

Figure 5. Aeromagnetic map of central and western New York (after Zietz and Gilbert, 1981).

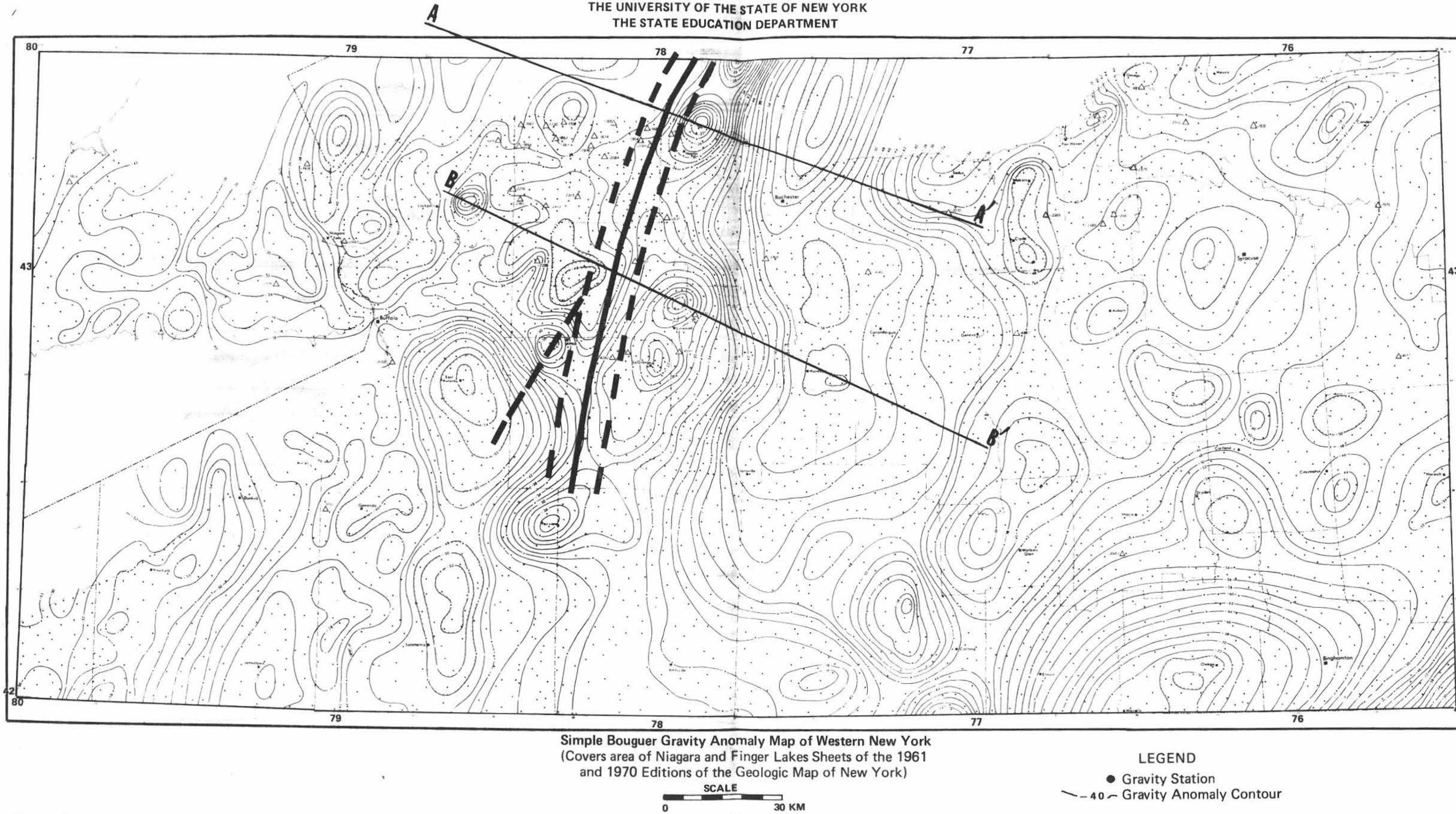

Figure 1. Bouguer Gravity map of central and western New York (after Diment and Revetta, 1971).

The gravity anomalies in western New York and Lake Ontario are due to an extension of the Precambrian rocks of the Bancroft-Madoc area of southern Ontario, Canada. The rocks in the area are part of the Ottawa River remanent, the most extensive and thickest metasedimentary-metavolcanic belts known in the Grenville province. The rocks exposed in the Bancroft-Madoc area include mafic plutons intruded into the Tudor mafic volcanic sequence. These in turn are surrounded by a marble rich metasedimentary group and granitic batholiths. The mafic plutons intruded into the mafic-metavolcanic sequence produce the gravity highs in western New York and Lake Ontario, while the granite batholiths cause the gravity lows. The intermediate gravity values are thought to be due to the marble rich metasedimentary group. This interpretation of the gravity anomalies is supported by several wells that have penetrated the Precambrian basement. The lithologies from the wells confirms that granite occurs in areas of low gravity.

The gravity anomalies of central New York between 75°30' and 77°45'W (Fig. 1) cover larger areas and have lower magnitudes and gradients than those in western New York. They indicate the Precambrian basement complex of central New York (Finger Lakes Sheet) is more homogeneous and consists of rock types that vary little in density. The area contains no major structural features such as the Clarendon-Linden fault and there is a relative absence of seismic activity. These characteristics indicate the basement complex of western New York is distinctly different from that in central New York.

Two gravity profiles were drawn across the northeast trending gravity anomalies; the Hamlin profile (AA') and the Linwood profile (BB'). The Hamlin gravity profile is shown in Figure 2. It lies along the south shore of Lake Ontario near the village of Hamlin, New York. This profile shows two prominent anomalies, a gravity high and a gravity low. The gravity high has an elliptical shape with gravity values reaching a high of -18 mgals. East of the Hamlin gravity high is a gravity low with Bouguer values as low as -62 mgals. A gabbro stock emplaced in mafic metavolcanics and extending to a depth of at least 5 kms below the basement would satisfy the observed anomaly. A magnetic high coincides with the Hamlin gravity high so the anomalous body has both high density and magnetization. The Victor gravity low is probably due to a granitic batholith. Steep gravity gradients occur along the western perimeter of the batholith at its contact with dense mafic metavolcanics. Figure 3 shows profile BB' across the Linwood gravity high. The anomaly is elliptical in shape with gravity values ranging from a high of -16 milligals to a low of -56 milligals. The gradient of the anomaly on its east and west flanks is 1.5 milligals per kilometer. The Linwood gravity high is attributed to a mafic pluton emplaced in mafic metavolcanics and extends to a depth of 5.0 kms. Based on these profiles and principally the gravity map a lithology map of the Precambrian basement is shown in Figure 4.

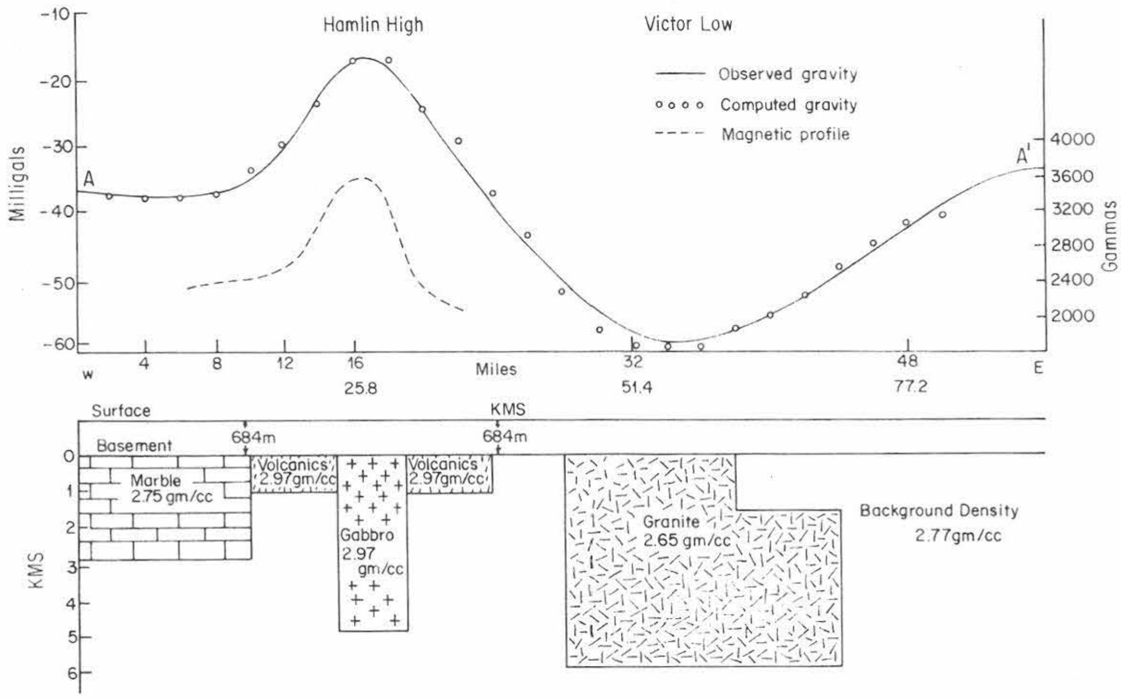

Figure 2. The Hamlin two-dimensional gravity profile AA'.

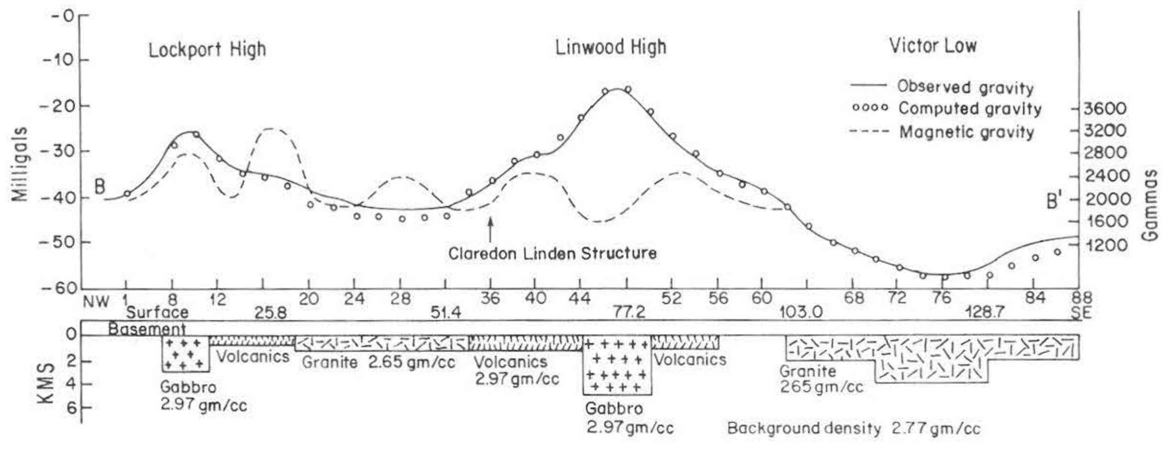

Figure 3. The Linwood two-dimensional gravity profile BB'.

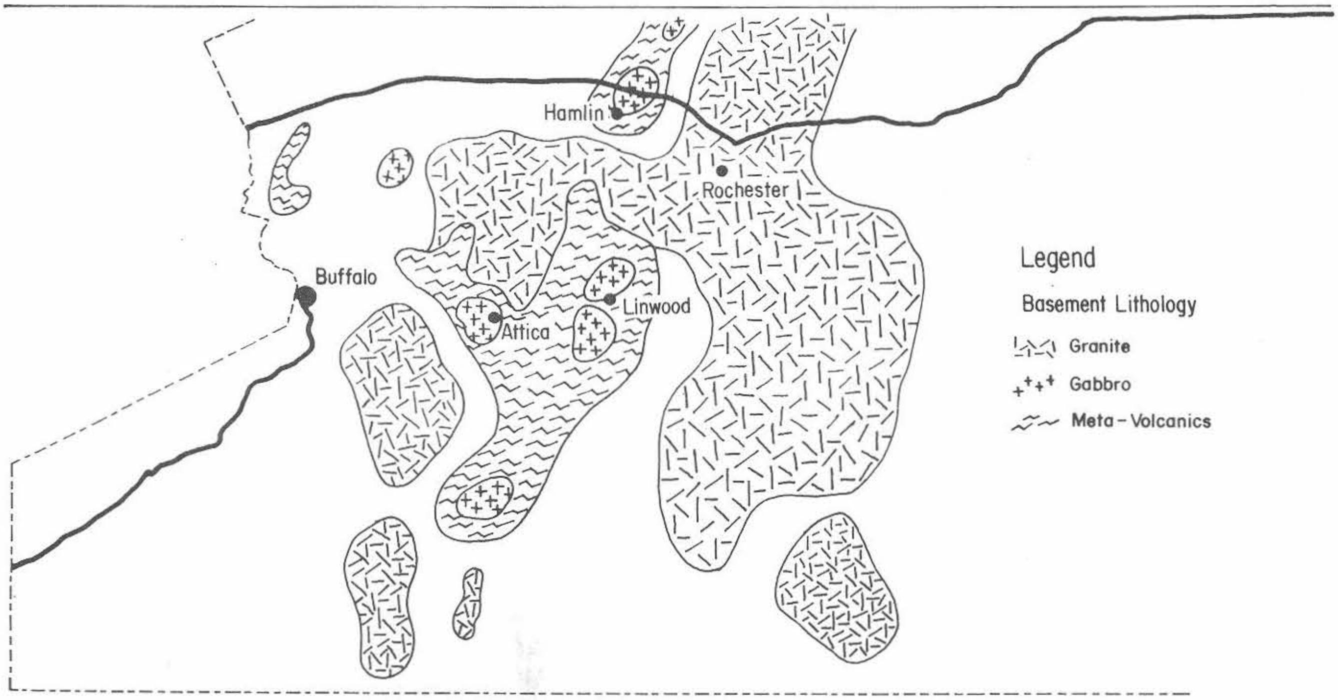

Figure 4. Lithologic map of Precambrian basement base on gravity and magnetic anomalies.

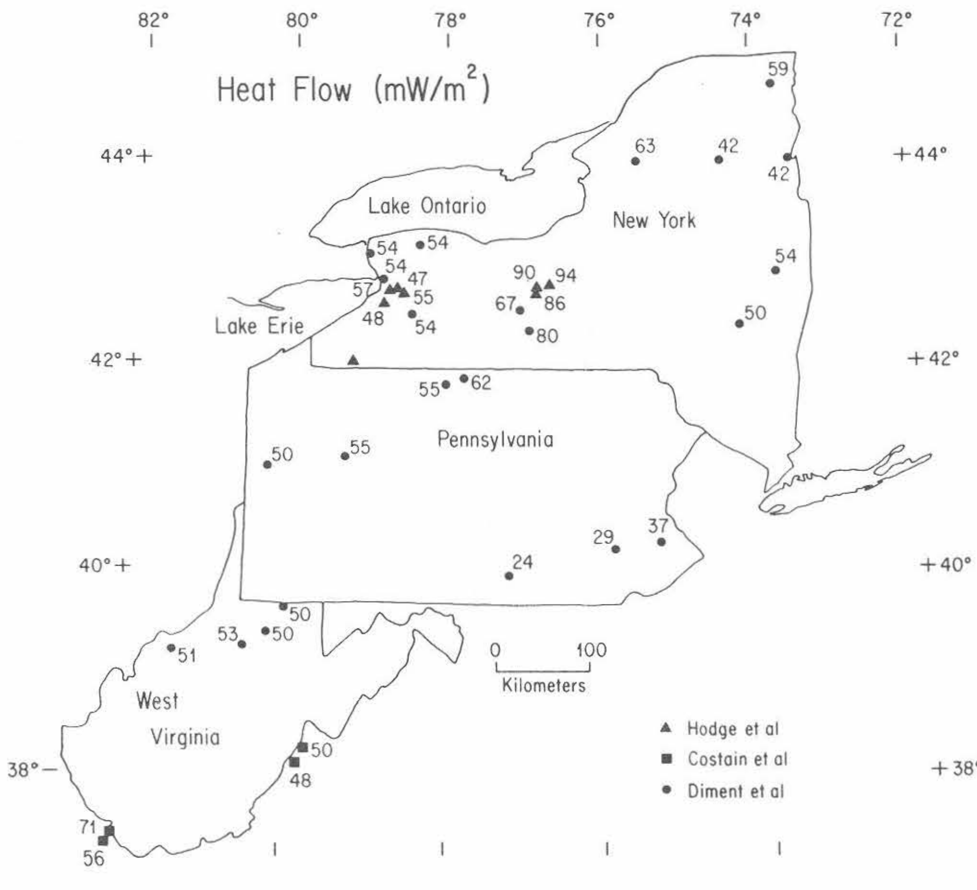

Figure 6. Heat flow determinations in New York, West Virginia, and Pennsylvania (after Hodge, et al., 1981).

AEROMAGNETIC MAP

The aeromagnetic map (Fig. 5) is modified from Zietz and Gilbert (1981). The magnetic highs generally correspond to the positive gravity anomalies that are associated with gabbro intrusions. The negative gravity anomalies that are underlain by granite in the basement, show broad low aeromagnetic anomalies. The pattern of magnetic anomalies west of the Clarendon-Linden fault zone shows numerous short wavelength anomalies whereas east of this zone the anomalies are broader. This contrast in the magnetic field suggests that the Clarendon-Linden fault zone separates distinctive basement terrains.

CLARENDON-LINDEN FAULT ZONE

The Clarendon-Linden fault is the most prominent structural feature in the Paleozoic rocks of western New York (Fig. 1). Chadwick (1920) suggested the existence of a large fault between the towns of Clarendon and Linden and found a displacement of the Onondaga and Niagara escarpments with the western side farther north. In the vicinity of Linden, the formations of the west were at a much lower elevation than those to the east. The southern portion is a monocline, the Linden monocline, and the northern part a fault called the Clarendon fault.

The Clarendon-Linden fault lies along the western flank of a series of gravity highs and the eastern flank of a magnetic low. These gravity and magnetic anomalies are due to Precambrian basement rocks so it is believed that the Clarendon-Linden fault is controlled a Precambrian basement fault at depth.

Gravity anomalies associated with the structure extend into Lake Ontario and this fact together with similar faults occurring on the north shore of Lake Ontario led Revetta (1970) to suggest a possible northeastward extension of the structure across Lake Ontario. Recently, more than 400 km of high resolution seismic and magnetic data collected by Hutchison and others (1979) indicate a possible lakeward continuation of the Clarendon-Linden fault of western New York. Their geophysical evidence suggest a west facing bedrock ridge known as the Scotch Bonnet Rise is a continuation of the Clarendon-Linden structure beneath the Lake.

HEAT FLOW AND TEMPERATURE GRADIENTS

Summary of heat flow of the U.S. indicates that the eastern U.S. has a lower average heat flow compared to the more tectonically active western U.S. In New York, Urban (1970) reported heat flow values of approximately 50 mW/m^2 (1.2 HFU) near Buffalo and two values of 54 mW/m^2 (1.29 HFU) and 67 mW/m^2 (1.6 HFU) in the Finger Lakes region. In the crystalline rocks of the Adirondacks, heat flow values range from 38 mW/m^2

(0.9 HFU) to 54 mW/m^2 (1.29 HFU) with the lowest value found over anorthosite rocks. In an area of sedimentary rocks in northern Pennsylvania, Joyner (1960) determined near normal heat flow values of 55 mW/m^2 (1.3 HFU) and 62 mW/m^2 (1.5 HFU), (Figure 6). Lateral variation in heat flow in New York and Pennsylvania ranges from 33 mW/m^2 (0.8 HFU) to 88 mW/m^2 (2.1 HFU).

In regions of gas and oil production temperature estimates of gradients are obtained from bottom-hole and surface temperatures. The American Association of Petroleum Geologists in cooperation with the USGS published a temperature gradient map of the U.S. using 25,000 temperature gradients; 125 sites were used to establish the gradient map in New York. Hodge et al. (1980) evaluated the temperature gradient using over 1490 sites in western and central New York. Temperature gradients, corrected for a drilling temperature disturbance, range from 24°C/km to over 38°C/km in local areas (Fig. 7). Temperature gradients may be strongly affected by thermal conductivity changes in the outer layer of the crust and there seems to be a distinct correlation between thermal conductivity and stratigraphic lithology in central and western New York. A detailed temperature log is shown in Figure 8 that illustrates this relationship; the heat flow at this site is 57 mW/m^2 (Urban, 1970). Surface temperatures for the gradient calculation in this study were estimated from mean annual temperature compiled at 73 NOAA recording stations located throughout the state. The temperature gradients were calculated by taking the BHT's (Bottom hole temperatures) temperatures measured during routine logging runs, minus the estimated surface temperature, divided by the well depth. Because of transient temperature disturbances during drilling, the bottom-hole temperatures recorded by routine electric logging are likely to be slightly lower than the undisturbed equilibrium temperatures of the host rock due to recent circulation of fluids and air associated with drilling processes. A correction procedure was adopted (Hodge et al., 1981) and figure 7 is the temperature gradient contour map. Contoured gradients show values for the East Aurora, Cayuga, and Elmira regions to be 36, 38, and 36°C/km respectively. These regions show higher than normal temperature gradients.

SEISMICITY OF WESTERN NEW YORK

Western New York, isolated from the stress that accompanies being located near a plate boundary, has nevertheless experienced numerous seismic events. These events have ranged from microearthquakes to strong earthquakes as high as VIII on the Modified Mercalli Intensity Scale (MMS). An earthquake of intensity VIII occurred near Attica, N.Y. in 1929. This earthquake and the frequency of seismic events indicate in western New York is an area of moderate seismicity. Frequency of events and location near a fault system, the Clarendon-Linden Fault system.

The map of epicenter locations in the Northeast U.S. and adjacent parts of Canada, Fig. 9, indicates several zones of seismic activity. These include the Hudson River-Lake Champlain zone, St. Lawrence River Valley

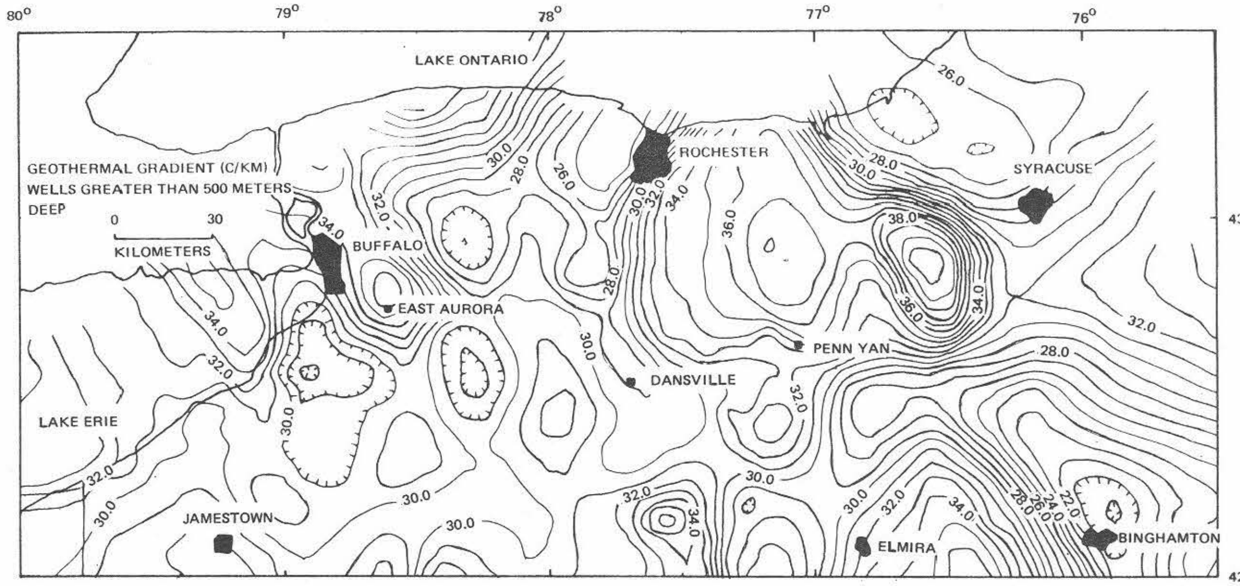

Figure 7. Contoured temperature gradients (°C/km) for wells with depth greater than 500 meters assuming a drilling disturbance correction (after Hodge, et al., 1981).

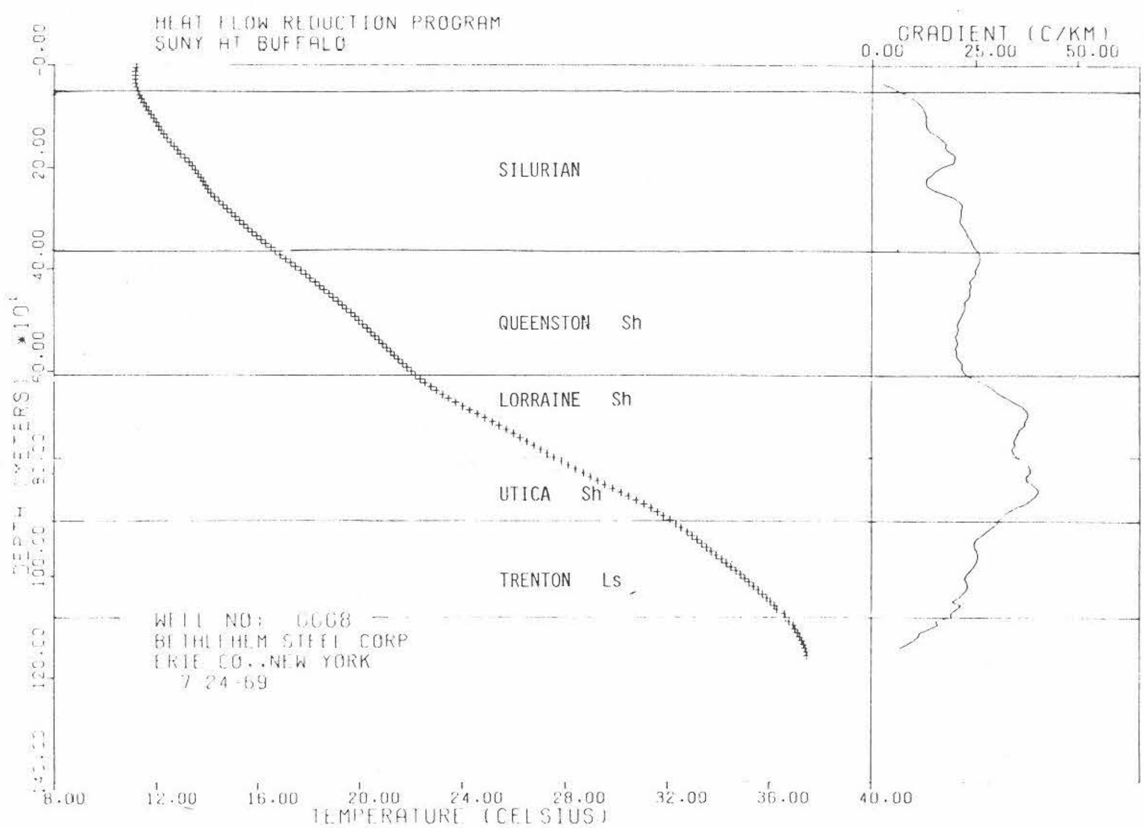

Figure 8. Temperature-depth plot of well #6668, Bethlehem Steel Corp., Buffalo, New York.

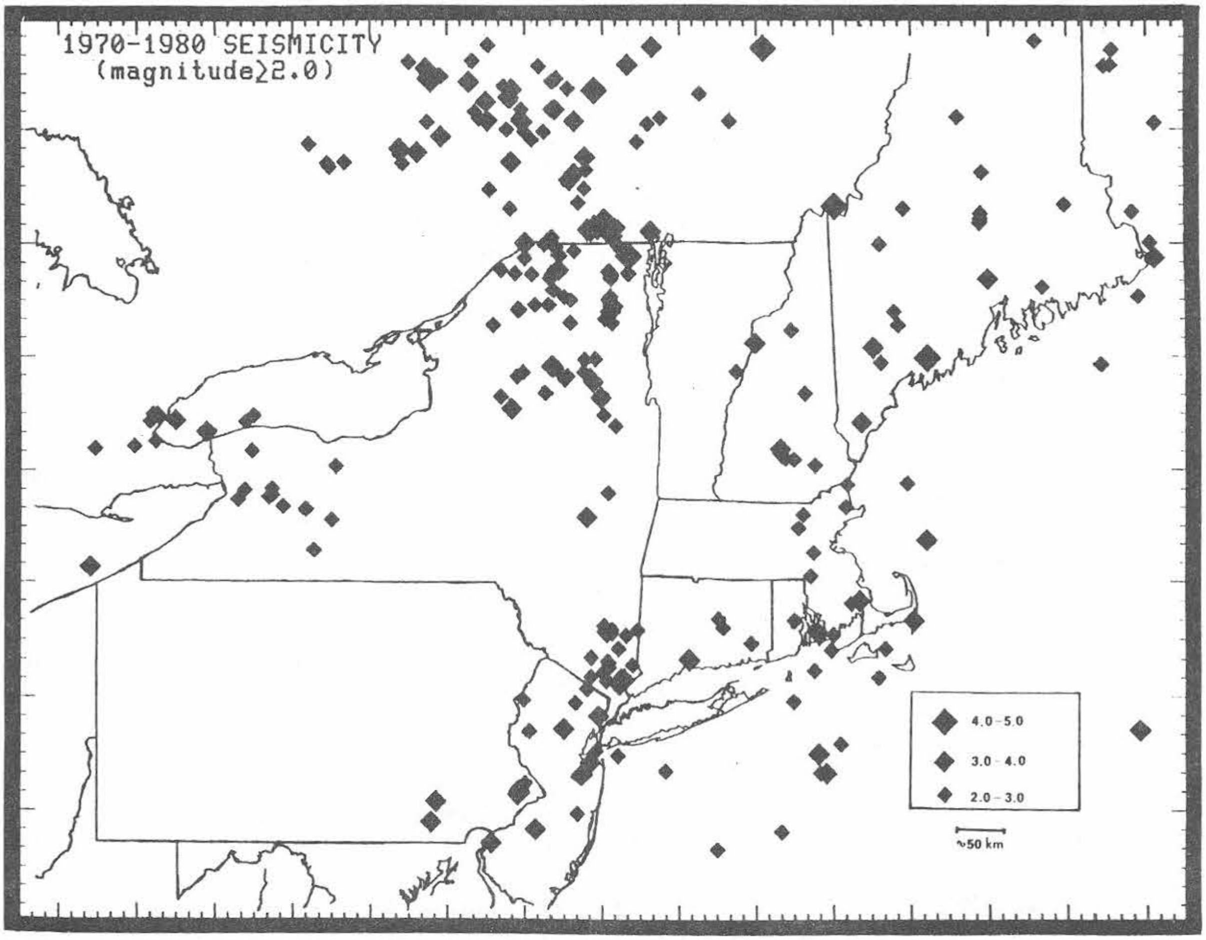

Figure 9. Location of earthquake epicenters for eastern U.S. from 1970-1980 after Schlesinger-Miller and Burstow, 1980).

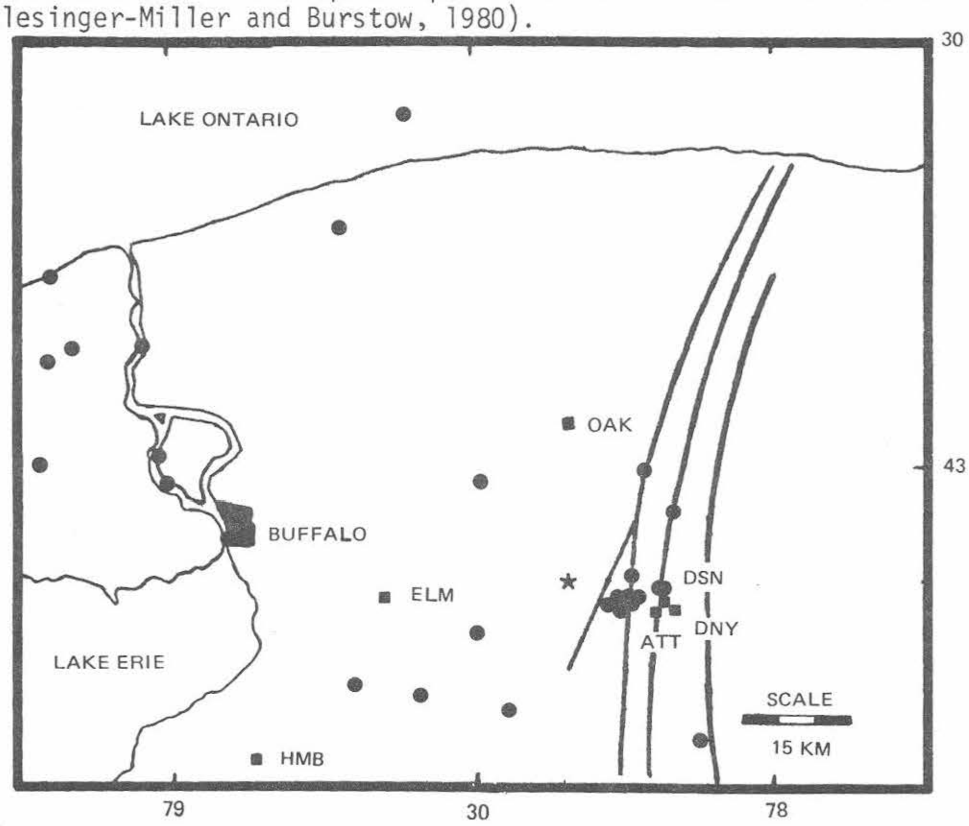

Figure 10. Location of earthquake epicenters for western New York. The 1929 earthquake is noted by a star (after Fletcher and Sykes, 1977).

region and the Buffalo-Attica region. The St. Lawrence River region has had the strongest shocks as well as the most far reaching in terms of affected areas. Earthquakes near Quebec in 1883 and 1925 had intensities of IX or X (MMS) or approximately 7.0 on the Richter Scale. The Buffalo and western New York area experienced intensities of IV (MMS) from these shocks. The St. Lawrence region has experienced at least nine other earthquakes of intensities greater than VIII all of which affected the western New York area as well.

The western New York region has not experienced earthquakes of intensity VIII as frequently as the St. Lawrence region, however, several quakes with intensities greater than VI have been centered in Attica and Niagara Falls. Figure 10, shows the distribution of epicenters in western New York and as might be expected, there is a concentration of events along the Clarendon-Linden Structure near Attica. The strongest shock in western New York (star on Fig. 10) occurred near Attica on August 12, 1929. An intensity of about VIII (MMS) was determined and the quake affected an area of about 50,000 square miles. Further shocks of less intensity occurred in 1929, 1935, 1955, 1966, and in 1967 indicating continued activity in this region (see table 1).

Smaller earthquakes have occurred between Attica and Buffalo and into the Niagara Peninsula of Ontario, Canada. The Lockport earthquake of 1857 is included in this group of shocks and it attained an intensity of VI (MMS). As of yet, no major earthquake has occurred along the Buffalo-Attica zone west of the Clarendon-Linden structure, however, since the area is far more populated than other parts of western New York, the potential risk of extensive damage must be considered.

TABLE 1. EARTHQUAKES AFFECTING THE WESTERN NEW YORK REGION (1928-1959)*

340. 1928 March 18. 15:20. V-VI. M_L 4.1 (determined from a Shawinigan Falls seismogram.) 44°.5N, 74°.3W. Saranac Lake, NY. Perceptible over 12,000 square miles.

411. 19[illegible] November 2. 14:31:58. M_L 5.4, 47°14'N ± 18', 78°10'W ± 15. About 56 miles northeast of Timiskaming, Que. This tremor was felt in Ontario (at Kitchener, North Bay, Owen Sound and Toronto), in New York (at Buffalo and Rochester) and in Pennsylvania (at State College).

448. 1929 August 12. 6:00 to 7:00. III. 42°.2N, 77°.3W. Between Little Valley and Binghamton, NY. Probably a foreshock of No. 449.

449. 1929 August 12. 11:24:8. VIII. M_L 5.3. 42°52'N, 78°21'W. (a) Near Attica, New York. The shock was felt in the northeast corner of Ohio, northern Pennsylvania, most of New York and parts of New Hampshire, Vermont, Massachusetts and Connecticut.

453. 1929 December 2. 22:14. IV. 42°.8N, 78°.3W. Attica, NY.

454. 1929 December 3. 12:50. IV. 42°.8N, 78°.3W. Attica, NY.

457. 1939 January 17. p.m. 1. 42°.8N, 78°.3W. Attica, NY.

475. 1931 April 22. IV. 42°9N, 78°.9W. "A strong earthquake was felt in Buffalo, NY, and surrounding cities on April 22nd, causing panic."

478. 1931 June 7. 0:00. II. 43°.2N, 77°.6W. Rochester, NY.

561. 1939 February 24. 0:20. III. 42°52'N, 78°17'W. Attica, NY.

619. 1943 March 9. 3:25:34. M_L 5.5. 42°.2N, 80°.9W. Lake Erie.

628. 1944 January 16. 10:00. II 43°09'N, 77°37'W. Rochester, NY.

652. 1946 November 10. 11:41:23. M_L 3.1. 42°52'N, 77°27'W ± 18'. About 25 miles south-southwest of Rochester, NY.

715. 1955 August 16. 7:35. V. 42°52'N, 78°17'W. Attica, NY.

* From the Canada Department of Mines and Technical Surveys. Observatories branch. Publications of the Dominion Observatory, Ottawa. Volume XXXII, No. 3. Earthquakes of Eastern Canada and Adjacent Area 1928-1959. W.E.T. Smith.

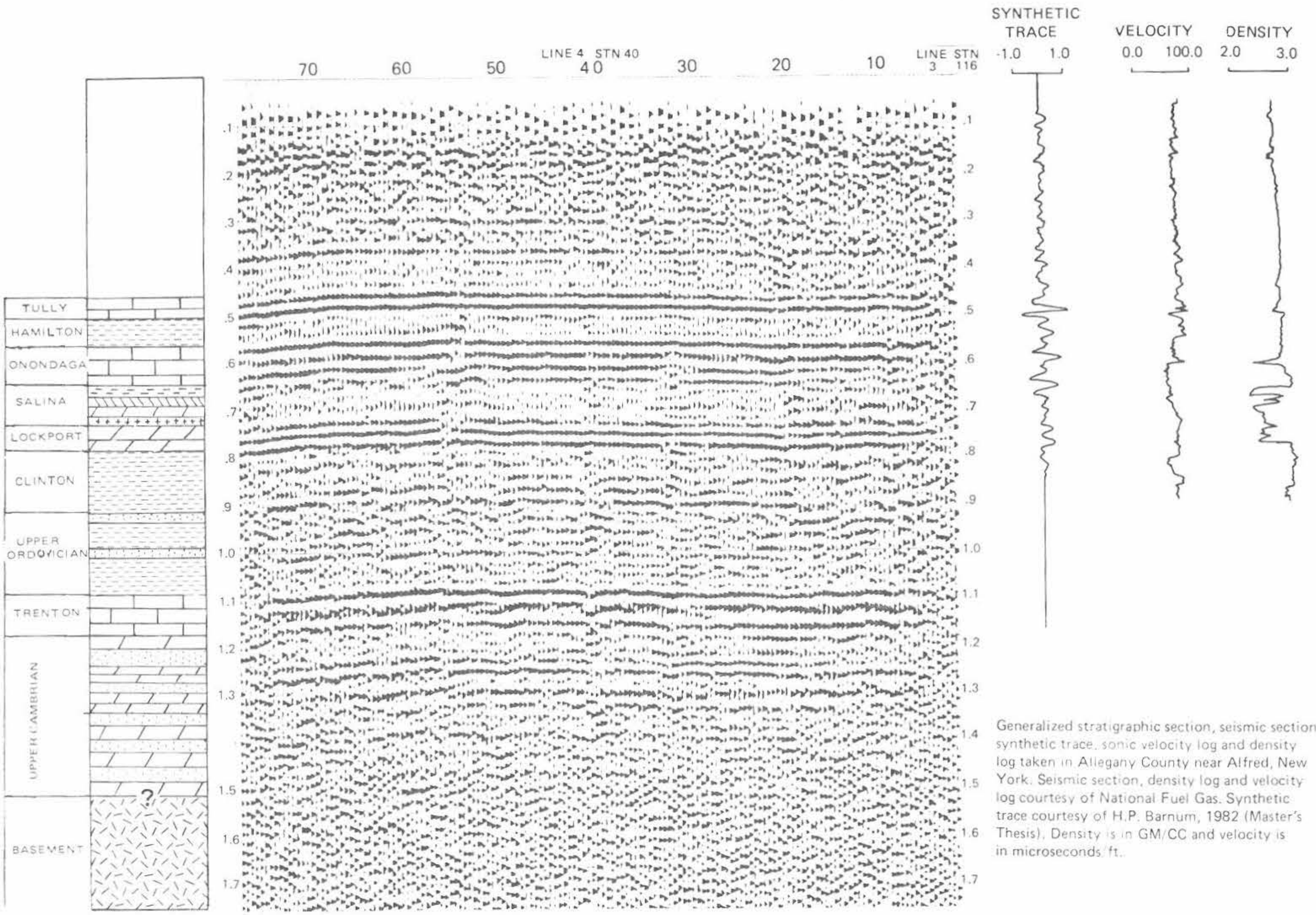

Figure 11. Seismic reflection cross section from Alleghany Co. N.Y. that shows that seismic stratigraphy in the Paleozoic rocks. Vertical density and sonic velocity distribution from well #6213, Alleghany county is shown compared to the seismic time section. A synthetic seismogram using these logs is also shown (Barnum, 1982).

SEISMIC STRATIGRAPHY

The seismic stratigraphy of western New York is relatively uniform throughout the region and there is a direct correlation to the lithologic units. Seismic exploration has become increasingly important in western New York (WNY) with conventional exploration techniques such as Vibroseis yielding good seismic sections.

In Figure 11 a seismic section, courtesy of National Fuel Gas, is from a survey done near the town of Alfred in southern WNY. This survey was done using the Vibroseis method. The principle reflections shown by the dark bands that occur at various times throughout the section are produced by the limestones and dolostones that are present in the upper Cambrian, Ordovician, Silurian, and Lower Devonian units. Lithologies and names of some units are shown to the left of the seismic section.

Also shown in Figure 11 to the right of the seismic section are a synthetic trace, velocity log, and density log. The velocity and density logs are courtesy of National Fuel Gas and were taken from a well in Almond township, Allegheny County. The synthetic waveform is produced by relating rock density and seismic velocity throughout a vertical section. The acoustic impedance of a layer can be defined as the product of density (ρ) and p-wave velocity (v) and when a seismic wave strikes an interface between two media with differing acoustic impedances, part of the wave is reflected and part is transmitted. The synthetic seismogram of vertical incidence reflections matches well with the strong reflections corresponding to the previously mentioned limestones and dolostones.

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A-1

UPPER MOSCOW-GENESEE STRATIGRAPHIC RELATIONSHIPS IN WESTERN NEW YORK: EVIDENCE FOR REGIONAL EROSIVE BEVELING IN THE LATE MIDDLE DEVONIAN

CARLTON E. BRETT
University of Rochester
Rochester, New York 14627

GORDON C. BAIRD
SUNY College at Fredonia
Fredonia, New York 14603

INTRODUCTION

A widespread regressive episode, immediately preceding the late Middle Devonian Taghanic onlap, has long been recognized in eastern North America (Grabau, 1917; Cooper, 1930; Cooper and Williams, 1935; Heckel, 1973; Dennison and Head, 1975; Rickard, 1975). Regional erosive overstep of Hamilton beds by Tully and Genesee strata was documented in the Finger Lakes area (Cooper and Williams, 1935) and was recognized west of Canandaigua Lake (Cooper, 1930), but the nature of pre-Taghanic unconformity in western New York remained obscure due to supposed lack of stratigraphic controls in the upper Moscow (Windom) shales. Indeed certain workers (Stover, 1956; Fulreader, 1957) believed that upper Windom beds were relatively little affected by erosional truncation.

Contacts of the Middle Devonian Windom Shale Member, previously examined and discussed (Brett, 1974a; Brett and Baird, 1975), have received continued study by the present authors; results of new work on the Windom-Genesee paraconformity and adjacent strata in the Erie County region are discussed herein.

Detailed correlation and study of Windom Shale component zones has been undertaken. At least fourteen mappable Windom marker beds and intervals have been correlated across all, or part, of Erie County; these are summarized in the present text. Mapping of Windom units immediately underlying the Genesee Formation permits detailed inference as to the nature of this unconformable contact. These beds also provide insight into the processes of marine mud sedimentation during deposition of the Windom and its influence on fossil community distribution. Particularly notable is the discovery that the thickest Windom sections in the county are the least stratigraphically complete and the thinnest generally display the full complement of marker beds and zones.

Clearly, the regional configuration of Windom Shale units results from the interplay of two processes: westward stratigraphic condensation of the section, and progressive erosive overstep of upper Windom beds. As will be shown here, the magnitude of erosional beveling of Moscow strata increases to the north as the depositional hinge of the Appalachian Basin is approached.

In addition, the Moscow-Genesee discontinuity and associated erosion lag deposits have been reexamined. A newly discovered discontinuity at the base of the Genundewa Limestone (basal Upper Devonian?; see Oliver, et al., 1981) is found to have some similarity to the North Evans Member ("conodont bed") and is probably coeval with it. The sedimentology and mode of origin of the North Evans and Leicester pyrite members is further discussed. Finally, new data are synthesized in a chronological outline of events affecting western New York during Late Middle Devonian time.

DEPOSITIONAL SETTING

Western New York Shelf

The Hamilton Group of New York includes Middle Devonian sediments deposited at or near the northern margin of the Appalachian Basin; they accumulated in the northern arm of an inland sea, the deepest part of which was developed southeast of the study area. The northern and western boundaries of the basin bordered low-relief, cratonic shelf regions; these areas supplied relatively little detrital sediment to the basin as compared with actively rising tectonic source terrains to the southeast of the basin. This accounts for the thin deposits in western New York (Dennison and Head, 1975). A broad, gently south-sloping, muddy shelf existed across most of central and western New York during Hamilton deposition (Cooper, 1957; McCave, 1967, 1973; Grasso, 1970, 1973; Heckel, 1973).

The eastward thickening Hamilton clastic wedge is a result of Acadian tectonic events, including uplift to the east and southeast (Cooper, 1957; Heckel, 1973; Oliver, 1977). It is the initial expression of the Catskill Deltaic Complex which expanded greatly during the Late Devonian. Upper Hamilton formations are composed largely of detrital sediment; in eastern and central New York they record basin filling and general westward migration of the eastern shoreline. In western New York, the Hamilton Group is markedly different, consisting of thin shelf sediments which do not record simple shallowing-upward sequence.

Upper Hamilton sediments are characteristically fossiliferous; the rich biotas of the Ludlowville and Moscow Formations in Erie County have been a source of study for paleontologists for more than a century (Hall, 1843; Grabau, 1989-1899; 1899; Cooper, 1929, 1930, 1957;

Buehler and Tesmer, 1963; Beerbower, et al., 1969; Oliver and Klapper, 1981). The Hamilton sea apparently was relatively shallow and had near-normal salinity, water temperatures, and circulation as evidenced by the presence of diverse stenotopic benthic organisms.

In contrast to the fossilferous Hamilton beds in Erie County, overlying Genesee shales contain relatively low diversity assemblages which are dominated by pelagic taxa. Only in the paraconformity-related Leicester and North Evans Members is the fossil diversity greater, but many of these fossils may have been reworked from the underlying Moscow Formation (see below).

Slope and Basin Environments

The western New York Shelf was bounded to the south by a more actively-subsiding central region of the Appalachian Basin during the Middle Devonian. A southward trending regional slope is recognized for Onondaga (Eifelian) carbonates based on extensive study of subsurface drill cores and well log data (Kissling and Moshier, 1981; Koch, 1981). Brachiopod/coral associations reflect southward increasing depth within the Onondaga and there is major southward thinning of the whole Onondaga carbonate package across the southern tier of western New York (Koch, 1981).

Fossiliferous, gray mudstones and thin limestones of the upper Hamilton Group are correlated southward with the Millboro Member, a thick sequence of dark gray and black shale developed in western Pennsylvania, southeastern Ohio, and West Virginia (Dennison and Hassan, 1976). Similarly, the Tully Formation, which is a compact, laterally extensive carbonate unit in central New York grades southward and westward across Pennsylvania into a thick sequence of calcareous shale and finally into black shale (Burket Member) near the Pennsylvania-Maryland border (Heckel, 1973). The Moscow-Genesee paraconformity of western New York is coextensive with disconformities of decreasing magnitude in the Tully Formation in central New York, this hiatus disappearing southward as the Tully thickens and grades into black shale. Thus, in the region of greater subsidence, the upper Middle Devonian sequence is characterized by deeper water deposits, as indicated by the dysaerobic-anoxic mudstones, and is apparently more complete.

In the present paper, it is argued that apparent eastward erosional truncation of Windom zones and marker beds is really a northward overstep effect, this also reflecting the subtle influence of depositional hinge effects or on differential subsidence to the south. This and other north-south depositional changes are discussed in the text.

During Genesee (latest Givetian to early Frasnian) time general deepening of shelf waters occurred in western New York, as a part of a widespread Taghanic Onlap (Johnson, 1970; Dennison and Head, 1975). The consequent development of dysaerobic and anaerobic basin-type conditions in western New York resulted in deposition of dark gray and black, sparsely fossiliferous muds (Sutton, et al., 1970; Thayer, 1974; Bowen, et al., 1974). Anoxic bottom conditions inhibited development of benthos. However, as will be shown, significant bottom currents were present to the degree that reworked fossils and diagenetic debris were sorted into lenses within the basal Genesee sequence.

General Stratigraphy

The Middle Devonian (Givetian) Hamilton Group and the Middle to Upper Devonian Genesee Formation (U. Givetian-Frasnian) are eastward thickening wedges of terrigenous sediment which are predominantly marine except in east-central and eastern New York. This aggregate sequence, ranging from 88 m (290 ft) at Lake Erie to more than 1220 m (4000 ft) at the Catskill Front, is composed of detrital sediments which coarsen to the east and southeast. In western New York , the Genesee beds are highly condensed stratigraphically, thinning westward from over 305 m (1000 ft) in the Seneca Lake region to as little as 3 m (10 ft) at the Lake Erie Shore, the Erie County region being characterized by particularly slow sedimentation. Hamilton and Genesee marine shales in western New York are displaced eastward by siltstone and sandstone facies in central New York, these to be succeeded, in turn by redbed floodplain deposits and fluvial sandstones in east-central and eastern New York State (Rickard, 1975).

Westward thinning of marine Hamilton and Genesee strata is associated with appearance of widespread discontinuities and condensed sedimentary units. Most significant of these is the regional paraconformity separation Hamilton-Tully beds and the overlying Genesee Formation. This discontinuity, originating in the uppermost Tully Formation in Seneca County (see Huddle, 1981) becomes progressively more significant westward as a stratigraphic gap; several upper Hamilton faunal zones are overstepped progressively to the west beneath the break (Baird and Brett, in press). However, there is corresponding westward depositional onlap of overlying Genesee beds; the basal Geneseo Shale rests on Windom in Genesee County whereas only the upper Penn Yan and finally Genundewa Members rest directly on the Hamilton in western Erie County (deWitt and Colton, 1978). This interpretation is corroborated by conodont (Huddle, 1981; Klapper, 1981) and goniatite biostratigraphy (Kirchgasser, 1973).

STRATIGRAPHY, THICKNESS VARIATION AND EROSIONAL BEVELING OF THE WINDOM SHALE MEMBER (MOSCOW FORMATION)

Detailed Stratigraphy

The Windom Shale Member (Grabau, 1917) comprises the uppermost unit of the Moscow Formation and of the Hamilton Group as a whole. In Erie County localities (Fig. 1), the Windom Member is composed primarily of medium to dark gray, variably calcareous mudstone with several thin persistent argillaceous limestones, concretionary beds, and pyritic horizons (Buehler and Tesmer, 1963; Fig. 2). It is bounded by discontinuities at its base and top (Brett, 1974a; Baird, 1978, 1979) and ranges in thickness from 2.2 m (7 ft) at the southwestern section near Pike Creek on the Lake Erie Shore to perhaps as much as 16 m (53 ft) near the Genesee-Erie County boundary. Previous workers (Cooper, 1930) have noted an anomolous area of eastward-thinning of the Windom Shale in Genesee County, to about 10.7 m (35 ft) near Darien. As will be demonstrated, the thickness pattern of the Windom reflects a complex interplay of westward sedimentary thinning and northward regional, erosional truncation.

The most complete and best exposed Windom section is in the abandoned Penn-Dixie cement quarry near Bay View, New York (loc. 7), which provides an excellent reference section (Fig. 2). Component fossil assemblage zones from this locality are described in detail elsewhere (Brett, 1974b; Baird and Brett, in press). The following section summarizes the characteristics of the various Windom fossil zones and marker beds in ascending order.

Unit 1: *Ambocoelia umbonata* Beds. Soft, friable, medium to dark-gray shales at the base of the Windom contain very abundant specimens of the brachiopod *Ambocoelia umbonata*, as well as chonetids (Stover, 1956; Brett, 1974b). This zone is only 16 cm (6 in) thick at the Lake Erie Shore and 20 cm (8 in) at Penn-Dixie Quarry but it thickens abruptly eastward from about 0.5 m (2 ft) at Smoke Creek to 2.6 m (18.5 ft) at Buffalo Creek and over 7.5 m (24 ft) at West Alden near the Erie-Genesee County border.

Unit 2: Bay View Coral Bed (Baird and Brett, in press). This unit, equivalent to the Moscow Coral Bed of Grabau (1899) and *Spinatrypa spinosa* bed of Brett (1974b), consists of soft to somewhat indurated calcareous shale containing a diverse fauna of at least 50 species of fossils including small rugose corals, diverse brachiopods, pelmatozoan debris and trilobites. The Bay View bed thickens eastward from 5 cm (2 in) at Lake Erie to 2 m (7 ft) near Buffalo Creek. Correspondingly, the density of fossils declines strikingly and the single band breaks into an interval of shell-coral rich layers separated by barren, calcareous mudstones. Coincident with this lithologic transition is a slight faunal change. Western sections of the Bay View bed

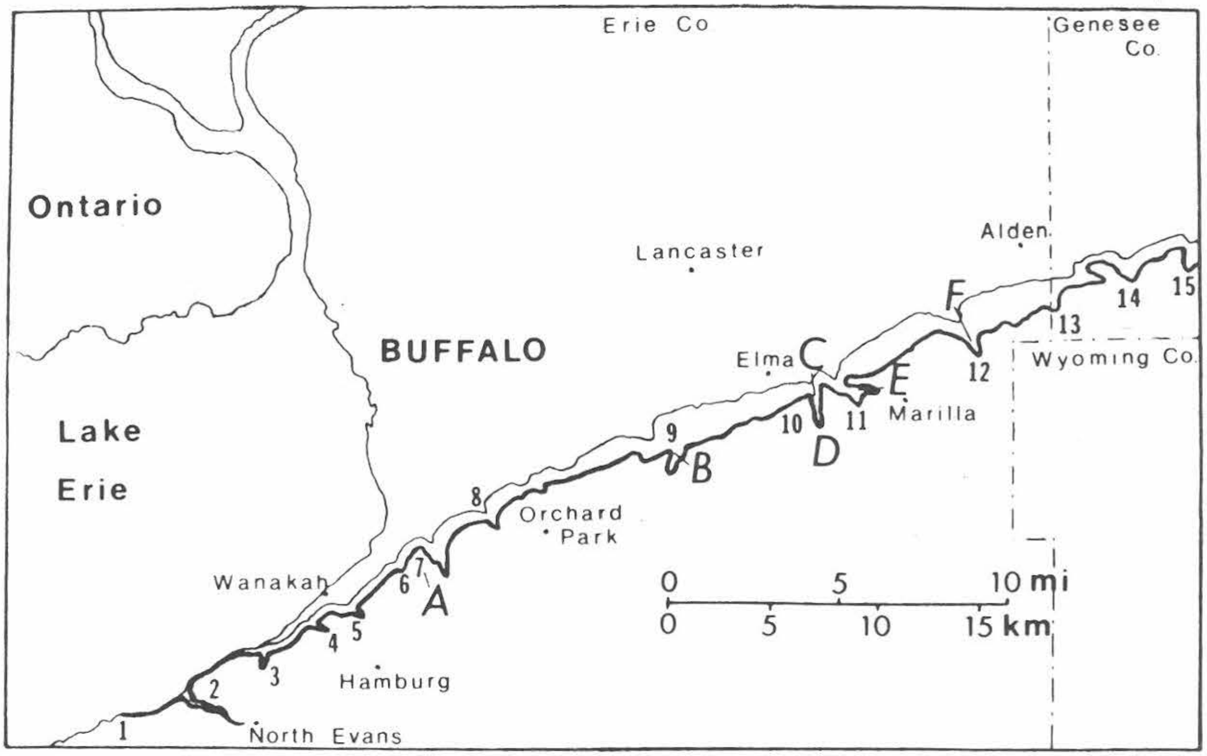

Figure 1. Study area. Outcrop belt of Middle Devonian Windom Shale Member and localities examined; light line denotes base of Windom Member; heavy line denotes base of Genesee Formation. Numbered sections include: 1) Pike Creek; 2) Eighteen Mile Creek; 3) unnamed creek at Weyer; 4) unnamed creek at Amsdell; 5) Cloverbank Shale Pit (Bethlehem Steel Co.); 6) unnamed creek south of Big Tree Road; 7) Bay View Shale Pit (formerly Penn Dixie Co.); 8) South Branch, Smoke Creek; 9) Cazenovia Creek; 10) Buffalo Creek; 11) Little Buffalo Creek; 12) Cayuga Creek; 13) Durkee Creek; 14) Eleven Mile Creek; 15) Murder Creek. Field trip stops, in capital letters, include: A) Bay View Shale Pit; B) Cazenovia Creek; C) Buffalo Creek (lower Windom); D) Buffalo Creek (upper Windom-Genesee Fm.); E) Little Buffalo Creek; F) Cayuga Creek.

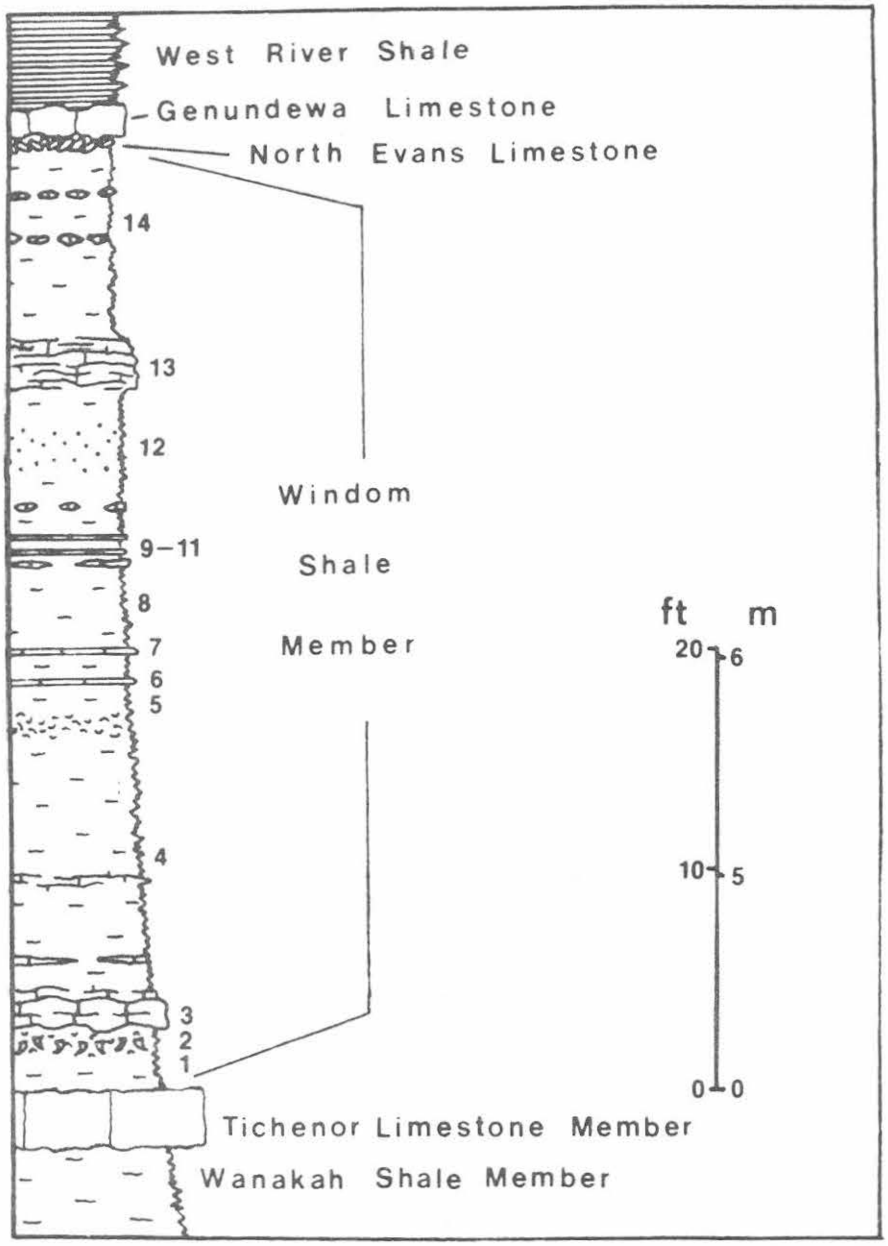

Figure 2. Stratigraphic subdivisions of the Windom Shale Member; standard section Bay View Quarry and unnamed creek near Big Tree; Units include: 1) Ambocoelia umbonata beds; 2) Bay View coral bed; 3) Smoke Creek bed; 4) barren shale interval; 5) Big Tree bed; 6,7) A-B limestones; 8) Buffalo Creek pyritic beds; 9-11) C,D, and E limestones; 12) Penn Dixie pyritic beds; 13) Amsdell bed; 14) upper Ambocoelia? praeumbona-bearing shales.

contain local biostromes of the large rugose corals Cystiphylloides, Heliophyllum and Heterophrentis, east of Smoke Creek. These corals are replaced by a suite of smaller stereolasmatid corals (Amplexiphyllum, Stereolasma), auloporids and Pleurodictyum. Similarly, certain brachiopods such as Spinatrypa spinosa are restricted to the western coral-rich facies whereas bivalves become abundant farther east.

Unit 3: Smoke Creek Bed. Formerly termed the "coral-trilobite" bed (Brett, 1974b), the Smoke Creek bed (Baird and Brett, in press) constitutes a very persistent, ledge-forming calcareous interval. It maintains a relatively uniform thickness of 2-75 cm (8-30 in) in Erie and Genesee Counties, and consists of irregular to blocky fracturing, light-gray calcareous shale and argillaceous limestone. The Smoke Creek bed is one of the most widespread and distinctive markers in the Windom; it is traceable from Lake Erie Shore eastward to Canandaigua Lake (Baird and Brett, in press). This unit typically contains an abundance of small rugose corals, brachiopods including Pseudoatrypa, Mucrospirifer consobrinus and Ambocoelia umbonata. The trilobites Phacops and Greenops may occur as clusters of complete specimens on certain bedding planes.

Unit 4: Shales overlying the Smoke Creek bed are medium to olive gray and are sparsely fossiliferous or completely barren, yielding, at most, scattered trilobites and chonetid brachiopods. From Buffalo Creek (loc. 10) eastward these shales contain two or more calcareous-concretionary horizons which are particularly well displayed at Cayuga Creek (loc 12). This interval ranges from 0.8 m (2.6 ft) at Pike Creek (loc. 1) to over 5.0 m (16 ft) in central Erie County. It appears to correlate with a more fossiliferous zone, containing abundant trilobites (Greenops), pyritized nuculid bivalves and nautiloids, which occurs near the upper contact of the Windom in eastern Erie County (see Stop 5 description).

Unit 5: Big Tree Bed. A thin (5-10 cm) pyrite-rich, fossil horizon occurs at the top of Unit 4. This horizon is characterized by abundance of the brachiopods Pseudoatrypa and Mediospirifer, typically as highly compressed specimens, small rugose corals and crinoid columnals. Weathered exposures of the bed in Penn-Dixie Quarry (loc. 7) have also yielded an abundance of pyritized sponges, blastoids and a variety of molluscan steinkerns. A similar fauna (including the blastoids) has been recognized in Cazenovia and Buffalo Creeks. The Big Tree bed may correlate with the Fall Brook bed, a coral-rich horizon in the Genesee and Wyoming Valleys (Baird and Brett, in press). However, this correlation is tentative due to erosional removal of this sequence in western Genesee and eastern Erie Counties and to apparent regional facies change at this level.

Units 6, 7: A and B Limestones. Immediately overlying the Big Tree bed is an interval containing two thin (3-4 cm), but persistent bands of barren, hard, argillaceous limestone, spaced about 50 cm apart.

These marker bands are simply designated the A and B limestones. Locally, the upper (B bed) may be missing altogether and these beds may break up into a zone of flattened concretions particularly at Buffalo Creek. These bands resemble an overlying set of limestone beds (C, D, and E limestones; see below).

Unit 8: Buffalo Creek Bed. Separating the A or B and C limestones is an interval approximately 1.5-2.4 m (5-8 ft) thick of soft, fissile, medium-gray, pyrite-rich shale. Large (20-30 cm) irregular masses of pyrite occur near the base of this interval at Buffalo Creek (loc. 10). This interval is nearly barren in the westernmost localities (locs. 6-9); however, at Buffalo Creek (loc. 10) it contains *Ambocoelia umbonata*, chonetids and scattered pyritized burrows, nuculid bivalves, nautiloids and other fossils. At Little Buffalo Creek (loc. 11) *Pseudoatrypa*, *Mediospirifer* and *Devonochonetes coronatus* also occur in these shales suggesting eastward increase in faunal diversity. Because this interval is absent in Genesee County due to erosive overstep, the facies change is inferrential.

Units 9-11: C, D and E Limestones. A series of three, regularly-spaced, hard, argillaceous limestone bands, resembling the A and B beds occur at the top of the Buffalo Creek interval. Each band is uniformly 3-4 cm (1-2 in) thick. Limestone beds D and E are invariably slightly closer together (30-38 cm) than C and D (39-46 cm). As with the A and B limestones one or more of these beds may locally grade laterally to a horizon of tabular concretions. Scattered concretions also occur above or below each of the three beds. These limestones are easily recognizable markers between Penn-Dixie Quarry (loc. 7) and Buffalo Creek (loc. 10). In the vicinity of Eighteen Mile Creek (loc. 2) two or three of the beds appear to merge into a single 20 cm thick calcareous band.

Unit 12: Penn-Dixie Beds (= "small *Tropidoleptus*" beds, Brett, 1974b). Overlying the E limestone bed is a 1-3 cm (3-10 ft) interval of dark-gray, friable, pyritic shale, closely resembling the Buffalo Creek bed. This shale generally lacks concretionary horizons, but a layer of small oval concretions, many containing pyritic cores is developed at Smoke, Cazenovia and Buffalo Creeks (locs. 8-10). These concretions are associated with scattered large masses of pyrites and, near Lake Erie, thin crusts of pyrite occur at the same level. Unit 12 shalesare characterized by an abundance of small specimens (juveniles?) of *Tropidoleptus carinatus* and *Ambocoelia* cf. *A. nana*; these diminutive brachiopods are not found elsewhere in the Windom of western New York. At Penn-Dixie Quarry, where Unit 8 is best exposed, these beds also yield an abundance of pyritic fossil steinkerns, including nuculoid clams, gastropods, nautiloids, ammonoids and enrolled trilobites. This fauna, reminiscent of the Alden pyrite fauna of the Ledyard Shale (Fisher, 1951), is currently being studied in detail (Dick and Brett, 1982).

Unit 13: Amsdell Beds (= "praeumbona beds", Brett, 1974b). This unit comprises 30-75 cm (12-29 in) of light-gray, slightly concretionary, argillaceous limestone and medium-gray calcareous shale, with abundant specimens of the brachiopods *Ambocoelia? (Crurithyris?) praeumbona*, *Leiorhynchus? quadricostatum*, chonetids and trilobites. In the western most exposures the Amsdell beds are expressed as a series of two or three hard calcareous bands separated by shale. However, at Cazenovia Creek (loc. 9) the Amsdell bed forms a single blocky band of ledge-forming argillaceous limestone. The Amsdell beds are not present in the outcrop belt east of Cazenovia Creek.

Unit 14: From Eighteen Mile Creek northwestward to Smoke Creek the uppermost Windom consists of 1-2.4 m (3-8 ft) of soft, gray shale with three or four horizons of flattened ellipsoidal to bedded concretionary limestone bands. The shale and associated concretions are well exposed in the abandoned Bethlehem Steel quarry near Cloverbank. This interval is generally sparsely fossiliferous, but some concretions contain *Ambocoelia? praeumbona*, *Schizobolus truncatus*, *Allanella tullius*, chonetids and trilobites.

Westward Thinning and Condensation

In western New York, the Windom Shale exhibits abrupt southwestward thinning (Figs. 3, 4A). At Cazenovia Creek (loc. 9) the Windom (Units 1-13) is about 14.4 m (47 ft) thick, whereas at Pike Creek on Lake Erie Shore, 30 km to the southwest, the corresponding stratigraphic interval is only 2.2 m (7 ft). Moreover, subsurface data indicate that the Windom pinches out altogether a few kilometers south of this region (Rickard, pers. comm. 1981). Associated with thinning is pronounced stratigraphic condensation; notably, the thin C-E limestone bands (Units 9-11) appear to merge into a single calcareous band at Eighteen-mile Creek. Similarly the lower *Ambocoelia*-rich shales (Unit 1) thin dramatically from nearly a meter to 16 cm and, locally, pinch out (Fig. 3).

Various subunits within the Windom exhibit unequal westward thinning. Notably, most of the thickness variation involves shale packages, while thin carbonate units show only minor thickness changes. For example, the Smoke Creek bed maintains a thickness of 70-75 cm from Buffalo Creek eastward through all of Genesee County and only thins to 20 cm in the most condensed section at Pike Creek. The A-E limestones are 3-4 cm thick at all outcrops examined. Similarly, the Amsdell bed only ranges from 30-70 cm. These observations suggest that the carbonate-rich mudstones were generated by processes largely independent of those producing the differentially thickened packages of terrigenous sediment. The fact that all of these limestone bands grade into beds of discrete concretions further suggests that the carbonates are diagenetic, having formed through peculiar geochemical conditions existing in the sediment.

Moreover, westward thinning of shale units is not strictly proportional, but, rather temporally imbricate (Fig. 3). There appears to be a westward progression of the area of maximum thickness in successively higher shale packages within the Windom. Unit 1 is thickest near West Alden (loc. 13) and thins to a feather edge at Lake Erie shore; Unit 4 obtains maximum thickness at Buffalo Creek (loc. 10), Unit 8 at Cazenovia Creek (loc. 9), Unit 12 at Smoke Creek (loc. 8) and Unit 14 near Cloverbank (loc. 5).

Slight variations in lithology accompany thickness changes. Shales are characteristically darker and more pyritic in thicker areas, and more calcareous in the thinner regions. The thickness variations also coincide with subtle changes in the fossil contents of the beds. In every case fossils are the least common and diverse in areas of maximum shale thickness. In fact, Units 4, 8, and 14 are very nearly barren at and near their thickest areas. Marginward from each depocenter, these shales contain increasingly diverse assemblages, typically beginning with Ambocoelia-chonetid associations which give way to those containing larger brachiopods such as Pseudoatrypa, Mediospirifer and Devonochonetes coronatus. As noted above, slightly increased fossil diversity, associated with thinning, is characteristic of the Bay View coral bed.

Association of faunal and lithologic changes with stratigraphic thickening suggests that increased mud deposition was accompanied by other environmental changes such as a decrease in oxygen levels and/or environmental energy. We suggest that the westward progression of thicknesses in the Windom is a reflection of westward migrating diastrophic ridges ("swells") and basins. If so, it appears to mirror on a small scale a trend which has previously been noted for the Ludlowville and Moscow formations as a whole (Baird, 1979; Baird and Brett, 1981). There is an apparent westward shift of the area of greatest thickness in successively higher units in the Ludlowville and Moscow Formations: The King Ferry Shale is thickest near Cayuga Lake, the Jaycox Shale thickest near Seneca Lake, the Deep Run Shale at Canandaigua Lake, and the Kashong Shale in the Genesee Valley. In each case paleontologic and sedimentologic evidence suggest a coincidence of deepest water conditions with the depocenter. We can only speculate that this pattern was generated by a standing wave-type progression of submarine fold axes, these possibly propagating westward from the tectonically active Acadian region in eastern New York.

Based on the general eastward thickening of most component zones in the Windom Shale one might predict that the thickest Windom sections should exist east of Erie County. Indeed subsurface data (Rickard, pers. comm., 1981) indicate Windom thicknesses in excess of 18 m (60 ft) south of Darien in Genesee County. However, the Windom is considerably thinner (9.5-11 m; 31-36 ft) in measured sections along the outcrop belt in Genesee County (Figs. 3, 4A). This is a result of a separate process discussed below.

Erosional Beveling of the Windom

Detailed examination of the Moscow-Genesee contact reveals evidence for erosive overstep of the Hamilton in a northeastward direction (Fig. 4). Thus, eastward thickening is partially counteracted by northeastward truncation of upper Windom beds probably due to post-Hamilton erosion.

Although fourteen distinctive marker beds and fossil assemblage zones are present in the thin Windom section of western Erie County (locs. 1-7), only the basal three or four of these are present in thicker sections near the Erie-Genesee County border. Thus, ironically, the stratigraphically most complete sections are generally the thinnest. At Amsdell Creek (loc. 4), 2.8 m (9 ft) of shale containing three distinct concretion horizons overlies the Amsdell bed (Unit 13). At Smoke Creek, only 1.5 m (5 ft) of this shale, including only the lower two concretion horizons, is observed. By Cazenovia Creek, the Amsdell bed is in direct contact with Leicester Pyrite or Penn Yan Shale (Figs. 3-5); locally, northward along the same creek the Amsdell bed is completely truncated (see below). Continued northeastward truncation brings successively lower Windom beds against the Genesee contact: the Penn Dixie beds form the contact at Buffalo Creek, the C or D limestone bed at Little Buffalo Creek and apparently the Big Tree Bed at Cayuga Creek near Clinton Road. Finally, at Eleven Mile Creek (loc. 14) in Genesee County the Smoke Creek bed (Unit 3) is less than 2 m below the Genesee contact (Fig. 3). The Smoke Creek bed is at or near this contact along most of the outcrop belt in Genesee County. Locally, at Bowen Creek, the northernmost outcrop, the

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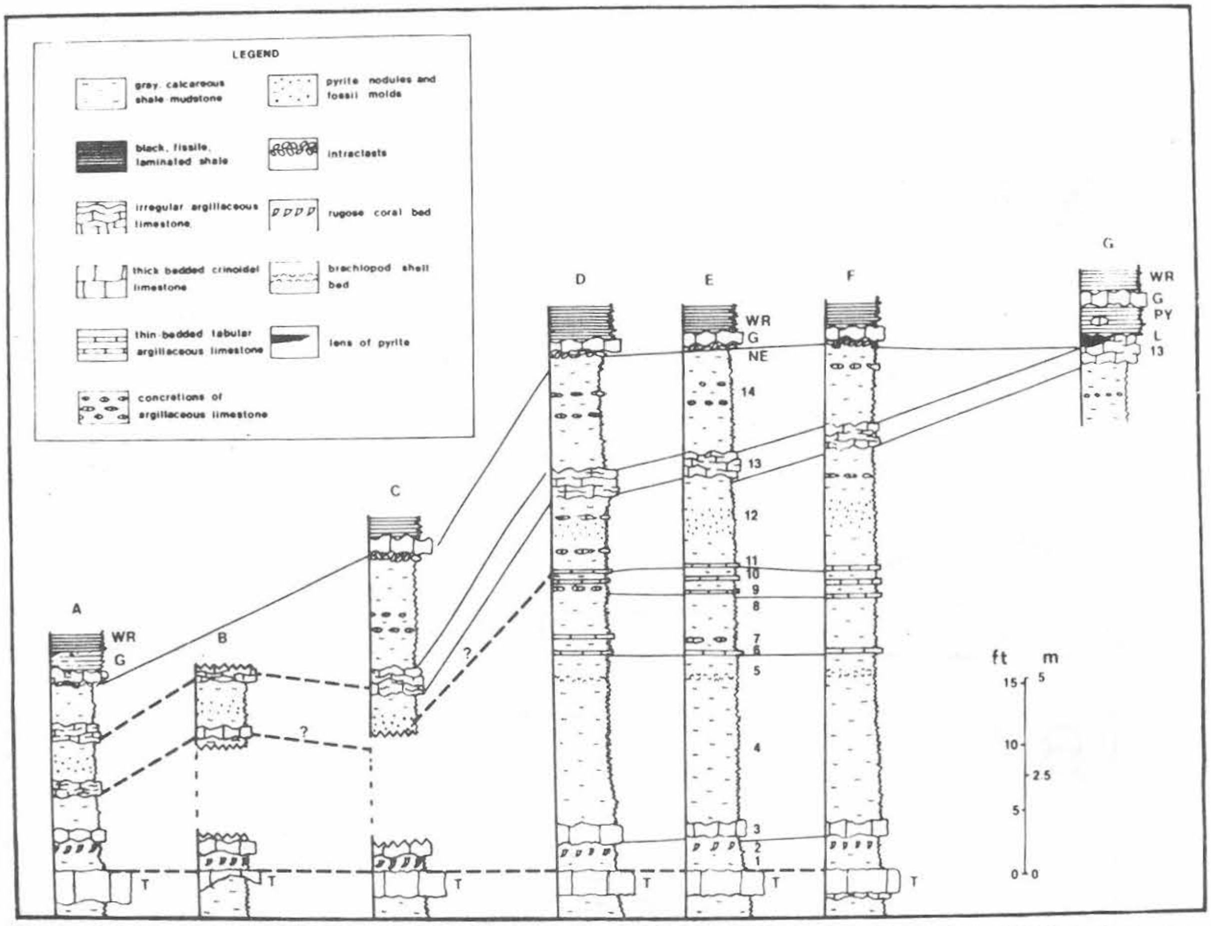
Figure 3. Stratigraphy of the Windom Shale Member in Erie and western Genesee Counties. Columnar sections include: A) Eighteen Mile Creek; B) unnamed creek at Weyer; C) unnamed creek at Amsdell; D) unnamed creek near Big Tree Road; E) Bay View Quarry (shale pit); F) South Branch Smoke Creek; G) Cazenovia Creek (south section); H) Cazenovia Creek (north section); I) Buffalo Creek; J) Little Buffalo Creek; K) Cayuga Creek; L-1, L-2) first west-flowing tributary of Cayuga Creek north of Clinton Road; L-3) second tributary of Cayuga Creek; M) Durkee Creek; N) Eleven Mile Creek. Stratigraphic units include: 1) Ambocoelia umbonata beds; 2) Bay View Coral Bed; 3) Smoke Creek bed; 4) lower barren shales; 5) Big Tree pyritic bed; 6,7) A,B limestones; 8) Buffalo Creek pyritic beds; 9,11) C,D and E limestones; 12) Penn Dixie pyritic beds; 13) Amsdell beds; 14) upper shales and concretions. Symbols: T) Tichenor Limestone; M) Menteth Limestone; K) Kashong Shale; L) Leicester pyrite; NE) North Evans Limestone; PY) Penn Yan Shale; G) Genundewa Limestone: WR) West River Shale.


LEGEND
gray, calcareous shale mudstone
pyrite nodules and fossil molds
black, fissile, laminated shale
intraclasts
irregular argillaceous limestone
rugose coral bed
thick bedded crinoidal limestone
brachiopod shell bed
thin-bedded tabular argillaceous limestone
lens of pyrite
concretions of argillaceous limestone
A
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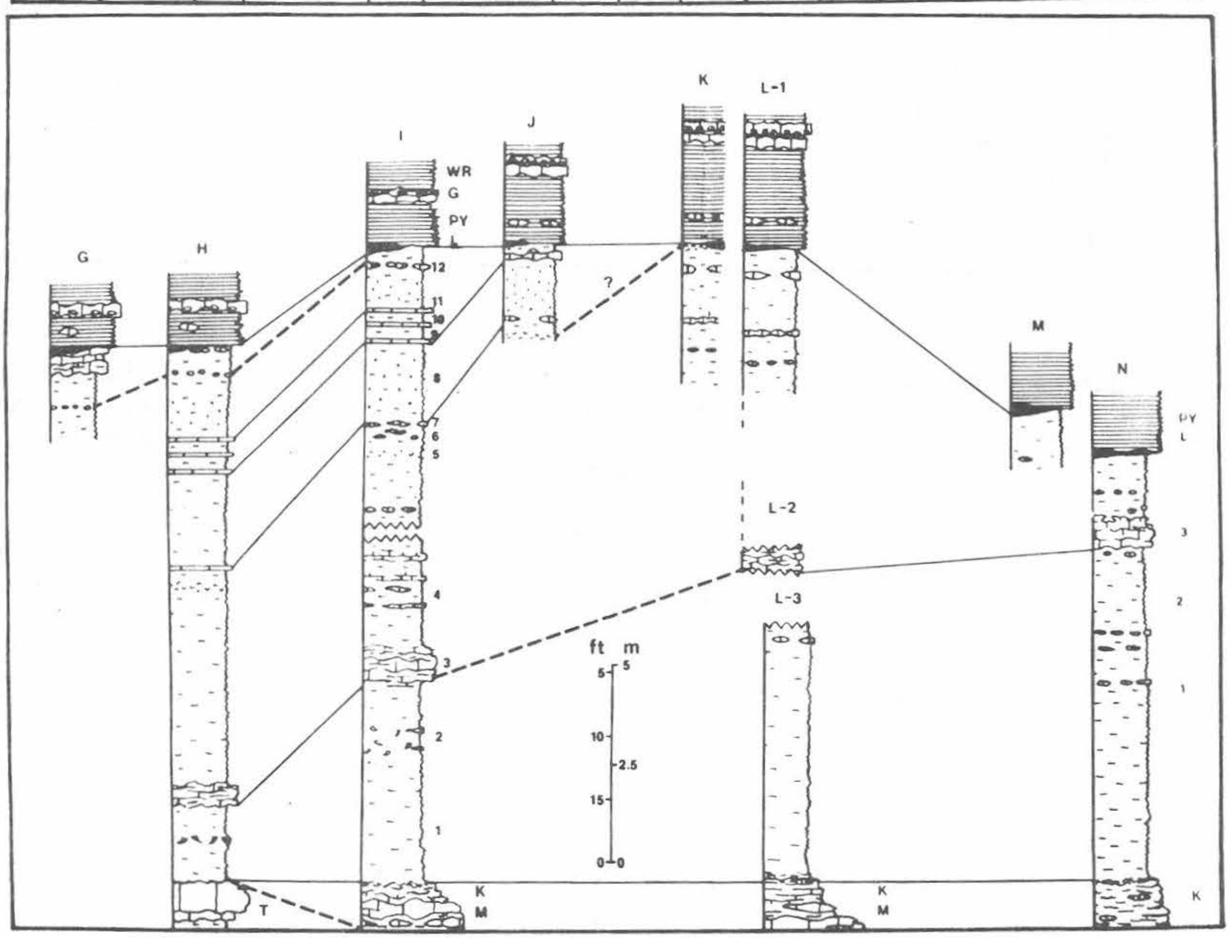
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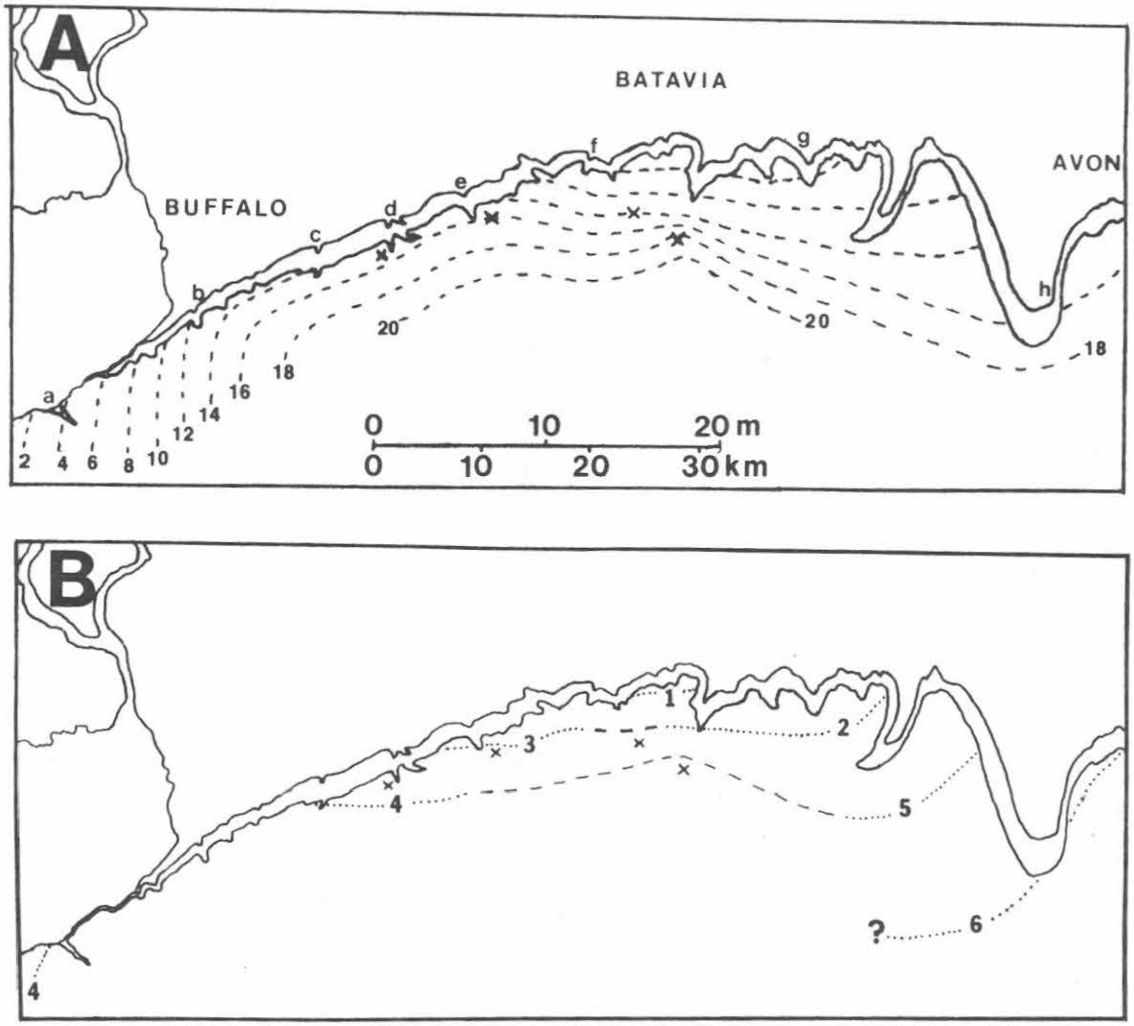

Figure 4. Regional Windom truncation and thickness trends. A) isopach map of Windom based on outcrop and subsurface data; isopleths in meters. B) inferred "paleooutcrop" trends of Windom beds along top Hamilton erosion surface that would be observed if Genesee and younger strata were removed. Localities in A include a) Eighteenmile Creek; b) Bay View shale pit; c) Cazenovia Creek; d) Buffalo Creek; e) Cayuga Creek; f) Elevenmile Creek; g) White Creek; h) Fall Brook. X's denote well sites. In B, strike trends include: 1) Smoke Creek bed; 2) Fall Brook Coral bed; 3) Big Tree bed; 4) Amsdell bed; 5) <u>Ambocoelia? praeumbona</u> beds; 6) uppermost Windom black shales with <u>Alanella tullius</u>.

Smoke Creek bed is also partially beveled. However, southeast of Pavilion, New York, higher Windom zones reappear beneath the contact (Fig. 4B). Thus, only the lower three or four units of the Windom (Ambocoelia beds-Smoke Creek bed) can be traced continuously between Erie and Livingston Counties. As noted above, these beds exhibit relatively little change in lithology or fossil content across this region and can be traced readily into the Finger Lakes region. However, correlation of the higher units with upper Windom beds in Erie County is difficult because, unlike the lower beds, these units exhibit pronounced facies changes across this region, and this change is not observable in outcrop.

Initially, we suspected that the trunction of upper Windom beds in Genesee County reflected an area of differential uplift ("Alexander Arch") which ran roughly north-south through the center of Genesee County. In actuality, this pattern is an artifact of the configuration of the outcrop belt, relative to erosional and depositional strike. The outcrop belt forms a northward arc; it trends northeastward across Erie County, then eastward to Pavilion, and finally southeastward into the Genesee Valley (Fig. 4). Most Windom sections in Genesee County exhibit similar levels of truncation (contact lies within 1-2 m above the Smoke Creek bed) and similar thicknesses. Higher beds appear beneath the Genesee contact both east and west of this area at approximately the regions where the outcrop belt begins to curve southward. This suggests that the direction of maximum truncation is not eastward, but northward.

To test this hypothesis we examined long sections of Windom exposed continuously on north-flowing creeks in Erie and Genesee counties, where overstep could be measured in outcrop (see Figs. 5B and D). The Windom/Genesee contact is exposed nearly continuously along a one kilometer north-flowing stretch of Cazenovia Creek (Stop 2). Local northward truncation is observable along this section; at the southernmost exposure (loc. 9b) the complete Amsdell bed and a few centimeters of overlying shale were found to occur beneath the Genesee unconformity, but near Northrup Road bridge, 1 km farther north, the Amsdell bed is largely truncated and fragments of calcareous shale containing Ambocoelia? praeumbona occur in the base of the Leicester Pyrite. A more precise estimate of difference in relative truncation was obtained by measuring the distance between the Windom-Penn Yan unconformity downward to a distinctive concretion zone in the upper part of Unit 12. This concretion band was found to be 84 cm beneath the contact at location 9a and 155 cm below this same contact at 9b, giving a total northward truncation of 71 cm in one kilometer.

Similarly, rapid northward truncation of the Big Tree fossil bed can be observed along a short north-flowing section of Cayuga Creek near Clinton Road (Stop 5A).

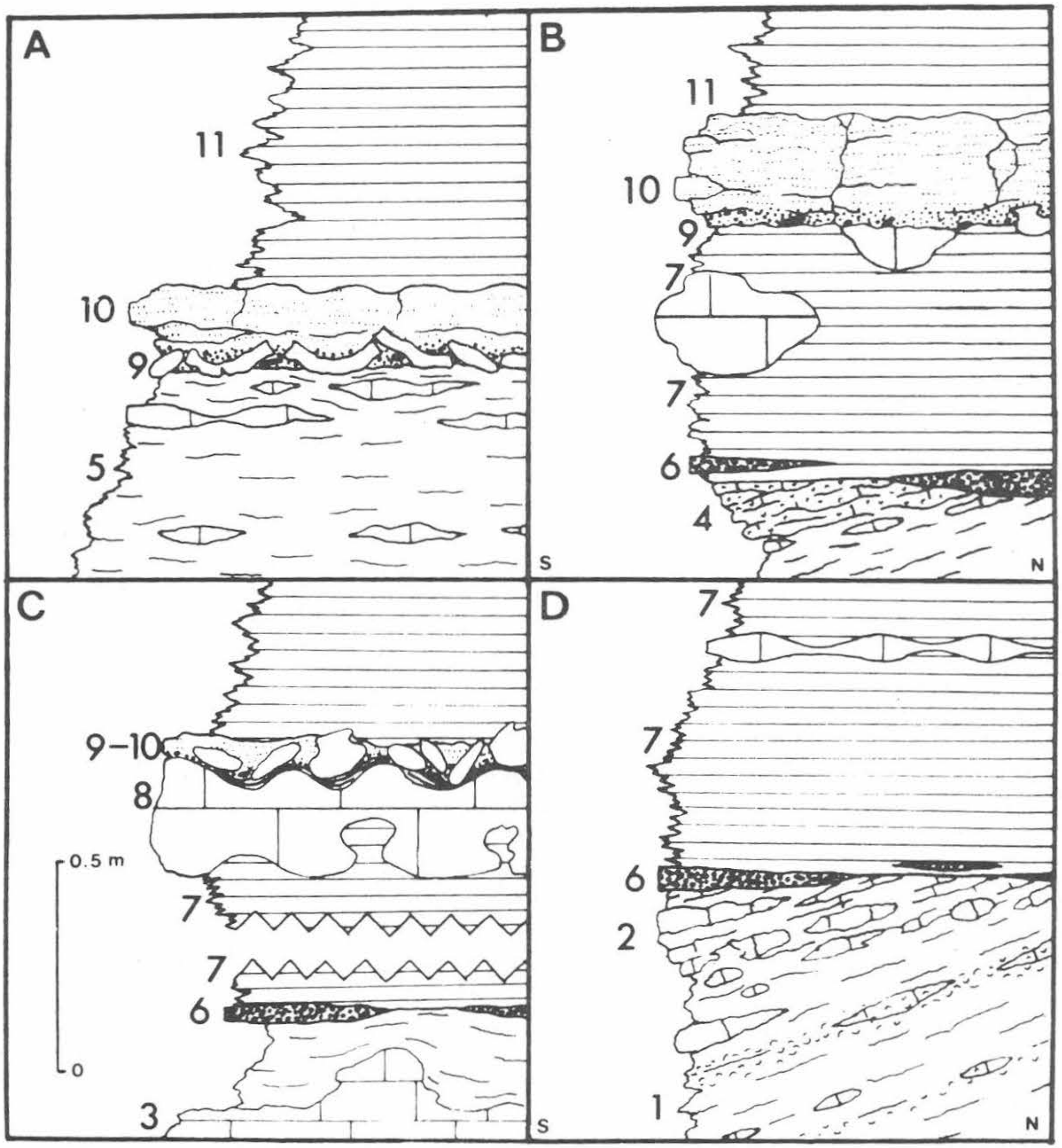

Figure 5. Moscow/Genesee contact and associated beds. Northward erosive overstep of Windom beds is shown schematically in B and D. Localities include: A, Big Tree Creek (Loc. 6); B. Cazenovia Creek (Loc. 9); C. Little Buffalo (Loc. 11); D. Bowen Brook. Units include: 1. Bay View Coral bed; 2. Smoke Creek ed; 3. C-D limestones; 4. Amsdell bed; 5. uppermost Windom Shale (Unit 14); 6. Leicester Pyrite Member; 7. Lower Genesee black shales (Geneseo-Penn Yan members); 8. upper concretionary Penn Yan bed; 9. North Evans Member and eastward equivalent; 10. Genundewa Limestone Member; 11. West River Shale Member.

Such rapid truncations, though suggestive, could result from local undulationsof Windom beds rather than from regional overstep. However, further corroboration of northward truncation is provided by detailed correlation of gamma-ray logs for the Moscow Formation in several wells in Erie and Genesee Counties (Rickard, pers. comm., 1981). Thicker calcareous Windom units, most notably the Smoke Creek bed, were recognized and correlated on these profiles. Wells drilled near Buffalo Creek, Cayuga Creek and Eleven Mile Creek (Fig. 4B) closely match those measured sections and provide outcrop control for gamma-ray patterns. Of particular interest is the Stedman Well south of Darien in Genesee County. It exhibits a combined Moscow Formation thickness substantially greater than that observed at Murder Creek (loc. 15) 5.6 km to the north. Furthermore, in this log profile the peak corresponding to the Smoke Creek bed occurs 8.5 m below the Genesee contact while at Murder Creek this bed is only 1.5 m below the Genesee-Windom contact. This equates to a loss of 1.25 meters per kilometer due to northward truncation. Also, peaks corresponding to the A-E limestones occur below the contact in the well profile; these limestones are missing entirely in the outcrop belt immediately to the north. Hence these subsurface data provide substantial support for the hypothesis of northward truncation of Windom beds.

In summary, Windom zones and beds display not only westward convergence due to sedimentary condensation, but northward-northeastward progressive truncation from the top down. This is best illustrated in Figure 4 where inferred isopach lines for the upper Windom (Units 5-14 interval) trend at nearly right angles to the subsurface erosional strike of several Windom units in western Erie County. Only in eastern Erie and Genesee Counties are these more nearly of coincident strike. The pattern clearly shows that the local condensation of the western Erie County Windom is a process that was largely independent of post-Windom diastrophic events, later erosion having been superimposed on Windom deposits of variable thickness.

STRATIGRAPHY AND CONTACT RELATIONSHIPS OF THE LOWER GENESEE FORMATION

The Genesee Formation in Erie County ranges in thickness from 12 m (40 ft) near the Genesee-Erie County line to 3 m (10 ft) at Lake Erie (Figs. 1, 3). Three component members, Penn Yan, Genundewa, and West River, in ascending order, are traceable across the county. A lower black shale unit, the Geneseo Member, is absent in Erie County (deWitt and Colton, 1978). The Penn Yan and West River Members, composed of dark gray and black bituminous shale, account for the westward thinning, while the Genundewa Member, a pelagic limestone, maintains a nearly constant 15-36 cm (0.5-1.2 ft) thickness range. Two additional Genesee units, Leicester and North Evans Members, are erosional lag concentrations which occur on the Genesee-Windom paraconformity; these have a more restricted distribution in Erie County (see Brett, 1974a; this paper). All Genesee units except the West River will be discussed herein.

Leicester Pyrite

General Character and Stratigraphy. At the base of the Genesee Formation, in most western New York localities, are widely separated lenses of Leicester Pyrite (Sutton, 1951). This distinctive deposit composed of pyritized clasts and fossil steinkerns occurs as a lag concentration both on and immediately above the Windom-Genesee unconformity from central Erie County east to Gage Creek in the Canandaigua Valley. Farther east the Leicester is apparently represented by discontinuous lenses of encrinite in the basal Geneseo Shale. The westernmost locality at which the Leicester can be observed is Cazenovia Creek (loc. 9; Stop 2), where it is well exposed in lenticular pods at the base of the Penn Yan Shale Member (Fig. 5B).

The Leicester Member has previously been interpreted as a condensed lateral equivalent of the Tully Formation (Heckel, 1973; Rickard, 1975). However, lenses of typical Leicester Pyrite overlie the westernmost tongue of the Tully Limestone (Carpenters Falls Bed of Heckel, 1973) at Gage Gully and Leicester-equivalent encrinites occur above the Tully at Seneca Lake (Fulreader, 1957; Huddle, 1981; Klapper, 1981). Conodont studies summarized by Huddle (1981) also indicate a post-Tully age for all Leicester lenses. Moreover, the Leicester is markedly diachronous; westernmost exposures contain conodonts indicative of the lowermost *Polygnathus asymmetricus* subzone, while lenses east of Honeoye Valley belong to the older *Schmidtognathus hermani-Polygnathus cristatus* zone (Huddle, 1981; Klapper, 1981).

The Leicester is composed mainly of granule-to pebble-sized nodular to tubular pyrite with subordinate amounts of pyritized fossil molds, occasional shells, phosphatic nodules, and fish plates. Most of the originally calcareous fossils are preserved as pyritic internal molds; even some pelmatozoan columnals occur as pyritized stereom void-fillings. Most pyrite nodules are irregular but many are distinctly tubular, resembling pyritized burrow fillings which are common in the underlying Hamilton Group (Dick, 1982; Dick and Brett, 1982).

East of the Genesee Valley Leicester pyritic clasts are surrounded by dark gray to black mud matrix. However, western Leicester lenses in Genesee and Erie Counties are characterized by calcite cement. At Cazenovia Creek (loc 9; Stop 2), Little Buffalo Creek (loc. 11), and Cayuga Creek (loc. 12; Stop 5), the Leicester is often a pyrite clast grainstone, nearly lacking intergranular mud.

Sedimentology, Paleontology and Depositional Settings. Evidence of current transport is frequently manifest in Leicester lenses. Tubular clasts may be strongly aligned and foreset beds are present in certain lenses (Fulreader, 1957; this paper). Parallel, linear ridges or furrows also occur on the bases of many Leicester pyrite lenses, although other lenses have smooth, flat bases. Fulreader (1957, p. 37) noted that the long axis alignment of clasts generally parallels the strike of the ridge and groove features. These azimuths trend northeast-southwest in the Canandaigua Valley region, but become more nearly north-south in eastern Erie County. Fulreader interpreted the ridges at the base of the Leicester as casts of ripples on the upper surface of Windom muds. However, bedforms are rarely, if ever, observed in clay-sized sediments (Friedman and Sanders, 1978, p. 86-93). Furthermore, there is evidence that the Windom muds were already lithified or semi-lithified by Geneseo or Penn Yan times and not of a consistency to be rippled; Windom rip-up clasts are common in the Leicester at many localities. Rather, it appears that the ridges are groove casts; indeed, they closely resemble features observed on the soles of turbidites (Sutton, et al., 1970). The grooves thus represent scour features produced in cohesive Windom muds by turbulence and transport of debris by bottom currents (Fig. 6). Parallelism between such sole marks and internal Leicester fabric probably reflects transport and alignment of particles parallel to current direction rather than normal to it as inferred by Fulreader (1957; p. 36).

The Leicester contains a diverse fossil assemblage dominated by brachiopods and mollusks (see Loomis, 1903; Fulreader, 1957, for faunal lists). Loomis (1903) interpreted the fauna as composed of "stunted" or "dwarfed" taxa based on the small size of many forms. Subsequent work by Fulreader (1957) and the present authors shows that a significant size range exists for Leicester fossils and that normal adult individuals are present. Uniformly small size of particular taxa in several lenses is apparently the result of current sorting.

Several Leicester taxa including varcus zone conodonts, are clearly derived from the underlying Windom Shale (cf. Huddle, 1981, p. B8-9). As Windom fossil assemblage zones are overstepped regionally from the Canandaigua Valley northwestward into Genesee County, there is corresponding sequential appearance of fossils from these zones in the Leicester. Similarly, northeastward overstep of the Amsdell bed and overlying shales at Cazenovia Creek (loc. 9; Stop 2) is associated with the appearance of calcitic shells of the diagnostic brachiopod Ambocoelia? praeumbona in the pyrite lenses.

Other Leicester fossils may be of post-Hamilton age: fish bones, and conodonts, in particular, may be of Late Taghanic age (hermanicristatus to lowermost asymmetricus conodont subzones; Klapper, 1981). Although many large brachiopods and bivalves are Hamilton forms and appear to have been exhumed from the Windom, it is probable that some

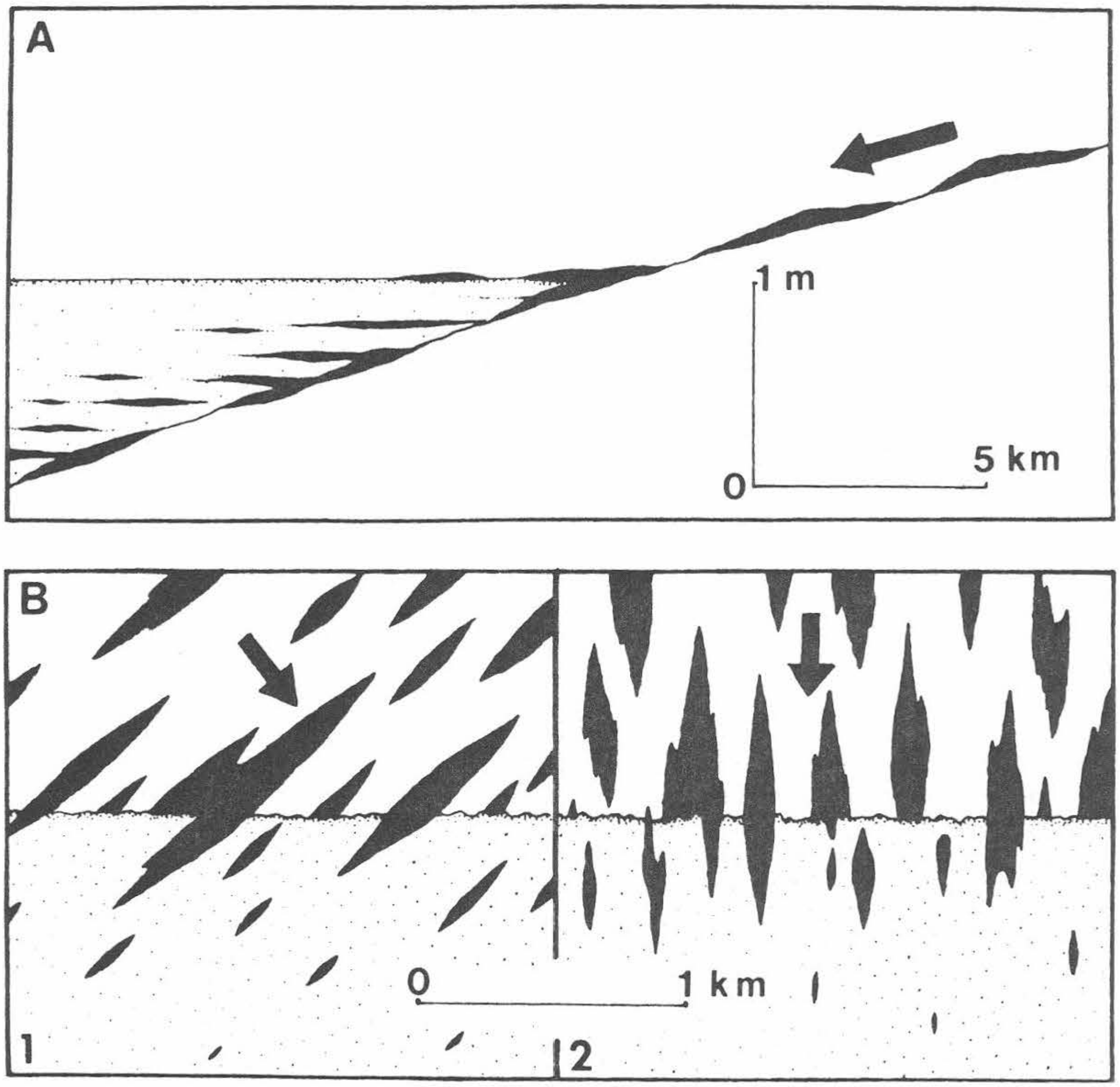

Figure 6. Deposition of Leicester pyrite lenses. A shows hypothetical downslope transport of Leicester reworked debris through current traction. Some lenses migrate out over Genesee mud deposits. Depositional onlap of Genesee muds results in diachronous imbrication of lenses. B shows possible modes of tractive lens migration; in 1, lenses are moved as pyrite granule-sand waves aligned perpendicular to current flow; in 2, lenses are aligned as flutes parallel to current flow.

smaller, easily transported shells may have come from post-Windom habitats ("Tully" shelf or other upslope aerobic settings) at some distance from present Leicester sections.

A key question to be considered is whether the pyritic tubes, nodules and steinkerns were exhumed and reworked as pre-pyritized material or whether the pyritization occurred mainly after lenses were formed and buried (cf. Huddle, 1981, p. B14). It is apparent that much intergranular pyrite formed after Leicester deposition (Park and Weiss, 1972). However, certain pyritic clasts strongly resemble early diagenetic pyrite of the Hamilton Group (Dick, 1982). The interval of time between deposition and submarine re-erosion of Windom muds is clearly of a magnitude such that early diagenetic pyrite would have long been present in this sediment, prior to reworking. Recent studies indicate that much pyrite mold formation may occur within tens to hundreds of years of burial (Berner, 1969, 1970; see Dick, 1982, for review). Moreover, the anaerobic bottom setting suggested by Genesee sediments would favor stability of exhumed pyrite on the sea floor.

Reworking of at least some pre-pyritized material is suggested by the abundance of the aforementioned tubular clasts in the Leicester. These resemble pyritic burrow structures in the Windom, which occasionally protrude slightly above the Windom erosion surface, into the overlying Genesee. Additionally, pyritic fossil steinkerns occur in the Leicester, which display reoriented geopetal and compactional features, clearly indicating an earlier phase of fossil burial and diagenesis.

A more problematical feature of the Leicester is the scarcity of calcareous shells and reworked Windom carbonate debris, even though Windom concretion beds and coral-brachiopod zones are regionally truncated (Baird and Brett, in press). Except for the Leicester-equivalent encrinites in eastern Ontario County (Bellona and Gorham sections) which are largely calcareous, the Leicester rarely contains calcareous fossils. It is probable that much calcareous debris exhumed during post-Windom erosion had already been destroyed by abrasion and bioerosion by the time the Leicester lenses accumulated. Hiatus concretions and large exhumed fossils are similarly scarce along other discontinuities where erosive overstep is indicated such as the base of the Tichenor Member and Tully Formation.

However, the nearly total lack of even calcitic shells and preponderance of pyritic clasts suggests a further mechanism for the destruction of calcareous material. A key to the absence of primary carbonate is the inferred paleoenvironment of the lower Genesee Shales. The Geneseo and, to a lesser extent, the Penn Yan are interpreted as deeper water, anaerobic bottom muds (Thayer, 1974; Bowen, Rhoads and

McAlester, 1974). A highly reducing, low pH environment could have accelerated submarine dissolution of reworked calcareous debris. Thus, the Leicester could, in part, represent a chemical residue of formerly thicker, calcareous-pyritic and phosphatic lag deposits.

Penn Yan and Geneseo Shales

Above the Leicester Pyrite in Erie County is the Penn Yan Shale Member which thins from 3 m (10 ft) at the Genesee-Erie County line to .6 m (2 ft) at Cazenovia Creek (Stop 2). It pinches out southwest of Stop 2 and the overlying Genundewa Limestone Member and subjacent North Evans Member rest directly on the Hamilton from Smoke Creek southwest to Eighteen Mile Creek (Buehler and Tesmer, 1963; Brett, 1974a). At the Lake Erie Shore, southwest of Eighteen Mile Creek, a 25-30 cm (10-20 in) thick black shale reappears below the Genundewa, which may be a black shale wedge within or immediately above the North Evans.

The Penn Yan consists of a dark gray to chocolate-brown, calcareous mudstone with one or more beds of calcareous septarian concretions. Fossils include numerous small Devonochonetes, Styliolina, wood fragments, and both orthoconic and goniatitic cephalopods. The depauperate biota reflects dysaerobic to anaerobic bottom conditions combined with aerobic surface waters capable of supporting planktonic organisms.

The base of the Genesee Formation is diachronous; in western Erie County it is formed by the Genundewa Limestone (Fig. 5A), near Cazenovia Creek by the Penn Yan Shale (Fig. 5A). In Genesee County a still older unit, Geneseo Shale Member (Fig. 5D), appears at the base of the Formation and progressively thickens eastward (Kirchgasser, 1973; Rickard, 1975; deWitt and Colton, 1978).

The Leicester Pyrite must be similarly diachronous as many pyrite lenses occur slightly above the unconformity within the basal Geneseo in the east or the Penn Yan in the west (Fig. 5B). This is corroborated by conodont studies of Huddle (1981) and Klapper (1981), cited above.

Leicester lenses within the black shales reflect lateral current transport during the time of black mud deposition and imply the presence of bottom currents associated with the anaerobic setting. It is suspected that during and following the Taghanic transgression, the deeply submerged relict Hamilton substrate was gently sloped to the east and south; although the surface was progressively covered by onlapping Genesee sediments, first in the deeper basinal areas and later in Erie County, a long period of time persisted when bottom currents scoured the exposed and sloped Windom substrate. The importance of bottom currents is well documented in the deep sea (Heezen and Hollister, 1971), many of these being strong boundary currents which scour large regions of the sea floor. Within the Geneseo black shale there is abundant evidence for bottom currents; Styliolina shells, orthoconic cephalopods, and wood fragments are commonly current aligned. Similar features are likewise known from Mesozoic black and dark gray shale sequences (see Brenner and Seilacher, 1978).

Erosion and transport of Windom fossils and diagenetic structures, accompanied by possible dissolution of shell and nodule carbonate is envisioned to have produced the Leicester lenses (Fig. 6). Transport of some debris across the upslope margin of Genesee sediment accumulation is believed to have produced the secondary, and usually thinner, lenses in the basal Penn Yan and Geneseo; this hypothetical reconstruction is shown in Figure 6.

North Evans Member ("Conodont Bed")

Overlying the Windom Shale from Smoke Creek southwestward to Eighteen Mile Creek is a thin 2-18 cm (1-7 in) lag concentration of hiatus-concretions, bone fragments, pelmatozoan fragments, and conodonts (Figs. 3, 5A) which rests on the Moscow-Genesee unconformity (Hussakoff and Bryant, 1918; Brett, 1974a). Unlike the Leicester, the North Evans is conspicuously carbonate-rich. Pyrite is a variable but minor component, occurring as tubular clasts identical to pyritic burrow tubes in the underlying Windom. The typical crinoid-rich North Evans is present where Penn Yan shale is absent or very thin; southwest of Eighteen Mile Creek where Penn Yan-equivalent black shale reappears beneath the Genundewa, the North Evans changes laterally into a more pyrite dominated "Leicester" type of unit.

The North Evans is a striking deposit which is described in earlier papers (see Grabau, 1898-1899; Hussakoff and Bryant, 1918; Brett, 1974a). It is one of the richest units for conodont yield in the world. Furthermore, the bed contains a mixture of conodonts indicative of three to four distinct subzones (lower asymmetricus, lowermost asymmetricus, upper hermani-cristatus and probably upper varcus subzones) providing evidence for sedimentary condensation (Huddle 1981; Klapper, 1981). Also conspicuous in the North Evans are tabular, glauconite-coated hiatus concretions and limestone fragments derived from the Windom (Fig. 5A). In sections east of Cazenovia Creek, hiatus-concretions are derived from the Penn Yan. Windom limestone fragments commonly contain brachiopods, bryozoans, and trilobites. Mixed throughout the conodont-crinoid calcarenite between the fragments are abundant hybodont and ptyctodont fish teeth, placoderm armor, and unidentifiable bone fragments. The North Evans Member can be sampled at the Penn Dixie (Bay View) Quarry (Stop 1); fish and conodont material can be easily etched from slabs with dilute acid.

From Cazenovia Creek (loc. 9; Stop 2) eastward at least to Eleven Mile Creek (loc. 15), a thin basal portion of the Genundewa Limestone contains some crinoid fragments, large fish bones, conodonts and reworked, glauconite-coated hiatus concretions (Fig. 5B,C). The similarity of this debris with typical North Evans west of Smoke Creek suggests that the two units are coextensive. In turn, this implies that a widespread post-Penn Yan discontinuity exists. This surface may overstep the Moscow/Genesee unconformity in the west.

Any interpretation of the temporal relationships of Leicester Pyrite, Penn Yan Shale, and North Evans Limestone must account for several factors, as follows: First, some reworking of North Evans calcarenites evidently took place contemporaneously with dark mud deposition, as the two facies are locally interfingering (Brett and Baird, 1975). Very probably, the limit of black mud deposition was an upslope boundary (Fig. 6). The region of Genundewa-North Evans-Windom juxtaposition has been referred to as a local structural axis or arch in western Erie County (Brett, 1974a; Brett and Baird, 1975). Shoaling in this area is also indicated by evidence of local high energy (storm?) conditions such as upended rip-up clasts in North Evans calcarenite. Such storm conditions could also have transported North Evans debris layers out over adjacent black muds to produce interfingering. The occurrence of a shale wedge between thin North Evans and Genundewa along Lake Erie Shore also proves that in some places remanié sediments were buried by dark muds.

Second, there are two distinct erosional hiatus surfaces within the Genesee in sections east of Cazenovia Creek (loc. 9): the lower, Windom-Genesee unconformity marked by lenses of Leicester pyrite and an upper, pre-Genundewa discontinuity of lesser magnitude, characterized by a thin North Evans-like calcarenite on eroded Penn Yan Shale. These two discontinuities evidently merge southwest of Cazenovia Creek into the single Windom-North Evans boundary (Fig. 7).

Third, although the Leicester and North Evans both appear to represent submarine erosion-lag deposits, there are distinct differences between them. The North Evans is a sublenticular to tubular unit, rich in calcareous debris and containing only sparse pyrite clasts; in contrast the Leicester is dominated by pyritic clasts, depicted in carbonate and distinctly lenticular. These distinctions evidently reflect subtle differences in the depositional settings of the two units. An explanation for the difference can be seen in the two different Genesee units which overlie the two discontinuities; the Leicester is overlain by dark gray shale while the North Evans is normally overlain by limestone (cf. Huddle, 1981, p. B14). The Penn Yan Shale clearly appears to be an anaerobic type unit as discussed earlier. The Genundewa Limestone similarly lacks evidence

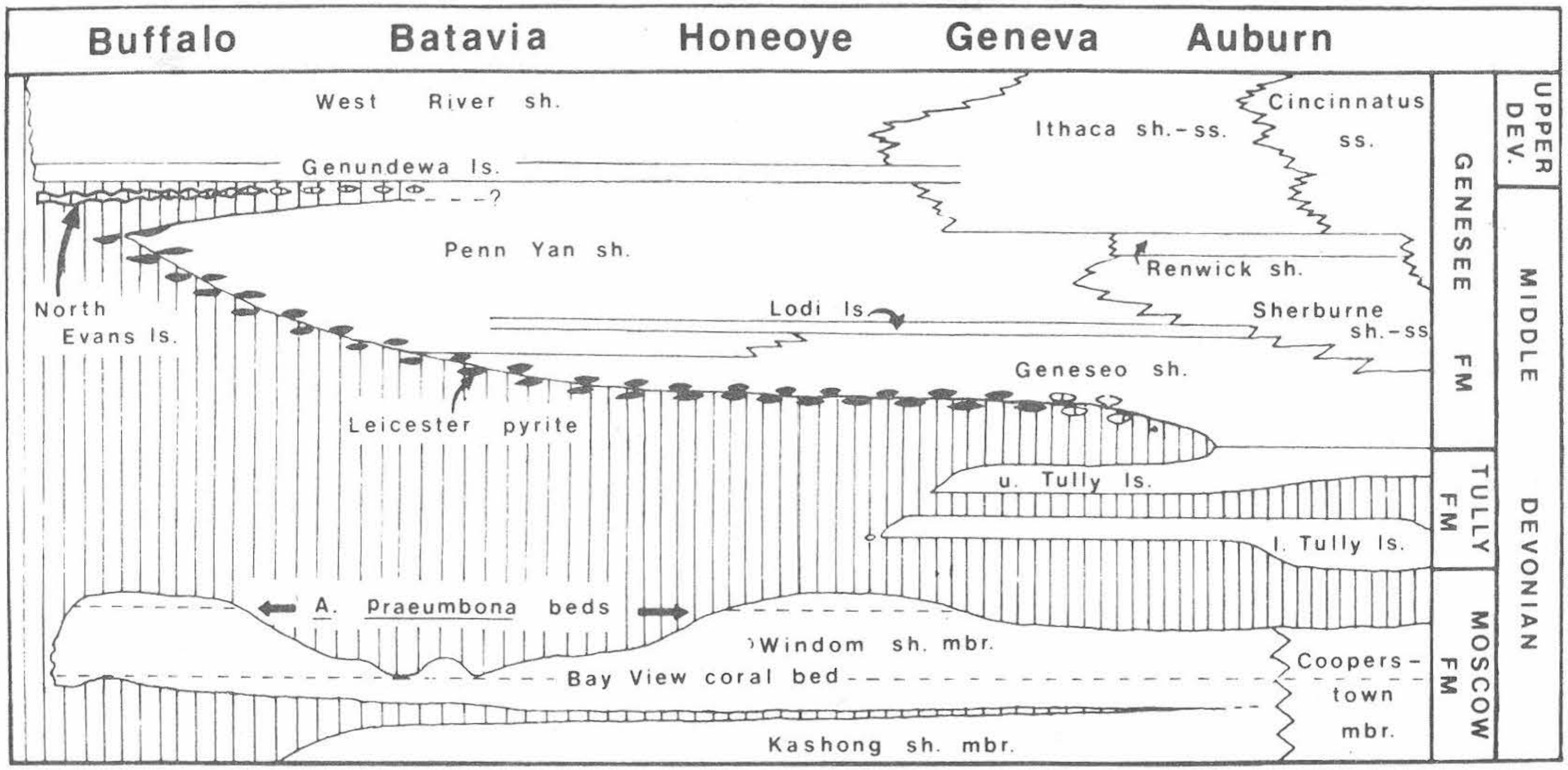

Figure 7. Correlation chart for Moscow Tully, and lower Genesee strata in western New York State. Note proposed revisions. Eastward extent of sub-Genundewa erosion surface is not known at present. Modified from Rickard (1975).

for benthic fauna, but it is calcareous. This suggests somewhat greater oxygen enrichment of the bottom, but still insufficient to support bottom organisms. Such an interpretation is strongly supported by lack of hard surface borers or encrusting organisms on North Evans limestone fragments. It appears that erosion, transport, and burial of the hiatal carbonate material took place within a dysaerobic depositional setting. Both North Evans and Genundewa deposits suggest that the bottom was not so strongly reducing as during deposition of contemporaneous and earlier black muds; a dysaerobic outer shelf and/or slope setting is thus envisioned for the North Evans.

Genundewa Limestone Member

Above the Penn Yan is the thin, but regionally widespread Genundewa Member (Fig. 5a-c); it extends in outcrop from the Lake Erie Shore eastward through the Canandaigua Valley (see Clarke, 1903; Sass, 1952; deWitt and Colton, 1978). In Erie County it is typically a compact 1-50 cm (0.5-1.8 ft) thick resistant bed. In Erie and western Genesee counties, the base of the Genundewa is everywhere abrupt and appears to be erosional.

The Genundewa is composed of a dense concentration of small conoidal shells of the problematic organism *Styliolina fissurella* with lesser amounts of pelmatozoan columnals, goniatite conchs, wood fragments, and fish fragments. Thin-shelled bivalved organisms *Pterochaenia* and *Buchiola* occur in the unit, and, in Ontario County sections, colonies of the crinoid *Melocrinites* have been found associated with wood fragments by Baird; these last may have attached to floating logs during life (Seilacher, *et al.*, 1968; McIntosh, 1978) or they may have grown on waterlogged wood on the sea floor (Kauffman, 1978).

The Genundewa fits closely in general characteristics with certain Devonian pelagic limestones in Europe (see Tucker, 1973, 1974). These latter, described from tectonic rises in Variscan eugeosynclinal deposits of Germany and France (Tucker, 1974), are similarly characterized by abundant *Styliolina*, goniatites, wood fragments, and posidoniid bivalves. The physical setting for the Genundewa is markedly different, however, from the eugeosynclinal setting in Europe and it is questionable that the Genundewa is abyssal as is claimed for the Variscan carbonates. Nonetheless, the Genundewa appears to represent slow pelagic sedimentation in a deeper-water intracratonic shelf-basin environment. It could represent a prolonged effect of decreased sediment influx; hence, the rain of planktonic *Styliolina* and other shelled forms could produce a relatively pure, condensed carbonate blanket instead of the usual black shale.

CHRONOLOGICAL SUMMARY: DEVELOPMENT OF WINDOM/GENESEE DISCONFORMITIES

Figure 7 summarizes the inferred chronological relationships among the several stratigraphic units discussed herein. The following sequence of events is envisaged for the Windom-Genesee unconformity in Erie County, New York. Certain stages are presently hypothetical and the entire sequence is under continuing study.

1) Deposition of Windom Shale sediments includes minor fluctuations of litho- and biofacies (e.g. Ambocoelia-rich dark gray shale to coral dominated, soft gray mudstones). These facies changes are cyclic to some extent and are thought to reflect minor oscillations of depositional environments due to variations in relative sea level. Deeper basinal dark shales record minor transgressions; coral beds reflect minor regressions, coupled with low sediment rates. Facies belts migrated north or south during transgressions and regressions paralleling the northern shoreline.

Largely independent of this cyclic facies variation is a general westward thinning of the Windom, presumably reflecting increasing distance from eastern clastic source areas. Local variation in thickness of particular shale packages within the Windom suggests a migrating axis of differential subsidence perhaps recording diastrophic instability during the late Middle Devonian.

2) Following deposition of the highest Windom unit (Unit 14 and probably higher beds which have subsequently been eroded) a major regression occurred. This was coupled with diastrophic upwarp of the western New York shelf, resulting in gentle, regional southward tilting of Hamilton beds. This event may coincide temporally with upwarp of the Chenango Valley High which, according to Heckel (1975), provided a barrier to clastic sedimentation during Tully deposition.

3) Erosive beveling of Hamilton sediment caused progressive loss of higher units toward the margins of the Appalachian Basin (i.e. toward the northeast in western New York). Overstep of beds probably resulted from submarine erosion and downslope transport of muds. In the process Windom mud was winnowed resulting in a blanket of remanié sediments (fossils, burrow fills, etc.).

The Windom-Genesee unconformity in western New York may actually reflect several periods of submarine erosion. Certainly, a large part of the erosive truncation was accomplished prior to deposition of the late Middle Devonian Tully Limestone, as a Windom truncation surface can be traced beneath the lowest bed of the lower Tully Member (Cooper, 1930; Heckel, 1973). However, Heckel (1973) also recognized a mid-Tully erosion event which destroyed parts of the previously deposited lower Tully (see Fig. 6). Quite probably it further eroded the Windom in areas where the Tully was removed. Finally, a minor post-Tully erosion surface is evidenced by local truncation of upper Tully beds (e.g. Bellona coral bed) west of Cayuga Lake.

Thus, the Windom/Genesee unconformity in western New York (from Canandaigua Lake to Cazenovia Creek) may actually be a composite hiatus surface resulting from the merging of pre-, syn-, and post-Tully disconformities. West of Cazenovia yet another erosion surface merges with these (see below).

4) Following most truncation of the Windom-Tully beds (uppermost varcus subzone time) minor diastrophic upwarp produced a local shoal area (Buffalo Arch) in southwestern Erie County; erosion of upper Windom beds took place here, episodically, during severe storms.

5) During latest Tully (Filmore Glen?) to latest Penn Yan deposition (hermani-cristatus to lowermost asymmetricus subzone) conodonts and fishbones accumulated along with older, erosion-derived relict debris on the truncated Windom surface. This period of prolonged non-deposition and minor erosion coincided with the Taghanic onlap which began with widespread deepening of waters over western New York. These remanié sediments (Leicester lenses) were variably reworked by deep bottom currents, forming starved ripple and/or flute-like lenses. In anaerobic areas, low pH conditions may have resulted in dissolution of carbonates yielding a pyrite-phosphorite enriched residue. In somewhat shallower (upslope), dysaerobic regions (e.g. Buffalo Arch and areas north of the present outcrop belt) calcareous North Evans-type remanié sediments persisted.

6) Input of fine clastics, beginning in latest Givetian time, resulted in deposition of black Geneseo muds in deeper portions of the Appalachian Basin. These laminated muds grade upward into dark gray concretionary Penn Yan sediments; Genesee detrital sediments did not completely onlap the study area until late Penn Yan (lower P. asymmetricus subzone) time. In exposed areas remanié sediments (Penn Yan-age Leicester) continued to be reworked intermittently even after initial deposition of the dark muds nearby. At times fine debris layers were spread out over nearby mud deposits. During this time interval the Buffalo Arch remained an area of extremely slow sedimentation and eposodic erosion. Here younger conodonts (of the lower P. asymmetricus subzone) were ultimately added to the palimpsest deposits. Sediments underwent synsedimentary lithification locally but were then reworked and broken into lithoclasts. Interfingering of black muds with North Evans near shoal margins (e.g. Eighteen Mile Creek) reflect lateral storm transport of the encrinite-lag debris off the Buffalo Arch. From Eighteen-mile Creek southwest to the Lake Erie Shore, intermediate Leicester-NorthEvans-type lenses and encrinites occur beneath a thin black shale which is placed in the Ancyrodella rotundiloba or lower asymmetricus zone (equivalent to upper Penn Yan Shale;

see Huddle, 1981, p. B15). The relationship between this shale and the Penn Yan Shale and Leicester Pyrite east of the Buffalo Arch is uncertain, although the thin shale and true Penn Yan may be continuous south of the outcrop belt.

7) A regressive episode occurring in earliest Frasnian (lower asymmetricus) time resulted in a Penn Yan-Genundewa disconformity in western New York. West of Cazenovia Creek this erosion surface merges with the older pre-, syn-, and post-Tully disconformity forming a composite unconformity of considerable magnitude between the Windom Shale and North Evans limestone. During the pre-Genundewa event erosion removed upper Penn Yan muds and possibly western equivalents of the Leicester pyrite. Erosion reworked the blankets of North Evans relict sediments and ripped up calcareous Windom clasts in the area of the Buffalo Arch. The older condensed sediments were thoroughly homogenized forming a "composite lag deposit" (Huddle, 1981, p. B8). In areas farther east the same erosional event disinterred and reworked syngenetic concretions from the upper Penn Yan Shale. A thin layer of encrinite, fish fragments, and conodonts (North Evans eastern correlative unit) was also deposited over the eroded upper surface of the Penn Yan. This material may have been derived from adjacent exposed shoals, including the Buffalo Arch.

8) Rapid transgression with accompanying clastic sediment starvation resulted in deposition of a condensed Styliolina-rich, carbonate ooze (Genundewa Limestone) which accumulated as a continuous blanket over the entire erosion surface.

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ROAD LOG FOR DEVONIAN MOSCOW/GENESEE UNCONFORMITY
IN ERIE COUNTY, NEW YORK

TRIP

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
00.0	00.0	Begin trip at Buffalo Marriott Inn. Turn right (south) onto Millersport Highway and proceed to entrance for Youngmann Highway (I-290).
00.4	00.4	Take entrance ramp for I-290 east-bound, and proceed eastward on Youngmann Highway.
2.9	2.5	Roadcuts in cherty limestone of the Clarence Member, Onondaga Limestone (Middle Devonian). This is the site of the Vogelsanger Quarry; prior to construction of the highway, this quarry exposed a well developed reef in the Edgecliff Member (now covered).
3.4	0.5	Junction Route I-90, New York State Thruway; keep right and merge onto westbound (actually southbound) lane, toward Erie, Pennsylvania.
4.9	1.5	Overpass of Route 33, Kensington Expressway (Interchange 51).
6.5	1.6	Overpass of Walden Ave. (Interchange 52).
6.9	0.4	Overpass of New York Central Railroad tracks
7.6	0.6-0.7	Overpass of I-90 over combined Erie-Lackawanna and Lehigh Valley Railroad tracks
9.3	1.7	Interchange 53; I-190; continue on I-90 west.
10.1	0.8	Cross Buffalo River; 0.3 mile west of junction of Cayuga and Buffalo Creeks.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
11.0	0.9	Interchange 54 for Routes 16 and 400.
11.9	0.9	Cross Cazenovia Creek.
12.8	0.9	Overpass of Route 16, Seneca Street, Interchange 55.
13.1	0.3	Cross North Fork of Smoke Creek.
13.3	0.2	Overpass over Pennsylvania Railroad tracks.
13.6	0.3	Toll Booth, receive ticket; proceed to first exit after booth.
14.5	0.9	Take exit 56 for Mile Strip Road (Blasdell); bear right around ramp.
14.85	0.35	Toll Booth, pay $.15; then proceed straight ahead to intersection of Mile Strip Road.
14.9	0.05	Junction of Mile Strip Road. Go straight across onto Route 299.
15.1	0.2	Junction of South Park Road (Route US 62). Turn left (south).
16.2	1.1	Junction of Big Tree Road. Turn right (west).
16.6	0.4	Junction of first road on left, after West Avenue, which cuts across intersection of Big Tree and Bay View Roads. Turn left onto the cutoff road and park vehicles. Walk north across Big Tree Road and proceed into entrance to Penn Dixie Quarry; continue on foot for about 0.3 mile and turn into shale pit on the east side of quarry road.

STOP 1 (1 1/2 hours): BAY VIEW (PENN DIXIE) QUARRY. This abandoned shale pit, previously described in detail (Brett, 1974b), provides an excellent reference section for the Windom Shale and its upper contact with the Genesee Formation. Here the Windom Member is approximately 13 m (42 ft) thick; 14 units described in the text can be recognized at this location. The disconformable basal contact with the Tichenor Limestone is excellently exposed in a low domal outcrop in the northeast corner of the shale pit.

The basal Ambocoelia umbonata beds, Bay View Coral bed (type section) and Smoke Creek bed are accessible in this area. Barren shales of unit 4 occupy much of the flat floor in the center of the shale pit. Near the top of this interval is a narrow zone containing abundant fossils including highly compressed brachiopods (Pseudoatrypa, Mediospirifer, Mucrospirifer consobrinus) and rare pyritized sponges and blastoids; this unit is termed the Big Tree bed (unit 5). The A-E limestones (units 6-11) crop out as uniform, white-weathering bands which trend roughly east-west across the quarry. Soft shales above the highest limestone band (unit 12) yield abundant pyritized burrow fillings and fossil steinkerns including nuculid clams, gastropods and cephalopods. This interval of approximately 2 m (6 ft) thickness, termed the Penn-Dixie pyritic beds, is characterized by small specimens of the brachiopods Tropidoleptus carinatus and Ambocoelia cf. A. nana. The calcareous Amsdell bed (unit 13; = "Praeumbona bed") caps a low bench near the southern end of the quarry. Beyond this the Penn-Dixie shale pit terminates in a low (2-3 m high) bank which exposes uppermost Windom beds (unit 13) and the contact with the overlying Genesee Formation. The upper beds of the Windom comprise soft,gray shales with concretionary beds near the top; the top of the Windom is marked by a concretionary layer of 20-30 cm (8-12 in) thick containing Ambocoelia? praeumbona, followed by a thin band of argillaceous limestone which has been partially torn up and incorporated in the overlying North Evans Member.

The North Evans Limestone of the basal Genesee Formation comprises 3-10 cm (1.5-4 in) of buff-weathering, dark-gray crinoidal, calcarenitic limestone, which contains very abundant conodonts of mixed zones (lower asymmetricus subzones). Angular intraclasts, up to 10 cm (4 in) across, derived from the upper Windom occur abundantly in the North Evans. Black shales (Geneseo and Penn Yan) are absent from the lower part of the Genesee Formation and the Genundewa Member, a thin Styliolina-rich limestone rests directly on the North Evans.

At Penn Dixie Quarry as at most localities in western Erie County, the Windom is thin, but stratigraphically quite complete. However, ironically, in these sections unlike those farther east, the upper contact exhibits evidence for erosional scouring. Presumably the scouring of the upper Windom beds and presence of pebble imbrication within the North Evans remanié sediments reflect a late episode of erosion, associated with the pre-Genundewa erosion surface. This event post dates the development of a south dipping truncation surface,prior to Tully deposition, which had left upper Windom beds relatively intact in southwestern Erie County sections.

16.6		Leave Penn Dixie Quarry and return to transportation. Proceed across cutoff road to the intersection of Bay View Road.
16.65	0.05	Intersection of Bay View Road. Turn left (southeast).
17.45	0.8	Junction Route U.S. 62. Proceed straight across on Bay View Road.
17.55	0.1	Junction Route U.S. 20, Southwestern Blvd. Turn left (northeast).
17.85	0.3	Overpass over New York State Thruway (I-90) and Rush Creek.
18.25	0.4	Junction Route 20A, at 6-way intersection. Proceed straight on Route 20.
19.55	1.3	Junction Abbott Road. Proceed on US 20.
19.95	0.4	Entrance to Rich Stadium.
20.15	0.2	Cross south branch of Smoke Creek.
20.25	0.1	Junction California Road. Continue on US 20.
20.7	0.5	Overpass of Route 219 (Southern Tier Expressway).
22.15	1.45	Cross north branch of Smoke Creek.
22.3	0.15	Junction Route 277 (Union Road).
23.15	0.85	Junction Michael Road.
24.0	0.85	Junction Reserve Road.
24.1	0.10	Junction Angle Road

24.9	0.8	Junction Leydecker Road
25.2	0.3	Curve in Route 20 to Junction with Route 78 (Transit Road); prepare to turn right.
25.45	0.25	Junction Kingsley Road. Turn right (east).
25.95	0.5	Junction Northrup Road. Turn right (south).
26.55	0.6	Park near location where power lines cross Northrup Road. At utility poles on east side of road turn right and walk down dirt path to the base of cliff on the flood plain of Cazenovia Creek. At bottom of path turn left and walk north for about 600 ft, following the base of cliff (swampy ground), to exposures along the west bank of Cazenovia Creek.

STOP 2 (1 1/2 hours): CAZENOVIA CREEK SECTION. This east facing cliff provides an excellent, long section of the Windom/Genesee contact. The Genundewa Limestone is nearly at water level at the southern (upstream) end of this section; however, downstream about 1/10 mile some 2 m (7 ft) of Penn Yan and upper Windom shales are exposed beneath the Genundewa. Lowest beds, exposed in the creek floor, are the Penn Dixie pyritic beds of the upper Windom. Diagnostic small specimens of Tropidoleptus carinatus can be obtained in abundance at this location. Overlying beds contain a zone of distinctive ellipsoidal concretions, typically with pyritic cores. The Amsdell bed (unit 12) forms a prominent, light-gray weathering calcareous band about 1 m thick near the top of the Windom. Near the southern end of the section this bed forms a small waterfall in Cazenovia Creek. Note the absence of upper shales (unit 13) in contrast to Penn Dixie Quarry (Stop 1).

The contact between the Windom and the overlying Genesee is sharp and planar with little evidence for erosive scour. Four or five lenses of the Leicester Pyrite, ranging from 1-15 cm thick and up to 2 m across, occur along the Windom/Penn Yan contact at widely spaced intervals. The pyrite contains abundant, reworked?, pyritic burrows, diminutive brachiopods, mollusks, fish bones and wood. One lens along this section exhibits interfingering with the black Penn Yan shale. Lower surfaces of the pyrite lenses exhibit parallel ridge and furrow marks suggestive of current scour of the underlying Windom muds. At this locality about 0.5 m (1.6 ft) of barren, black, laminated Penn Yan Shale overlies the Windom; the upper portion of this shale contains large septarian concretions that locally protrude up into the overlying Genundewa. The Genundewa forms a prominent overhanging ledge about 40 cm (15 in) thick. The lower surface of this ledge is undulatory and is marked by a veneer of encrinite, conodonts, carbonized wood, and abundant fish bones and rare hiatus concretions. This unit probably is correlative in part with the typical North Evans a few miles west of this locality. The upper contact of the Genundewa with the silty, dark-gray West River Shale is gradational. Lower West River beds here yield abundant Pterochaenia, wood, chonetids, pyritized goniatites and rare crinoid columns. Higher units exposed in the upper cliff face include the black Middlesex Shale and gray, concretionary Cashaqua Shale, members of the Sonyea Formation.

26.55		Return to vehicles and reverse direction, proceeding north along Northrup Road.
27.45	0.9	Northrup Road bridge over Cazenovia Creek. Note falls in creek over the Tichenor Limestone of the basal Moscow Formation, just upstream (east) of the bridge. High cliffs visible in the distance, upstream from the falls, expose about 12 m (40 ft) of Windom shale, capped by the Genesee Formation. The Smoke Creek bed is near the base of the cliff. At the upper contact, the Amsdell bed has been largely eroded away.
28.35	0.9	Junction Route 16 (Seneca Street). Turn right (southeast).
28.95	0.6	Junction Rice Road. Turn left (east).
29.45	0.5	Overpass for Route 400 (Aurora Expressway). Note exposures of black Rhinestreet Shale (upper Devonian).
30.95	1.5	Junction Bowen Road. Turn left (north).
32.05	1.1	Junction Bullis Road. Turn right (east).
33.65	1.6	Junction Girdle Road. Continue on Bullis Road.

33.85	0.2	Intersection with old Bullis Road (loops to south over old bridge). Turn right onto old road.
34.10	0.25	Park near old bridge over Buffalo Creek. Walk out onto bridge.

STOP 3A (1/4 hour): BUFFALO CREEK AT OLD BULLIS ROAD BRIDGE. From bridge, view exposures along Buffalo Creek. Downstream (north) and just south of new bridge a series of ledges crop out in the creek floor, representing a condensed upper Ludlowville/lower Moscow section; the lowest ledge is a basal bed of the Jaycox Member, followed by Tichenor Limestone, and calcareous ledges of condensed Deep Run, and basal remnant of Menteth Members followed by highly calcareous Kashong Shale Member. The Windom/Kashong contact, marked by scattered phosphatic nodules is exposed in the floor of the creek directly beneath the old bridge. Looking upstream (south), note high bank exposures of gray, lower Windom Shale. A light-weathering calcareous band about 4 m (13 ft) about the creek floor represents the Smoke Creek bed; concretionary shales beneath this level for about 1 m (3 ft) comprise a local manifestation of the Bay View Coral bed. Note the thickening of the lower _Ambocoelia_-rich shales, here about 3 m thick, compared to about 15 cm (6 in) at Penn Dixie Quarry (Stop 1). At the next stop we will see higher beds of the Windom Member, the lowest of which are slightly higher than the top of this cliff section.

34.10		Return to vehicles and retrace route to new section of Bullis Road over Buffalo Creek
34.35	0.25	Turn right and proceed east on Bullis Road.
34.7	0.35	Intersection Stolle Road. Turn right (south).
35.1	0.4	Turn left on small road leading down into gravel pit; drive to its end and park. Continue on foot for about 1200 ft, downhill and through the woods to bank of Buffalo Creek. Walk south along northeast side of creek to high bank of Windom Shale.

STOP 3B (1 hour): BUFFALO CREEK, UPPER WINDOM SHALE SECTION. This bank exposes about 5 m of upper Windom overlain by Genesee Formation. The lowest shales exposed in the creek bed contain abundant _Zoophycos_ spreiten and thin layers rich in the chonetids, _Pseudoatrypa_, _Mediospirifer_, and other brachiopods. Pyritized fossils including nautiloids and one cluster of blastoid _Hyperoblastus_ have been discovered at this level. These shales appear to represent the Big Tree fossil bed (unit 5). They are overlain by two or three horizons of large, flattened ellipsoidal concretions probably equivalent to the A and B limestone bands farther west. Large irregular masses of pyrite occur associated with these concretions. Above the concretions is an interval of friable, dark gray shales, with abundant pyritic burrow tubes and scattered pyritized nautiloids, nuculid clams, wood and other fossils, herein designated the Buffalo Creek pyritic beds (unit 8). The C,D and E limestone beds (units 9-11) are clearly visible in the upper part of the bank. At the far south end of the section it is possible to examine the upper contact of the Windom with the overlying Genesee. The last meter of soft shale immediately beneath the contact contains abundant, small Tropidoleptus diagnostic of the Penn Dixie beds (unit 12). Small ellipsoidal concretions just beneath the contact may correlate with those in the upper Penn Dixie beds at Cazenovia Creek. Note the absence of Amsdell and upper shale beds (units 13 and 14) in contrast to Cazenovia Creek and localities farther west. These upper units have been truncated prior to deposition of the overlying Penn Yan Member. Lenses of Leicester Pyrite are present at the contact in several locations, but are not readily accessible, although the Leicester can be observed in fallen blocks. The Genundewa Limestone ranges from 0 to 10 cm in thickness; Styliolina limestone and glausonite coated hiatus-concretions occur in depressions on an undulatory erosion surface at the top of a ledge-forming concretionary band 1.6 m above the base of the Penn Yan.

35.1		Return to vehicles and drive from gravel road back to Stolle Road.
35.2	0.1	Turn left (north on Stolle Road and retrace route to Bullis Road.
35.6	0.4	Junction Bullis Road. Turn right (east).
37.6	2.0	Junction Two Rod Road (Route 358). Turn left (north).
37.7	0.1	Cross Little Buffalo Creek.
37.8	0.1	Park along roadside opposite the home of V.I. Boldt, who has kindly provided access to this creek. Walk eastward through field behind house and barn about 1000 ft to creek bank.

STOP 4 (1 hour): LITTLE BUFFALO CREEK, UPPER WINDOM SECTION. This small creek exposes an excellent section of the Upper Windom, Leicester, Penn Yan Shale (here about 1.8 m thick) and Genundewa Limestone. Lowest exposures include the richly fossiliferous Big Tree bed in the Windom, slightly below a layer of flat concretions. This fossil bed yielded atrypids, Spinocyrtia and other species. The Buffalo Creek pyritic beds (unit 8) are well exposed in the floor of the creek upstream and here contain Devonochonestes coronatus and rare Pseudoatrypa in addition to pyritic tubes and nautiloids characteristic of the bed farther west. The uppermost beds of the Windom include a band of argillaceous limestone followed by shale and irregular large septarian concretions. These units comprise the C and D limestone beds respectively. The upper E unit as well as the Penn Dixie beds are missing at this locality. In places, the septaria of unit D protrude upward and deform the basal Penn Yan Shale: lenses of Leicester Pyrite lap directly onto these concretions. Upstream from the Windom/Penn Yan contact a spectacular bed of grotesquely shaped concretions occurs in the upper Penn Yan Shale immediately below thin Genundewa Limestone. Here the Genundewa contains a basal veneer of North Evans lithology, including glauconite and pyrite-coated hiatus concretions. These concretions, characterized by internal lamination, have been completely disinterred from the Penn Yan and rolled into new orientations. This pre-Genundewa disconformity is believed to correlate westward with the North Evans limestone.

37.8		Return to vehicles and proceed northward along Two Rod Road.
38.75	0.95	Junction Route 354 (Clinton Road). Turn right (east).
39.75	1.0	Intersection of Four Rod Road. Continue straight on Route 354.
40.75	1.0	Intersection of Three Rod Road. Continue straight on Route 354.
41.3	0.55	Turn left into driveway just before intersection with Eastwood Road on right. Park at end of the drive and proceed on foot to lower section of Cayuga Creek. Walk back upstream toward Clinton Road bridge.

STOP 5A (1/2 hour): CAYUGA CREEK, UPPER WINDOM SECTION. The upper 3.5-4 m (11-13 ft) of the Windom Shale are exposed beneath the bridge. There are two zones of ellipsoidal calcareous concretions, separated by about 2 m (6.5 ft) of nearly barren pyritic fissile dark gray shale. The lower concretion band is well exposed in the creek floor north of Clinton Road; surrounding shales contain abundant small brachiopods and a pyritic fauna with large masses of pyrite. This interval resembles superficially the Buffalo Creek pyritic bed (unit 8) but is apparently coextensive with lower barren shales of unit 4 seen farther west.

An upper concretionary band occurs 15-30 cm (6-12 in) below the upper contact of the Windom. Shales immediately above this concretion level are highly fossiliferous and contain abundant chonetids, Mediospirifer, Spinocyrtia, Athyris and Pseudoatrypa. This band appears to be coextensive with the Big Tree bed (unit 5) and it may also represent a western remnant of the Fall Brook Coral bed seen in Genesee County and farther east.

The upper contact of the Windom is sharply defined and exhibits local truncation; the Big Tree bed is observed to approach the contact as it is traced northward along this section of Cayuga Creek. The Windom is abruptly overlain by black fissile shale (possible westward limit of the Renwick Member); it contains oriented Styliolina and patches of chonetid brachiopods. Large lenses of rusty-weathered Leicester pyrite occur at the contact near the Clinton Road bridge. These have yielded well-preserved pyritized wood, fish bones, and other fossils.

41.3		Return to vehicles. Turn left from driveway and proceed eastward along Clinton Road.
41.5	0.2	Cross Cayuga Creek and prepare to turn right.
41.55	0.05	Turn right into driveway on east side of bridge and park at end of drive. Walk to falls and remains of concrete dam across Cayuga Creek.

STOP 5B (1/4 hour): CAYUGA CREEK, LOWER GENESEE SECTION. This section has been thoroughly described in previous field trip guidebooks (see Kirchgasser and Brett, 1981, p. 18). The Genesee Formation here includes 1.3 m (4.3 ft) of black and dark gray (Renwick?) Shale, followed by 1.8 m (6 ft) of gray, concretionary Penn Yan Shale. A prominent line of concretions occurs near the top of the section and is probably correlative with that seen in Little Buffalo Creek. A ledge of Genundewa Limestone about 20 cm (8 in) thick caps a low waterfall near the remains of an old concrete dam. The base of this ledge, again, contains encrinite and black hiatus concretions indicating an eastward extension of the North Evans type lithology associated with a pre-Genundewa erosional unconformity.

41.55		Return to vehicles. Turn left (west) onto Clinton Road and proceed westward.

42.6	1.05	Junction Three Rod Road. Turn left (south).
43.7	1.10	Junction Bullis Road. Turn right (west).
52.9	9.2	Junction Route 28, Transit Road. Turn left (south).
53.4	0.5	Junction 400, Aurora Expressway. Take entrance ramp for Route 400 north (=west).
58.4	5.0	Junction I-90 (NY State Thruway). Take exit for I-90 eastbound (actually northbound in this section).
66.5	8.1	Junction I-290, Youngmann Highway (Exit 50). Exit onto I-290 west.
69.	2.5	Junction Millersport Highway (Route 263). Take exit for Millersport Highway, north.
69.4	0.4	Return to Marriott Hotel.

A-2

CARBONATE FACIES OF THE ONONDAGA AND BOIS BLANC FORMATIONS NIAGARA PENINSULA, ONTARIO

By

MARY ROSE CASSA
Gulf Research and Development Company
Pittsburgh, Pennsylvania

and

DON L. KISSLING
ERCO Petroleum Services, Inc.
Houston, Texas

INTRODUCTION

Devonian bioherms of the northern Appalachian Basin have been an important source of hydrocarbons since the 1967 discovery of gas from an Onondaga pinnacle reef in Steuben County, New York. Onondaga bioherms are represented in the outcrop belt by small organic buildups which have been studied for at least three-quarters of a century (Grabau, 1903; Stauffer, 1913, 1915). More recent studies of these buildups and their components enable us to better understand the paleogeographic distribution of the buildups, their internal facies structure, and their importance in hydrocarbon exploration.

STRATIGRAPHY

The Onondaga Limestone crops out in a narrow belt across New York (Figure 1) from Albany to Buffalo and continues the length of the Niagara Peninsula of Ontario as far as Hagersville, where it becomes lost beneath glacial drift. South of the outcrop belt the Onondaga is present everywhere in the subsurface.

Within the last 25 years, Onondaga stratigraphy and nomenclature have been modified and refined (Oliver, 1954, 1956a,b, 1966, 1976). Because of their similarity to Onondaga Limestone members

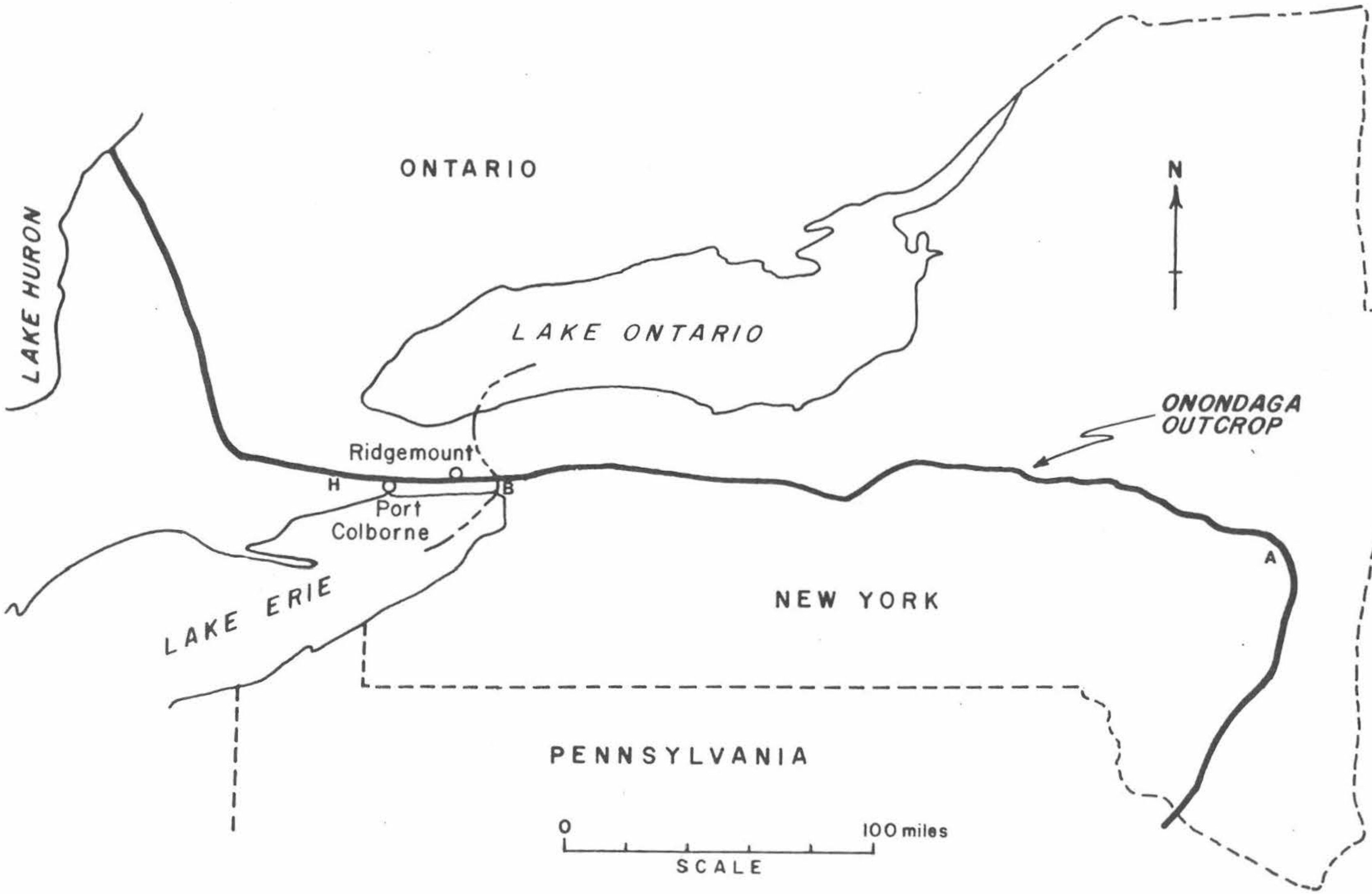

FIGURE 1. Map showing Onondaga outcrop belt and locations of field trip stops. A=Albany; B=Buffalo; H=Hagersville.

of western New York, stratigraphic equivalents on the Niagara Peninsula of Ontario are commonly referred to by New York terminology rather than terminology of central-southwestern Ontario and Michigan. Figure 2 shows approximate relationships between the stratigraphy in southwestern Ontario, New York and Pennsylvania.

Onondaga County, New York, is designated as the type locality for the Onondaga Limestone. The unit consists of four members (in ascending order): Edgecliff, Nedrow, Moorehouse, and Seneca. In western New York, a fifth member, the Clarence, replaces the Nedrow or lies between it and the Edgecliff. In western New York and Ontario, the Onondaga may be underlain by Middle Devonian Bois Blanc Limestone, Lower Devonian Oriskany Sandstone, or Upper Silurian Bertie Dolomite. The Silurian-Devonian unconformity is characterized by a broadly undulatory surface displaying local relief up to 1 m and development of joints and fractures which do not extend into overlying beds (Kobluk and others, 1977). Infilling of some of these solution-widened joints by "Oriskany-type orthoquartzitic sand and conglomerate" indicates post-Oriskany exposure and erosion (Kobluk and others, 1977, p. 1157). In many places, a thin bed of dark green or gray glauconitic, calcareous shale separates the light olive gray, finely crystalline Bertie Dolomite from overlying Bois Blanc or Edgecliff beds.

The Blois Blanc Formation consists of basal Springvale Sandstone and Bois Blanc Limestone. In outcrop, it ranges in thickness from zero at Buffalo to nearly 35 ft at Hagersville. Oliver (1966) discussed the multiple unconformity at the base of the Devonian in New York and on the Niagara Peninsula. In addition to faunal evidence, he described six sequences in which sand may be present or absent at the base of the Onondaga or Bois Blanc formations. The sand is believed to be reworked basal Devonian sandstone and has been designated "Springvale Sandstone," extending the original Ontario usage (Stauffer, 1913) into New york. Because the name applies to two sandy horizons of different ages, this terminology is confusing. In this paper, the original terminology of Stauffer (1913) is adhered to and the term Springvale is applied only to the lower sandstone member of the Bois Blanc Formation.

The Springvale Sandstone and the Bois Blanc Limestone characteristically contain sufficient glauconite to give both units a greenish color. The Springvale also commonly contains phosphate nodules and clasts of Silurian dolomite. The Bois Blanc Limestone is typically medium dark gray, medium-bedded, abundantly cherty skeletal wackestone in which fossils are concentrated in layers. Brachiopods are the dominant constituent, however bryozoans, trilobites and corals are also common. Wispy laminae are characteristic in the Port Colborne area, but at Hagersville, bioturbation has homogenized the clay content of the carbonate. Quartz sand-filled burrows occur in the upper portions of

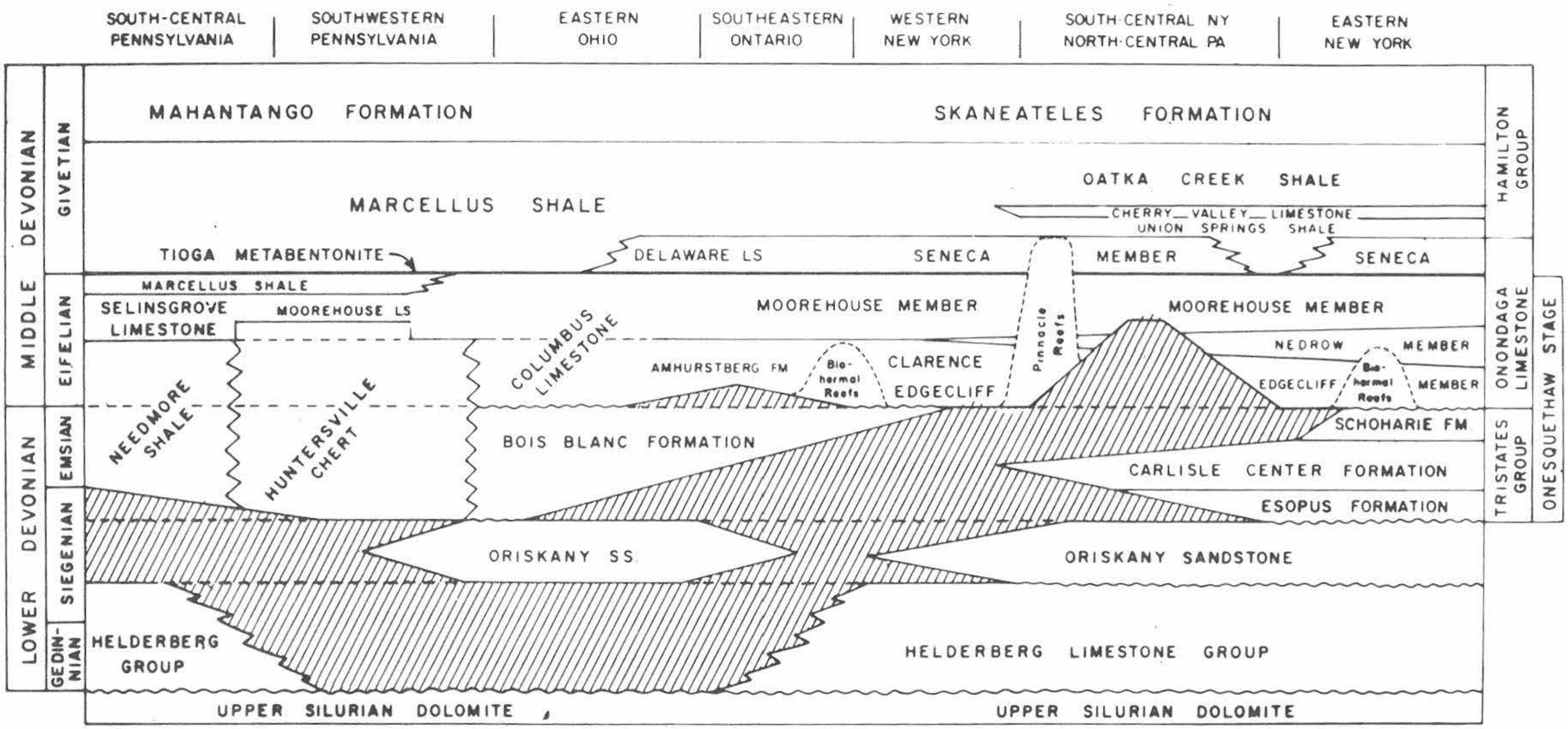

FIGURE 2. Middle and Lower Devonian stratigraphy in the northern Appalachian basin.

some beds. Chert is common in the Bois Blanc in the form of nodules, silicified fossils, and siliceous lime mudstone. Several beds of chert are found at Hagersville. At Port Colborne a bed of dolomitic, glauconitic chert breccia, in which angular chert clasts float in a matrix of siliceous lime mudstone, occurs at the base of the Bois Blanc Limestone.

In many places, a Bois Blanc-Onondaga disconformity can be readily observed; in other places, it is not easily detected. Oliver (1960) recognized two distinct coral assemblages at the base of the Onondaga in western New York. He found the lower (Amphigenia) zone to be lithologically and paleontologically distinct from the Edgecliff Member and suggested it represents "the eastern feather edge of the Bois Blanc Formation" (p. B173), separated from the upper assemblage by a disconformity, which does not represent subaerial exposure.

In the type area the Onondaga is approximately 70 ft thick, and its members exhibit greatest differentiation. It thickens westward and eastward in outcrop from the Syracuse area, exhibiting lateral changes in fauna and lithology, which require alternative criteria for differentiating members and the introduction of an additional member in the west.

Along the outcrop belt of New York and the Niagara Peninsula, the Edgecliff Member is a light gray, coarsely crystalline, massive-bedded limestone, dominantly a crinoid packstone or wackestone containing numerous rugose and tabulate corals. Biohermal facies occur locally. Oliver (1954, 1966, 1976) described many of the bioherms of western New York and Ontario in which colonial rugose corals are profuse. The Edgecliff is thinner in the Buffalo region (5 ft or less) than in the type area, but thickens westward to 12 ft at Hagersville. Light gray, nodular chert is common in the Edgecliff, especially in the upper portion.

Overlying the Edgecliff in the type area is the Nedrow Member; an argillaceous, cherty limestone containing a sparse fauna characterized by platycerid gastropods. In the east, the relative increase of terrigenous sediments, represented by the sharp lower contact of the Nedrow, is interpreted as a time plane (Oliver, 1976). To the west, faunal and lithologic characteristics of the unit overlying "typical" Edgecliff strata change sufficiently to warrant designation (Ozol, 1964) of the Clarence Member. The exact relationships between the Clarence, Nedrow and Edgecliff Members are not clear. Oliver (1966) suggested that in western New York and adjacent Ontario "the Clarence is roughly equivalent to the Nedrow Member, but may include some of the lower Moorehouse Member and uppermost Edgecliff as well" (p. 40). According to Ozol (1964), in western New York the Nedrow Member "does not overlie the Edgecliff, but is present higher in the section" (p. 4145). This relationship is diagrammed by Rickard (1964). The Clarence is the least fossiliferous member of the Onondaga Limestone in western New York and Ontario. It is an olive gray, argillaceous mudstone containing a sparse fauna, including corals, brachiopods, and bryozoans. Dark chert nodules are extremely abundant.

The Nedrow (and Clarence ?) grades upward into the Moorehouse Member, a fine-grained, massive-bedded limestone containing diverse fauna and representing a return to more Edgecliff-like conditions. In many places it resembles the Edgecliff in lithology and fauna, but lacks abundant corals and biohermal facies. It consists of interbedded, massive, olive gray and light olive gray crinoidal wackestone, packstone, and grainstone with sparse tabulate and rugose corals and abundant nodular chert.

Above the Moorehouse is a thin clay bed, the Tioga Bentonite, known throughout the Appalachian basin. It forms a prominent break in outcrop, separating the Seneca and Moorehouse Members which are similar in lithology and fauna, although the Seneca is darker and somewhat more argillaceous. In western New York and Ontario, the upper Onondaga is rarely exposed in outcrop, but subsurface data show that it grades upward and southward into dark shales of the Marcellus Formation.

Chert is ubiquitous throughout the Onondaga. In fine-grained facies it occurs most commonly as lumpy, ellipsoidal nodules, their long axes parallel to bedding. In coarser grained facies (for example, in the Edgecliff), nodules are of highly variable shape and dimensions and are sometimes anastomosing. Many chert nodules contain preserved carbonate skeletal material. In addition, secondary carbonate phases also occur in the form of ferroan and nonferroan forms of dolomite and calcite (Pfirman and Seleck, 1977). Biogenic origin for the chert is indicated by the presence of preserved radiolaria tests and sponge spicules within chert nodules. The high concentration of dissolved silica which resulted in the large volume of Onondaga chert is believed to be related to the advanced erosion of the land area to the east, providing a large proportion of dissolved silica relative to terrigenous detritus (Pfirman and Seleck, 1977).

TECTONIC SETTING

Tectonic elements which influenced Devonian sedimentation in the eastern Great Lakes region were dominated by the northeast-trending Appalachian basin, a center of subsidence which accumulated thick sedimentary sequences. Positive, more stable areas (shallow marine areas when not exposed) were the Cincinatti-Algonquin arch system to the west and the Adirondack uplift in northeastern New York (Figure 3). During deposition water depths increased eastward and southward off the platform into the basin.

Onondaga sedimentation began in eastern New York, following Schoharie deposition (Lindholm, 1969). Glauconitic horizons and phosphate nodules found locally above the Schoharie and Carlisle Center formations suggest a depositional hiatus. Post-Bois Blanc sedimentation probably followed a hiatus as well, except on the Niagara Peninsula, where Oliver (1976) suggested that the greenish, argillaceous beds at Port Colborne might fill the time gap between Bois Blanc and Onondaga deposition.

Linear facies belts (Figure 4) indicate that Onondaga sedimentation took place on a carbonate ramp; an inclined platform extending basinward without a pronounced break in slope (Ahr, 1973). Characteristics of this model include concentric arrangement of facies belts about the basin axis and dominance of finer grained facies basinward of coarser, less muddy facies. This model differs from carbonate platform models (for example, Purdy, 1963; Enos, 1977) in the distribution of progressively lower energy facies basinward and the absence of abrupt high-energy basin-margin shoals. In the ramp model the highest energy zone is close to shore, in the platform model, it lies on the shelf margin.

The ramp model is also characterized by the occurrence of patch reefs or bioherms rather than continuous reef trends. Isolated large bioherms developed in south-central New York and north-central Pennsylvania, and the bioherms of the Buffalo-Port Colborne area developed landward in somewhat shallower water. Location of these high-energy areas is probably influenced by subtle irregularities on the seafloor, reflecting relict topography or zones of tectonic adjustment. Currents deflected over and around such irregularities provided necessary circulation and food source for carbonate-producing biota.

ONONDAGA BIOHERMS

Small bioherms are common in the Edgecliff and have been the focus of numerous studies. Grabau (1903) described a bioherm exposed near Williamsville, New York which has since been largely removed by freeway construction. In the early 1950's over 20 bioherms were found in east-central New York (Oliver, 1956a), and in recent years several of these buildups have been studied individually (Mecarini, 1964; Bamford,

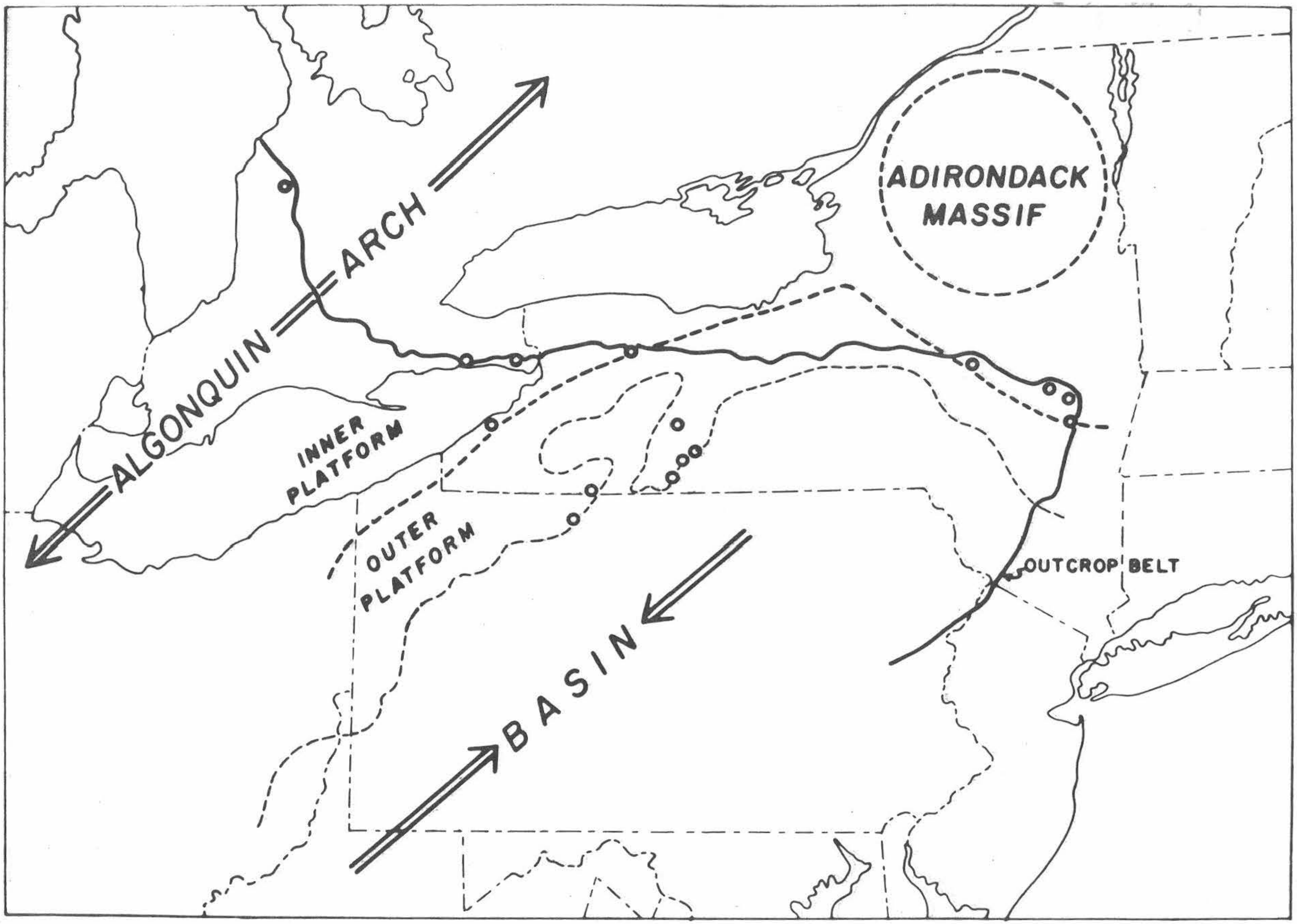

FIGURE 3. Map of tectonic elements which influenced Devonian sedimentation in the northern Appalachian basin, showing position of the platform upon which Onondaga bioherms developed.

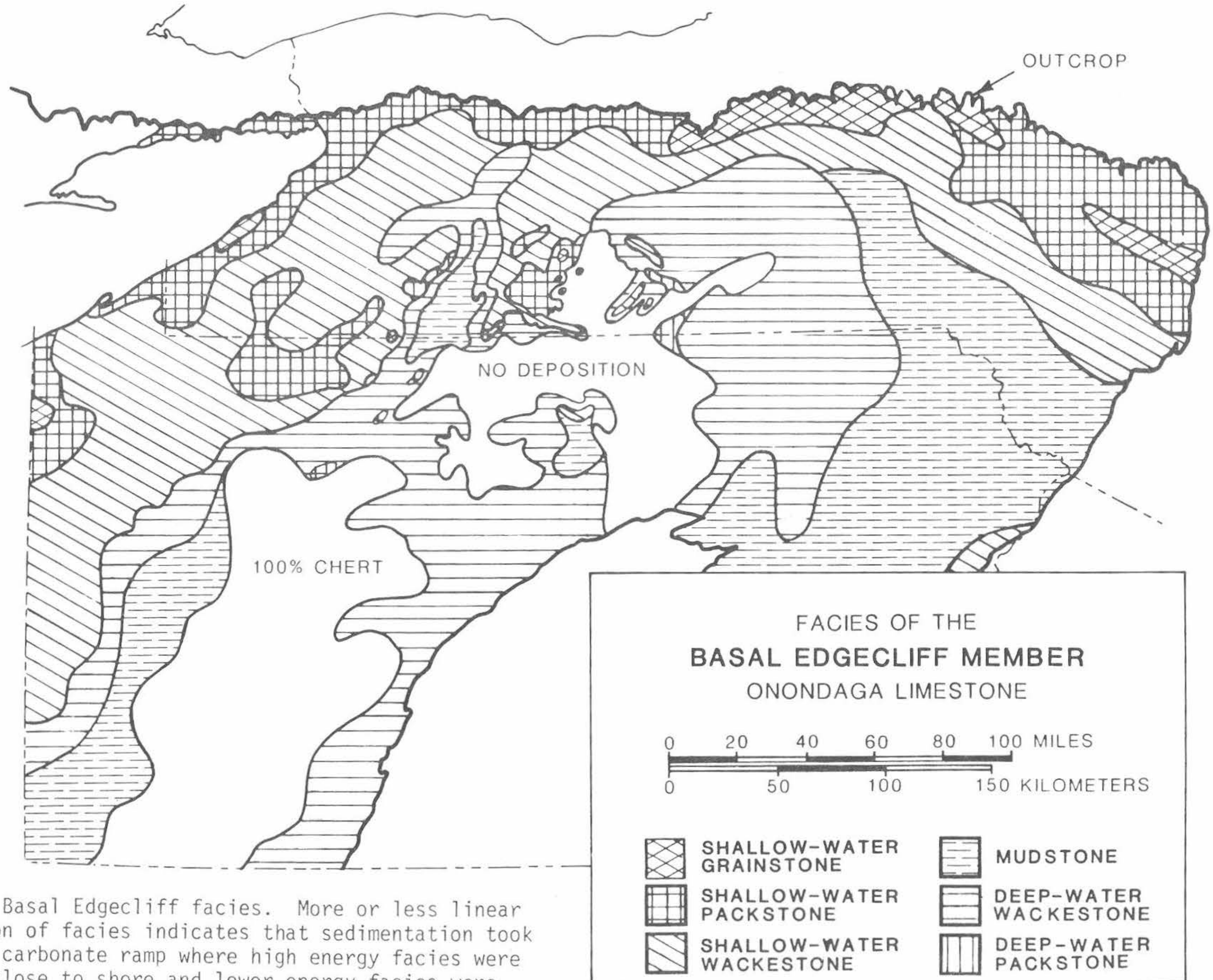

FIGURE 4. Basal Edgecliff facies. More or less linear distribution of facies indicates that sedimentation took place on a carbonate ramp where high energy facies were deposited close to shore and lower energy facies were deposited basinward.

1966; Williams, 1977, 1979; Polasek, 1978). In addition to the eastern buildups, several buildups are known from western New York (Poore, 1969; Crowley and Poore, 1974; Oliver, 1976; Coughlin, 1981), from the Niagara Peninsula of Ontario (Telford and Tarrant, 1975a,b; Oliver, 1976; Cassa, 1979) and from southwestern Ontario (Fagerstrom, 1961; Mayo, 1965). The buildups are stratified and lack a rigid framework. In many cases only slightly dipping marginal beds and prolific coral growth give any clue to the presence of a buildup. In some quarries, massive core rock remains undisturbed because of the difficulty in quarrying rock without well developed bedding and joints. Reconstructed dimensions range up to several hundred feet in diameter and 80 ft in thickness.

In contrast to outcrop bioherms, subsurface "pinnacle reefs" of south-central New York and north-central Pennsylvania (Mesollela and Weaver, 1975; Kissling and Coughlin, 1979; Coughlin, 1981; Kissling, 1980, 1981; Kissling and Moshier, 1981; Kissling and Polasek, 1982) cover several hundred acres and comprise over 200 ft of Onondaga sedimentation. Like their surface counterparts, these bioherms contain an abundant coral fauna. Gamma ray-neutron logs record low radioactivity and extremely high porosity.

Subsurface Facies

The subsurface Onondaga is dominated by moderately deep-water facies which grade southward into the Needmore and Marcellus shales and Huntersville Chert. Facies represent relative bathymetry. Presumably deep-water calcisiltite mudstone, stylioline wackestone, and thin-shelled brachiopod packstone facies delineate the epicratonic Appalachian basin in south-central New York and most of Pennsylvania. Wackestone, packstone, and grainstone facies, dominated by crinoids, bryozoans, and robust brachiopods outline an arcuate platform that bounded the basin on its northern and western sides. Ooliths, peloids, carbonate intraclasts, coated grains, and calcified or stromatolic algae are completely absent, indicating that if shallow, nearshore environments had existed, they once lay north of the present outcrop belt, perhaps marginal to the Algonquin arch and Adirondack massif.

During deposition, water depths increased progressively throughout the basin, as reflected by northward shift of facies comprising successive members, and by northward migration with time of the Marcellus Shale. Basin and platform were joined in south-central New York and northwestern Pennsylvania by a south-sloping ramp dissected by troughs and surmounted by isolated banks. All seven gas-bearing Onondaga bioherms discovered to date were initiated as Edgecliff coral-crinoid mounds on the seaward or southeastern margin of these banks.

Subsurface Stratigraphy

The Onondaga ranges from 8 to 215 ft thick and is readily divisible into the Edgecliff, Nedrow, Moorehouse and Seneca Members. Of these, only the upper part of the Moorehouse Member is present throughout the basin. Other units are absent in places as a result of nondeposition or submarine scour or because of lateral gradation with the Marcellus Shale.

The upper contact of the Onondaga is characterized by an abrupt decrease in radioactivity in gamma-ray profiles (Figure 5). The Tioga Bentonite, separating the Seneca and Moorehouse members, is typically moderately radioactive. Because as many as three bentonites occur in the upper part of the Onondaga, in areas where the Seneca is very thin or absent the others could be misidentified as the Tioga. The Moorehouse displays relatively low radioactivity on gamma ray logs, except for the bentonites. The upper Moorehouse is commonly somewhat more porous than the lower Moorehouse.

The argillaceous Nedrow Member can be identified by a slightly shaly kick in the gamma-ray profile and corresponding moderate neutron count. However, this relationship does not always occur, nor does the Nedrow always appear as shaly as expected.

Both the Clarence and the Edgecliff Members are uniformly low in radioactivity, displaying little character in gamma-ray profiles, and are relatively low in porosity. The contact between the two members appears to be transitional and cannot be accurately placed in many cases. Extrapolation from outcrop has resulted in an estimate of where the contact should be and is the basis on which Edgecliff and Clarence lithofacies have been mapped.

The Bois Blanc generally contains one or more shaly layers and is of high porosity, making it fairly easy to pick with consistency. The basal Devonian unconformity almost always stands out as a thin shaly kick. The Oriskany Sandstone and Bass Islands Group are characteristically of low radioactivity and high porosity.

Subsurface Bioherms

Subsurface Onondaga buildups (Figure 6), popularly known as pinnacle reefs, are similar in fauna and facies to bioherms known in the outcrop belt; however, in contrast to outcrop bioherms, they are separated paleogeographically, are generally far larger (118 to 207 ft in thickness and 3,900 and 10,500 ft in diameter), and continued their growth throughout Onondaga deposition. Like their surface counterparts,

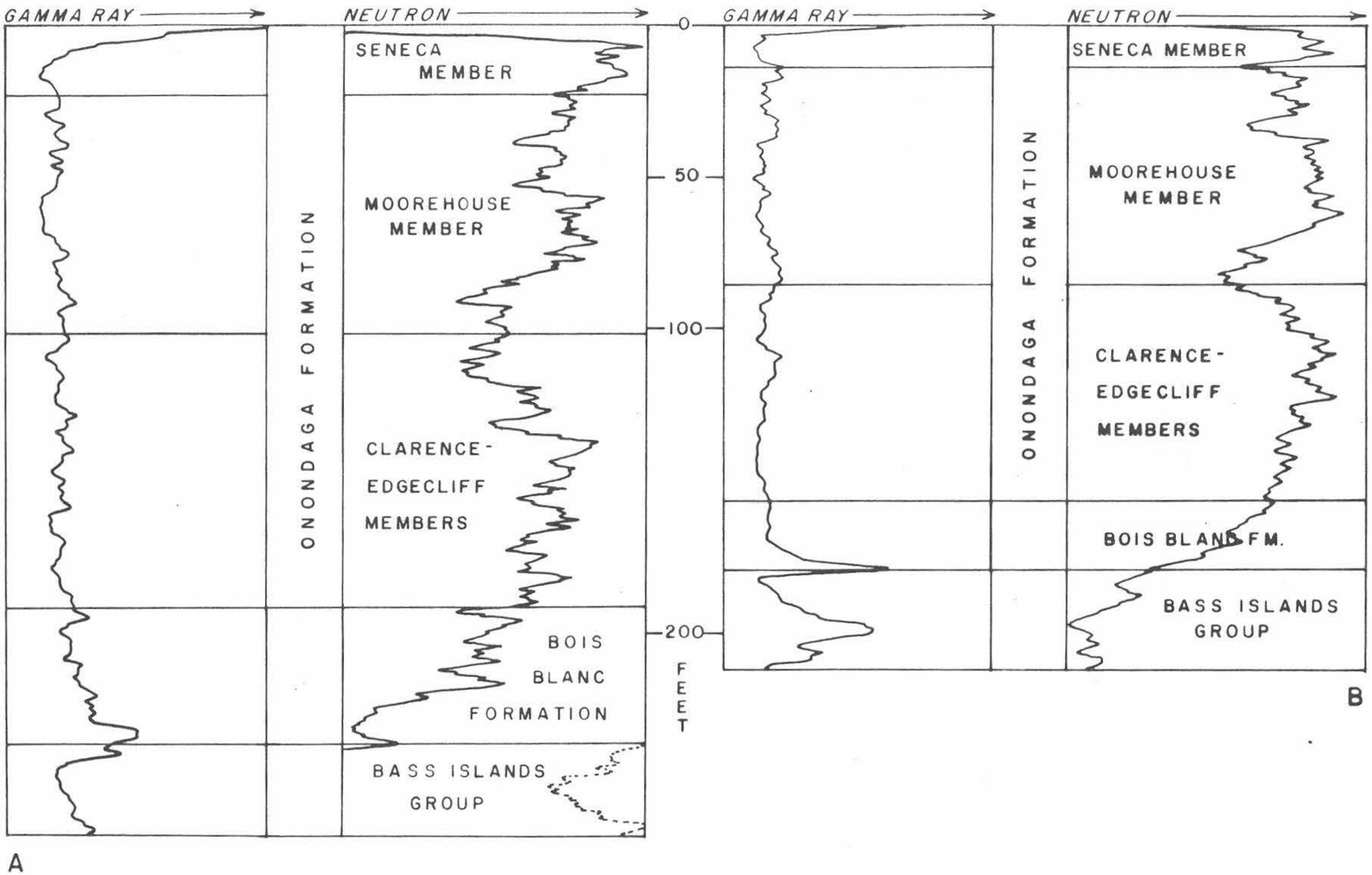

FIGURE 5. Gamma ray-neutron logs of the Onondaga Limestone. A. Apache Corp.-Hica Corp. H. Carnahan No. 1, Chautauqua County, New York. B. New York State Natural Gas Co. Jones No. 1, Erie County, New York.

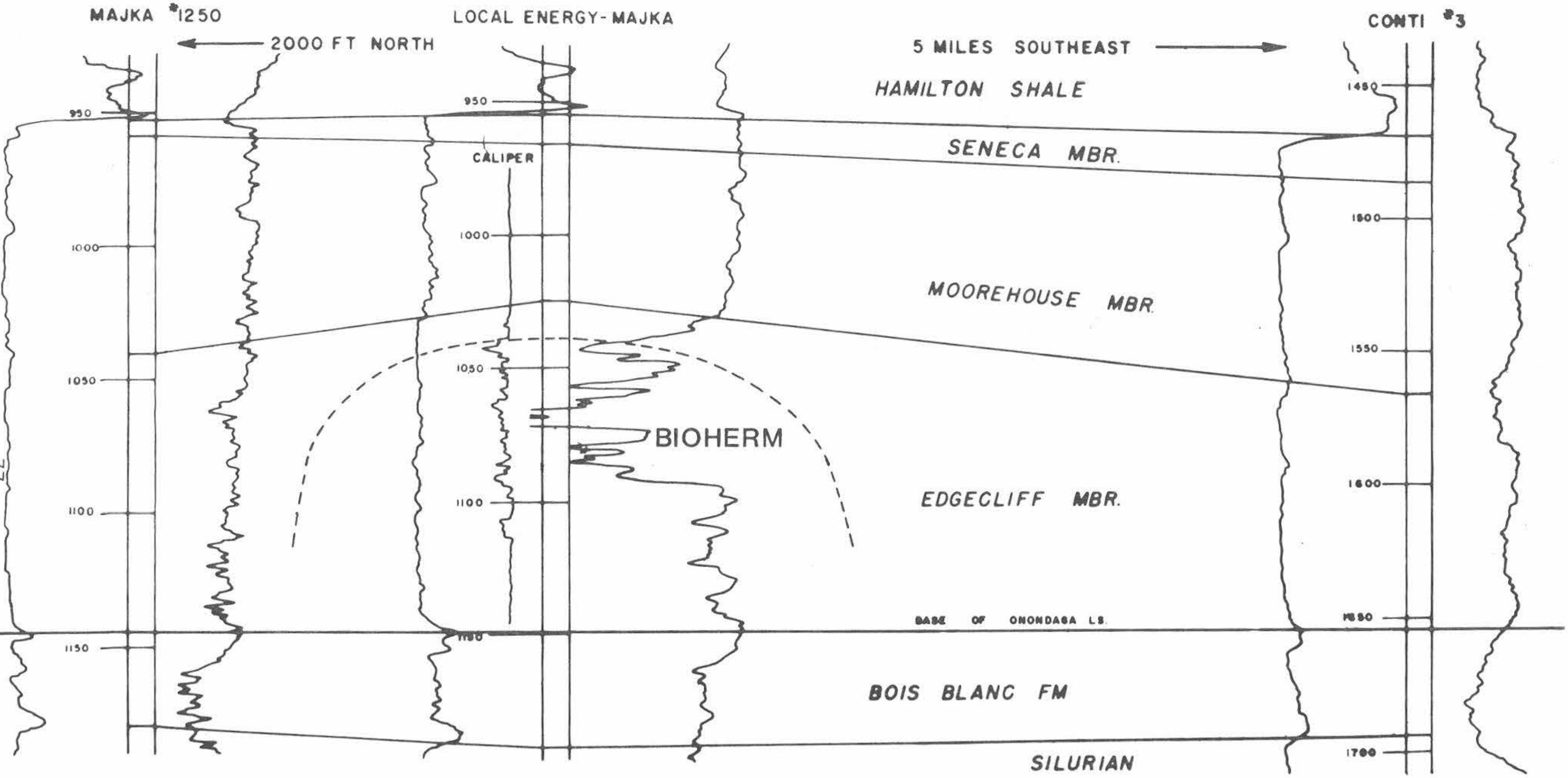

FIGURE 6. Gamma ray-neutron logs showing biohermal (center) and nonbiohermal characteristics of the Onondaga Limestone in Erie County, New York.

subsurface bioherms exhibit broadly domed structures that consist of coral bafflestone in crinoid-coral fragment packstone matrix. Initial Acinophyllum-Cladopora coral thickets were succeeded by alternating thickets of Cylindrophyllum and Acinophyllum-Cladopora, and were capped by bryzoan-Cladopora wackestone. Proximal bioherm flanks consist of branching coral mudstone in crinoid packstone and grainstone; distal flanks are coral floatstone rich in favositid tabulates. Although surrounded by deep-water Moorehouse and Seneca facies upon subsidence of the supporting banks, these bioherms continued vertical and lateral growth until terminated by anoxic waters accompanying northward-encroaching euxinic Marcellus deposition. Neither outcrop nor subsurface bioherms were wave-resistant reefs. They formed at considerable depths, probably below effective photic zone. Calcified algae and stromatoporoids, primary components of Devonian reefs elsewhere, are virtually absent, despite the equatorial paleolatitude of the northern Appalachian Basin. Extinct buildups remained as submarine knolls for millenia until onlapped by the Marcellus and Skaneateles Shales.

Most existing porosity consists of primary intraskeletal, intergranular, and growth-framework voids. Effective pore-reducing processes, which characterized the early diagenetic history, included internal sedimentation, aragonite-to-calcite inversion, pressure-solution, and associated calcite cementation and neomorphism. Subsequent development of fractures and microfractures furnished permeability needed for migration of liquid hydrocarbons into the reservoirs, and led to significant hydrocarbon-induced leaching. Late diagenetic events such as stylolitization, pore plugging by bitumen residue, and silica replacement further reduced bioherm porosity to the present average of 4 to 5%. Primary intercoralline and intraskeletal voids are the most common pore types observed from core slabs and thin sections. Many intercoralline voids apparently represent original sheltered cavities. Solution-enlarged fractures constitute the major pore-communication system in Onondaga reservoirs. To date, six gas-producing and one shut-in Onondaga bioherm fields are known from New York and Pennsylvania. As of January, 1981, nine wells have produced more than 18 bcfg.

COMPARISON WITH OTHER DEVONIAN BUILDUPS

Devonian bioherms have nearly worldwide distribution and, as a group, are perhaps best known among Paleozoic carbonate buildups. Despite their widespread occurrence they are remarkably similar in many respects, including facies types, faunal components and development.

The most striking difference between Onondaga bioherms and Devonian bioherms elsewhere is the role of stromatoporoids as frame builders and constructors. Generally, the stromatoporoid-dominated community so characteristic of most Devonian bioherms represents a highly specialized community which may have been particularly sensitive to changes in the environment. Stromatoporoids, however, are not a appreciable component of Onondaga fauna. They are virtually absent in bioherms of eastern New York, but thin, laminar forms become relatively common in biohermal and non-biohermal facies of western New York and the Niagara Peninsula. Formosa bioherm, 90 mi northwest of Hagersville, is composed primarily of tabular stromatoporoids and rugose corals, which suggests it developed under more turbulent and perhaps somewhat shallower conditions than had existed to the east.

A second important difference between Onondaga bioherms and most other Devonian buildups is the conspicuous absence of calcified algae. In most Devonian reefs algae play an important role as binders and encrusters, as well as contributing clay-sized sediment in the manner of modern codiaceans. However, depositional features of Onondaga sediments suggest that sedimentation took place in water sufficiently deep to inhibit the growth of green algae.

Thus, despite many similar characteristics, Onondaga bioherms do not seem to fit the mold of other Devonian buildups. Upper Middle Frasnian bioherms of the Dinant basin (Lecompte, 1959) exhibit stratified, nonrigid structures similar to those of Onondaga buildups and lack signs of subaerial exposure. Flank beds of reef-derived talus and indications of wind and current influence such as elongation or asymmetry are also lacking. Like many Onondaga bioherms, the initial phase of bioherm development is represented by impure, argillaceous limestone. However, instead of fasciculate corals characteristic of Onondaga bioherms, lamellar forms such as Alveolites dominate initial assemblages in Dinant basin bioherms. Stromatoporoid reefs were superimposed on the small, stratified Dinant basin mounds when they grew into the zone of turbulence, illustrating an environmental potential different from that of the lamellar corals.

Embry and Klovan (1972) analyzed facies relationships of an Upper Devonian buildup in the Canadian Arctic and established absolute water depths based on paleoecological zones. They determined that locally prolific coral growth in relatively deep, quiet water resulted in a small biogenic bank. As the bank grew up toward wave base, tabular stromatoporoids became the dominant fauna. In many cases, this pattern resulted in development of extensive reef tracts analogous to modern examples such as the Great Barrier Reef of Australia and reefs of the Yucatan and Belize shelves (Klovan, 1974). The latter stages of this pattern, however, are not represented in Onondaga bioherms.

MODERN ANALOGUES

No direct analogues for Onondaga buildups can be found today, however, several partial analogues may be of help in establishing parameters for the Onondaga depositional environments. Cold- and deep-water coral banks and patches composed primarily of ahermatypic corals together with a diverse invertebrate assemblage exist in water 600 to 900 ft deep off the coast of Norway and elsewhere in the Atlantic Ocean (Teichert, 1958). They show no evidence of erosion, however, flanking debris may be composed of organically fragmented skeletal material. Because the environment of these banks is generally deep, calcareous algae are absent.

Several submerged reef banks are found on the Campeche shelf off Mexico, a modern carbonate ramp (Logan and others, 1969). They are oval to round in outline and show little indication of influence by action of wind or waves, although situated in only 30 ft of water. They do, however, possess flanks composed of algal nodules or encrusting coralline algae.

Boo Bee patch reef, on the Belize shelf (Halley and others, 1977) is one of many flat-topped lagoonal mounds founded on relict Pleistocene topographic highs. Unlike Onondaga buildups, these mounds developed behind a protective barrier reef. Resemblance to Onondaga bioherms is found in the occurrence of lime silt and mud at the base and coralline lime sand and rubble which comprise the major portion of the mound. Like Onondaga buildups, Boo Bee patch reef lacks internal structure and lithologic or faunal zonation. It represents a local proliferation of coral growth which was protected from wave destruction and kept pace with rising sea level. Relatively sudden deepening of the water could result in cessation of coral growth by relative increase in fine-grained terrigenous and carbonate sedimentation associated with decreased rate of carbonate production. Such sediments would closely resemble the Clarence Member which overlies Onodaga bioherms.

Rodriguez Bank (Turmel and Swanson, 1976) on the Florida reef tract near Key Largo is a Holocene mudbank having sedimentological characteristics similar to those of Onondaga buildups. Absence of rigid coral-algal framework and abundance of lime mud are indicative of calm-water development. Mud, which probably was produced in place by disintegration of green algae and other skeletal material, was trapped and stabilized by marine grasses and small branching corals, perhaps similar to the initial Acinophyllum beds of Onondaga buildups. With sea level rise and wider inundation, the quiet-water environment of early bank development changed to more open-water circulation favorable for skeletal sand and gravel production, indicated by vertical increase in average grain size. Packstone and grainstone beds in the Onondaga buildups may be similarly interpreted.

ACKNOWLEDGEMENTS

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ROAD LOG AND STOP DESCRIPTIONS FOR CARBONATE FACIES OF THE BOIS BLANC AND ONONDAGA FORMATIONS, NIAGARA PENINSULA, ONTARIO

[Routes and locations of stops are shown in Figures 7 and 8.] Passports are not required of U.S. citizens to enter Canada or return to the United States. However, proof of citizenship must be carried - birth or baptismal certificate, voter registration card. Naturalized citizens should carry naturalization papers. U.S. resident aliens must have an Alien Registration Receipt Card. Citizens of countries other than the U.S. or Canada must have a valid passport and visa.

Road log starts from from Marriott inn.

Cumulative Mileage	Miles from last point	Route and Stop Description
0.0	0.0	Leave Marriott Inn and turn right onto Millersport Highway (NY 263).
0.2	0.2	Pass under railroad tracks and bear right onto Interstate 290 west.
6.7	6.5	Keep left and continue south on Interstate 190.
8.8	2.1	Tonowanda Channel of Niagara River on right. Grand Island is visible across the river, behind us. At the south end of Grand Island, the Tonowanda and Chippewa Channels join and Fort Erie, Ontario can be seen across the river.
11.2	2.4	Buffalo District, U.S. Army Corps of Engineers on right.
12.0	0.8	Schaefer Brewery on left.
12.2	0.2	Toll booth.
12.4	0.2	Lock in Black Rock Canal on right.
13.0	0.6	Pass under Peace Bridge.
13.1	0.1	Bear right onto Porter Avenue exit ramp.

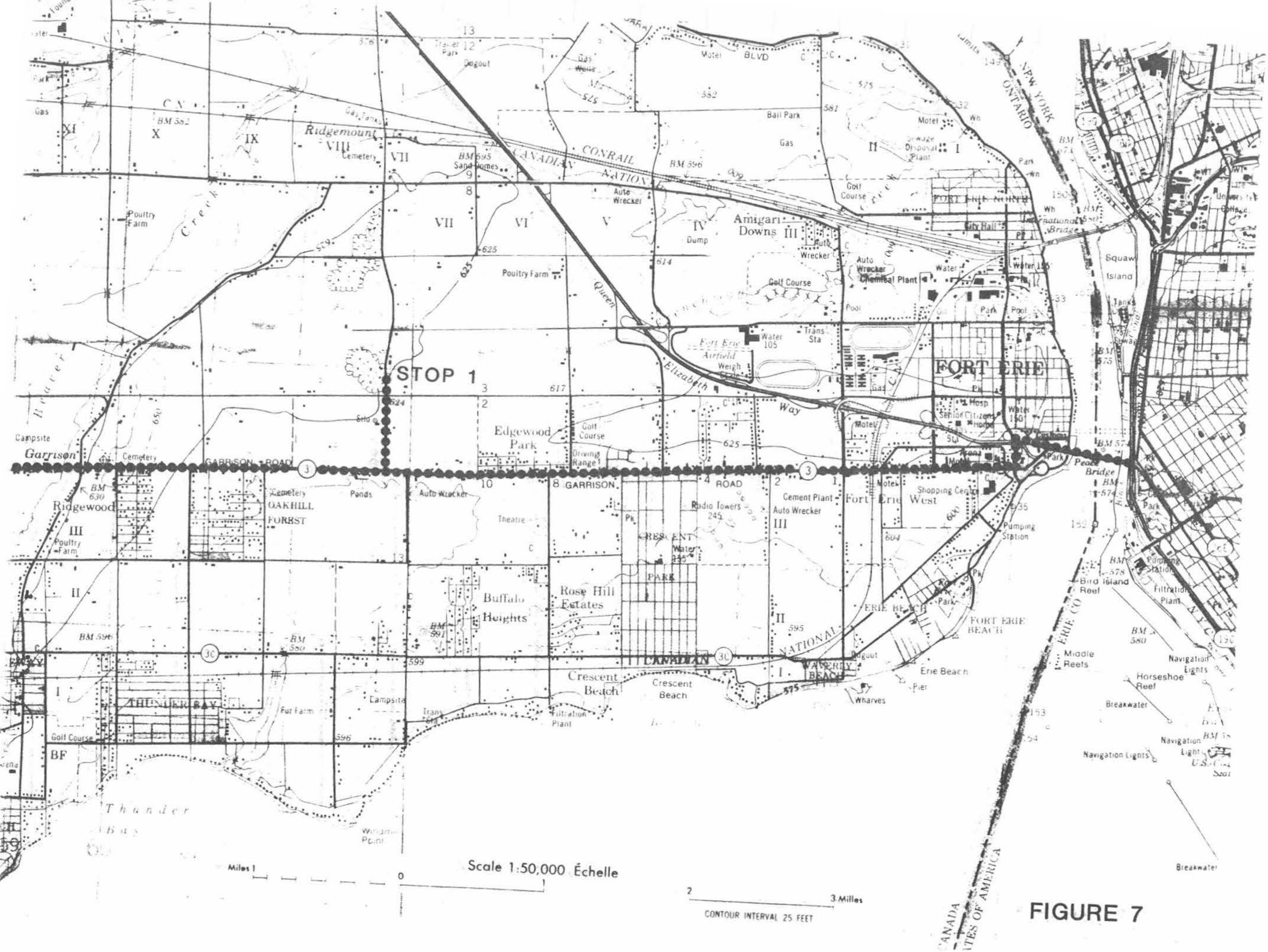

STOP 1
FORT ERIE
Fort Erie West
GARRISON ROAD
Garrison
Ridgewood
Ridgemount
OAKHILL FOREST
Edgewood Park
Buffalo Heights
Rose Hill Estates
CRESCENT PARK
Crescent Beach
THUNDER BAY
Thunder Bay
Amigari Downs
Fort Erie Airfield
Queen Elizabeth Way
CONRAIL
CANADIAN NATIONAL
FORT ERIE NORTH
Peace Bridge
Squaw Island
NEW YORK
ONTARIO
FORT ERIE BEACH
ERIE BEACH
WAVERLY BEACH
Erie Beach
Bird Island Reef
Middle Reefs
Horseshoe Reef
Beaver Creek
CANADA
UNITED STATES OF AMERICA
Scale 1:50,000 Échelle
Miles 1
0
2
3 Miles
CONTOUR INTERVAL 25 FEET

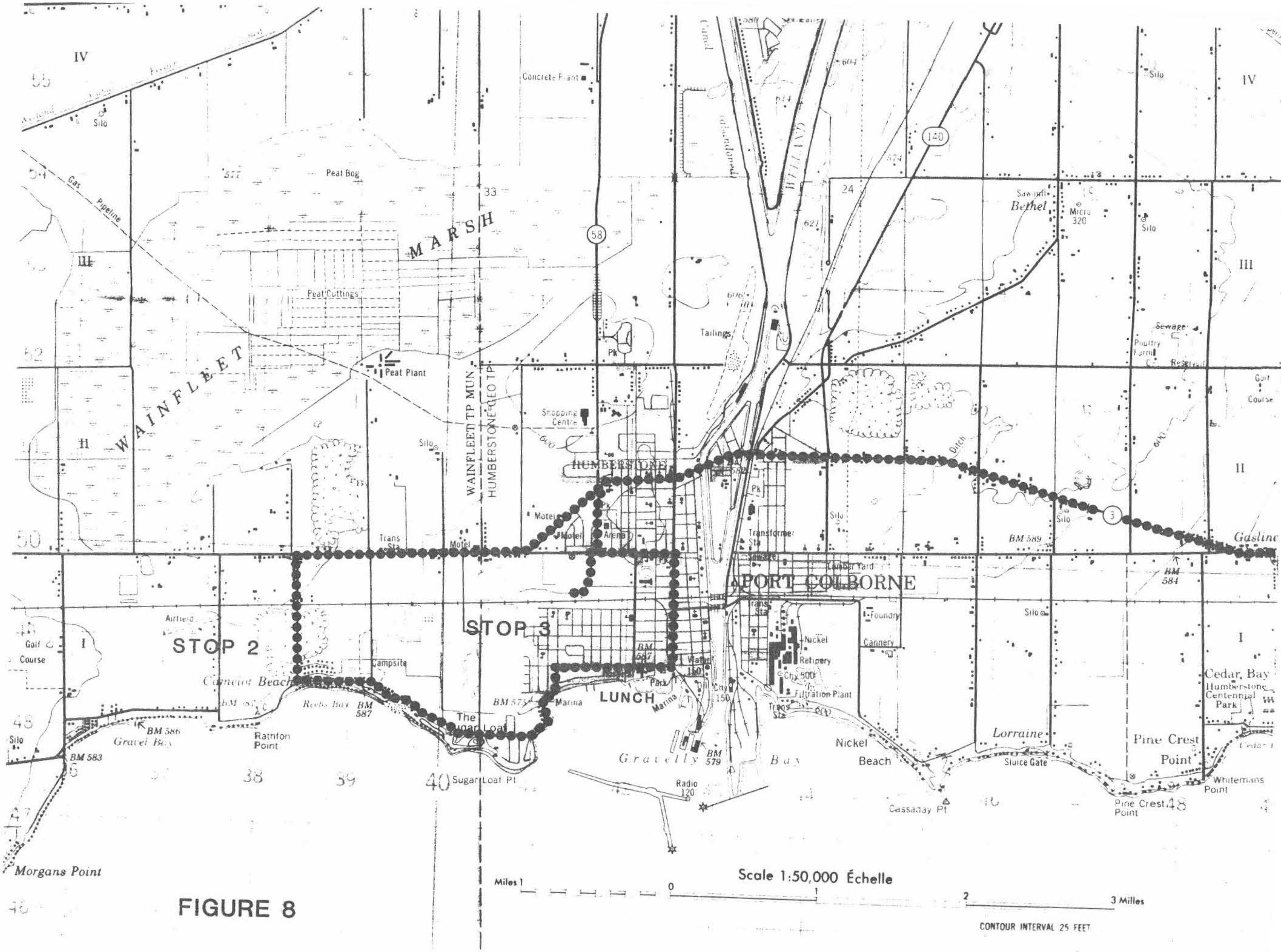
FIGURE 7

13.6	0.5	Stop. Turn left onto Porter Avenue.
13.8	0.2	Turn left at traffic light, following signs to Peace Bridge.
14.1	0.3	Toll booth - Peace Bridge.
15.0	0.9	Customs.
15.1	0.1	Bear right at exit to Ontario Highway 3, toward Windsor and Crystal Beach.
15.5	0.4	Turn right at traffic light onto Ontario Highway 3 west (Garrison Road).
19.6	4.1	Stonemill Road on left.
19.8	0.2	Turn right onto Ridgemount Road (Regional Road 120).
20.5	0.7	Turn left into driveway of Ridgemount Quarries, Ltd.

STOP 1. RIDGEMOUNT QUARRY

Hard hats must be worn. Please exercise caution near quarry walls; watch for overhangs and loose rock.

An excellent exposure of the Silurian-Devonian unconformity is found in the north wall of this quarry. Several small bioherms are exposed in the northern part of the quarry and in an "embayment" a few hundred feet west of the entrance. The mounds are characterized by abrupt thickening of lower Edgecliff strata which contain abundant solitary and colonial rugose corals, tabulate corals, and crinoid material (Figure 9). Dominant coral genera include Acinophyllum, Cladopora, Heliophyllum, Cystiphylloides, Syringopora, and Emmonsia. Internal structure is stratified; facies are represented as successive layers which pinch out laterally and drape over one another to give the characteristic mounded appearance. Stratified structure and absence of rigid framework indicate that turbulence was low during deposition. Biohermal strata contrast markedly with laterally equivalent and onlapping crinoidal packstone and wackestone beds of non-biohermal Edgecliff.



WAINFLEET
MARSH
Peat Bog
Peat Cuttings
Peat Plant
WAINFLEET TP MUN
HUMBERSTONE GEO TP
HUMBERSTONE
PORT COLBORNE
STOP 2
STOP 3
LUNCH
Camelot Beach
The Sugar Loaf
Sugar Loaf Pt
Gravel Bay
Rathfon Point
Gravelly Bay
Nickel Beach
Lorraine
Pine Crest Point
Cedar Bay
Humberstone Centennial Park
Whitemans Point
Morgans Point
Cassaday Pt
Bethel
Nickel Refinery
Filtration Plant
Shopping Centre
Tailings
Concrete Plant
Gas Pipeline
Golf Course
Airfield
Campsite
Marina
Sluice Gate
Scale 1:50,000 Échelle
Miles 1
0
1
2
3 Milles
CONTOUR INTERVAL 25 FEET

FIGURE 8

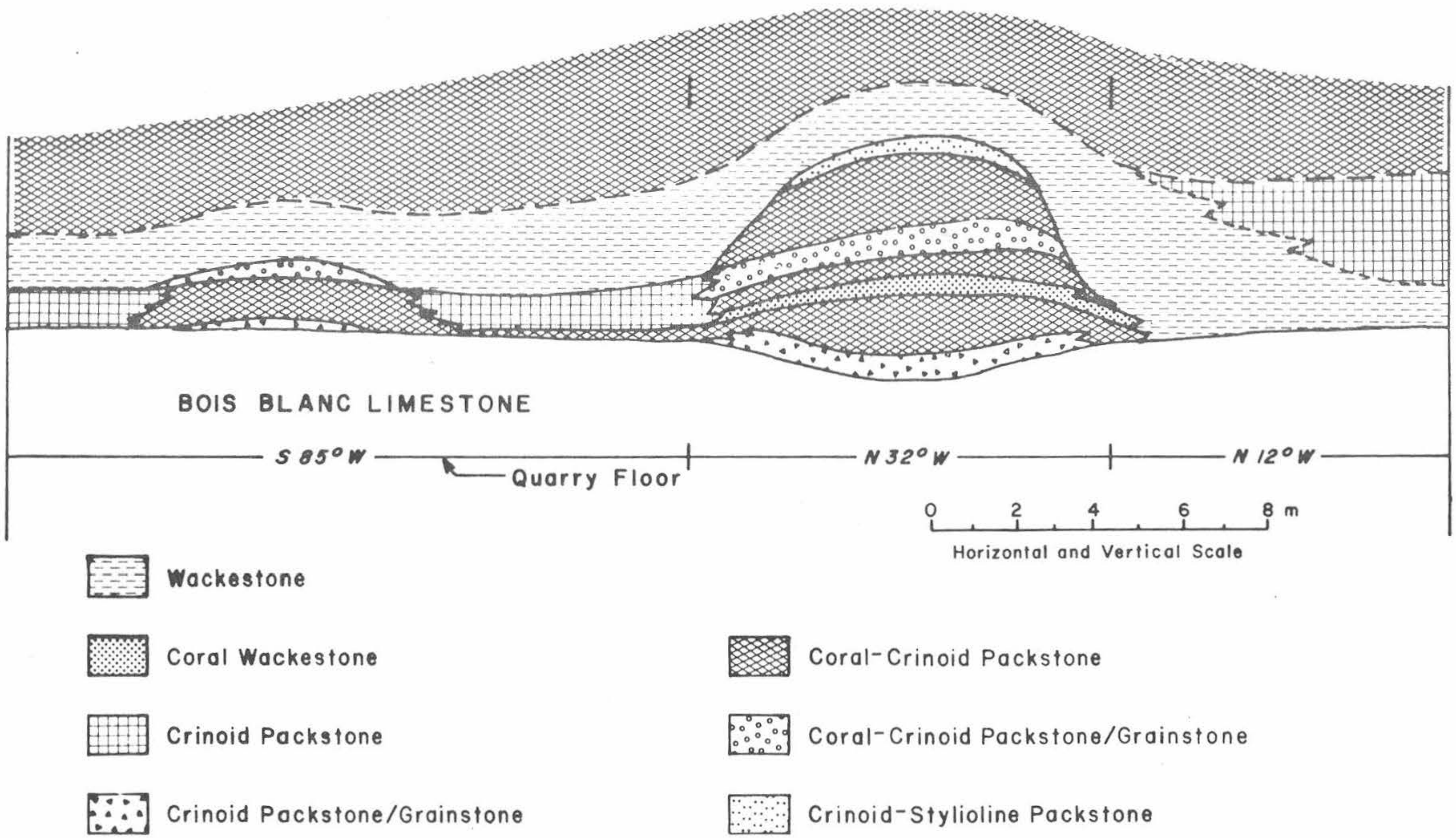


FIGURE 9. Edgecliff facies profile, Ridgemount bioherm.

Sub-biohermal Strata

Mottled olive-gray Bertie Dolomite is unconformably overlain by 0.4 m of yellowish-gray conglomeratic Springvale Sandstone, containing large, flat dolomite clasts. The Bois Blanc consists of 3.2 m of light olive-gray, cherty, fine-grained limestone, containing trilobites, bryozoans, and abundant brachiopods. The mounds are separated from the Bois Blanc Limestone by a thin, shaly zone.

Biohermal Strata

The base of the larger mound sags markedly into underlying strata. Basal beds of the two mounds are similar, consisting of 0.2 to 0.5 m of yellowish-gray, crinoid packstone. Abundant in-place *Syringopora* and other corals are found in the basal beds of the smaller mound.

The overlying facies, 1.0 to 1.5 m of crystalline coral-crinoid packstone, can be traced between mounds. It contains *Acinophyllum*, *Syringopora*, *Cladopora*, *Heliophyllum*, and *Cystiphylloides*, as well as large crinoids whose columnals measure up to 2 cm in diameter. The smaller mound is capped by coral-crinoid packstone-grainstone, containing abundant *Acinophyllum*, common *Heliophyllum*, and sparse *Cystiphylloides*.

In the larger mound, the coral-crinoid packstone is overlain by 0.5 m of coral-dominated wackestone, which, in turn, is overlain by 0.5 m of packstone containing *Acinophyllum* and branching tabulate corals, *Cladopora* and *Aulopora*. Above that, *Acinophyllum*, *Heliophyllum*, and *Cystiphylloides* are found in coral-crinoid grainstone or packstone matrix. This facies is porous, with bitumen-lined interparticle and intraskeletal pores. The mound is capped by up to 0.5 m of crinoid-stylioline wackestone-packstone, containing few corals. Rare chert nodules in biohermal beds are light-colored and highly irregular in shape. Overlying the mounds are olive-gray, argillaceous crinoid-stylioline wackestone and mudstone, containing abundant dark chert nodules and many rugose and tabulate corals.

Non-Biohermal Strata

The overthickened, richly fossiliferous strata of the bioherms grade laterally into somewhat thinner, non-biohermal beds of the lower Edgecliff which consist of medium-grained crinoid grainstone and packstone containing few corals. This facies, approximately 0.8 m thick, is overlain by 7.6 m of crinoid packstone and grainstone which contain many Heliophyllum and Acinophyllum colonies. In his description of this locality, Oliver (1976) assigned these beds to the Clarence Member, perhaps based on the presence of chert nodules or on the darker, more argillaceous nature of laterally equivalent beds which overlie the mounds. Although the exposure could represent a local, fossiliferous facies of the Clarence, Oliver's (1976) suggestion that the Edgecliff and Clarence interfinger may be borne out here in the occurrence of light gray biostromal beds characteristics of the Edgecliff in association with darker olive gray, argillaceous beds containing fewer corals.

Leave quarry and return to Ontario Highway 3.

21.2	0.7	Turn right.
23.1	1.9	Battlefield and museum on right. Plaque on museum building reads: "This house stood on the battlefield during the Fenian raid June 2, 1866. John Teal's family cared for the wounded soldier shot near the front verandah." Monument is memorial to "Queen's Own Rifles, 13th Hamilton Battalian Caledonian and York Rifle Companies of Haldimand in defense (sic) of country June 2, 1866."
27.1	4.0	Port Colborne town line.
29.8	2.7	Village of Gasline.
31.3	1.5	Humberstone International Speedway on right.
33.3	2.0	Port Colborne Quarries, Ltd. to the north. These quarries are floored in the Silurain Bertie Dolomite.
33.5	0.2	Junction, Ontario Highway 140. Robin Hood Flour mill is visible to the northwest. Port Colborne is the second largest flour milling city in Canada.
34.1	0.6	Cross Welland Canal - Lock No. 8. The Welland Canal connects Lake Ontario (St. Catherines) with Lake Erie, an elevation difference of 99.5 m (326.5 ft). It was built in the early 1800's and had to be widened, straightened, and deepened several times as ships increased

		in size and traffic volume grew. This lock, the guard lock, was the longest in the world when it was built in 1933.
34.4	0.3	Cross abandoned canal channel.
34.6	0.2	Sunbeam shoe factory on left. Sunbeam is the largest manufacturer of bowling shoes in Canada.
35.1	0.5	Junction, Ontario Highway 58, McDonalds on NW corner.
36.0	0.9	Wainfleet town line.
37.0	1.0	R.E. Law Crushed Stone Quarry on right.
37.3	0.3	Turn left onto Quarry Road.
37.6	0.3	Cross Canadian National Railroad tracks. Pull off and park immediately south of the tracks.

STOP 2. PORT COLBORNE WEST QUARRY

Hard hats are optional. Please exercise caution near quarry walls, both from below and from above.

A portion of a broad biohermal bank is particularly well exposed in the east wall of this quarry (Figure 10). Original reconnaissance and detailed study were carried out here and in the R. E. Law quarry, approximately one mile north. The Law quarry is floored in Silurian dolomite and exposes Springvale and Bois Blanc strata, including greenish, *Acinophyllum*-rich beds of the upper Bois Blanc and lower Edgecliff, which represent initial stages of bioherm development. This southern quarry is floored in the upper Bois Blanc (Oliver, 1976). Lower Edgecliff strata are characterized by a coral assemblage which includes *Cystiphylloides*, *Cylindrophyllum*, and *Heliophyllum*. The tabular "core" facies contains a variety of colonial rugose corals and replaces bioclastic beds which are more characteristic of the Edgecliff. Massive favositid tabulate corals are more abundant in the upper, less argillaceous beds.

Glacially scoured surfaces are exposed around the perimeter of the quarry, particularly at the south end. The smoothed surfaces reveal glacial striations and offer spectacular plan-view exposures of the fossil bioherm community.

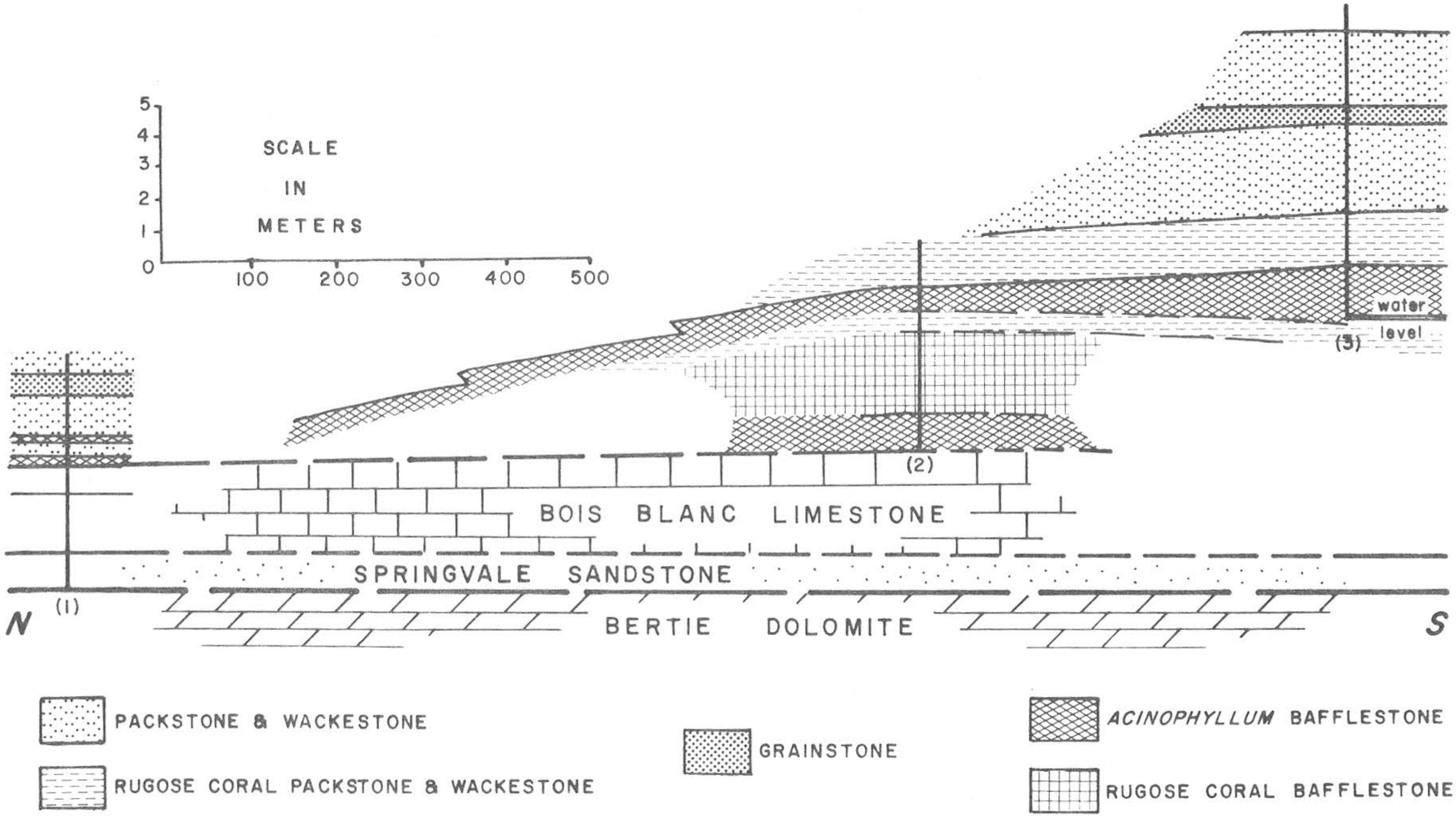


FIGURE 10. Edgecliff facies, Port Colborne bioherm. Three sections, measured in 1978, are represented: (1) R. E. Law quarry; (2) "Port Colborne west" quarry, north face; (3) "Port Colborne west" quarry, southeast corner.

Sub-Biohermal and Bioherm-Equivalent Strata

Calcareous sandstone and sandy limestone and chert of the Springvale Member unconformably overlie the Bertie Dolomite (exposed in the Law quarry). Glauconite gives the Springvale its characteristic greenish color. Cherty, fossiliferous, fine-grained limestone makes up most of the remaining Bois Blanc, which is characterized by the brachiopod Amphigenia elongate (Telford and Tarrant, 1975; Oliver, 1976), in addition to rugose and tabulate corals, trilobites and bryozoans. The upper part of the Bois Blanc differs in consisting of greenish, shaly limestone which grades upward into overlying Edgecliff beds. Oliver (1976) drew the boundary between the formations based on occurrence of Acinophyllum species. Beds which contain A. stokesi and A. simcoensis are considered Bois Blanc; those containing A. segregatum are Edgecliff. He suggests that these units may represent the missing time interval between Bois Blanc and Edgecliff deposition elsewhere. The lithologic transition suggests that similar depositional conditions prevailed throughout late Bois Blanc and early Edgecliff sedimentation. Acinophyllum biostromes which developed under these conditions represent incipient stages of bioherm development.

Biohermal Strata

Two thin (20 cm) beds of Acinophyllum bafflestone mark the earliest phases of Edgecliff bioherm development. These beds and argillaceous limestone containing diverse coral fauna, including rugose corals, Cystiphylloides, Cylindrophyllum, Heliophyllum, and the tabulate favositid coral Pleurodictyum, are exposed in Law quarry and in the northern part of the inactive (southern) quarry. Total thickness is 4.2 m. Overlying are 0.8 to nearly 2.0 m of massive, light-colored limestone dominated by Acinophyllum in association with rugose corals, Cystiphylloides and Heliophyllum, tabulate corals, Favosites and Emmonsia, and sparse stromatoporoids. Intraskeletal porosity is well-developed; black bitumen stains are common. Matrix consists of crinoid grainstone and packstone and contrasts markedly with that of the underlying, more argillaceous Cystiphylloides and Acinophyllum beds.

The massive-bedded Acinophyllum facies is overlain by 7.6 m of medium-bedded packstone, wackestone and grainstone which contains a varied and abundant fauna, including colonial rugose corals, Acinophyllum and Heliophyllum; solitary rugose corals, Cystiphylloides and Siphonophrentis; massive tabulate corals, Favosites, Lecfedites, and Emmonsia; branching tabulates, Cladopora, Thamnopora, Syringopora, and Aulopora; and laminar stromatoporoids. Crinoid debris is abundant; brachiopods are sparse. Thin sections from the upper 4 m of the buildup reveal abundant fenestrate and encrusting bryozoans, the latter occurring in association with crinoid holdfasts. Chert is common throughout the section, but is absent from the *Acinophyllum* and *Cystiphylloides* beds. The buildup is overlain by argillaceous, sparsely fossiliferous mudstone of the Clarence Member, characterized by abundant dark chert nodules.

Vertical extent of the buildup cannot be determined exactly, but a thickness of about 10 m is estimated. Estimating lateral extent is even more difficult, due mainly to the tabular nature and lack of "typical" biohermal zonation and distinct flank facies. Although the position of the exposure in relation to the buildup as a whole is unknown, absence of bioherm facies in the Law quarry suggests the buildup is on the order of 1 km broad.

Leave Stop 2 and continue south toward Lake Erie.

38.0	0.4	Turn left onto Lakeshore Road.
38.3	0.3	Large, stabilized sand dunes on right.
38.7	0.4	Dunes migrating over road surface.
39.3	0.6	Sugarloaf Hill on right. This dune rises more than 100 ft above lake level and has been a landmark to Lake Erie travellers since the days of the earliest explorers.
40.0	0.7	Bridge over drainage canal.
40.1	0.1	Curve left; Lakeshore Road becomes Rosemount Avenue.
40.3	0.2	Stop. Turn right onto Sugarloaf Street.
40.8	0.5	Turn right into park.

LUNCH STOP.

The park offers a view of Port Colborne's harbor and the Welland Canal's port of entry. Over 6,600 ships and nearly 800 pleasure craft passed through the canal in 1980. Inco Metals Company nickel refinery is Port Colborne's largest employer. Canadian nickel, in Inco-developed alloys, was used in the main engines and other components of the space shuttle *Columbia*.

Leave park.

41.4	0.1	Cross Sugarloaf Street and proceed north on Elm Street.
41.5	0.4	Cross railroad tracks.
41.8	0.3	Turn left at traffic light onto Killaly Street (Regional Road 5).
42.3	0.5	Junction, Regional Road 64 (Ontario Highway 58) on right.
42.4	0.1	Turn left.
42.5	0.1	Cross railroad tracks.
42.7	0.2	Rubble pile on right. Park on concrete pad just beyond. Walk SW into abandoned quarry complex.

STOP 3. PORT COLBORNE EAST QUARRY

No detailed studies of this quarry complex have been published since the days of Stauffer (1915). It was once operated by the Canadian Portland Cement Company and included a large plant, of which only the foundations remain. The quarries were developed on a low, ENE-trending anticline. Telford and Tarrant (1975) mapped only Edgecliff at this locality, however, in places, strata have a distinctively Clarence character - argillaceous, fine-grained, abundant chert - particularly in the north and west walls. This is probably another manifestation of the interfingering nature of Edgecliff and Clarence facies as described at the Ridgemount quarry. Beds at water level are packstone or grainstone; skeletal material decreases and mud increases upward. The "island" in the southern end of the quarry is a bioherm. Large blocks of light-colored Acinophyllum bafflestone are piled to the east of the island, along the roadway which leads southward into the quarry. Near the island, a railroad trestle crosses a cut which connects the northern quarry with a smaller one south of the tracks. In this area are exposed approximately 4 m of massive bedded, light gray, crinoid grainstone and packstone containing abundant large favositids, common Cystiphylloides, and fewer Thamnopora and Syringopora. These strata have all the aspects of proximal flanks observed in the subsurface bioherms.

This stop offers excellent fossil-collecting, either from the rubble piled east of the quarry, or from around the quarry itself. As at the western quarry, glacially-smoothed surfaces provide two-dimensional exposures of the fossil bioherm community.

Leave stop 3 and return to Killaly Street.

43.0	0.3	Turn right.
43.1	0.1	Turn left onto Regional Road 64/Ontario Highway 58.
43.5	0.4	Turn right onto Ontario Highway 3/Regional Road 3.
44.5	1.0	Cross main channel, Welland Canal.
61.5	17.0	Peace Bridge. Return to Marriott Inn by best route.

WPC(#1348)

EURYPTERIDS, STRATIGRAPHY, LATE SILURIAN-EARLY DEVONIAN OF WESTERN NEW YORK STATE AND ONTARIO, CANADA

SAMUEL J. CIURCA, JR. - Rochester, New York

INTRODUCTION

The eurypterid horizons of the Bertie Group in New York State (Ciurca, 1973, 1978a) continue into adjacent Ontario, Canada bearing their distinctive eurypterid faunas (Ciurca, 1967a, 1976). In addition, Helderbergian strata (Clanbrassil Fm.), between Byng and Haggersville, show the presence of post Bertie-Akron strata in this region in a manner analogous to that of the Honeoye Falls Fm. described for similarly positioned strata discovered south of Rochester, New York (Ciurca, 1967b, 1973).

The purpose of this paper is to present a preliminary account of the stratigraphy and paleontology of this interesting eurypterid-bearing sequence of rocks and to stimulate research into the origin and significance of the waterlimes and associated lithologies in which eurypterids are peculiarly characteristic and abundant.

STRATIGRAPHY AND PALEONTOLOGY

Detailed field studies in New York and Ontario, Canada were undertaken with the hope of providing a better framework for understanding the occurrence and distribution of eurypterid faunas and waterlimes in the regions, and to facilitate correlation. In these studies, the waterlime units have been utilized as the basis of a stratigraphic framework. They have been given an importance similar to that given coal beds in cyclothemic sequences and it is hoped that this will help tracing and understanding of the cyclic sequences present in our Late Silurian-Early Devonian rocks. The distribution of the waterlimes of the Bertie Group, and of the new stratigraphically higher waterlimes, is shown in Figure 2. Heretofore, the only well known waterlime was that of the Williamsville from which much of the eurypterid material in collections was knowingly or unknowingly obtained (i.e. *Eurypterus remipes lacustris* Fauna).

The following units, beginning with the Salina Group, are present in western New York. At least some of these have been traced westward as far as Haggersville, Ontario, Canada.

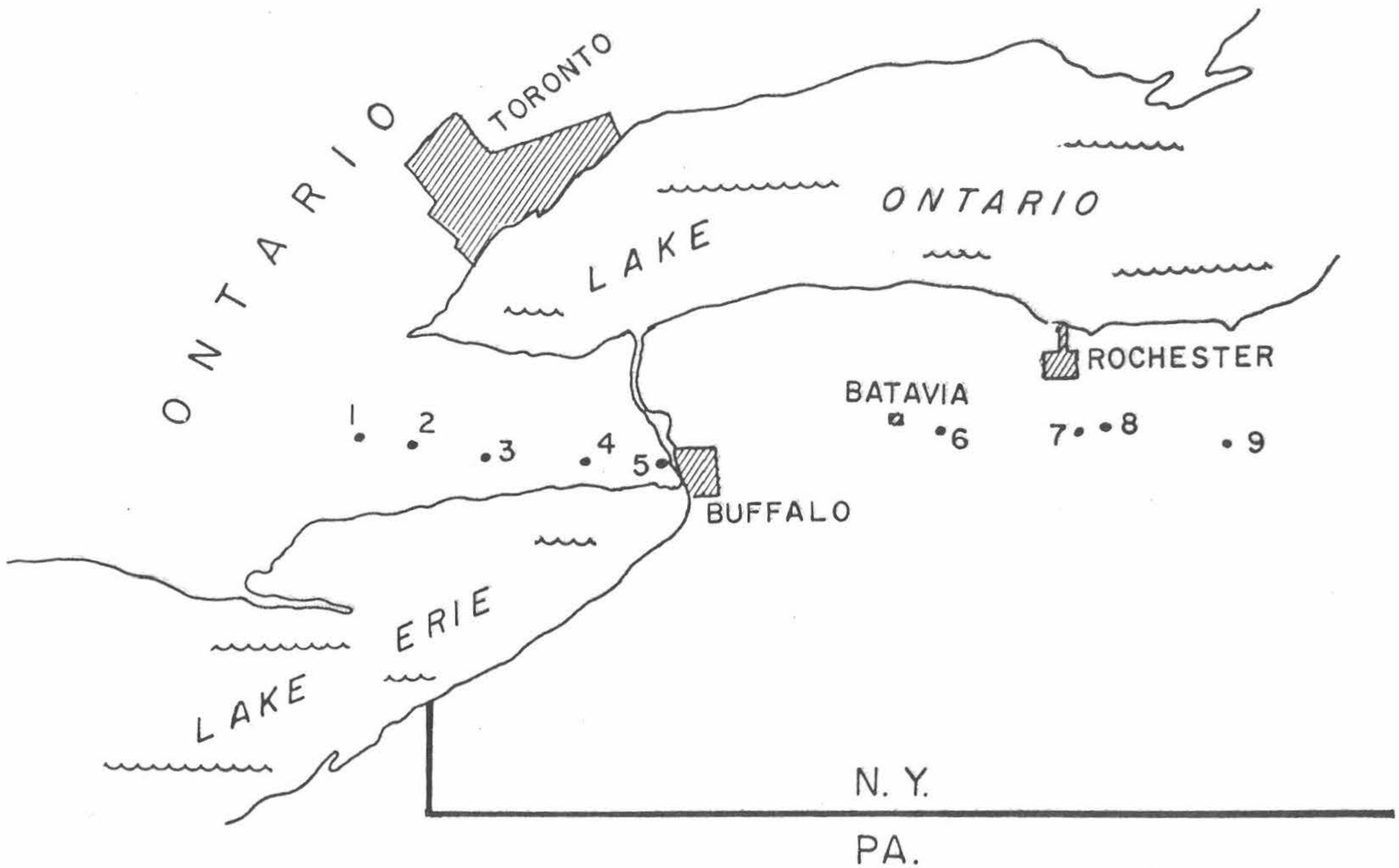


FIGURE 1 Areas (1-9) having exposures of Late Silurian-Early Devonian strata:

1. Haggersville
2. Cayuga
3. Dunnville
4. Port Colborne
5. Fort Erie
6. Le Roy
7. Honeoye Falls
8. East Victor
9. Phelps

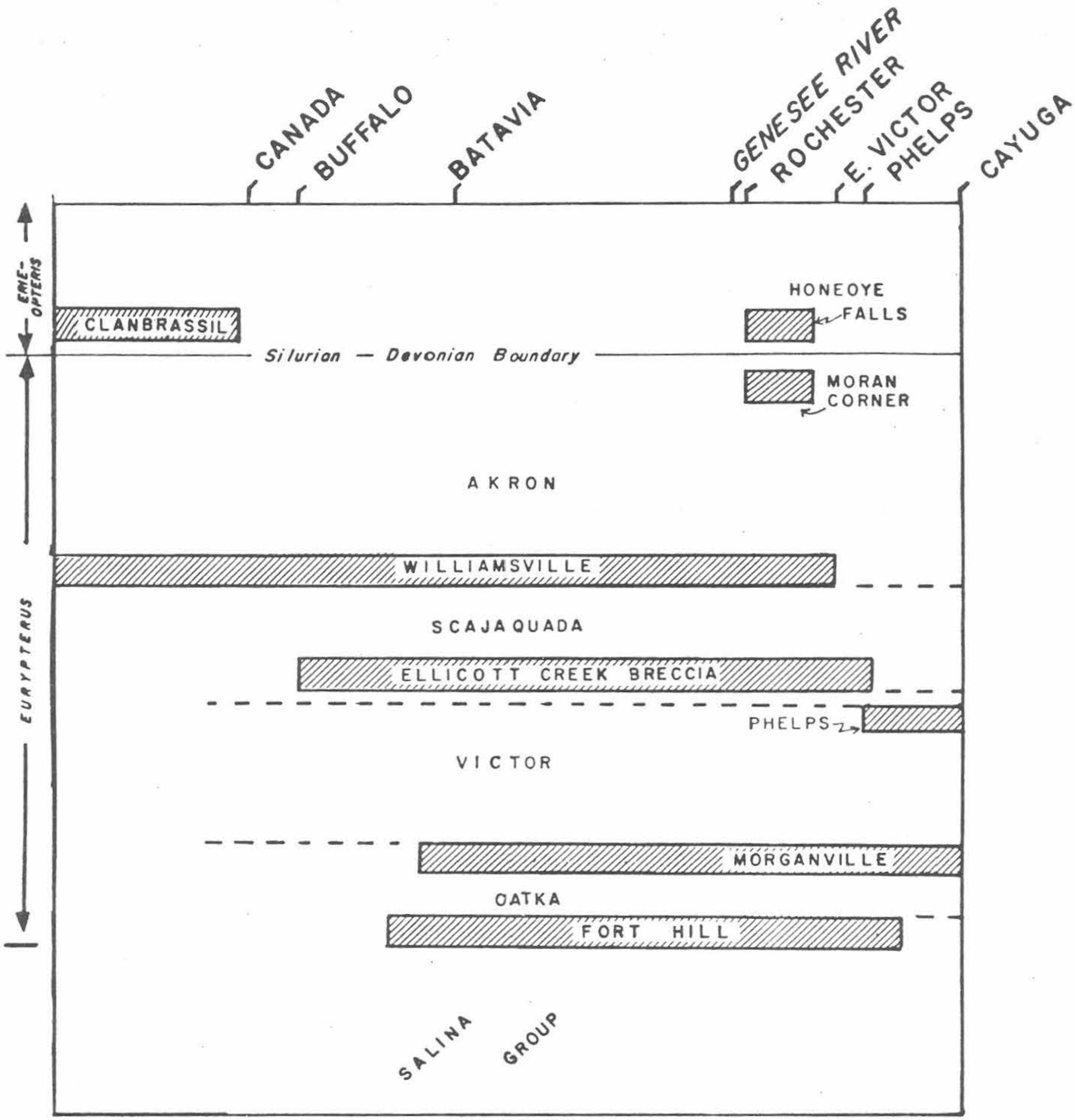


FIGURE 2 Distribution of Late Silurian and Early Devonian Eurypterid-bearing Waterlimes (not to scale).

Salina Group

Much of western New York is underlain by dolomitic rocks and evaporites of various types; red and green shales, halite, anhydrite, and gypsum, hematite, etc. The basal Vernon Fm., in its typical red and mottled facies, is well exposed along the Barge Canal at Pittsford, New York. Nothing is known of the outcropping of the Vernon Formation westward.

The Syracuse and Camillus Formations are exposed in the Oatka Creek area north of Le Roy, at Black Creek near Batavia, and along Tonowanda Creek on the Tonowanda Indian Reservation. In Canada the Salina Group is rarely exposed and this interval ought to be studied further.

Except for the eurypterid fauana of the Pittsford Shale, which is found between the underlying Lockport Dolostone and the typical red Vernon Fm. near Rochester, little is known about the paleontology of the Salina Group in this region. The only eurypterids known are from the Syracuse Fm. near Batavia. Here *Waeringopterus* (see Figure 4d) occurs in resistant dolostones associated with salt hoppers.

Eurypterid fragments (*Eurypterus*? sp.) were observed in outcrops of waterlime (Syracuse Fm.) along the Oatka Trail, but again, outcrops are so few and so small that it is difficult to learn of the fauna contained in this interval.

The contact of the Salina Group with the underlying Lockport dolostones is not exposed in western New York. A study of the lithofacies and paleoenvironments of the Lockport Fm. was recently provided in a thesis by Domagala (1982).

Bertie Group

The Bertie Group is best known because of the eurypterids, i.e. the *Eurypterus remipes lacustris* Fauna, found in one unit, the Williamsville Formation. Because of the confusion attending the use of the terms "Bertie Waterlime" and "Bertie Limestone" in identifying the horizon from which the *Eurypterus remipes lacustris* Fauna was obtained, it is suggested that these quoted terms not be used. The Williamsville Fm. is only one of several waterlimes that are found within the Bertie Group. These Bertie waterlimes appear to carry distinctive faunas and every effort should be made to clearly identify the horizon from which specimens are obtained.

The oldest Bertie waterlime occurs in western New York and was termed the Fort Hill Waterlime (Ciurca, 1973). The youngest Bertie waterlime is represented by the Williamsville Formation.

Fort Hill Waterlime. The Fort Hill Waterlime carries an *Eurypterus* fauna associated with ostracods and large salt hoppers. It is a thin unit occurring at the base of the Oatka Shaly Dolostone. While this waterlime has been traced from Phelps, New York westward, it has not been observed in the Buffalo area and no outcrops exposing this interval have been seen in Ontario, Canada.

Oatka Formation. The Oatka Formation is an inconspicuous unit seen only where exposures descend below the massive Fiddlers Green Formation. It is thus far barren of fossils. Like the underlying Fort Hill Waterlime, little is known of its extent westward into the Buffalo or Ontario, Canada areas.

Fiddlers Green Formation. Throughout much of the outcrop belt east and west of Phelps, New York (type section of the Phelps Waterlime), the Fiddlers Green Fm. consists of a central unit of massive dolostone and limestone (Victor Mbr.) sandwiched between two eurypterid-bearing waterlimes. The upper waterlime (Phelps Member) contains the *Eurypterus remipes remipes* Fauna so well known from the Herkimer County area (Ciurca, 1965, 1973). The lower waterlime (Morganville Member) also contains an *Eurypterus* fauna. These waterlimes appear to be attenuated in a westerly direction, especially in Ontario, Canada. Exposures, especially of the lower parts of the sequence, are rare and make tracing difficult. Additionally, the massive Victor Member dominates most sections. Currently, both the Morganville and Phelps Waterlimes are believed to be present in Ontario, Canada. The eurypterid fauna of the Phelps Waterlime is listed in Table 1.

Ellicott Creek Breccia. In Ontario, below the Scajaquada Formation, occurs a peculiar brecciated waterlime. This waterlime was traced into the New York sections and is termed the Ellicott Creek Breccia. The type section is at Ellicott Creek at Williamsville, New York

The Ellicott Creek Breccia is a highly variable unit generally having a more massive middle unit. At the type locality it is 2.3-2.5 meters thick. Variously colored, banded and straticulate waterlimes occur and these are rich in eurypterid remains and cephalopods. In Canada the Ellicott Creek Breccia becomes much thinner but retains a tripartite structure. Eurypterids preserved in the upper waterlimes are quite flattened while those preserved in the lower waterlime appear to be uncrushed. Salt hoppers are common.

The more massive middle unit is probably the result of 'reefy' algal masses and the variability in the thickness of the unit is probably due to variation in algal mound development.

TABLE 1. ARTHROPODS OF THE BERTIE GROUP

Phelps Waterlime	Williamsville Waterlime
EURYPTERIDS	
Eurypterus remipes remipes DeKay	Eurypterus remipes lacustris Harlan
Acanthoeurypterus wellsi Kjellesvig-Waering	Acanthoeurypterus dekayi (Hall)
Pterygotus (Acutiramus) macrophthalmus macrophthalmus Hall	Pterygotus (Acutiramus) macrophthalmus cummingsi Grote & Pitt
Pterygotus (Pterygotus) juvenis Clarke & Ruedemann	Pterygotus (Pterygotus) cobbi Hall
Paracarcinosoma NEW SPECIES	Paracarcinosoma scorpionis (Grote & Pitt)
Dolichopterus herkimerensis Caster & Kjellesvig-Waering	Dolichopterus macrocherius Hall
Dolichopterus jewetti Caster & Kjellesvig-Waering	Dolichopterus siluriceps Clarke & Ruedemann
X	Buffalopterus pustulosus (Hall)
Alloeurypterus linsleyi Kjellesvig-Waering	X
Clarkeipterus testudineus (Clarke & Ruedemann)	X
SCORPIONS	
Proscorpius osborni (Whitfield)	
Archaeophonus eurypteroides Kjellesvig-Waering	
XIPHOSURANS	
Pseudoniscus clarkei Ruedemann	Bunaia* woodwardi Clarke
PHYLLOCARIDS	
Ceratiocaris aculeatus Hall	Ceratiocaris acuminata Hall
	Ceratiocaris maccoyana Hall

* may be junior synonym of Pseudoniscus
(list compiled by S. J. Ciurca, Jr. for Buffalo NYSGA, 1982)

Besides the type locality, the Ellicott Creek Breccia is well displayed in the quarries at Port Colborne, Ontario. At Phelps, New York it appears to overlie the Phelps Waterlime. However, it has not been recognized eastward. A large eurypterid-bearing slab of brecciated waterlime, found on the shore of Cayuga Lake near Cayuga Junction, may indicate the presence of the Ellicott Creek Breccia in this area.

At most localities chert and sphalerite are especially characteristic of this unit. In Ontario, a thin black shale occurs at the base and may represent a hiatus between the Ellicott Creek Breccia and the underlying Victor Dolostone (or Phelps Waterlime if present).

Scajaquada Formation. Underlying the Williamsville Formation is a unit of argillaceous dolostones containing small chert nodules, the Scajaquada Formation. This unfossiliferous unit appears to be, at least in part, an extension of the gypsiferous Forge Hollow Formation of central New York. In western New York and Ontario, the Scajaquada Fm. is distinctively different in lithology than the Forge Hollow Formation. Beds are thicker and more massive. Mudcracks are present and evaporite (crystal) casts have been observed.

The Scajaquada Fm. is about 1 meter thick at Williamsville, New York. The unit is also exposed at Scajaquada Creek, and in several quarries in Ontario, Canada, particularly in those at Port Colborne. See Figure 3.

Williamsville Formation. In the Buffalo-Williamsville area, and in adjacent Ontario, Canada, the Williamsville Fm. has been of considerable interest to those who appreciate the great variety and quantity of eurypteroid arthropods preserved in this unit.

While little attention has been given to the detailed stratigraphy of the eurypterid-bearing sequence of the region, it seems clear that in order to understand many facets of Late Silurian sedimentation, the migration of environments resulting in cyclic sequences, and the evolution of eurypterid faunas, much more information is needed.

Although a relatively large fauna has been reported from the "Bertie" over the years, little is known about the exact stratigraphic position from which much of the material was obtained. Even now material is still being labeled and sold as simply "Bertie Waterlime" or "Bertie Limestone" and there is no doubt that specimens currently in the 'marketplace' have been obtained from various horizons.

Fortunately, most material I've studied in museums and private collections comes from the Williamsville Formation. The majority of specimens labeled Buffalo or Williamsville, New York came from the "waterlime" quarries and originated in the Williamsville Formation. Specimens labeled Black Rock or other localities need further study.

The type locality for the Williamsville Formation is the exposure in Ellicott Creek at Williamsville, N. Y., although originally, exposures just to the west in the waterlime quarries, were what was intended to represent this unit.

At Ellicott Creek the Williamsville Fm. forms part of the falls just north of NY 5 and part of the walls on both sides of the ravine. Here it is 1.58 m. to 1.8 m. thick and contains the characteristic _Eurypterus remipes lacustris_ Fauna.

To the east the Williamsville Fm. is present at numerous localities where this interval is exposed. It is 2.2 m. thick at Akron Falls where only a single specimen of _E. r. l._ has been found in several years of searching. At Oakfield and at Batavia only a portion of the Williamsville Fm. is present, the unit being in direct contact with the Silurian-Devonian Unconformity at the base of the Onondaga Ls., or the Bois Blanc Ls. where present. Near Le Roy it is present at the Nied Road Quarry, but no eurypterid remains have been observed there, only phyllocarids (Ciurca, 1973).

In the Buffalo-adjacent Ontario region, the Williamsville Formation is basically a tripartite unit here termed A, B, and C. In addition, a transitional unit is also included and is termed Williamsville D.

Williamsville A and C are quite eurypterid-bearing and in the Fort Erie area, most of the specimens I have obtained were from Williamsville A. Early collections (e.g.) those preserved in the Buffalo Museum of Science) contain numerous specimens from the transitional unit D, in which large ostracods are often present, and probably from Williamsville C. The brachiopod _Eccentricosta jerseyensis_ was found at Fort Erie in Williamsville B and this unit may be simply a tongue of a more marine phase present elsewhere.

The Williamsville Fm. is readily traced through Port Colborne to Byng where it is well displayed in a number of quarries. In this area the Williamsville Fm. is quite fossiliferous. _Eurypterus remipes lacustris_, _Pterygotus macrophthalmus cummingsi_, phyllocarids, ostracods, and other forms have been found. In the Port Colborne area, north to south facies changes appear to take place between the Scajaquada and Williamsville Formations. To the north the Williamsville Formation appears to be replaced by shaly dolostones typical

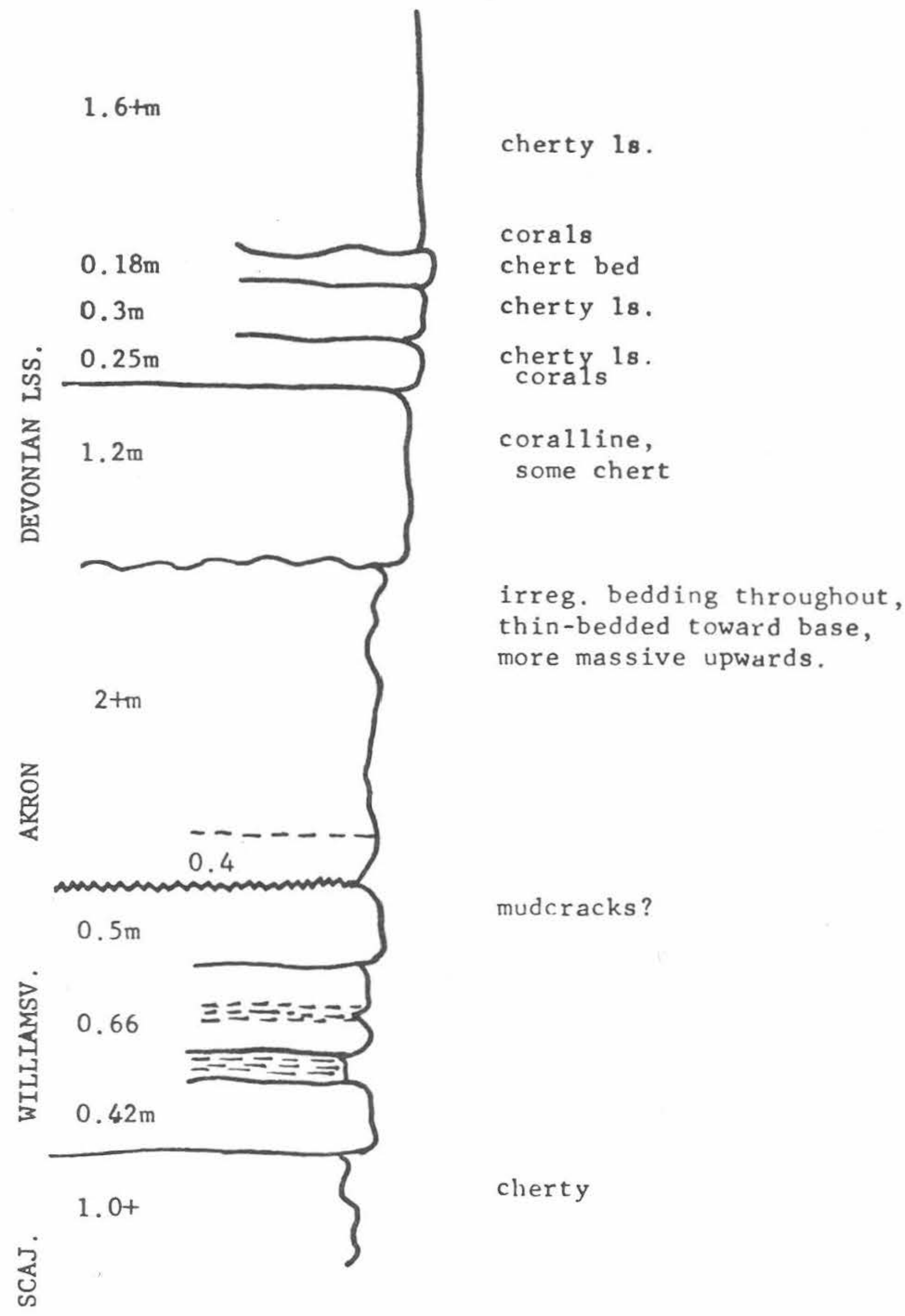


FIGURE 3 Section at Williamsville, New York, Ellicott Creek, west side, tributary showing Scajaquada through Akron Formational relationships.

of the Scajaquada Formation. The eurypterid fauna of the Williamsville Fm. is listed in Table 1.

Akron-Cobleskill Formations

Overlying the Williamsville Fm. is a unit of mottled, generally massive dolostone referred to as the Akron Formation. The thickness of this unit is usually assigned eight feet. However, in the area of the type locality, an unconformity occurs at the top of the unit. Had another area been chosen, it would have been realized that the unit is actually thicker. In eastern New York, the type Cobleskill Fm. has been assigned 6 to 9 feet of fossiliferous strata and workers have generally tried to maintain this uniformity in thickness across the state in relating the Cobleskill/Akron units.

There is no doubt that the two units are related, however the exact relationship is still unknown (Ciurca, 1978a). I believe that the Akron Fm. in western New York is younger than the Cobleskill Fm. of the type area. This suggests that the type Williamsville Fm. correlates with some portion of the Cobleskill Fm. of eastern New York.

In Ontario, the Akron Fm. is relatively unfossiliferous. Ostracods and brachiopods are the most common forms found. Salt hoppers, not yet seen in western New York exposures of this unit, occur in the Akron Fm. near Fort Erie and indicate a much more restricted 'marine' influence in this region.

Recently, Belak (1978, p. 5, Fig. 3) studied the Cobleskill Fm. and attempted to define "lithologic units" A, B, C, across the state. Unfortunately, the Cobleskill Fm. is a complex unit not amenable to such a simple subdivision. As pointed out previously, a variety of facies resulting from reef-associated sedimentation, is present in the Cobleskill/Akron units (Ciurca, 1978a). Additionally, the limestones occur at various stratigraphic positions and are of various types. It seems probable that eurypterid and cephalopod-bearing dolostones, usually assigned to the Williamsville Fm., grade eastward into eurypterid and cephalopod-bearing limestones so that at Forge Hollow, N. Y. only the Oxbow Waterlime of Rickard (1962) remains between the Forge Hollow and Cobleskill Formations.

Belak (1978, p. 152, 153) misidentified strata at several sections. No Williamsville Fm. is known to occur at the Chittenango Falls exposures, nor at the stream bed to the north. The Forge Hollow Fm. he identified is Chrysler Fm., and the Williamville Fm. above this is actually a waterlime bed in the Chrysler Formation (see Ciurca, 1978a, Figure 5).

Also, Williamsville Fm. (Belak, pp. 157-158) equals the uppermost Cobleskill Fm. (transitional to Chrysler Fm.); Williamsville Fm. (Belak, 1978, p. 167) equals Cobleskill Fm.; Cobleskill/Akron Fm. (Belak, p. 162) equals the Fiddlers Green Fm.; Rondout/Chrysler (Belak, p. 162) equals the Phelps or Williamsville Waterlime (both are exposed in the vicinity).

Clanbrassil Formation

In the region from Byng to Haggersville, Ontario, a sequence of fine-grained dolostones, the Clanbrassil Fm., overlies the Akron Formation. These rocks were formerly referred to the Bertie Group by many authors (e.g. Hewitt, 1971) but are stratigraphically higher and do not occur in the area of the type Bertie Group.

The type section at Clanbrassil exhibits an eurypterid-bearing facies that can be confused with other units because of lithologic similarities. The stratigraphic position, and the occurrence of the genus Erieopterus, distinguish this sequence from the rocks below.

The Clanbrassil Fm. is present at Cayuga and at Haggersville in the large quarries. At these localities, Erieopterus is abundant, particularly in the lower half of the unit. No other fossils have yet been found.

Erieopterus-bearing units with which the Clanbrassil Fm. correlates, at least in part, include the Honeoye Falls Fm. south of Rochester, N. Y., the Chrysler and Olney Formations of central New York, and only the Erieopterus-bearing portions of the sequence exposed on the Bass Islands of Lake Erie.

EURYPTERID BIOSTRATIGRAPHY

In New York State it is quite clear that recurring eurypterid-bearing lithologies, the result of Late Silurian cyclic sedimentation (Ciurca, 1973, 1978a), appear at numerous stratigraphic levels. More importantly, each contains at least one characteristic eurypterid or eurypterid fauna that should eventually prove useful in regional correlations. In western New York, two eurypterid horizons have been well known for about 80 years. One, the lowest, contains the Hughmilleria socialis Fauna near the base of the Vernon Shale (Salina Gp.). The other well known horizon, the Williamsville Fm. (Bertie Gp.), bears the Eurypterus remipes lacustris Fauna and, while many specimens continue to be misidentified, or their exact stratigraphic horizon recorded inaccurately, it remains an important horizon that is geographically widespread.

To these faunas and horizons for western New York and

Ontario, Canada can be added several new ones. The current status of forms and distributions follow. Eurypterid zonation is shown in Figure 4. Several characteristic forms are shown in Figure 5.

Tylopterella boylei Fauna

In 1884, Whiteaves described a peculiar eurypterid from the Guelph Fm. of Ontario, Canada (Clarke and Ruedemann, 1912). The eurypterid, Tylopterella boylei, is known only from a single specimen. It is important because it is a very distinctive form and is not known from succeeding rocks. It is hoped that more material from this horizon will be found in the future.

Hughmilleria socialis Fauna

In green, greenish-black and black shales in the Pittsford Shale near the base of the Vernon Fm. (Salina Gp.) occur abundant specimens of Hughmilleria socialis Sarle and an associated fauna of rarer forms including Eurypterus pittsfordensis, Pterygotus, and Mixopterus. Recent excavations at Pittsford have provided hundred of specimens for study and illustrate the richness of this stratigraphic interval in this portion of New York State. The extended geographic distribution of at least the Eurypterus pittsfordensis portion of this fauna was recently described (Ciurca, 1978b; Hamell, 1978).

Waeringopterus cumberlandicus apfeli Fauna

Little is known about the surface outcropping of rocks of the Salina Group in western New York. However, a fine exposure of massive dolostone and waterlime at Black Creek, east of Batavia, undoubtedly represents a portion of the Syracuse Formation. From these beds I obtained several specimens of the characteristic genus, Waeringopterus. This form was previously described from the Syracuse Fm. in central New York (Leutze, 1961; Kjellesvig-Waering and Leutze, 1966). As is often the case, the new specimens are intimately associated with salt hoppers.

Eurypterus sp. Fauna

The Fort Hill Waterlime at the base of the Oatka Formation has yielded numerous specimens pertaining to a small Eurypterus sp. associated with salt hoppers up to several inches on a side (Ciurca, 1973). This horizon remains unknown in the Buffalo area and westward and little is known at present of the associated fauna.

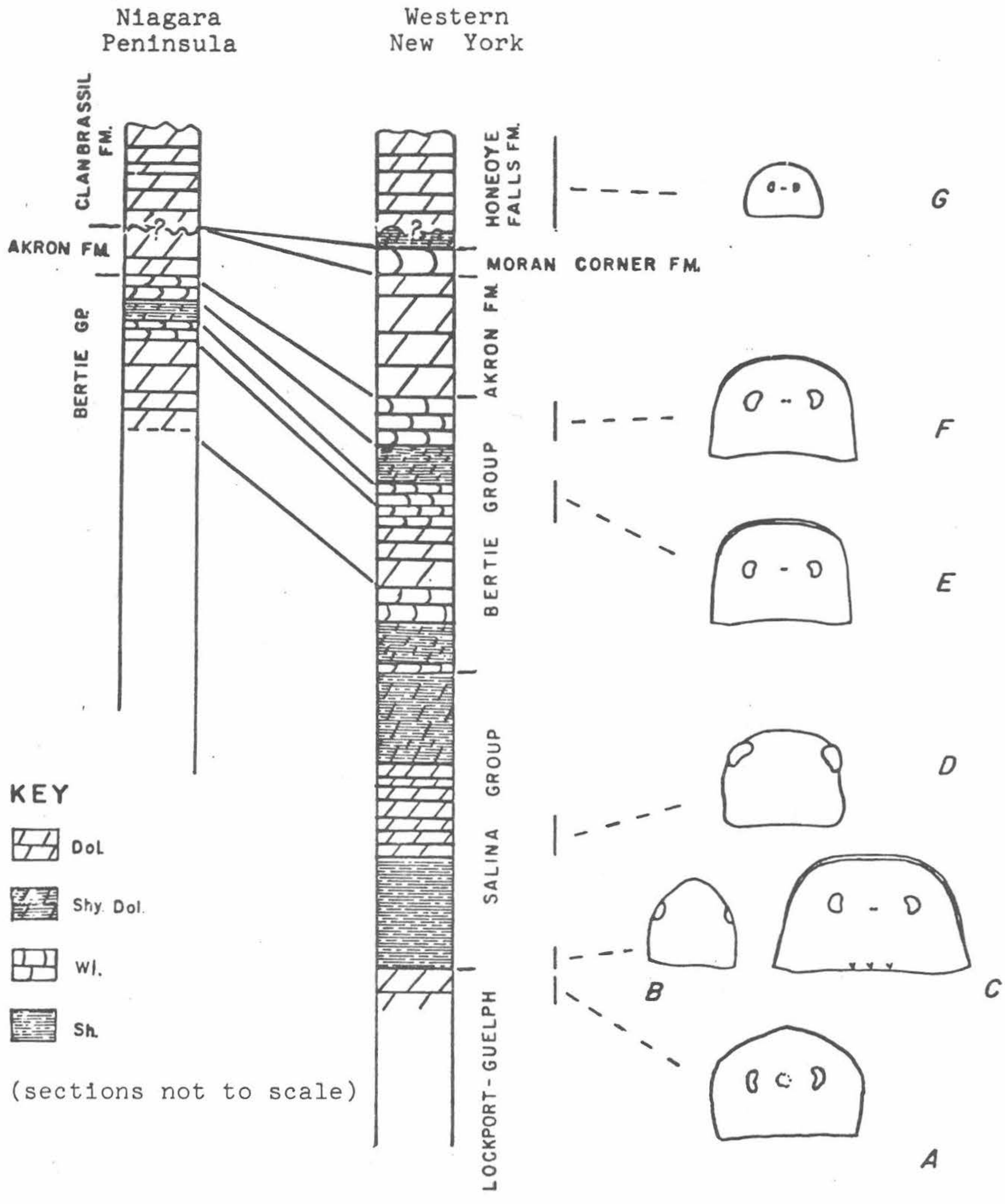


FIGURE 4 Eurypterid Biostratigraphy: Zonation based on stratigraphic distribution of characteristic Silurian and Devonian eurypterids, a) Tylopterella, b,c) Hughmilleria socialis, Eurypterus pittsfordensis (respectively), d) Waeringopterus, e) Eurypterus remipes remipes, f) Eurypterus remipes lacustris, and g) Erieopterus.

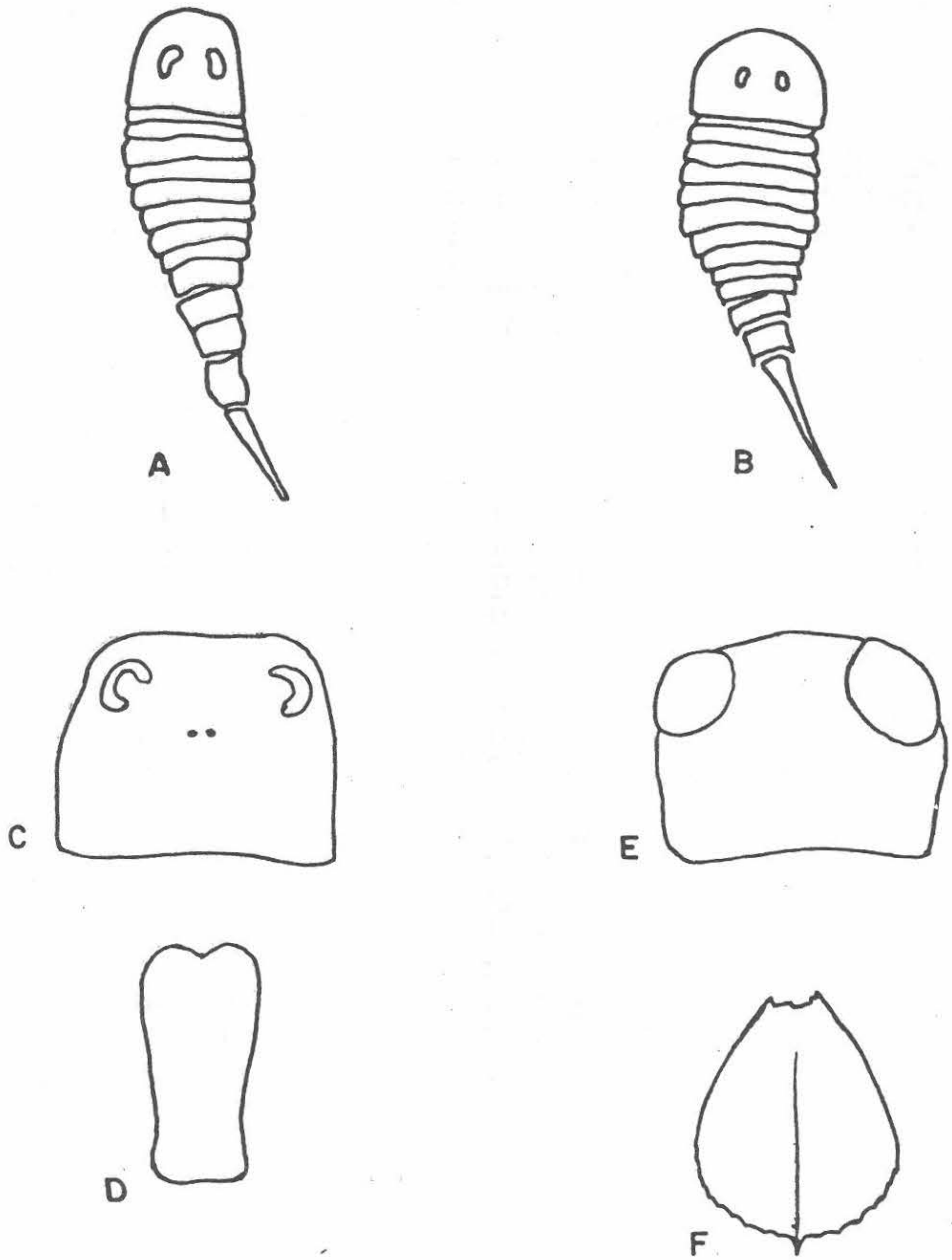


FIGURE 5 Characteristic Eurypterids of Southern Ontario (Niagara Peninsula), a) Eurypterus remipes remipes, Ellicott Creek Breccia, Port Colborne, Ontario; b) Erieopterus microphthalmus ssp., Clanbrassil Fm.; c,d) carapace, metastoma, Dolichopterus macrochirus, Williamsville Fm., Fort Erie, Ontario; e,f) carapace, telson, Pterygotus macrophthalmus cummingsi, Williamsville Waterlime, Fort Erie, Ontario.

Eurypterus remipes remipes Fauna

The well known Eurypterus remipes remipes Fauna of central-eastern New York is apparently represented in the Ellicott Creek Breccia of western New York and in Ontario, Canada. It may also be present in a thin waterlime just below this unit. Associated forms include Dolichopterus sp., Pterygotus sp., and in Canada, probably Clarkeipterus (Ciurca and Gartland, 1976). This fauna is intimately associated with casts and impressions of halite crystals.

Eurypterus remipes lacustris Fauna

The best known fauna in the region is that of Eurypterus remipes lacustris. The eurypterid collections of the Buffalo Museum of Science are rich in specimens of this characteristic species. The associated fauna includes Pterygotus macrophthalmus cummingsi, Dolichopterus sp., and Paracarcinosoma sp. (Table 1). The E. r. l. Fauna has been found as far west as Haggersville, Ontario and as far east as East Victor, southeast of Rochester, New York (Ciurca, 1973). Paracarcinosoma scorpionis, if the species is correctly identified, continues to be represented eastward in the Williamsville Waterlime as far as Jamesville, N. Y., southeast of Syracuse, where it is still associated with Lingula and phyllocarids (Ciurca, 1978a).

It is important to note that this fauna is associated, via Williamsville B, with Eccentricosta jerseyensis and other brachiopods. The brachiopod E. jerseyensis characterizes beds of Latest Pridoli age in New York, Pennsylvania, New Jersey, Maryland, and West Virginia (Berry and Boucot, 1970). E. jerseyensis has been found in the Williamsville Fm. at East Victor (Ciurca, 1973) and was also found just east of Clarence, New York. In Canada only Williamsville B has yielded this brachiopod at Fort Erie, Ontario.

Erieopterus Fauna

As pointed out previously, Erieopterus continues to be known from rocks believed to be Early Devonian in age and associated with the Manlius Group of central-eastern New York (Ciurca, 1978a). Erieopterus has been found abundantly near the base of the Honeoye Falls Fm. in the area south of Rochester (Ciurca, 1967b, 1973) and in Ontario, Canada, it occurs in the Clanbrassil Fm. (Ciurca, 1976). Both the Honeoye Falls and Clanbrassil Formations correlate with the Chrysler-Olney sequence of central New York and it appears that all occurrences of Erieopterus, from Thacher Park (near Albany, New York) westward to Ontario, Canada and Ohio, occur near or at the base of the Early Devonian.

EURYPTERIDS IN ONTARIO, CANADA

Like New York, Ontario rocks record the presence of a substantial variety of eurypterids during the Silurian. Both regions, however, need much more exploratory work, particularly in the Early and Medial Silurian rocks. Occurrences of Carcinosomatids and Pterygotids in the Niagaran rocks of Ontario are noteworthy, but generally based on scant remains. Particular attention needs to be focused on these earlier stratigraphic occurrences.

Copeland and Bolton (1960) have reviewed the occurrence and distribution of Canadian arthropods including eurypterids. For the Bertie Group, however, they note only the occurrence of the typical *Eurypterus remipes lacustris* (Williamsville Fm.). Careful tracing of the New York eurypterid horizons into Ontario and extensive collecting of fossils shows the presence in Ontario of most of the forms known from the Late Silurian (Cayugan) rocks of New York.

The following forms were recognized in the Ellicott Creek Breccia: *Eurypterus remipes remipes*, *Pterygotus*, *Dolichopterus*, *Clarkeipterus*, ostracods, cephalopods (common in the upper bed), and brachiopods (lowermost bed).

The following forms were found in the Williamsville Fm. in Ontario: *Eurypterus remipes lacustris*, *Pterygotus macrophthalmus cummingsi*, *Dolichopterus macrochirus*, *Eurypterus dekayi*, *Ceratiocaris*, *Lingula*, gastropods, cephalopods, and the brachiopod *Eccentricosta jerseyensis*.

The Clanbrassil Fm. has provided only one form, *Erieopterus microphthalmus*, that is common in the lower half of the unit.

Fragments of eurypterids and abundant ostracods and brachiopods occur in the massive dolostones of the Victor Member (Fiddlers Green Fm.), especially at Port Colborne, but little is known of the fauna.

UNCONFORMITIES

The nature of outcropping rocks is quite variable below the unconformity at the base of the Onondaga Limestone (or the Early Devonian Bois Blanc Fm. where present). See Ciurca, 1973, Figure 2, p. D-8A.

The effect of the unconformity on the nature of exposed strata for part of Ontario, Canada and western New York State is depicted in Figure 6. The contact of the Akron Dolostone with the Clanbrassil Fm. is indicated as probably unconformable. A biostratigraphic unconformity is suggested because

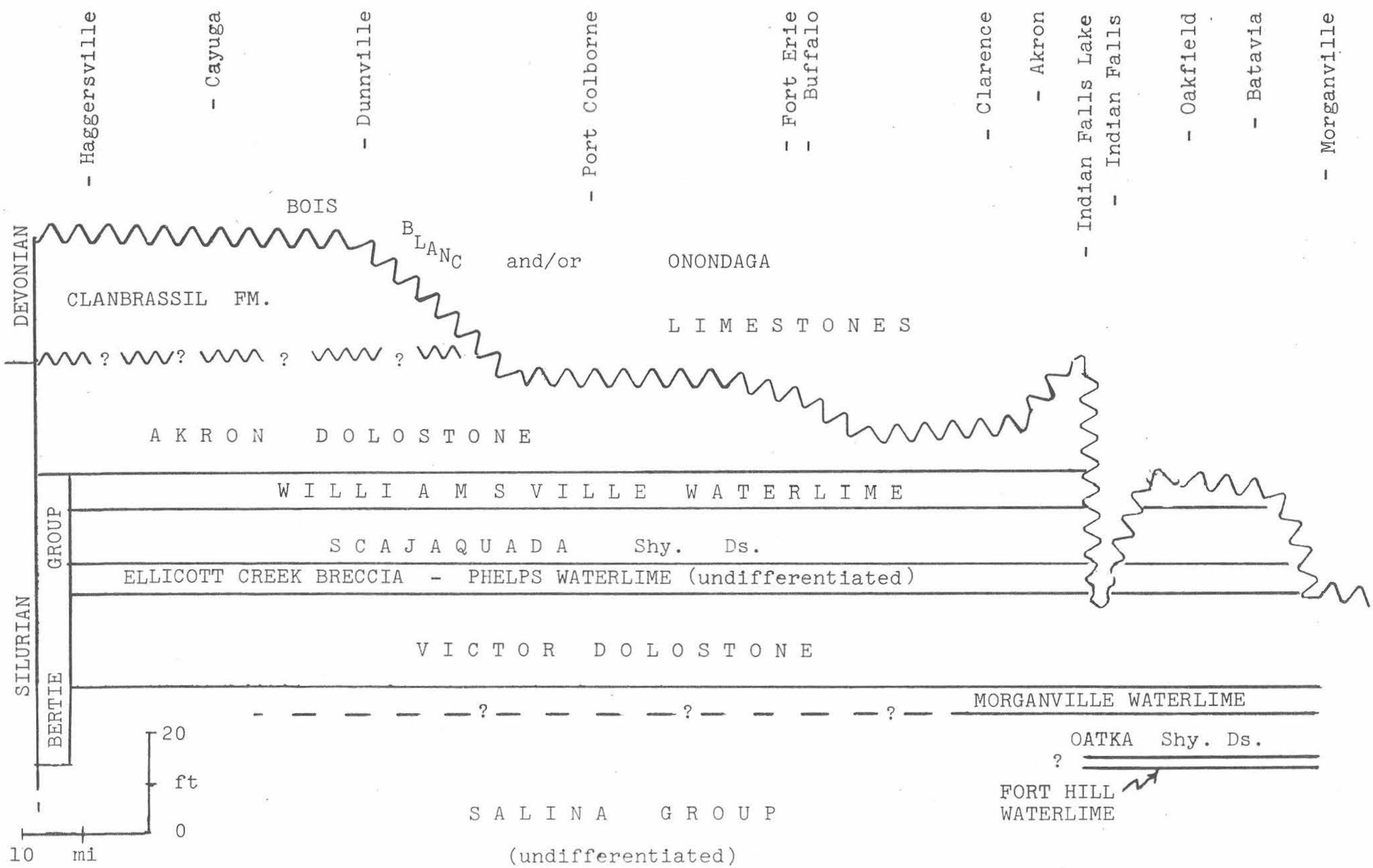


FIGURE 6 Lithostratigraphic cross section. Note the position and distribution of the Clanbrassil Fm., and also the effect of the unconformity on the nature of outcropping rocks (continuation of Fig. 2, Ciurca, 1973, p. D-8A).

of the replacement of an Eurypterus Fauna by an Erieopterus Fauna across this interval.

EURYPTERID COLLECTIONS

New York is fortunate in having a number of repositories in which large eurypterid collections are housed and, hopefully are readily available to anyone wishing to study this fascinating and problematical group of arthropods.

The Eurypterus remipes lacustris Fauna is particularly concentrated in the collections of the Buffalo Museum of Science (Clarke and Ruedemann, 1912; see also Bastedo, 1979), while the Eurypterus remipes remipes Fauna, especially material from the early collectors (see Clarke and Ruedemann, 1912) are concentrated in the New York State Museum at Albany. Collections from the Pittsford Shale horizons are maintained primarily at the New York State Museum and the Buffalo Museum of Science (see Clarke and Ruedemann, 1912).

Recent collections are rare. The largest is the Ciurca Eurypterid Collection (Ciurca, in preparation) at Rochester, New York which contains not only the E. r. l. and E. r. r. Faunas of the type areas (Buffalo and Herkimer "Pools" of early authors), but the entire interval geographically and vertically between these two extremes as well as considerable material from adjacent Ontario, Pennsylvania, Ohio, and Indiana.

Still another collection which has recently received attention is a collection emphasizing the E. r. l. Fauna of Ontario, Canada and the Buffalo area. The collection was assembled by Tony and Mike Sojka beginning in 1954 and was recently purchased by C. M. Seyfert (Seyfert and Seyfert, 1981).

Significant collections for study are also contained in the Smithsonian Institution, the American Museum of Natural History, and the Field Museum at Chicago. See also Fossil Invertebrates-Collections in North American Repositories 1976 (Paleontological-Society, 1977).

SUMMARY AND REMARKS

Eurypterid remains are found at several horizons in Late Silurian and Early Devonian strata of western New York and these have been traced into adjacent Ontario, Canada.

The Eurypterus remipes remipes Fauna occurs in the Ellicott Creek Breccia and the Eurypterus remipes lacustris Fauna occurs in the Williamsville Fm. as far as Haggersville, Ontario, Canada.

The replacement of an Eurypterus Fauna by and Erieopterus Fauna may indicate an unconformity at the Silurian-Devonian boundary in this region (Ciurca, 1978a, p. 245).

Examination of Late Silurian-Early Devonian strata in Ontario, Canada, and nearby states (Ohio, Indiana, Pennsylvania) indicates that cyclic sedimentation took place in these areas also. It is expected that several new eurypterid horizons and localities will be discovered in the future.

Recent use of cyclic sedimentation to interpret paleoenvironments within the Bertie Group has been provided by Hamell (1981). The research of Carl Stock (1979) on the stromatoporoids of the Cobleskill-Akron was recently published by the Paleontological Research Institution at Ithaca, New York. A study of the Cobleskill Fm. was attempted by Belak (1978, 1980). Eurypterus remipes remipes DeKay has been proposed as the State Fossil of New York (Fisher, 1982).

ACKNOWLEDGEMENTS

Richard D. Hamell helped considerably with the preparation of the manuscript and drafting of most of the figures. Steve Jarose provided help in drafting Figure 5.

Gene Gartland gave valuable field assistance, especially in Ontario, Canada.

Mark Domagala showed me outcrops in the 'jungle' near Oakfield, New York that I would never have found. These are important in showing the continuity of the Bertie Group in this area. I found the Onondaga Ls. to be in contact with the Williamsville Fm. in this area.

I thank those who have given me encouragement through the past several years and with whom I have had the benefit of many interesting discussions, especially the late Erik N. Kjellesvig-Waering.

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FIELD TRIP

The following stops are planned for the field trip but are subject to change if permission cannot be obtained. Please bring hard hats if you have them.

STOP 1 Dunnville Rock Products, Ltd.

Quarry at Byng, Ontario, Canada. This quarry produces "crushed aggregates-concrete stone, Armour rock for shoreline protection, clay fill".

STOP 2

Port Colborne Quarries, Ltd., Quarze at Port Colborne just north of Route 3.

STOP 3

Ellicott Creek at Williamsville, just north of NY 5 (called Glen Park).

GLACIAL GEOLOGY OF THE ERIE LOWLAND AND ADJOINING ALLEGHENY PLATEAU, WESTERN NEW YORK

PARKER E. CALKIN - Department of Geological Sciences, State University of New York, Buffalo, NY 14226

GEOMORPHIC SETTING

The Allegheny section of the Appalachian Plateau (Allegheny Plateau) in western New York may be divided into three physiographic areas; from the Pennsylvania border northward, these include: 1) the high and rugged, unglaciated Salamanca Re-entrant, south of the Allegheny River; 2) the glaciated southern New York Uplands with rounded summits and a network of "through valleys" and breached drainage divides (Cole, 1941; Muller, 1963); and 3) the Erie County portion north of the east-west Cattaraugus Valley. This area is furrowed by deep parallel north-trending troughs separated by broad interfluves and strongly developed to accommodate glacier flow (Donahue, 1972, Calkin and Muller, 1980). Bordering on the west and north, respectively, are the Erie and Ontario lowlands blanketed by glaciolacustrine and ice contact drift, and traversed by subdued, waterlaid end moraines. The Erie County portion of the plateau and the Lake Erie Lowland are the areas spanned by the accompanying field trip log (Fig. 1).

At the close of Paleozoic sedimentation in the Appalachian geosyncline (see Frontispiece I) and prior to glaciation, the area underwent epeirogenic uplift and gentle southward tilting of about 8 m km^{-1} (40 ft mi^{-1}. Initial southward consequent drainage on this surface was eventually reversed to a northwesterly obsequent system and episodic uplift resulted in deep entrenchment of these north-flowing rivers (Calkin and Muller, 1980). The most prominent drainage lines headed south of the present Cattaraugus Valley and included from west to east: 1) the preglacial Allegheny, flowing westward past the Salamanca Re-entrant and northward through Gowanda to Lake Erie along the path of the present Conewango and lower Cattaraugus valley (Ellis, 1980; Frontispiece II); 2) the Connisarauley and 3) the Buttermilk rivers (LaFleur, 1979; D. Hodge, personal communication, 1980) which flowed northward along the present paths of South and North Branches, respectively, of Eighteenmile Creek (Fairchild, 1932); and 4) the Preglacial Cazenovia River (Calkin and others, 1974) which extended 70 km (43 mi) northward from Ischua near the east edge of the Re-entrant, through the present East Branch Cazenovia Creek and East Aurora toward Buffalo (Frontispiece II).

GLACIATION OF THE ALLEGHENY PLATEAU

The Southern Uplands Area and Pre-Lake Escarpment Glaciations

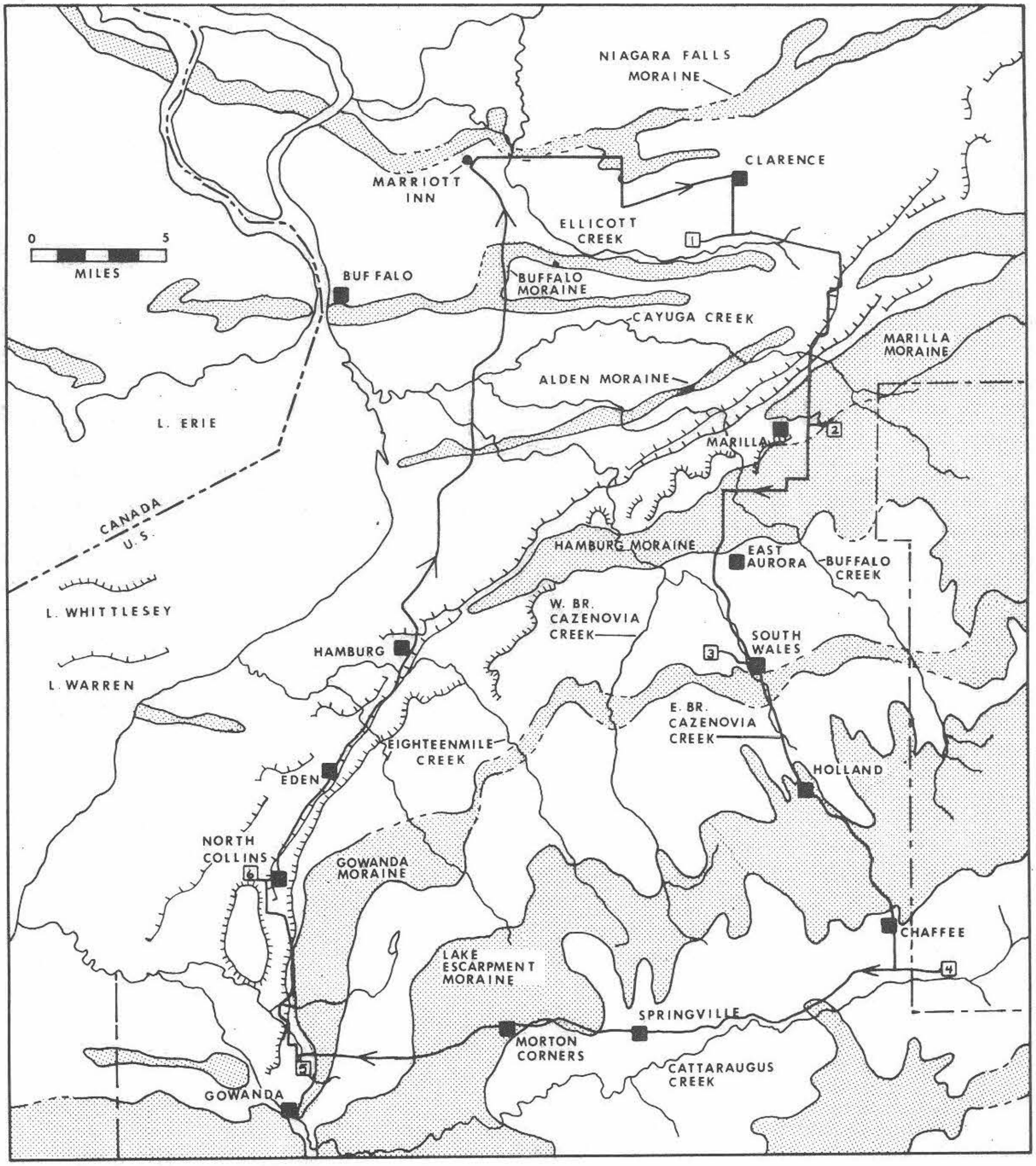

Figure 1. Map of Erie County, New York showing major end moraines, strand lines of the major Glacial Great Lakes, and the field trip route with stops.

Ice sheets moving through the Erie and Ontario troughs abutted repeatedly against the Salamanca Re-entrant (MacClintock and Apfel, 1944; Frontispiece II). Weathered till and gravel spotted around the margins may be of Illinoian age and represent the only pre-Wisconsin drift exposed in New York. However, several events suggest a longer history of multiple glaciation. One of the more interesting is the impondment and diversion of the Allegheny River to its present southerly path to the Ohio River (Muller, 1975; Philbrick, 1976). A second major change in topography appears to be the product of ice marginal superposition and entrenchment which allowed Cattaraugus Creek (Frontispiece II and Fig. 1) to cut westward to Lake Erie across the Tertiary north-trending valley system (Fairchild, 1932; Calkin and others, 1974).

Several drift sheets have been distinguished across western New York (Fig. 2; Muller, 1975, 1977a; LaFleur, 1979, 1980; Calkin and others, 1982). Frontispiece II, at the beginning of this guidebook, shows the major end moraines distinguished. The oldest drift ends southward at the Olean Moraine, the Wisconsin terminal along the northeastern side of the Salamanca Re-entrant. MacClintock and Apfel (1944) considered this end moraine to be more deeply weathered and topographically subdued than the terminal moraine on the northwestward flank of the Re-entrant which is assigned to the younger Kent Drift. The drab, Olean Drift is presently correlated with a Middle Wisconsin advance (Muller, 1977a; Calkin and others, 1982) based on subsurface stratigraphic relations (see below) tied to events near Titusville, Pennsylvania (Chapman and Craft, 1976). Other geologists tracing the Wisconsin drift border northward from Pennsylvania have correlated the Olean with the Late Wisconsin (Crowl, 1980).

The Kent Moraine and associated drift to the north and west displays prominent topography with relatively fresh unweathered deposits often characterized by large percentages of far-traveled stones in the lowlands (LaFleur, 1980). Far-traveled (exotic) stones typically include crystalline rocks transported from the Canadian Shield and tough red, green or gray sandstones and carbonate rocks derived from the Niagara escarpment area. Gadd (1981) has recognized purple-weathered anorthosite boulders in the Cattaraugus Basin that appear to be derived from areas northeast of Montreal. The Kent Moraine marks the maximum advance of the Late Wisconsin (Muller, 1977a) and its age can be bracketed by radiocarbon dates. The advancing ice margin had not crossed the buried St. Davids Gorge of the Niagara by 22,800 yr B.P. (Hobson and Terasmae, 1969; Calkin and Wilkinson, 1982, this volume) nor Rush Creek in the Genesee Basin at 25,300 yr B.P. (Muller and others, 1975) but had reached its maximum throughout the Erie Basin by 20,000 yr ago (Dreimanis and Goldthwait, 1973).

Multiple till exposures and subsurface data at the Gowanda Hospital and at Otto interstadial sites in southwestern New York (Frontispiece II and Figs. 1-3) provide evidence of extensive Early Wisconsin glaciation followed in Middle Wisconsin time by a long, cool interstadial before the Kent glaciation (Muller, 1964; Calkin and others, 1982). At Gowanda

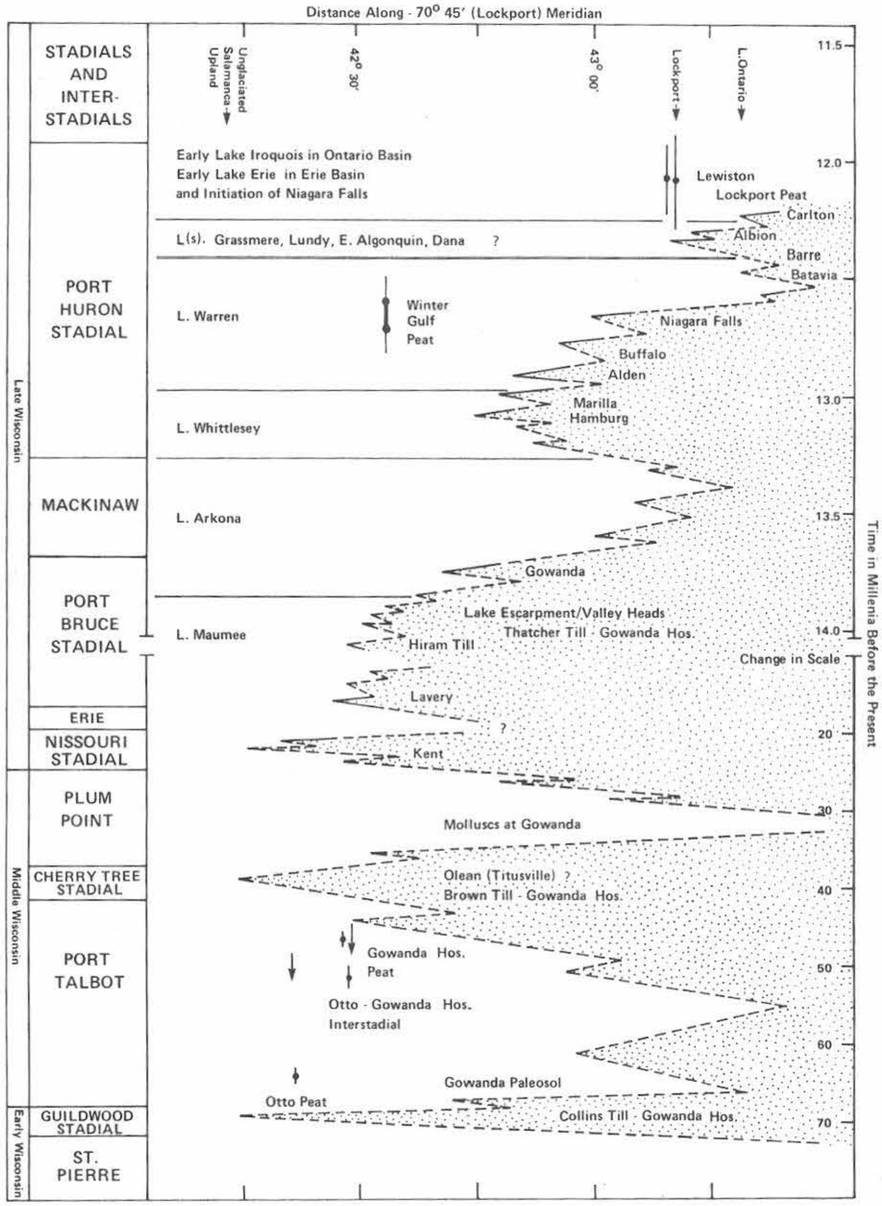

Figure 2. A working chronology of Wisconsin glacial events in western New York. Modified after Muller (1977a).

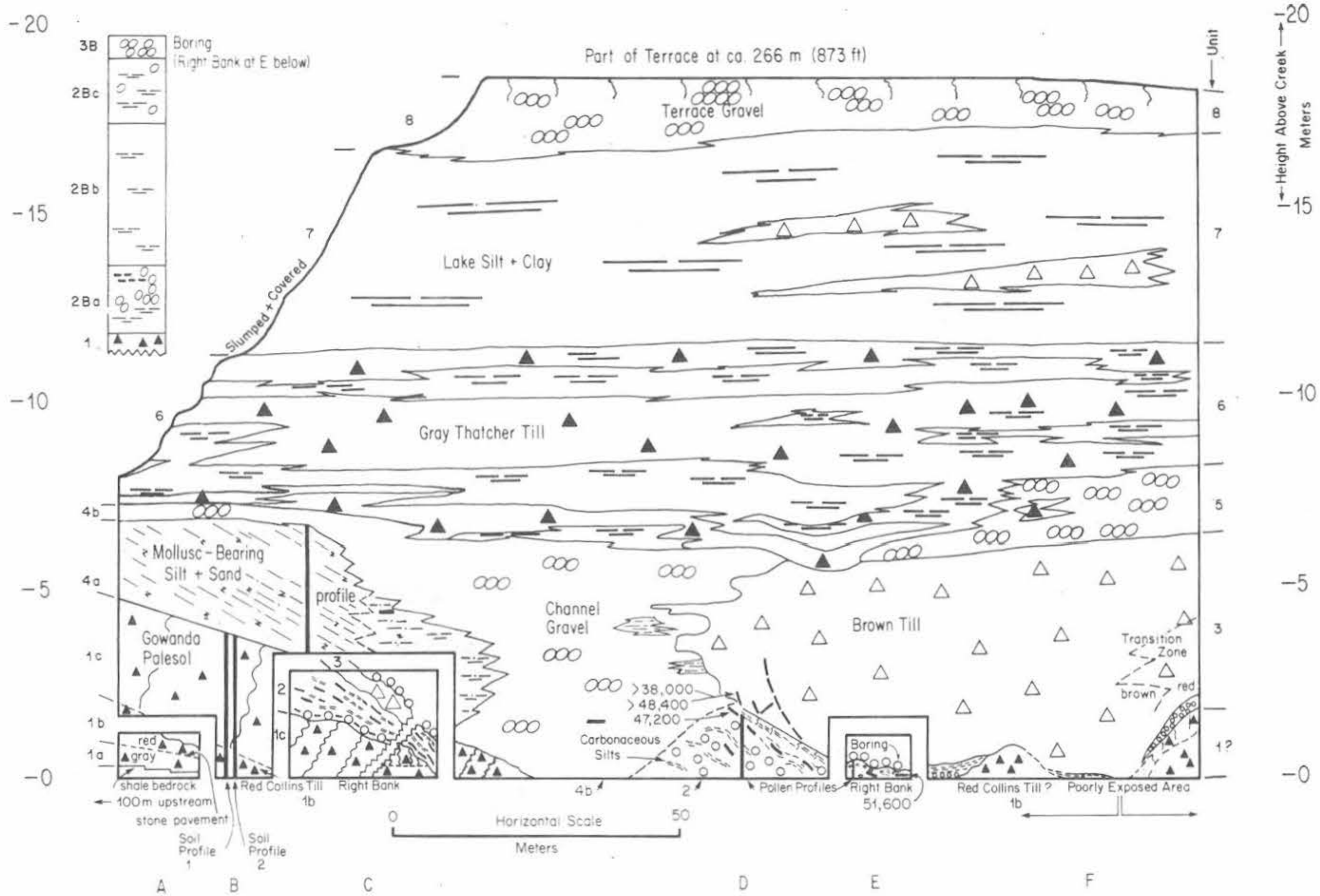

Figure 3a. Composite stratigraphic section along right and left banks of Clear Creek at the Gowanda Hospital Interstadial Site, Erie County, N.Y. From Calkin and others (1982).

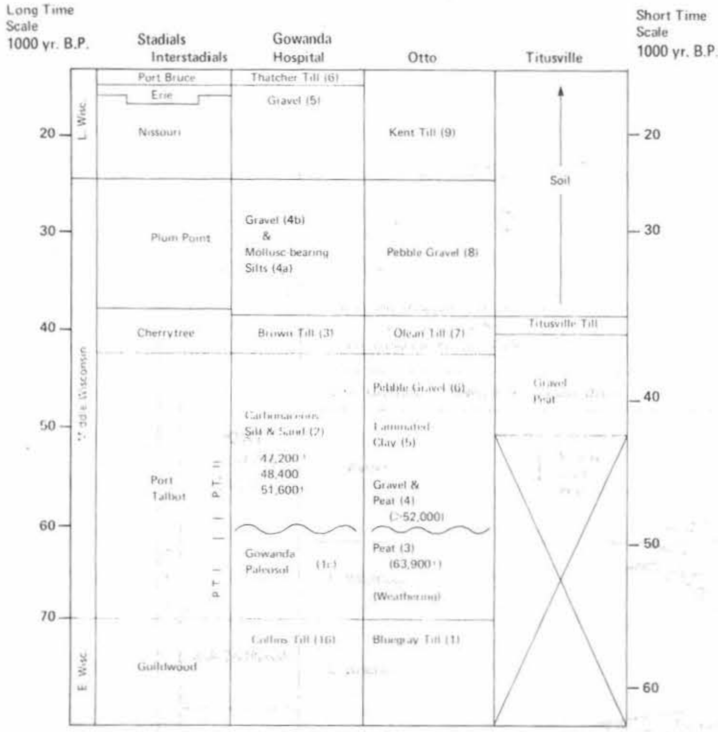

Figure 3b. Correlation of stratigraphic sections at the Gowanda Hospital and Otto interstadial sites, N.Y., and a section at Titusville, Pennsylvania. Modified after Calkin and others (1982).

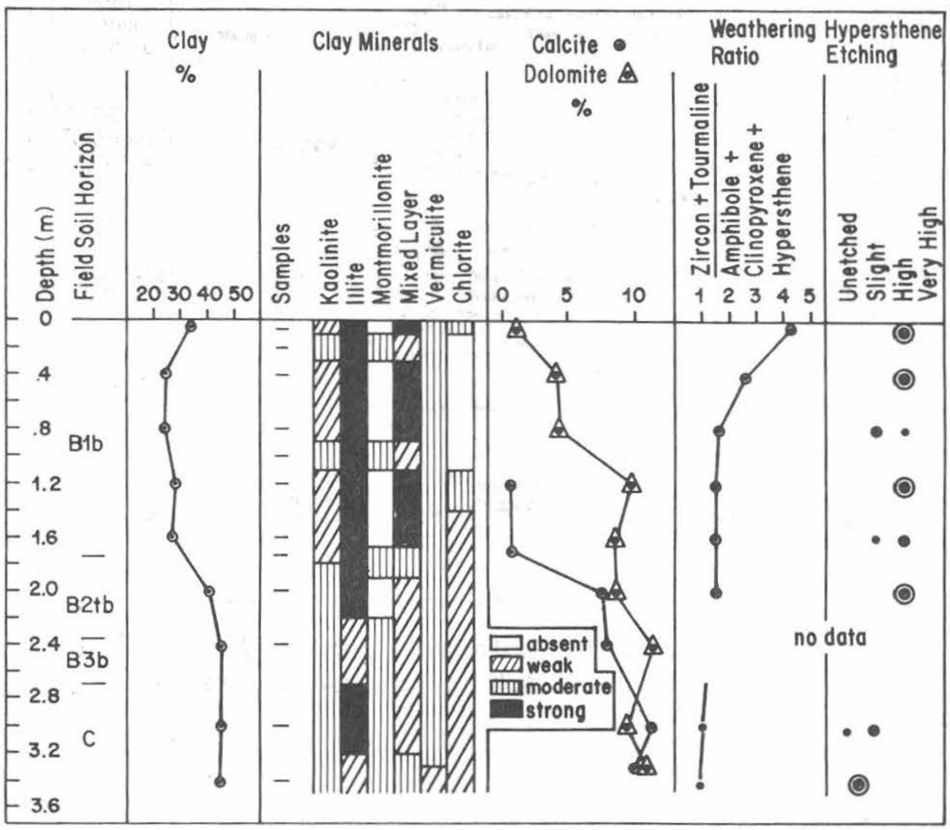

Figure 3c. Properties of the Gowanda Paleosol with: clay % < 2000 μm; clay mineral variation with depth expressed as indication of mineral presence; calcite and dolomite % of < 63 μm fraction; heavy mineral ratio for indicated depths relative to that for lowest two samples (3.0 and 3.4 m); and etching of three grains per sample. From Calkin and others (1982).

Hospital (Fig. 3), a red till (named the Collins Till) derived from a southwestward-moving ice lobe, bears a deep soil profile (Gowanda Paleosol). This is overlain in turn by gravelly organic silt, a brown basal till which incorporates some of the silt, and gravel bearing a terrestrial mollusc assemblage indicative of cold forest-tundra conditions and free drainage. The organic silt carries a spruce-rich pollen spectrum and wood radiocarbon-dated to a probable finite age of 51,000 yr B.P. This sequence is correlated (Fig. 3b) with a similar one at the Otto site (Muller, 1964) and with the eastern Great Lakes glacial chronology on the basis of ^{14}C dates and pollen data (Calkin and others, 1982).

Following the Kent glaciation and during the succeeding interval correlated with the Erie Interstade, the ice front retreated from southwesternmost New York and possibly northward into the Ontario Basin allowing formation of Lake Leverett in the Erie Basin (Morner and Dreimanis, 1973; Fullerton, 1980, note 29). LaFleur (1979, 1980) has described subaerial erosion surfaces and stream gravels in the upper Cattaraugus Valley that seem to support this retreat as well as lowering of baselevel to near that of the present Lake Erie.

Clayey tills which overlie the "typical" Kent drift along the Plateau margin in Chautauqua County (Muller, 1963) as well as farther east in the Cattaraugus Basin (Fig. 4), have tentatively been correlated with the Lavery and succeeding Hiram drifts of Ohio and Pennsylvania (Fig. 2; Muller, 1963, 1975, 1977a). These represent major glacial readvance with uptake of clay from proglacial lakes developed in the Erie Lowland or north-draining plateau troughs during the preceding retreat. The thick Lavery tills of LaFleur (1979, 1980) form the burial medium at the former nuclear fuels reprocessing plant in West Valley.

The Lake Escarpment Glaciation

South of the Village of Gowanda, and along the drainage divide north of Cattaraugus Creek are the very massive gravelly ridges, of Leverett's (1902) Lake Escarpment Moraine System (Frontispiece II, Fig. 1). These mark an oscillating stand of the ice margin behind the position of the Lavery and Hiram moraines and resulted in deposition of the Thatcher Till of western New York (Calkin and others, 1982). The Lake Escarpment Moraine is referred to as the Ashtabula Moraine in western Pennsylvania and Ohio (Muller, 1963; White and others, 1968) and is correlative with the equally massive Valley Heads Moraine System in central New York (Muller, 1977a, 1977b).

During the Lake Escarpment glaciation, pitted outwash plains built southward into the Cattaraugus Basin, over valley fills that are up to 200 m thick in the northeast-trending (buried pre-glacial) valleys at Chaffee and Springville. The outwash was graded to a series of proglacial lakes (Fairchild, 1932; Calkin and Miller, 1977; Calkin and McAndrew, 1980) that drained west along the ice margin into the Coneweango Valley (buried preglacial Allegheny) and thence to the Allegheny River.

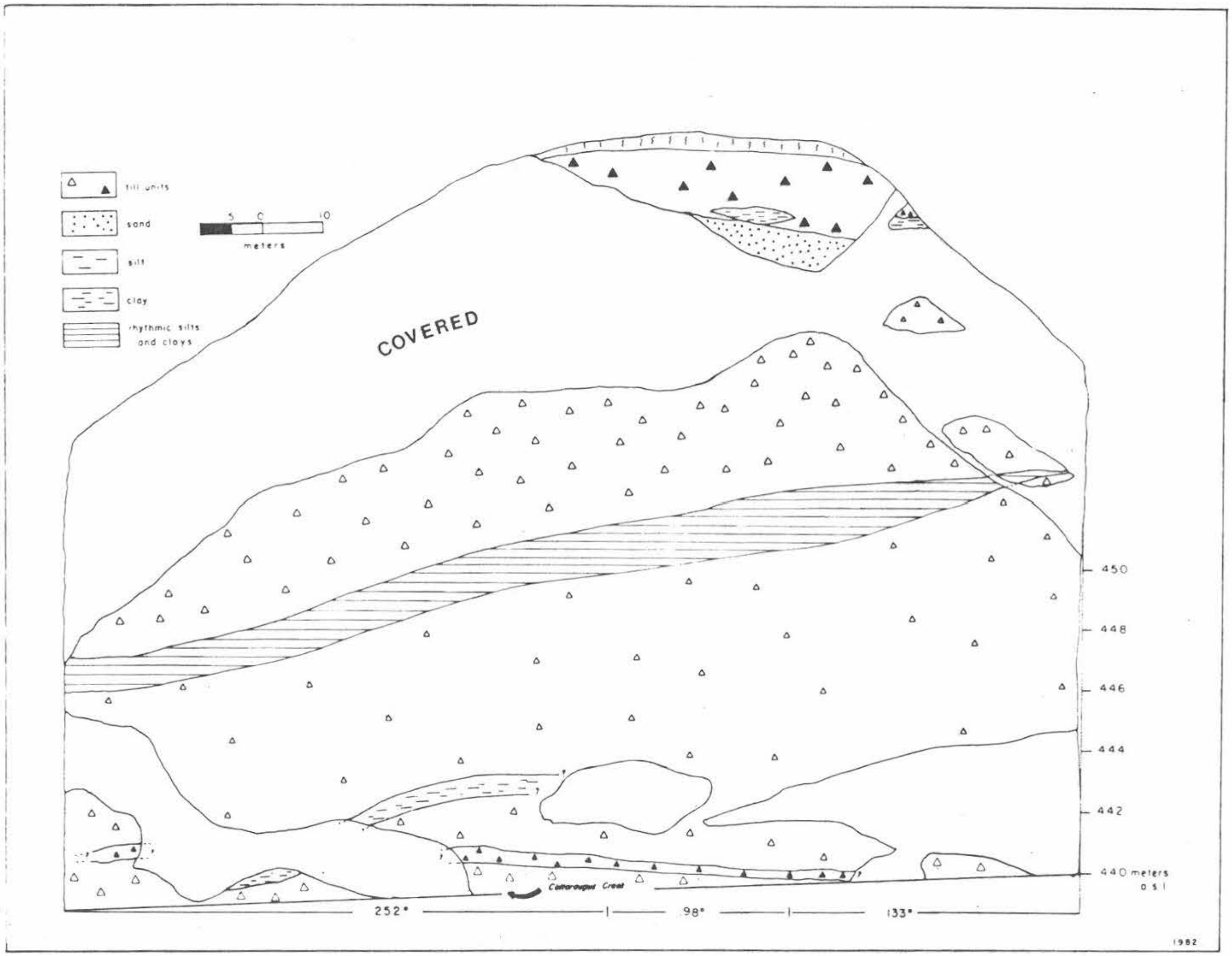

Figure 4. Stratigraphic section at the Arcade Site on Cattaraugus Creek. By Brenda Gagné, 1982).

Peat deposited directly on outwash at a mastodon locality along Nichols Brooks near Chaffee (Frontispiece II and Fig. 1) was considered to closely post date ice retreat when a basal peat date of 14,900 yr B.P. was obtained (Calkin and McAndrews, 1980). However, more recent resampling and study seems to confirm earlier suspicions that the dated peat included recycled (old) carbon and was too old by about 2000 yr (A. Morgan, pers. comm., 1982). Nevertheless, retreat on the order of 14,000 yr B.P. would be compatible with radiocarbon dates obtained from just south of the Valley Heads Moraine System to the east (Coates and others, 1971; Brennen, personal communication, 1981). The pollen at the Nichols Brooks Site indicated an open boreal spruce woodland vegetation without tundra; Morgan and Morgan (1977) have reported permafrost conditions recorded in adjacent Ontario at about 14,000 yr B.P.

A succession of short-lived proglacial lakes formed partly in the valleys trending northwestward from Chaffee, Springville, and Morton Corners. Initially, these drained southward cutting channels through the Lake Escarpment ridges and outwash into proglacial lakes of the Cattaraugus Valley. However, with continued ice retreat, these finger lakes spilled westward across the interfluves, cutting channels and building massive kame and lacustrine deposits (Owens and others, 1972; Hollands, 1975; Pryor, 1975) en route to a concurrently expanding glacial Great Lake in the Erie Basin (Calkin and Miller, 1977).

A longer pause in retreat and short readvance following the strong Lake Escarpment oscillation, is marked by the Gowanda Moraine (Fig. 1), a low but distinct ridge apparently tied closely to the Lake Escarpment ridges which extend between South Wales, North Collins, and the Lake Erie coast at Dunkirk (Frontispiece II). Unpublished geophysical exploratory data along the Lake Erie coast suggest possible correlation of the Gowanda with the Erie-Long Point (Norfolk) Moraine across Lake Erie. Retreat across western New York at this time is correlated with the Port Bruce Stadial or MacKinaw Interstadial on the basis of continuity of events across the eastern Great Lakes (Calkin and Miller, 1977; Fullerton, 1980). The marginal drainage was probably into Glacial Lake Arkona (Fig. 2) although no definitive Arkona strand lines have yet been traced north of Girard, Pennsylvania (Calkin, 1970; Schooler, 1974; Fullerton, 1980). The ice margin continued retreat north and eastward out of the Erie into the Ontario Basin and for a period of a few hundred years or less, lake levels fell considerably below the Arkona level, possibly allowing waters to drain northward across the Niagara Escarpment and eastward from the Ontario Basin through the Syracuse channels (Wall, 1968; Fullerton, 1980).

GLACIATION OF THE ERIE LOWLAND

Glacial Lakes and Ice Margins

At least ten stages of proglacial Great Lakes may be recorded in the New York portion of the Erie Basin during Late Wisconsin time (Fig. 2); however, only the highest, Lake Whittlesey, and two stages of the succeeding Lake Warren produced strand lines strong and continuous enough to be traced south or westward out of New York to type areas. Both Whittlesey and Warren Lakes stretched up to 480 km (298 mi) southwest from Buffalo to beyond Toledo, Ohio. These are believed to have been controlled by west-draining outlets through Michigan (Hough, 1963, 1966; Calkin, 1970; Muller, 1977a).

The Mackinaw Interstadial was terminated by the Port Huron advance in Michigan and in western New York, inducing a rise in glacial waters through at least 12 m (40 ft) to the Glacial Lake Whittlesey level in the Erie-Huron basins and the resubmergence of the Erie Lowland in western New York. A unit of glacial varved clay and thick superimposed basal till traced through several exposures along 32 km (20 mi) of Lake Erie bluff between South Buffalo and the Hamburg Moraine may record this advance. These relations and those elsewhere including at the Winter Gulf organic site (Fig. 5; Calkin, 1970; Calkin and McAndrews, 1980) suggest that the ice margin reached to, if not locally beyond, the massive Marilla Moraine or adjacent closely associated Hamburg Moraine (Frontispiece II; Fig. 1). If it reached much farther and into the Plateau to build or override the Gowanda Moraine (Taylor, 1939; Calkin and Miller, 1977) the ice must have been thin and the advance short-lived before recession to construct the Hamburg Moraine. Fullerton (1980) has linked the Port Huron advance with the Alden Moraine in New York but local field relations as presently interpreted do not appear to support this correlation. In southern Ontario, the Port Huron advance has been correlated with the margin of the Halton Till (Barnett, 1979; Feenstra, 1981) which occurs 20 to 30 km (12.4 to 18.6 mi) beyond the position of the submerged Port Maitland Moraine.

The strand of Glacial Lake Whittlesey is the strongest and most continuous of those in New York; it occurs principally as a single distinct gravel storm beach ridge of $\sim$6 m (20 ft) relief. However, pronounced beaches cut in bedrock are associated with the Whittlesey beach ridge in the villages of North Collins, Eden, and Hamburg (Calkin, 1970). The strength of the Whittlesey strand line must be a consequence of formation by rising waters and its location near a plentiful gravel supply at the plateau margin as well as the duration of the lake stand. The Lake Whittlesey strand line reaches its northern and easternmost expression near Marilla where weak, wave-cut features at an elevation of 277 m (910 ft) occur on the north flanks of the Hamburg Moraine. This and other relations (Calkin, 1970; Muller, 1977b) suggest that the Marilla Moraine was formed in Lake Whittlesey; however, lowering of Lake Whittlesey waters to Lake Warren I must have occurred during or soon after retreat from the Marilla Moraine (Calkin, 1970).

The strong development of beaches of all stages of Lake Warren on the north face of the Marilla Moraine indicate that at least 5 to 10 km of retreat occurred before renewed downcutting of Lake Warren's Grand River outlet in Michigan caused water levels to fall to the Lake Warren II level. Warren I and II strand lines as distinguished in western New York terminate in spits in the delta of Ellicott Creek, 5 km northeast of Alden and just south of the Alden Moraine (Fig. 1). A drop in lake

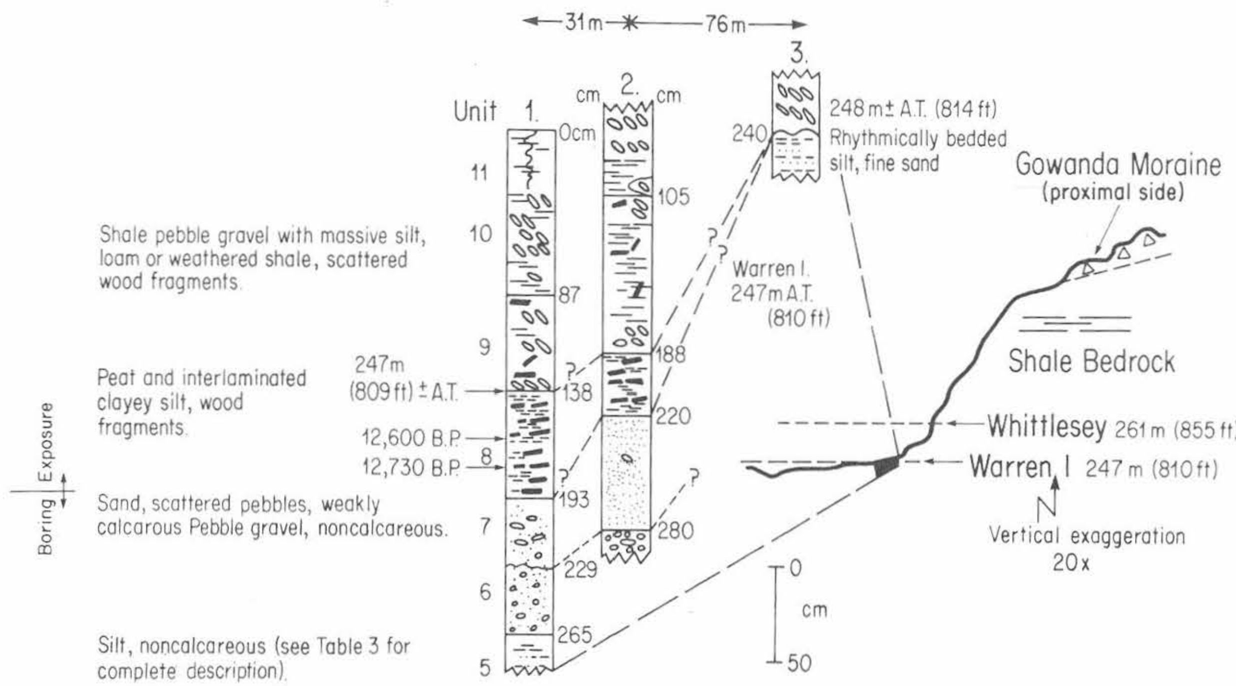

Figure 5. Stratigraphic section at the Winter Gulf Site, North Collins, New York. Units not shown include: 5, 55 cm noncalcareous silt; 6, 244 cm calcareous silt and interbedded sand; 3, 91 cm stony silt and bedded silt; 2, 1250 cm gray, sand-silt-clay till and interbedded stratified silt (Thatcher Till?); and underlying compact, gray-brown, silty sand till. After Calkin and McAndrews (1980).

level on the order of 5 m (16 ft) from the Warren II level corresponding to Lake Wayne of Michigan may have followed as the ice margin backed northward of Buffalo. Fullerton (1980) has suggested that active ice had withdrawn from the Batavia re-entrant, Syracuse channels, and Mohawk Lowland to allow eastward drainage at this time. Assignment of ridges to the Lake Wayne stage is very tenuous at best and field evidence has not corroborated or denied eastward drainage.

Subsequently, water levels rose and drainage returned clearly to the west again with the production of very strong ridges assigned to the Warren III (lowest Warren) stage. Contemporaneously, a readvance of at least 8 km overrode thick ice-contact lacustrine sediments in the Clarence area and deposited thick red basal till possibly culminating in formation of the Alden Moraine just south and west of Buffalo (Fig. 6; Calkin and Miller, 1977; Fullerton, 1980, p. 25). Glacial Lake Warren III persisted through recession from the Alden Moraine position, development of the succeeding Buffalo (= Fort Erie Moraine), Niagara Falls and Batavia moraines. The Batavia Moraine formed after a period of rapid reorganization of the ice front and southwestward readvance to truncate the Niagara Falls Moraine (Leverett, 1902). For the brief period following retreat from the Batavia Moraine, Lake Warren III drained eastward through the Syracuse channels to the Mohawk Valley (Muller, 1977b).

Extensive gravelly kame deltas formed in many areas of Lake Warren, including the distal margins of the Buffalo Moraine in downtown Buffalo and Clarence. Cross bedding, climbing ripples, and similar primary structures in these deposits record the strength and density of depositional currents moving off the ice margin into >60 m (197 ft) of water. The ubiquitous silt and clay blanketing most of the Erie Lowland must reflect the periods of lake ice cover that prevailed through a substantial, colder, part of the glacial lake year in this area.

The Glacial Lake Warren beaches are built out farther from the plateau margin than those of Lake Whittlesey and are best developed on broad deltaic deposits that occur along major stream mouths such as occur near Hamburg, Alden and Crittenden in Erie County (Calkin, 1970). In shallow shelf areas such as Brant and Eden in southwestern Erie County, the deltaic sands were spread out over extensive areas by waves. The Whittlesey and Warren beaches may be somewhat better developed in New York than Ontario (MacLachlan, 1938) and this observation with the orientation of spits and bars, suggests that westerly winds prevailed during strand formation.

Isostatic rebound has caused the Whittlesey and Warren strand lines to rise northeastward through New York at a gradient of about 0.5 m km^{-1} ($\sim$2.5 ft mi^{-1}) (Fig. 7). Approximately 9 m (30 ft) of uplift occurred between Whittlesey and Warren III time at Buffalo near the mouth of the Niagara River (and threshold of Lake Erie). Fifty-two meters (172 ft) of uplift has occurred here since the lowering of Lake Whittlesey (Calkin, 1970).

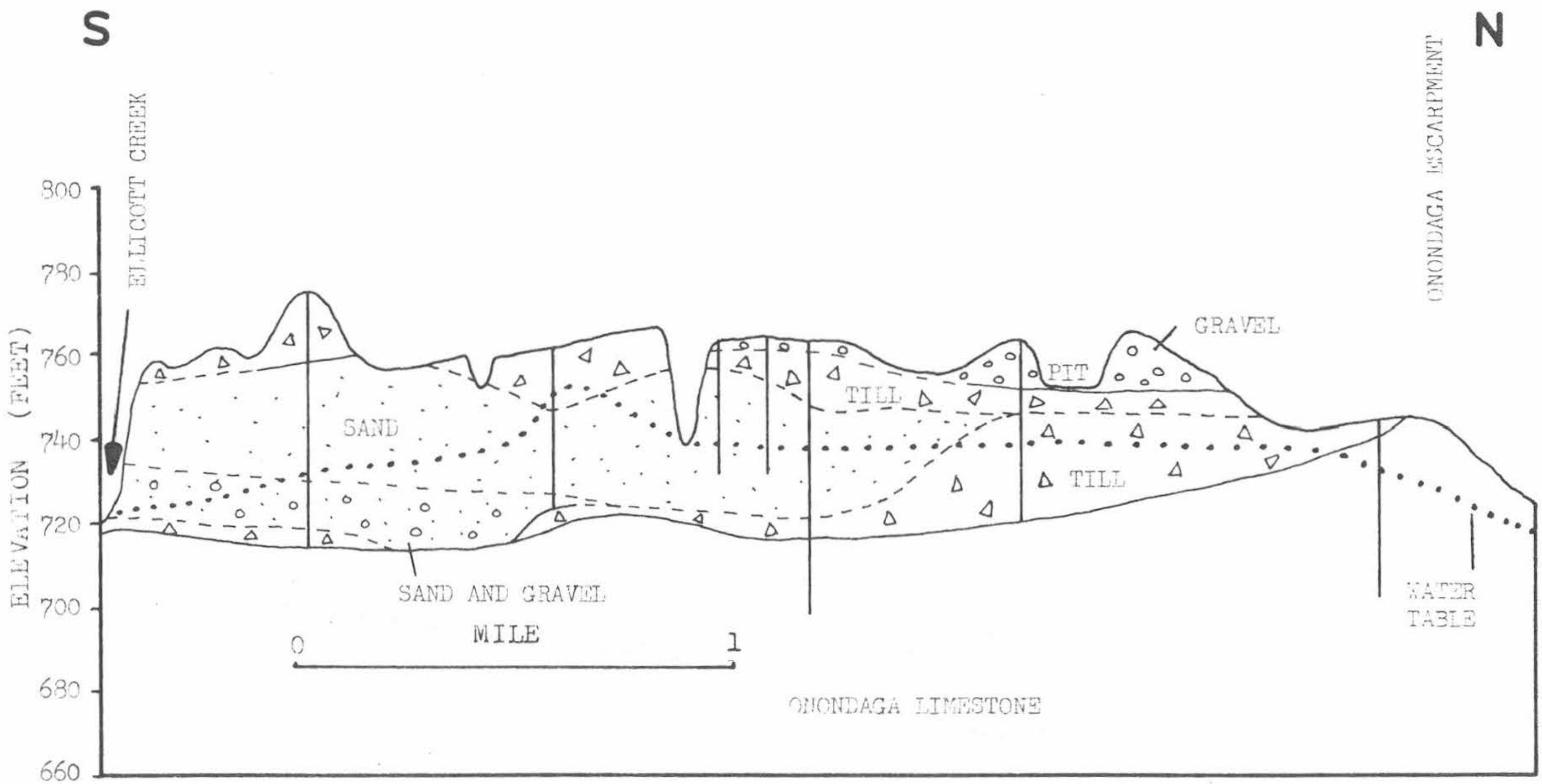

Figure 6. North-south stratigraphic section through the Buffalo Moraine area between the Onondaga Limestone Escarpment (N.Y. Route 5) Clarence, and Ellicott Creek, Town of Lancaster, New York. Anonymous (1982).

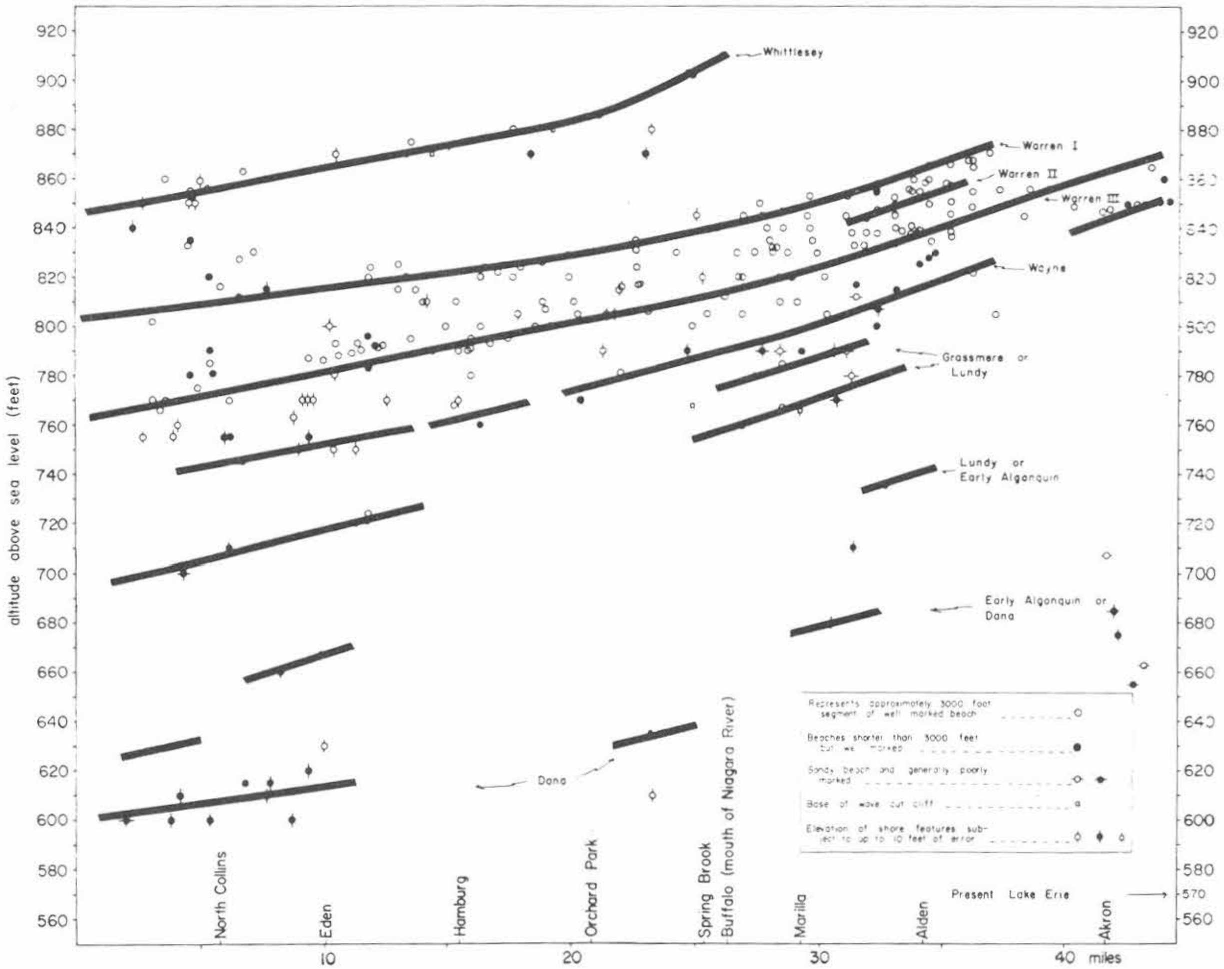

Figure 7. Tilted strand lines of the Glacial Great Lakes, Erie County, N.Y. Elevations from beach crests are projected to a line oriented N 24°E between Cattaraugus Creek at Vessailles and the town of Indian Falls. Heavy lines drawn by hand to suggest water planes have little control below Warren III level. From Calkin (1970, Fig. 4).

The beaches below those of Warren III in the New York portion of the Erie Basin are very discontinuous and display less than 3 m (10 ft) of relief. This is partly due to rapid isostatic uplift, but is also related to the limited beach materials on the clay-covered lake plain where the lakes shoaled and the short duration of each successive lake stand. In order of formation and decreasing elevation, the scattered beach segments have been correlated tentatively with glacial lakes Grassmere, Lundy, Early Algonquin, and Dana (Fig. 2; Hough, 1963, 1966; Calkin, 1970). Retreat from the Batavia re-entrant may have allowed eastward drainage of Lakes Grassmere and Lundy (Fullerton, 1980) but Early Algonquin if once existent in this area, was controlled by outlets to the west (Hough, 1966; Calkin, 1970).

During the post-Warren III lake stages the ice margin retreated at least 30 km northward, with short stands to form the Barre Moraine just above the Niagara Escarpment, and the Albion-Rochester Moraine locally just below it in Niagara County (Fig. 1). Lake Dana, which was confined to the northeastern end of the Erie Basin and southwesternmost portion of the Ontario Basin, drained eastward through the Syracuse channels to the Mohawk and Hudson Valleys following retreat from the Albion-Rochester Moraine (Calkin, 1970). It represented the slowly subsiding waters immediately preceding emergence of the Niagara Escarpment, nearly simultaneous formation of nonglacial, Early Lake Erie (Lewis and others, 1969) and development of Glacial Lake Iroquois in the Ontario Basin, and the initiation of Niagara Falls and Gorge (Calkin and Brett, 1978). A short readvance during initial stages of Glacial Lake Iroquois is marked by the Carelton Moraine near the present south coast of Lake Ontario.

Dating and Climatic Environment

The last glacial retreat and succession of glacial lakes across the lowland must have taken less than 1000 years. The Port Huron advance of Michigan and the rise to Lake Whittlesey level seem to be well dated at about 13,000 yr B.P. although data are not yet locally available to refine this further (Dreimanis and Goldthwait, 1973; Fullerton, 1980). The oldest date for initiation of Early Lake Erie is 12,650 ± 170 yr B.P. (Lewis, 1969) and for Lake Iroquois, 12,660 ± 400 yr B.P.; however, both samples may have been contaminated with recycled carbon and are too old (Calkin and Brett, 1978; Calkin and McAndrews, 1980). Other radiocarbon dates from the Lake Ontario and St. Lawrence areas suggest that Early Lake Erie and Lake Iroquois formed prior to about 12,200 yr B.P. (see Fullerton, 1980).

At the Winter Gulf site south of Buffalo (Figs. 1 and 5), dates of 12,730 ± 220 and 12,610 ± 200 yr B.P. were obtained on wood from a shallow water peat within the lake plain but below the Lake Whittlesey strand level. The peat is in turn separated by 4 to 5 m of lake sediments from an underlying till correlated with the Thatcher Till of the Lake Escarpment glaciation (Calkin and others, 1982). The wood dates are therefore minima for the last glacial retreat from this area, for lowering of Glacial Lake Whittlesey, and for northward retreat of the ice margin from the Hamburg or Marilla moraines in western New York (Calkin and McAndrews, 1980). The ice margin could have been as close as 21 km but was likely near the present coast of Lake Ontario, 80 km to the north when peat deposition began at Winter Gulf.

The pollen profile from both Winter Gulf and Nichols Brooks are dominated by spruce and other pollen which suggest that there was little warming at this time; however, an ecologically unbalanced biota may have occurred. The coleoptera (beetle) assemblage recovered from the peat at Winter Gulf (Schwert and Morgan, 1980) indicates that temperatures of deposition were considerably warmer than those suggested by the pollen. If such warming did occur, it records a rather sudden change from conditions of permafrost that may have existed in southwestern Ontario and nearby New York until the early phase of the Port Huron advance (Calkin and McAndrews, 1980, p. 305; Schwert and Morgan, 1980).

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ROAD LOG
GLACIAL GEOLOGY OF THE ERIE LOWLAND AND ADJOINING ALLEGHENY PLATEAU, WESTERN NEW YORK

(Route and location of stops are shown in Figure 1)

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE AND STOP DESCRIPTION
0.0	0.0	From Buffalo Marriott Inn turn left (north) onto Millersport Highway (Rt. 263 N).
0.2	0.2	Turn right, proceed east on Maple Rd.
2.9	0.7	Cross Hopkins Rd. with slight rise of Niagara Falls morainal ridge on right, Lake Tonawanda plain to left.

6.2	3.3	Turn right (south) onto Harris Hill Rd. with subdued hillocks of Niagara Falls Moraine partly mantled by Lake Warren silts. Rise up Onondaga Limestone Escarpment.
7.8	1.6	Turn left and proceed east on Rt. 5 (Main St.). Sandy soils on Escarpment edge here result from winnowing of older sediments by Lake Dana waters.
12.1	4.3	Descend Escarpment into river embayment at Clarence Village. Recent U.S. Geological Survey borings suggest that a large buried (drift-filled) valley connects the embayment southward with the present valley of Ellicott Creek near Stop 1.
12.5	0.4	Turn right onto Ransom Rd. Rise south back up Onondaga Escarpment onto gravel plain related to Buffalo Moraine and deposited in Lake Warren III. Beaches of lower stage--Lake Grassmere occur along Escarpment crest.
		Cross over N.Y.S. Thruway into Town of Lancaster.
15.0	2.5	Turn right (west) at traffic light onto Genesee St. (Rt. 33).
16.0	1.0	Turn left off Genesee St. into lot just east of Shisler Rd. intersection.

STOP 1. PINE HILL GRAVEL PIT.

Pit is cut into a portion of the Buffalo Moraine and is just on north margin of present northwest-flowing Ellicott Creek. Borings made for the study of the gravel aquifer (Fig. 6) in this area show two till sheets with sand and gravel between. This and several adjoining excavations expose the thick sand and gravel units deposited in at least 30 m (100 ft) of Lake Warren III or earlier stage. Thin red lake clay partings separate thick rippled sand units (varves?). The overlying till and sand/gravel unit shows evidence of flow and deformation (often irregularly interfingering). The till, laid down by thin, sometimes bouyant ice margin, can be traced at least to the Alden Moraine ∿5 mi to the south. Fabrics show moderate NE-SW maxima.

Return eastward on Genesee (Rt. 33E) through traffic light, through Millgrove to Walden Ave.

20.5	4.5	Turn left onto Walden and then immediately right (south) off Walden bending again left along crest of subdued sandy beach or off-shore bar of Lake Warren III.
20.9	0.4	Cross R.R. tracks on beach crest.
21.6	0.7	Bend right on Peters Corners Rd. onto strong Warren III beach ridge. This strand can be traced intermittently to Syracuse and south and west through Cleveland into Michigan.
22.1	0.5	Pit on right excavated in beach ridge. Cross several beach ridges obliquely.
22.6	0.5	Cross R.R. bed cut in Alden Moraine at left (source of beach material) and turn right onto North St. following a Warren III beach (moraine on left).
23.5	0.9	Bear left across Ellicott Creek (flows through U.B. campus) onto Sandridge Rd. and continue south.
24.4	0.9	Cross highest beach ridge of several oblique to Sandridge Rd.
25.2	0.8	Turn right (west) at W. Alden onto Broadway (Rt. 20) and cross several ridges-spits of Warren III. This is the area where Warren III beaches split off northward and Warren I and II? terminate south of Alden Moraine.
25.7	0.5	Turn left onto Four Rod Rd., still crossing Warren ridges.
26.3	0.6	Descend into floodplain and cross Cayuga Creek.
27.0	0.7	Rise across washed till (strand of Lake Warren I - highest Warren) at white farmhouse (∿860 ft [262 m]) and proceed south across Marilla Moraine. Moraine appears unaltered by Lake Whittlesey waters and hence formed before or soon after draining of Whittlesey to Warren I level.
28.2	1.2	Cross Clinton St. going south, still on Marille Moraine.
29.3	1.1	Turn left onto Bullis Rd. and stop along road near corner.

STOP 2. DRAINAGEWAY MARILLA-HAMBURG MORAINES.

View small esker and associated kames to northeast. To west of intersection, Bullis Rd. crosses delta surface formed in Lake Whittlesey and subsequently Lake Warren by water draining along ice margin here.

		Turn around and return to Four Rod Rd. south.
29.5	0.2	Leave Bullis on Four Rod Rd. south and cross Hamburg Moraine.
30.9	1.4	Not good till cut at far left along Little Buffalo Creek.
31.1	0.2	Moulin kame hillock on right behind farm consists of till and poorly sorted gravel. A few similar forms occur just to east out of sight.
31.6	0.5	Turn right onto Parker Rd., follow west to stop sign (in E-W-oriented meltwater channel.
32.5	0.9	Turn left (south) onto Two Rod Rd. (Rt. 358) crossing meltwater channel.
33.0	0.5	Turn right at riding stable onto Jamison Rd.; cross Buffalo Creek--its post-glacial position?
35.7	2.7	Turn left (south) at light onto Girdle Rd. Many gravel hillocks of moraine have been removed here.
36.3	0.6	Cross Aurora Expressway (Rt. 400); note deep cut in moraine.
37.1	0.8	Deep kettle hole and lake on left; hummocky ice-contact drift at distal side of Hamburg Moraine.
37.8	0.7	Turn right onto East Aurora-Porterville Rd. and immediately right again at bend onto Pine St. to East Aurora Waterworks. This is within partly buried valley of Preglacial Cazenovia River which trends northward through the Hamburg Moraine. Valley fill is major water source and buried valley a collector/conduit of groundwater.
38.7	0.9	Leave Waterworks at stop sign and proceed south on Pine St.

39.1	0.4	Cross Main St., East Aurora Village and Rts. 20A and 78. Proceed straight south on Olean St. (Rt. 16) (continuation of Pine). Note postglacial Cazenovia Creek has been glacially deflected to westward path.
40.9	1.8	Bear right, follow Rts. 16 and 400S. Note glaciated U-shaped valley cross-section modified by alluvial fans at margins.
43.1	2.2	Turn right at South Wales onto Emery Rd. and proceed to Emery Park.

STOP 3. EMERY PARK (Lunch Stop)

The Gowanda Moraine crosses the valley at this point. The park is on the west wall of the partly buried valley ($\sim$60-100 m of drift till) now occupied by East Branch Cazenovia Creek. Ice flow was approximately parallel to this and adjoining valleys. High level deltas graded to proglacial "finger lakes" line the east valley wall in this general area.

		Leave Park and pick up mileage at Rt. 16 intersection. Turn right southward.
48.7	5.6	Cross East Branch Cazenovia Creek. Creek is on bedrock at right, deepest part of valley is to left (east) of bridge. Pass through Village of Holland and rise southward onto proximal side of Lake Escarpment Moraine (bedrock actually fills slightly southward).
54.0	5.3	Turn right onto Hand Rd. at crest of Lake Escarpment Moraine.
54.3	0.3	Turn around at Chaffee solid waste disposal site (landfill). Although tills appear quite thick and dense here, the Chaffee Outwash Plain heads at north margin of the current landfill boundary and critically close to leachate drainage.
54.7	0.4	Right (south) onto Rt. 16. Note meltwater channels on right leading to head of outwash plain within few tenths of a mile.
56.2	1.5	Cross Grove St., Chaffee, on outwash plain. Note irregularly south-sloping surface, probably effected by differential settling. A nearby gas well reached bedrock at about 188 m (615 ft) depth (see Calkin and others, 1974).

57.7	1.5	Nichols Brook site (Calkin and McAndrews, 1980) on left.
58.0	0.3	At intersection with Rt. 39, turn left onto East Schutt St. (becomes Howe Rd.).
58.9	0.9	Cross R.R. track and pass out of Preglacial Cazenovia Valley toward Arcade.
59.9	1.0	Park at right (south side) of road just east of red farmhouse.

STOP 4. ARCADE SECTION.

The stratigraphy at this active cut-bank of Cattaraugus Creek is shown in Figure 4. The locality is south of the Lake Escarpment Moraine. At least three till units are exposed, separated by glacial lake deposits. The upper till is thin and has textural properties strongly influenced by underlying lacustrine deposits and appears to indicate deposition from thin ice. The lower tills are thicker and generally more homogeneous in their properties although the lowest unit carries clasts of red till and clay. The tills are tentatively correlated with Hiram, Lavery, and Kent drifts on the basis of sequence.

		Return westward along Howe and East Schutt Streets.
61.8	1.9	Turn left (south) onto Rt. 16S.
62.1	0.3	Turn left obliquely off Rt. 16 onto dirt road (before descending onto floodplain of Cattaraugus Creek) leading to gravel pit.

HESITATION STOP - GERNATT GRAVEL PIT

Good exposure (when working) of Chaffee Outwash at Cattaraugus Creek where it overlies till (generally not exposed but just above creek level). Near-horizontally bedded and very well sorted gravel overlies steeply inclined (foreset) units which on one visit appeared to be associated with till units.

		Pick up mileage at Rt. 16.
		Proceed north onto Rt. 16.
62.4	0.3	Turn left onto Rt. 39W.
62.8	0.4	Cross bedrock-cored ridge dividing buried valley into two channels. Deepest channel is on west side and connects valley to south of Cattaraugus Creek with present valley of East

		Branch Cazenovia Creek Valley (not with Buffalo Creek [Calkin and others, 1974]).
63.7	0.9	Bear left on 39W toward Springville.
65.9	2.2	Descend from outwash surface into Cattaraugus Creek Valley. Bluemont Ski area to south across creek.
66.6	0.7	Note slumped till of Lavery Till on right, Lord Hill Section of LaFleur (1979, 1980) on left at Creek. Strong fluvial dissection exposes till and lake sediments in the valley area.
70.7	4.1	Rise out of postglacial valley onto outwash tied to Springville Outwash Plain. Lake Escarpment morainal ridges are to north.
71.8	1.1	Cross Rt. 240.
72.9	1.1	At hospital, descend into Village of Springville occupying spillway that drained proglacial lake in North Branch Eighteenmile Creek and West Branch Cazenovia Creek. This is center of Preglacial Buttermilk Creek heading several miles to south.
73.6	0.7	Proceed straight (west) on Rt. 39. (The most scenic route would trend left via Waverly St. and Zoar-Gowanda Rd. but roads are narrow and steep, and bridge weight limits prohibit bus traverse.)
74.5	0.9	Cross Rt. 219 and leave outwash plain crossing onto distal ridge of Lake Escarpment Moraine.
78.8	4.3	Cross onto outwash plain (still on Rt. 39) at head of South Branch Eighteenmile Creek. Formed by meltwater of Lake Escarpment Moraine shed southward into buried Preglacial Connoisarauley River Valley.
80.0	1.2	Rise off outwash onto Lake Escarpment Moraine as it trends north-south.
80.7	0.7	Bear left continuing on Rt. 39.
85.2	4.5	Cross Rt. 75 at Collins Center and onto gravel valley till (underlain by lake clay and till respectively).

88.6	3.4	Cross Gowanda Moraine. Stratigraphic sections suggest moraine formed as ice advanced into proglacial lake.
90.1	1.5	Turn left (south) onto Rt. 62 within Clear Creek Valley (part of Preglacial Allegheny Valley) and park near corner. Reach exposure downstream from highway bridge 0.3 mi south.

STOP 5. GOWANDA HOSPITAL INTERSTADIAL SITE.

This site exposes three till units, a buried paleosol and overlying organic horizon. It is briefly described in the text and Figures 3a, b, and c. The right bank section is rarely exposed but the paleosol and wood horizons on the left bank are often clear after minor work. One of the most important aspects of the site is the Gowanda Paleosol, a buried, truncated (of its A horizon) soil whose substantial development and stratigraphic positioning appears to argue for a long mid-Wisconsin nonglacial interval prior to about 50,000 yr BP (Calkin and others, 1982).

One of the many problems of the stratigraphy is the age and correlation of the brown till unit. The till material appears to be derived locally by glacial erosion of the underlying, weathered red "Collins" till and therefore it is difficult to correlate regionally with other till units. Although it displays mudflow-like contacts with the channel gravel, it also shows good glaciotectonic structures including wood sheared from underlying deposits. The age of the wood suggests a mid-Wisconsin ice advance; however, there is some question whether the ice sheet reached past this area to Titusville, Pennsylvania, in Middle Wisconsin interval as suggested by current correlations (Figs. 2 and 3b).

		Proceed north on Rt. 62 from intersection with Rt. 39.
90.5	0.4	Turn left (west) onto Richardson Rd. rising onto Fairchild's (1932) Asylum (Gowanda State Hospital) Terrace graded to a local proglacial lake in the Cattaraugus Creek (Preglacial Allegheny) embayment of the Erie Basin. This surface is an extension of top gravel unit (8) at Gowanda Hospital site.
91.1	0.6	Turn right (north) onto Taylor Hollow Rd. View at left of Fourmile Level, a delta surface (on Cattaraugus Indian Reservation) formed in Lake Whittlesey and later Lake Warren.
91.4	0.3	Gravel pit at right displays large foreset bedding units in thickened section of Asylum Terrace seen at the Gowanda Hospital site (Fig. 3a, unit 8).

93.5	2.1	Bear left at intersection with Rt. 62 and proceed north with Gowanda Moraine parallel to right.
94.1	0.6	Cross Genesee Rd. at Lawton Corners. Rt. 62 follows along zone of weak strand lines of Lake Whittlesey and Lake Warren I. Lowland on left is an old buried tributary valley of Allegheny River.
95.8	1.7	Cross Winter Gulf (unofficial name) and organic site on left (Schwert and Morgan, 1980; Calkin and McAndrews, 1981). The section here is diagramed in Figure 5.
96.6	0.8	Turn left onto Milestrip Rd. (just south of a "Friends" cemetery built on a delta of Lake Whittlesey). Cross gravelly ridges (before low point) intermediate between Whittlesey and Warren (possibly Arkona?). Note ridge-spit extending from west (on right) graded first to Lake Whittlesey and later modified by Lake Warren waters.
97.6	1.0	Turn right onto Mile Block Rd. Spit visible (again) extending eastward is tied to well-developed Whittlesey beach (visible on left) which completely encloses bedrock-based high area here (plateau outlier).
99.1	1.5	Turn left (west) at intersection onto Rt. 249W. Cemetery at intersection spans a Warren ridge.
99.2	0.1	Turn left into gravel pit of Early Sunrise Construction Co.

STOP 6. NORTH COLLINS, WHITTLESEY BEACH EXPOSURE

The Glacial Lake Whittlesey beach with crest at 850 ft (260 m) here is almost totally composed of shale, eroded from the local bedrock (Calkin, 1970, Fig. 2). The good development of the ridge on this narrow-embayment-side of the island in Lake Whittlesey is of interest. Lake Whittlesey extended southwest beyond Toldeo and into the Huron Basin. Several ridges of Lake Warren occur below the Whittlesey ridge; many have been destroyed by years of farming or construction.

Return to Rt. 249E and proceed east (off Whittlesey island) to center of North Collins Village.

100.9	1.7	Turn left (north) at Rt. 62 (intersection at North Collins is on delta deposits of Lake Whittlesey). Rt. 62 follows along Lake Warren beach ridge. Lake Whittlesey beach is visible intermittently on incline to right (east), lower Warren ridges to left with deeper water silts and clays.
104.4	3.5	Whittlesey beach and wave-cut cliff exhumed in old gravel pit at right.
105.7	1.3	Pass main intersection in Village of Eden and exit to N.Y.S. Thruway.
107.4	1.7	Lake Warren beach ridges exposed in gardens on either side of road.
107.9	0.5	Cross South Branch Eighteenmile Creek. This and similar cuts are apparently post-glacial gorges.
109.1	1.2	View to right (northeast) of wave-cut beach and bluff of Lake Whittlesey. Rts. 62 (and 18) unite with Rt. 75N.
110.5	1.4	Cross North Branch Eighteenmile Creek. Cemetery on right and road itself still on Lake Warren beach.
111.4	0.9	Turn left at intersection in Village of Hamburg and follow Rt. 75N to N.Y.S. Thruway.
113.5	2.1	Turn onto N.Y.S. Thruway and keep left after toll barrier north to Buffalo.
120.4	6.9	Toll barrier at Lackawanna.
130.4	10.0	Exit right (#50) to Youngmann Hwy. (I-290) toward Niagara Falls.
131.4	1.0	Pass under Rt. 5 and through Onondaga Escarpment. Note south dip of strata. This was formerly a prize fossil collecting quarry.
133.4	2.0	Exit right for Millersport Hwy. (N.Y. 263N).
134.2	0.8	Turn into Buffalo Marriott Inn.

A-5

Recent Oil and Gas Developments on Public Lands in Western New York State

Peter Buttner, Director, Environmental Management,
New York State Park System

Manfred Wolff, Geology Department,
Hofstra University

Hyler Gray, Allegany State Park,
Salamanca, New York

INTRODUCTION

Three years ago, in response to a directive to either reduce the energy-related expenses of its recreational programs, or curtail recreational services, the New York State Office of Parks, Recreation and Historic Preservation (the Agency) began a program to find alternative resources to replace the energy used in many of its park facilities. The material included in this report is taken primarily from Buttner (1982).

With a staff of some 2500 permanent, professional staff, the Agency administers 146 state parks, 33 state historic sites and 64 state park campsites covering about one-half million acres of land throughout New York State. These lands, and associated facilities, include such diverse components as: nature and bike trails; woodland cabins and camping areas; cross-country ski trails and ski jumps; modern theaters; nature preserves; managed timberlands; historic sites; canal parks; watersheds and dams; coastal barrier islands; inland and coastal marshes; primitive hiking and camping areas; fresh and salt water bathing areas and a geological museum. About 45 million visitors use these facilities each year. Recently the cost of energy associated with the operations and programs available at these facilities has become a significantly large component of the Agency's budget. Since the Agency's operating funds come primarily from taxes, and reduction in energy costs will lessen the need to generate additional tax revenues.

In order to maintain the various recreational and supporting services that the Agency provides to the public, it had to find some method to reduce the energy cost component of those services. In 1979, the Agency instituted a program to gain access to public hydrocarbon resources known to exist beneath several tracts of parklands in central and western New York State. The main objective of this program is to obtain natural gas resources, in lieu of royalty payments, for use in park facilities in place of the oil and LPG purchased each year for those facilities.

Landscapes and terrains are often selected for addition to the State Park System because of their special features and characteristics. Once in the System, the Agency is committed to the protection of such attributes from any subsequent adverse effects. Such effects can result from any use, recreational or administrative, which might be inconsistent with the environmental setting of those lands. A typical, revenue-driven, oil and gas exploration and development program can cause significant changes in the environmental setting of any tract of land. The Agency is committed to an environmentally-sensitive, integrated resource management program with a limited, focused development driven by operational energy requirements. Developed as an environmentally-sensitive, resource management project, the Agency's program provides a strategy to meet some of its present and much of its future energy needs. The project is expected to have a twenty-year, self-sustaining term. The key element of this long-term program is that the Agency will be able to obtain energy for on-site use as part of its royalty payment credit. To date the project has been placed in operation at three locations in western New York State; the field trip will visit two of those locations, plus several other operations on private tracts. Using competitive bidding procedures, which initially produce a one-time, per-acre bonus payment, and then provide either a yearly rental or a royalty payment, the Agency has entered into lease arrangements with private interests for the development of the energy resources from beneath selected tracts.

Oil and gas production is, of course, a profit-making industry; protection of the environmental and of the public-owned surface resource is not the producer's chief concern. Therefore, procedures insuring this protection must be included in the lease, along with penalties for non-performance. Problems with the operation on oil and gas leaseholds involve either the physical aspects of the land or the style of program management. Examples of the physical aspects of an oil and gas program that require attention include: well-site selection; work-site arrangement and size; location and type of access ways and woods roads; the use of fabric roads; erosion and runoff control: timber handling and clearing of woodlands; wildlife displacement; noise; inclement weather during key operations; gas flaring and fires; alteration of slopes for roads and work sites; oil and salt-water spills; misuse of woodlands by sub-contractors; valves without locks; blowouts; drilling of service wells; storage tanks and tank heaters; crossing of streams and wetlands by operator's equipment; well-service activities; spacing requirements; informal ancillary equipment; oil-and gas-collection systems; gas-drying stations; handling of muds, salt waters, hydrofracturing chemicals and various toxic materials; and, the use of woodlands to purge garbage, trash and oil-field refuse.

Some important examples of the management aspects of an oil and gas program include: criteria for site selection; overall road and pipeline developments; degree of sunlighting needed per acre; subleases; easements and special use permits; types of rights being exercised; timed performance incentives; rents; terms; delay rental payments; bonus payments; royalty distributions; hydrocarbon types and amounts that leave the leasehold; logging contracts and limited timber sales; windfall profits and other taxes; resource inventory updates; detailed maps, boundaries and monuments; metering of gas and oil produced by each well; formation pressures; pumping rates; turnkey operations; field operations and emergency managers; road building alignment and site flagging; multiple use of roads and woodlands during operations; gates and locks; signs; hard-hat area restrictions; vandalism; constancy of production and market value; administration of trade secrets and confidential data; audit trails for problems; and, scheduling of work on a leasehold in relation to various public recreational activities.

The lease that was developed to deal with such aspects and situations consists of a framework of items common to all leaseholds on parklands together with appropriate tract-specific stipulations. For example, in addition to the usual components, a lease for one tract of parklands limited the maximum number of wells to be allowed on that tract to three. The framework of the general lease developed by Buttner for use throughout the Park System was assembled from the most appropriate parts of sample leases used by various state and federal agencies and other organizations. Into this framework was inserted a group of special provisions to provide long-term environmental safeguards for public parklands.

This lease requires the Agency to maintain an on-going analysis of the environmental effects of any oil and gas activity on the tract throughout the tenure of a program. This approach to the consideration of environmental aspects throughout the life of a project is distinguished from the usual one-time, environmental-impact review that is done prior to the start of a project. The lease also provides for changes over time in the direction and emphasis of the program.

In order to assure that maximum attention is given to environmental considerations at all phases of each exploration and development program, it is managed through the Office of the Agency's Director of Environmental Management. All bidding procedures, title search work and other related activities are conducted by the Office of General Services, as Administrator. The Division of Land Utilization, Office of General Services under provisions of Subdivision 4a of Section 3 of the Public Lands Law may lease to the highest bidder an interest in real property included, but not limited to; air rights, subterranean rights, etc., when such interests are not needed for present public use. This coordinated effort between an operating agency (Parks) and a service agency (General Services) has produced an especially comprehensive management scheme.

As noted previously, the operation of an oil and gas program on public lands can cause disturbance and modification of both the surface and the subsurface resources. In order to control and limit the effects of such programs on those resources, a set of rules, regulations and policies based on reasonable standards and procedures was developed. These guidelines had to be rigorous enough to control and limit operations yet not so constraining as to discourage private enterprise from investing in an exploration and development program on public lands. A leasehold owner will only invest in drilling and operating an oil or gas well if there is some chance that a reasonable profit will be obtained. Land managers who need to satisfy on-site energy requirements should also appreciate the profit considerations; the greater the profit to the investor, the more royalty energy will be available to the landowner-lessor. As energy resources are depleted and potential reserves become more valuable, managers of public lands will be subjected to increased pressure from both public and private interests.

OIL AND GAS HISTORY IN WESTERN NEW YORK

The oil springs of southwestern New York State were noted by a French missionary in 1627. These springs were located near the present border between Allegany and Cattaraugus Counties. By the late 1800's oil and gas exploration and development programs had developed into a vigorous industry in the region. Although not a major supplier of oil and gas, New York has produced approximately one-quarter to one-third of a billion barrels of oil and perhaps 180 billion cubic feet of natural gas during the last eighty years. In 1980, about 15.65 billion cubic feet of natural gas, valued at $29 million at the wellhead, and about 824,000 barrels of oil, with a wellhead value of $30.8 million, were produced by some 7500 wells in New York State.

The first producing gas well in New York State was drilled by William A. Hart at Fredonia, Chautauqua County in 1821. From this 70-foot deep well, Mr. Hart produced gas for the lamps of Fredonia, the first village to be illuminated by natural gas in the United States. In 1865, six years after the first oil-producing well was developed by Drake at Titusville, Pennsylvania, Job Moses completed the first commercial oil well in New York State in what is now Allegany State Park in Carrollton Township, Cattaraugus County.

Over the last hundred years, the oil-and gas-bearing rocks of western New York have been tapped by more than thirty thousand producing wells. Some of these wells continue to produce oil and gas while others have been abandoned once the easy-to-get hydrocarbons were removed. This primary production of easy-flowing hydrocarbons usually represented as little as 1/5 of the total amount of oil and gas trapped in the rock. This means that 4/5 of the once available original hydrocarbons might still reside in the rocks beneath some of the abandoned well sites.

This drilling, draining and abandoning activity was especially intense in the region that now contains Allegany State Park. Prior to the establishment of the Park, large tracts of its present acreage had been subjected to some rather intense and varied land use forces. Some of these tracts had been homesteaded and farmed, then abandoned; then opened and leased with a pattern of oil and gas wells with connecting pipelines, railroads and other scars of the day; later clearcut to produce raw materials for a variety of forest products that were manufactured by on-site factories; and finally, some of the tracts were set on fire and burned by both natural and other forces.

In the public parklands of central and western New York State, the Agency has discovered more than 250 abandoned drill holes of which 4/5 are located in Allegany State Park. Most of these well holes were abandoned with their lining of steel casing still in place, but, as the need for steel increased during World War II, many of the casings were scavanged. In a few years time, the surface of any abandoned well and well site becomes covered by forest litter, and abandoned wells are very difficult to locate. Beneath the surface cover, the inside walls collapse and soon a small cavern is formed. Developed to this stage, such abandoned, uncased well holes present a continuing hazard to people and animals and to the quality of the ground water and shallow aquifers. Such open conduits also allow valuable subsurface hydrocarbons to escape to the surface thus reducing reservoir pressures. The Agency has developed the equipment and a low-cost methodology that is technically adequate, to plug such abandoned and uncased wells. As new sites are discovered, these methods are used to plug any abandoned wells found at the site. The Agency's leasing program provides substantial protection to control such abandoned wells.

In general, significant oil and gas production in New York State thus far has included discoveries in Chautauqua, Cattaraugus, Allegany and Steuben Counties. These New York Counties directly adjoin the oil- and gas-rich area of northwestern Pennsylvania that includes both the Titusville and Bradford districts. Since political boundaries are transparent to the subsurface oil-and gas-containing rocks of this southwestern New York-northwestern Pennsylvania area, similar exploration and development techniques are used throughout the region. However, the oil and gas industry is regulated somewhat differently by each state.

The oil and gas producing companies operating in the region are small-to-medium-sized firms rather than the large, major oil-and gas-producing companies with world-wide interests. The availability of a well-developed, local oil and gas industry is especially useful to a land manager that is considering an exploration and development program. These development programs are usually based on a leasing arrangement between the administrator of such public lands, as lessor, and the private entity or operator that proposes to explore for and develop the hydrocarbon potential of those lands, as lessee. Each participant, lessor and lessee, has rights, privileges and responsibilities under such leasing agreements.

The concept of an oil and gas lease interest will be new to most students, some geologists and many land managers. A hydrocarbon exploration and development program is a capital-intensive, high-risk activity. In 1982 it costs about $100,000 to drill and complete for production 1500-foot deep gas well in New York State. It costs about half that amount if the well turns out to be a "dry hole". A dry hole is a well that either showed no hydrocarbons or else will not produce hydrocarbons at a rate that is presently economically feasible. (As market conditions become favorable, some dry holes may be re-worked and brought into production using various flow stimulation techniques.)

In order to assemble adequate funding for such exploration and development programs, a partnership is sometimes formed in order to finance both the purchase of leases and the drilling of one or more exploratory wells. Such oil and gas partnerships usually contain one or more general partners and a group of limited partners. For a fee, the general partners supply initial capital, management services, technical expertise, and, in some cases, the leases themselves. The general partners have full and exclusive discretion concerning the management and operations of the program; it is the general partner who will be directly responsible to land managers for maintenance of performance standards. The limited partners supply working capital for the program in exchange for a return of their investment plus a share of the profit after all expenses are paid. The point at which limited partners recover their full investment is called payout.

Although many kinds of financial arrangements are used in the oil and gas business, there are three main types of partial interest in oil and gas programs that should be understood. Usually when a tract of land is leased for an oil and gas exploration and development program, the landowner receives, in addition to other payments, a royalty interest in the program. This royalty interest is typically a one-eighth fractional share of the value of any hydrocarbons produced during the tenure of the program. The lessor's royalty interest is split from the production exclusive of the burden of any operating expenses whatsoever. It is this royalty interest that is designated by the lease to be made available for on-site use to support the Agency's recreational programs.

The remaining seven-eighths of the production forms the working interest and all expenses of the program are paid out of this interest. In order to entice investors to participate in an oil and gas program, the operator often sells off one or more fractional parts of the working interest. These fractional interests, generally called overriding royalty interests, are carved out of the working interest and are free of all operating costs. In some cases, in order to finance other ventures, the original operator might sell off any remaining working interest he holds in the program, sometimes retaining only a small overriding royalty interest in the venture. As with other interests, overriding royalty interests may be assigned and traded at any time during the term of the program's lease. In order to attract more investment capital, the operator of a lease might place news items in various newspapers and trade journals concerning the success of the program. Such publicity will call attention to the oil and gas exploration and development programs.

HABITAT OF OIL AND GAS

Crude oil and natural gas are hydrocarbon by-products of the decomposition of organic matter like that which accumulates today in freshwater and marine wetlands, shallow seas and other bodies of water. Most of these naturally-occurring deposits of oil and gas are at least one-half million years old; many deposits are as much as 400 million years old.

As organic sediment accumulates and is buried beneath successive layers of younger material, various organic and inorganic transformations take place. The processes involved, encouraged by the elevated temperatures and pressures associated with deep burial, convert the organic sediment into various types of hydrocarbon-rich deposits. Depending upon the type and volume of the original sediment, and the rates and scales with which the end result could be oil shale, coal, tar, oil, gas, graphite or a combination of by-products.

When deposited, the sediment source consists of a framework of grains and pore spaces. The grains are the larger pieces of organic matter and other debris while the pore spaces between grains will contain original fluids and finer organic debris. This sediment has two important, fundamental properties: the initial port space or porosity and the amount of connected pore space through which fluids can move called permeability. As the sediment is compressed and converted to rock and hydrocarbons, the pore space may be almost eliminated and the more mobile forms of hydrocarbons, driven by other pore fluids, may migrate from the source rocks to other rock units which have available porosity and permeability. Gradually, over long periods of time, the concentration of hydrocarbons in the host rock becomes significant and such rocks become reservoirs or pools of hydrocarbons.

The term trap may be applied to some types of subsurface features. A trap is a physical deterrent which limits the migration of the hydrocarbons from the host rock. In essence, a trap exists where a bubble of fluid or gas can not escape to the surface forming a pool. A field may consist of one or more reservoir pools. The important characteristics of reservoirs are: structure; texture and composition of the host rock; thickness and form; depth below the surface; porosity and permeability; character of the hydrocarbons; the variability of the properties of the host rock; and the relationship of the trapping mechanism to local and regional geological features. The lessee operator will usually hire a consultant geologist to evaluate the reservoir and the land owner may be able to receive copies of all reports and analysis.

Crude oil is measured in barrels, each of which contains 42 gallons. A barrel of oil produced from wells in the southwestern New York - northwestern Pennsylvania region can be expected, on the average, to yield: 12.5 gallons of gasoline; 9 gallons of distillate fuels; 11.5 gallons of lubricating oils; 1 gallon of wax; 7.5 gallons of residual fuels; and about .5 gallons of refinery waste.

The main types of wells that may be drilled on a leasehold include:

1. <u>New Field Wildcat Wells</u> - These are wells drilled on a geologic feature or in an area which had never before produced oil or gas. In 1980, thirty such wells were drilled in New York State; fifteen found gas and fifteen were judged to be dry holes. Obviously, such ventures can have the lowest probability of success. If a tract of land has never been associated with any oil or gas production, then land owners might find it especially difficult to entice commercial exploration.

2. <u>New Pool Wildcat Well</u> - This is a test well drilled outside the limits of a proven area or pool but in a geological area already productive. Depending upon the habitat of the hydrocarbons of the area, this type of wildcat well can be a less risky venture than the new field wildcats.

3. Deeper and Shallower-Pool Test Wells - These wells are drilled in an attempt to locate either deeper or shallower hydrocarbon resources inside the limits of an area or pool previously proven productive. If successful, such wells can increase the density of production from a particular leased tract.

4. Outpost or Extension Test Wells - These wells are drilled to test for possible distant extensions of a producing pool. Where existing programs have developed outside and close to land boundaries, such types of wells might be the first types considered by a leaseholder. Most of the recent successful oil and gas drilling programs in the northeastern United States has resulted from the extension of known producing fields or pools.

5. Development Wells - These wells are drilled to develop the production of a hydrocarbon field or pool. The growth of a development well field can be a function of many factors such as: state and other regulations; depth of the producing horizon; primary or secondary production; the ease with which the hydrocarbons can be brought to the surface; the complexity of the subsurface geology; the type of hydrocarbons being produced; the market value of the produced oil and/or gas; the availability of capital, equipment, product delivery systems and product purchase contracts; the optimum versus regulated spacing of wells; the availability of well sites on the surface; any stipulations that might be a part of the operator's lease with the landowner; and, other limiting or encouraging factors. As specified in the Agency's standard lease, all drilling plans must be approved prior to their use on park lands. This lease permits adjustment of the plans, well-site locations and well density based on the environmental characteristics of the terrain. In the southwestern New York region most of the active oil fields are between 500 and 2500 feet in depth and the spacing can be as tight as one well per ten acres. The producing gas wells of the region are usually from 2000 to 4000 feet deep and spaced about 1500 feet apart. In addition, wells are excluded from a corridor 660 feet wide inside, and parallel to, property boundaries in some fields.

6. Stratigraphic Test Wells - These wells are drilled to acquire information about the character of the subsurface. They are logged in detail, and carefully studied and evaluated, but they are not drilled for production. Land owners might have to provide a site and the ancillary operational features for such test wells. Since such test wells will provide no direct royalty-derived energy, land owners must consider requests to drill such wells carefully.

7. Service Wells - These wells are drilled to support the production program in an existing producing field. This may be used to stimulate production by injecting into the reservoir various substances such as water, steam, gases, detergents and other fluidizers. Such stimulated production is usually termed secondary production to distinguish it from the more easily obtained primary production.

Land owners must realize that leasing operations are long-term agreements which are extended by various stipulations in the lease. Although there are a variety of leasing arrangements, such agreements usually have a primary term of perhaps ten years that is then followed by a minimum secondary term of an additional ten years. Primary and secondary terms of a lease are not necessarily related to the primary and secondary production history of an oil or gas field. Some oil fields are extremely long-lived and have produced for a hundred or more years. One such field in southwestern New York has produced more than 150 million barrels of oil since first discovered about 100 years ago.

Applied secondary production technology can significantly enhance the overall production of an oil-producing region. The primary oil production of Allegany and Cattaraugus counties was about 80 million barrels over 70 years; secondary production techniques have nearly tripled the total production to some 230 million barrels of oil. It is expected that the oil-producing reservoirs involved still contain almost 14 times as much oil as the yield of primary production. Long-term land use and facility development planning by public land managers and other land owners must recognize the possibility of extended production on leased lands.

A COMPOSITE TEST WELL

Most oil and gas drilling is done to test the oil and gas potential of relatively shallow "pay" horizons. A typical gas well might extend 1,500 feet below the surface and cost between fifty and one-hundred thousand dollars. The drilling and completion of a typical gas well involves many complex operations but might be generalized as follows:

1. An 8" diameter "pilot" hole is drilled by cable-tool (impact) methods to a depth of about 150 feet. This "pilot" hole is then cased with a steel casing pipe of 7" diameter, and cement is placed in the annulus between the outside of the casing and the rock wall. This depth may vary from site to site, but an important benefit of this segment of the well is that near surface ground water is kept from flowing into the hole. This part of the work could take seven days to complete.

2. Next, a 6" diameter hole is drilled by rotary drill from the base of the pilot hole to perhaps 1,400' below the surface. A rotary drill takes about two days to drill 1,250 feet. The hole may then be cleaned and a log made of the rock types encountered. Then the well is encased with pipe of 4½" diameter and this tube is cemented to the rock.

3. The rock types encountered by the drill may then be logged again by using a special logging probe. These logs will give some indication as to where the "pay" horizon(s) might be found. A perforator is then placed in the casing, lowered to the "pay" horizon and activated. This device blows holes in the steel pipe, through the cement, and into the surrounding rock at the "pay" horizon.

4. Typically, the well is then hydrofractured to stimulate the flow of gas from the surrounding host rock to the well site. This process takes a few hours and involves the injection of a mixture of liquid nitrogen, surficant and sand under some 2,800 lbs. of pressure into the well and out into the surrounding rock for some 500 feet in all

directions. After this process, the gas pressure at the wellhead will be between 300 and 1,000 lbs. per square inch, typically.

5. A 1" diameter pipe is installed within the 4½" diameter casing, from surface to bottom, to allow any water that accumulates in the pipe (usually salt water) to be removed at the surface.

6. The various pipes, valves and a buried, marked plastic transmission line are then installed and natural gas is permitted to flow from the well to an on-site meter.

7. The site is restored and the surface piping, which would probably fill the area of the inside of a compact car, is fenced and signed.

OWNERSHIP PATTERNS OF OIL AND GAS RIGHTS ON PUBLIC LANDS IN NEW YORK STATE

In western and central New York State most of the lands now a part of the State Park System were tested by the wildcatter's drill bit for oil and gas resources prior to the acquisition of those lands into the system. This testing, and in some cases extensive development, was financed and conducted by private interests. Those private interests either owned, or else obtained through a lease arrangement the rights to explore for and remove any oil or gas from beneath the lands involved. Most likely, these oil and gas rights were obtained initially from the surface land owner. In New York State, an "ownership" state, a land owner may have obtained the oil and gas rights when the land was acquired. In some other states, subsurface oil resources belong to the landowner only after they are possessed via pumping.

In such ownership states as New York and Pennsylvania, subsurface oil and gas rights may be separated from the ownership and title of the surface. In some cases, as the lands of western and central New York changed ownership over the years, the subsurface oil and gas rights were carried along with the title to the land. As some of those lands were acquired for the Park System (except where the oil and gas rights were reserved by the owner) the public gained ownership of any subsurface hydrocarbons. However, in those cases where the drill bit had proven previously that the subsurface rights had some value, those rights were excepted and separated from the title and, in many instances, sold, traded or leased independent of surface ownership and use.

Managers of public lands and other land owners in ownership states will encounter two main forms of non-public rights: reserved and excepted. Reserved rights are usually easy to determine; they appear in the deed as part of the process which transferred the land to public ownership. Excepted rights were separated from the ownership of the land prior to the existing deed. Excepted oil and gas rights are almost always a problem to unravel and authenticate. One of the major efforts of the Office of General Services, as Administrator for the Agency's leasing program, has been to search the titles of all lands where oil and gas rights were excepted from public parklands prior to their acquisition. Such effort requires dedicated and professional "detective" work, and managers of public land and other land owners generally will need such services.

With the increased interest in New York State's oil and gas resources, there have been many exchanges of the private ownership of both excepted and reserved rights to the oil and gas resources beneath public lands. In more than a few cases such rights have been split into various arrangements of vertical, horizontal, time-related, resource-related and rock-type-related patterns. This often results in many entities having various levels of royalty interest in a single tract. The exercise by private entities of such excepted and reserved oil and gas rights on public parklands can take place at any time, and must be a concern of both the public and the managers of such lands. In western New York State, some 36,000 acres of the Park System have oil and gas exceptions and reservations. The exercise of such rights within the 700,000 acres, Allegheny National Forest of western Pennsylvania has caused significant disruption of that composite of public lands and private inholdings. Part of the National Forest adjoins New York's 64,000 acre Allegany State Park. Many of the oil-and gas-rich rocks that made the Bradford, Pennsylvania area a famous "oil patch", extend northwestward beneath both the National Forest and Allegany State Park. As a result, the subsurface oil and gas rights to some 92% of the Forest were long ago acquired by private interests and many tracts are now in various stages of development. (The situation at Allegany State Park will be discussed later.) The Forest Service has documented its experience with oil-and gas-rights management in a comprehensive handbook. Public land managers and other land owners should review the Service's handbook and associated materials. Note that the Forest Service is concerned primarily with the management of <u>private</u> oil and gas programs on public lands while the State Park System, although concerned primarily with the management of a public-sponsored leasing program conducted by private operators, is also involved in the management of private-sponsored programs on public lands.

Where subsurface rights are held by other than the surface land owner, that owner must recognize those rights. When oil and gas rights are held, the owner of those rights has a just claim or privilege to move on to the surface land and operate a hydrocarbon exploration and development program. These rights apply to activities on specific tracts, but they are not exclusive rights to surface use nor do they provide any general privileges on adjacent or proximal tracts where no rights are held. Because of this, any geological or geophysical studies, surveys, mapping, tests or other programs that are proposed by private entities to be conducted on public lands should have prior approval of the agency that manages the lands for the public. The reason for this is that there often exists a mixture of public and private interests to oil and gas rights on some public lands. Such programs could violate certain rights of those interests. Such approval is now needed for any programs which would use State Park roads, woodlands, lakes, islands and coastal zones.

Both the operator and the landowner have equal responsibility to consider and reasonably accommodate each other's interest and stewardship. Capricious, irresponsible or otherwise improper actions by either party are inconsistent with the rights, privileges and responsibilities of each party. It is the policy of the State Park System to apply the same general standards and procedures to all operators of oil and gas programs on public parklands, be they operators on state-leased tracts or operators on tracts via reserved or excepted rights. This policy is based on Agency's experience with both types of operators over the last several years.

For example, the future of present surface features of some 40% of the public lands within Allegany State Park (there are some private lands within the legal Park boundaries), are of special concern. The subsurface oil and gas rights of some of these public lands are held by private interests, and, with the increased value of oil and gas, those rights can be expected to be exercised at any time. The tracts involved, although concentrated in the southeast quarter of Allegany, are spread almost like a "crazyquilt" throughout the entire park. These rights are now far too expensive to be purchased by the public. Using a comprehensive lease as a management framework, the Agency has better control over how the public's surface lands will be treated by these private oil and gas interests. The Agency's oil and gas operation has provided experience with a public-sponsored program on certain tracts that have been valuable in managing similar, private-sponsored programs on its other lands.

As excepted and reserved rights are uncovered and authenticated, land managers and other land owners may wish to evaluate and classify each prospect. Once classified, it is then possible to estimate the thresholds for various factors that will cause such prospect tracts to be candidates for private oil and gas programs. Since excepted and reserved rights may be exercised at any time, independent of the plans and control of any entity or public agency that holds the surface rights, such evaluations are important. The indicator factors that one of the authors (Buttner, 1982) has used to develop such a classification system include:

1. Chain of Title - Is it distinct or diffuse? Are the rights excepted or reserved?

2. Operations - Were there previously, or are there now, oil and gas operations on the tract, on neighboring tracts or in hydrocarbon-bearing rock units or rock types which occur beneath the tract?

3. Resource Potential - Given the overall setting and regional production history, is it likely that the resource potential will be low, moderate or high?

4. Interest in the Tract - Have there been inquiries concerning the tract? Have programs that involve the tract been described in trade papers and journals? Are academic and/or industrial workers engaged in research that involves the tract?

5. Applications - Have applications for drilling regulatory agencies? (By agreement with the New York State department that

regulates well drilling, all applications for permits to drill on lands in the State Park System must first be approved by the Agency's Manager of the Oil and Gas Program.)

6. Proximal Transmission and REfining Facilities - Is the tract traversed by, or reasonably near, a natural gas transmission line? How far are the nearest refining facilities and, if pipelines are not used, can oil collection trucks reach any oil producing wells that might be operated on the tract?

7. Existing Road Systems - Are there any existing road systems, either external to, or within the tract, which might reduce road building and maintenance costs?

8. Terrain Characteristics - Are there unusual features of the tract such as special wood lots, perched bogs, eagle nesting areas, endangered species, steep slopes, earthquake hazards, extensive wetlands, nature study areas, unusual wildlife, ski jumps, beaver colonies, bike trails or other features that would conflict with oil and gas operations?

9. Extension or New Pool Possibilities - Is there some possibility that the tract might be a candidate site for either an extension well or a new pool wildcat?

10. Seismic Survey Activity - Have there been recent seismic surveys on private lands in proximity to the tract? Has there been a request by the subsurface-rights owner to conduct seismic or other studies on the tract?

11. Change in Ownership Patterns - Have changes in oil and gas rights or the purchase of the leasing privileges to exercise those rights taken place? Are these rights being assembled by a single entity?

12. Market Sensitivity - Given that there is a potential for oil and gas development to take place on the tract, how sensitive to market fluctuations are such operations and plans?

13. Overall Candidacy Rating - What is the overall rating for this tract given its characterization based on the above indicator factors?

Usually an operator, either leaseholder or rights holder, will contract for the various services needed to drill, test and complete wells, conduct various seismic and other studies, perform any well stimulation and hydrofracturing work that might be required, and provide a variety of other special well field services. It is important that land managers and other land owners require that all operators on public lands inform any contractors or subcontractors of the need to precisely conform with various provisions of any oil and gas management programs.

Access to a tract that is scheduled for development can be a particular problem for land managers. The Agency's program requires that the primary access to any tract of parklands be via existing park entrance roads and then continued, as necessary, via approved woodland roads and corridors. Otherwise the public parklands will end up with a maze or privately-controlled, informal entrances.

SUMMARY

This report provides guidance for those who are either considering, or expect to be required to consider, initiatives for the siting of oil and gas programs on the lands they manage or own. Such initiatives can be grouped into three types of situations. One type of situation involves the development of a landowner-sponsored proposal for the establishment of an oil and gas exploration and development program on public land. Following competitive bidding procedures, a lease is awarded to an operator-lessee to conduct a long-term, landowner-supervised, environmentally-sensitive oil and gas program on a tract of public land. In return for the grant of this privilege, the landowner-lessor receives various payments together with a royalty interest in any oil and gas produced from the tract. This fractional interest is exclusive of the burden of any production expenses and is taken by the landowner in the form of energy from the wellhead. Such free energy, usually natural gas, is used to replace some or all of the higher-priced energy purchased by the landowner to support various on-site programs, activities, services and facilities. In some jurisdictions, it may also be possible to use excess royalty payments to support other types of expenses.

A second type of situation arises when a non-public entity proposes to exercise the rights it holds for the exploration and extraction of any oil and/or gas beneath a tract of land presently in public ownership. Because of the capital-intensive, high-risk nature of the oil and gas business, such ventures will usually result in revenue-driven programs that include maximum technical development. Unless specifically negotiated in exchanged for some privilege, the public land owner has no right to any form of production-related compensation. Such privilege might include the right-of-way for a transmission pipeline which crosses public lands outside the tract under development, for example. Likewise, the landowner can not exercise the same level of control and supervision of activities on this tract as would be possible on a land owner-leased tract. Any powers the land owner exercises are derived from the land owner's responsibility to the public as the caretaker-steward of the land, of its natural resources, and of its recreational attributes. The land owner can only guide the activities of private enterprise on public land. Working with the private interests, the landowner can usually coordinate plans for such things as: roadway alignments; woodland rights-of-way; informal trails and work zones; timber cutting and management; erosion control and slope improvement; wildlife habitat enhancement; and the protection of special habitats. The exercise of private rights on public lands is an intrusion into the public's use of such lands. If such intrusions are possible, then managers of public lands and other interested parties must carefully prepare for them; to be caught without a plan of action will surely place a hardship upon the land.

The third type of situation combines a public land owner-sponsored, requirement-driven program with one or more private rights-sponsored, revenue-driven programs. Experience with a leasing program provides a keen basis for the active coordination and firmest control of any private program.

The field trip will attempt to show as many different types of oil and gas operations as possible by visiting a mixture of operations taking place on public and private tracts. Since it is impossible to forecast where these operations will be at the time of preparation of this report, the field trip itinerary will be provided to all participants on the day of the trip. Well logs, core and other materials will be available for discussion and examination.

In general, the trip will visit the Medina gas fields, the Allegany gas storage fields and various western oil and gas developments. We have provided a review of the stratigraphic units of western New York, some of which may be examined in the field and in various logs.

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As mentioned previously, in order to allow for flexibility in the selection of work sites and on-going field operations in October, a detailed road log and site descriptions can not be published with this article; an itinerary will be provided each participant. It is expected that the field trip will visit tracts under various stages of development. We would like to thank the following owners/operators for their support and cooperation: Abaterra Energy; Alden-Aurora Gas Company; Bowman Development Company; Clover Oil and Gas Company; Elcoex; Felmont Oil Corporation; H.L. Murry; National Fuel Gas Supply Corporation; N.Y.S. Natural Gas Corporation; Saxtet Energy; Summit Petroleum Corporation; U.S. Gold Corporation; WITCO Chemical; and, any others that might provide assistance.

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COMPOSITE PALEOZOIC STRATIGRAPHIC SECTION

FOR SOUTHWESTERN NEW YORK

PERIOD		GROUP	UNIT		THICKNESS	PRODUCTION
Penn.		POTTSVILLE	OLEAN	Ss, Cgl	75-100'	
Miss.		POCONO	KNAPP	Ss, Cgl	50-100'	
DEVONIAN	UPPER	CONEWANGO		Sh, Ss, Cgl	700'	
		CONNEAUT	CHADAKOIN	Sh, Ss	700'	
		CANADAWAY	UNDIFF. +	Sh, Ss	1100-1400'	Oil, Gas
			PERRYSBURG#	Sh, Ss		Oil, Gas
		WEST FALLS	JAVA NUNDA RHINESTREET	Sh, Ss	375-1250'	Oil, Gas
		SONYEA	MIDDLESEX	Sh	0-400'	Gas
		GENESEE		Sh	0-450'	
	?		TULLY	Ls	0-50'	Gas
	? MIDDLE	HAMILTON	MOSCOW LUDLOWVILLE SKANEATELES MARCELLUS	Sh Sh Sh Sh	200-600'	Gas
			ONONDAGA	Ls	30-235'	Gas, Oil
	LOWER	TRISTATES	ORISKANY	Ss	0-40'	Gas
		HELDERBERG	MANLIUS RONDOUT	Ls Dol	0-10'	
SILURIAN	UPPER		AKRON	Dol	0-15'	Gas
		SALINA	CAMILLUS SYRACUSE VERNON	Sh, Gyp. Dol, Sh, Salt Sh, Salt	450-1850'	
		LOCKPORT	LOCKPORT	Dol	150-250'	Gas
		CLINTON	ROCHESTER IRONDEQUOIT	Sh Ls	125'	
	LOWER		SODUS REYNALES THOROLD	Sh Ls Ss	75' 2-8'	
		MEDINA	GRIMSBY WHIRLPOOL	Sh, Ss Ss	75-160' 0-25'	Gas Gas
ORDOVICIAN	UPPER		QUEENSTON OSWEGO	Sh Ss	1100-1500'	Gas
			LORRAINE UTICA	Sh Sh	900-1000'	
	MIDDLE	TRENTON-BLACK RIVER	TRENTON BLACK RIVER	Ls Ls	425-625' 225-550'	Gas
	LOWER	BEEKMAN-TOWN	TRIBES HILL-CHUCTANUNDA	Ls	0-550'	
CAMBRIAN	UPPER		LITTLE FALLS	Dol	0-350'	
			GALWAY (THERESA)	Dol, Ss	575-1350'	Gas
			POTSDAM	Ss, Dol	75-500'	Gas
PRECAMBRIAN			GNEISS, MARBLE, QUARTZITE, etc			

STRATIGRAPHY

The lower Paleozoic strata in this region vary in thickness from 6000 feet near Lake Erie to over 12,000 feet in Allegany County, near the Pennsylvania border. They cover a time span from upper Cambrian to Pennsylvanian - about 200 million years. While most of the region is a southward dipping homocline, in the subsurface there are gentle NW-SE trending folds that plunge to the southwest. These are sometimes offset by reverse faults associated with doming anticlines that are related to deep seated thrusting - sometimes producing gas traps along the anticlines (Van Tyne and Foster, 1979).

The stratigraphic units described on the following pages represent a generalized section for Chautauqua, Cattaraugus and Allegany Counties, and are based on the correlations of Rickard (1975) with general descriptions compiled from various other sources.

STRATIGRAPHIC UNITS

There may be some gas potential in nearly all of the Paleozoic section. It begins with the Upper Cambrian dolomitic Theresa Sandstone which, near the Pennsylvania border, occurs about 11, 750 feet beneath the surface (Fisher, 1966). These sandstones are unconformably overlain by Middle Ordovician gray dolomitic limestones, over 700 feet thick, that form the Trenton and Black River Groups, and the Utica black shale - another 130-250 feet of strata. There are also some gas occurrences in the Upper Ordovician quartz siltstones and fine-grained sandstones of the Lorraine Group (550-700 feet), the Oswego Sandstone (100-650 feet) and the Queenston Formation (700-1000 feet). Though the Queenston appears as a fine-grained maroon shale near the surface, it becomes much sandier southward in the subsurface. Yet most of the oil-gas potential occurs in the younger strata above this sequence (Van Tyne, 1974), and these units will be briefly discussed on the following pages:

Lower Silurian-Medina Group (Vanuxem, 1837).

Though the Medina Group is a relatively thin sequence, the abundance of well-sorted sandstones with subordinate shales and the near absence of limestones and carbonate cement make it the principal reservoir for natural gas in this region (Fisher, 1966). The gas occurs only in certain units or facies of certain units, indicating that lateral changes in porosity and permeability are the chief factors controlling gas migration. Oil production potential is considered insignificant (Van Tyne, 1974). The important units are:

WHIRLPOOL FORMATION ("White Medina" of drillers).

Lithology: A white-light gray pure (95%) Quartz sandstone, medium-coarse grained, with well rounded and frosted surfaces embedded in quartz cement. The unit is medium-to thick bedded and planar cross-bedded and is unfossiliferous.

Thickness: 8 to 15 feet, increasing to 25 feet eastward toward Rochester, but it is absent east of Springville (Fisher, 1954).

Contact Relations: Forms a sharp disconformable contact with the maroon-red sandstones of the Upper Ordovician Queenston Formation, frequently with the presence of mudcracks (Fisher, 1966); forms a sharp but conformable upper contact with the gray, thin-bedded siltstones of the Grimsby eastward or dark gray shales and calcareous siltstones of the Power Glen Formation westward.

Age: Niagaran Series

Environment: Aeolian dunes and beach environment of Queenston-Juniata delta complex that was reworked into supratidal shoals and sandflats by transgressing seas.

POWER GLEN FORMATION

Lithology: A series of interbedded gray shale and siltstones becoming green-gray in the upper third of the unit. As with the Whirlpool Sandstone, the coarsest deposits are in the center of the unit. It grades westward into the Manitoulin Dolostone, and contains abundant pelecypods and brachiopods.

Thickness: It is 50 feet in Niagara Gorge and thins eastward and disappears near Gasport where the Grimsby overlies the Whirlpool.

Contact Relations: Forms a sharp lower contact with the white sandstones of the Whirlpool Formation, and grades laterally into the maroon sandstones of the Grimsby Formation (i.e. forms a sandwich between the "White Medina" and the "Red Medina" sandstones).

Age: Niagaran Series

Environment: Nearshore marine or lagoonal (low energy).

GRIMSBY SANDSTONE

Lithology: Consists of 3 facies - lowest (A) is a pink-white-pale green mottled fine-grained, rippled and cross laminated quartz sandstone and siltstone with interbedded maroon shale and sandstones. Middle (B) facies is a dark maroon hematitic sandstone with large scale planar crossbeds. The upper (C) facies is a deep red knobby shale with mottles of gray-green calcareous lenses (Fisher, 1966).

Thickness: The three facies are thickest in this area (100-125 feet) and then thin eastward and westward. They contain brachiopods and ostracodes.

Contact Relations: Forms a sharp but conformable lower contact with the green-gray shales of the Power Glen Formation or the white quartz-rich Whirlpool Sandstone; the upper contact produces a sharp disconformity with the gray, fine-grained sandstones of the Thorold Formation (Rickard, 1975).

Age: Niagaran Series

Environment - Facies A is nearshore marine (subtidal-intertidal) while Facies B represents the beach and sandflats and overwash of a barrier island (intertidal-supratidal). The C facies represents the low energy lagoonal environment behind the barrier chain.

Lower Silurian - Clinton Group (Vanuxem, 1837)

As the Medina Group, the lower two thirds of the Clinton is also a relatively thin sequence with several disconformities, and it can be clearly noted in the driller's logs. The white quartz sandstones (Thorold-Kodak Formation) are the only nearly continuous units that separate, with a sharp break, the red hematitic sandstones and shales of the Medina from the gray-fine-grained limestones of the upper Clinton (Rickard, 1975).

THOROLD SANDSTONE

Lithology: A light gray, fine-grained, thin bedded argillaceous (20% clay matrix) angular quartz sandstone. It appears to be a reworked sequence of the "B" facies of the Grimsby Formation, with the absence of red hematite cement from coastal winnowing (Fisher, 1966). Though discontinuous, it is correlative eastward with the Kodak Sandstone. Both are similar in appearance and contain lingulid brachiopods and ostracodes. It forms the base of the Clinton Group.

Thickness: The member varies in thickness from 0-12 feet, with 2-4 feet being the average.

Contact Relations: The gray fine-grained sandstone forms sharp disconformable contacts between the adjacent units (Rickard, 1975) and when present, forms a key bed between the Medina and upper Clinton Group.

Age: Niagaran Series

Environment: The numerous regional disconformities and the appearance of the thin-bedded quartz sandstones suggest reworking along the transgressing Niagaran Sea.

NEAHGA FORMATION

Lithology: An irregular bedded, knobby, calcareous gray-green shale and siltstone. Though discontinuous, it is correlative with the Maplewood Shale eastward, based on the presence of conodonts and ostracodes.

Thickness: It occurs as an elongate lens and varies in thickness from 0-6 feet with 2-3 feet being the average.

Contact Relations: It also forms a sharp disconformity, with slight relief, between adjacent members.

Age: Niagaran Series

Environment: The position and lithology suggest the presence of shallow transgressive lagoonal deposits in association with the coastal sands of the Thorold Formation.

HICKORY CORNERS LIMESTONE

Lithology: It occurs as a dark gray, thin-bedded shaley or crystalline limestone. The 2-4 inch layer of shale clasts and phosphate nodules (Kilgour, W.J., 1963) suggests a disconformity. Worn brachiopods (Pentamerus) and bryozoan are common, though ostracodes are the important index fossils.

Thickness: It varies in thickness from 0-8 feet, with 4-5 feet being common, and it may be an important marker bed.

Contact Relations: Though conformable with adjacent members in the Rochester area, it occurs along disconformable contacts in western N.Y. The lower contact may occur on the Neahga, Thorold, or Grimsby Formation; the upper contact is against the Rockway Formation. Near Buffalo it is overlain by the gray-tan, medium-crystalline Merritton Limestone Member. The contact contains a 6-inch zone of rounded chert pebbles, black phosphate sands, and green glauconite with pyrite - suggesting another disconformity. The unit varies in thickness from 2-3 feet and contains some Pentameroides brachiopods indicating its younger age.

Age: Niagaran Series

Environment: The Hickory Corners Member was deposited in a low energy restricted (lagoonal?) environment by the transgressing Niagaran Sea; the Merritton Member was deposited later, in a similar environment and was later reworked and extensively eroded. The chief reason for the lowered sea level and extensive disconformities during the Lower Silurian is now believed to be related to the period of continental glaciation across the Gondwana continents near the south polar region (McKerrow, etal. 1979)

Upper Silurian - Clinton Group (Vanuxem, 1837)

The upper third of the Clinton Group also transgresses the boundary into the Upper Silurian, and provides a more continuous section of strata in western N.Y.

IRONDEQUOIT LIMESTONE - Rockway Dolomite Member

Lithology: Unit consists of a tan-gray sparsely fossiliferous, massive, fine-grained dolomitic limestone with a few interbedded gray-brown shale layers, large brachiopods and conodonts.

Thickness: Comprises 12 feet of buff weathered dolomitic limestone and interbedded shales.

Contact Relations: The lower contact contains worn limestone and chert pebbles with pyrite over the Merritton Limestone; the upper is gradational into the subtidal facies of the Irondequoit Limestone.

Age: Niagaran Series

Environment: The association of thin-bedded dolomitic limestone and gray shales suggests an intertidal (lagoonal) carbonate environment.

IRONDEQUOIT LIMESTONE

Lithology: Forms a light gray, coarsely crystalline crinoidal limestone with thin seams of calcareous shale. Occurs in biohermal "reef-like" masses with abundant crinoids, corals, bryozoans and brachiopods.

Thickness: Consists of 6-8 feet of fossiliferous, crystalline limestone.

Contact Relations: Gradational with the fine-grained dolomitic Rockway Member; upper contact is sharp with the brownish-gray shale of the Rochester Formation.

Age: Niagaran Series

Environment: High energy subtidal "reef-like" shoal with common wave agitation and selective winnowing.

ROCHESTER FORMATION

Lithology: While the lower 10 feet are a brown-gray shale, it grades upward into dark blue-gray fossiliferous, calcareous shale with frequent thin-bedded dolomitic limestones that become more prevalent higher in the section.

Thickness: The unit is about 55 feet thick across this region and contains brachiopods, bryozoan, cephalopods and an important ostracode zone.

Contact Relations: Forms a sharp but conformable lower contact with the coarsely crystalline Irondequoit Limestone; the upper contact is a gradational contact into the Decew Dolomite.

Age: Niagaran Series

Environment: The Rochester Shale is the marginal offshore marine shales that grade eastward into the coastal plain deposits represented by the Herkimer Sandstone.

DECEW DOLOMITE

Lithology: An irregular bedded dolomitic shale and fine-grained dolomite with convolute slump structures and some brachiopods.

Thickness: The unit varies in thickness from 8-15 feet in this area and forms the top of the Clinton Group.

Contact Relations: The lower contact is gradational with the Rochester Shale; the upper is sharp but conformable with the Lockport Formation.

Age: Niagaran

Environment: Intertidal and subtidal carbonate mudflats.

Upper Silurian - Lockport Group (Hall, 1839)

The groups consists of four formations, most of which are buff colored, medium-thick bedded replacement dolomites with Stylolites, carbonaceous partings and mineralized cavities. Time correlations are now based upon important conadont zonations (Rickard, 1975).

Lithology: A. basal Gasport Fm. - a coarse-grained, rickly fossiliferous, biohermal limestone and dolomite with brachiopods, corals, bryozoans and stromatopoids (Zenger, 1965). It is 15-30 feet thick.
B. Goat Island Fm. - a sugary, massive, dolomitic limestone with abundant crinoids, brachiopods, and white chert nodules (20-25 feet).
C. Eramosa Fm. - a dark gray, silty, thin-medium bedded bituminous dolomitic limestone with brachiopods and cephalopods (15-20 feet).

D. Guelph - Oak Orchard Fm. - a buff-dark gray, medium to thick bedded, massive bituminous dolomitic limestone with strunatolitic algae corals and oolite horizons (120-140 feet).

Thickness: The Lockport averages 160-170 feet across the region.

Contact Relations: The lower contact with the Decew is sharp, but conformable; the upper contact (drill data) with the Salina Group is conformable.

Age: Niagaran Series

Environment: The association of bioherms, oolites, crinoids suggests a subtidal patch reef environment with well agitated crinoid and oolite banks.

Upper Silurian - Salina Group (Dana, 1863)

The presence of a restricted shallow seaway in the Trade Wind Belt south of the equator caused intense evaporation and the precipitation of dolomite, anhydrite and halite along with red and green shales and siltstones. Although the group is 400-700 feet thick, less than 100 feet are exposed in surface outcrops and most of the information comes from the subsurface analysis of Rickard (1966).

VERNON SHALE

Lithology: The unit consists of red shale and buff dolomite with anhydrite; and some red-green shale and siltstone near the top. The proportion of dolomite, anhydrite and halite increases west of Rochester.

Thickness: Approximately 200 feet of red-green shale occur in the subsurface. The middle contains brachiopods, molluscs, eurypterids and cyathaspid fishes.

Contact Relations: The basal contact is conformable with the dolomitic shales of the upper Guelph Formation of the Lockport; the upper contact is conformable with the Syracuse Formation.

Age: Cayugan Series

Environment: Deposits accumulated in and along the edges of the evaporite marine "Salina Basin" in Illinois, Pennsylvania, New York and Ontario.

SYRACUSE FORMATION

Lithology: Consists of gray shale, buff dolomite and anhydrite, with some salt layers.

Thickness: The Syracuse Formation, though not exposed, is recognized as being about 100 feet thick in this region (Kriedler, 1957).

Contact Relations: The lower contact is sharp but conformable with the Vernon Shale, the upper is conformable with the Camillus Shale.

Age: Cayugan

Environment: Chemical precipitates of anhydrite and halite in the basin, with thin bedded dolomite and anhydrite along the supratidal sabkha along the basin margin.

CAMILLUS FORMATION

Lithology: The Camillus consists of knobby green-maroon mudstone and shale with some zones of dolomite and anhydrite. Red shales with eurypterids occur eastward.

Thickness: The unit is about 80-100 feet thick in this area.

Contact Relations: While the basal member is gradational with the underlying Syracuse Formation, lack of exposure makes recognition of the upper contact difficult, but it may be disconformable with the Bertie Formation.

Age: Cayugan

Environment: Clastic shale and mudstone deposited from the east as nearshore tidal flats into the Salina Basin with the precipitation of evaporites.

BERTIE FORMATION

Lithology: The Bertie consists of three members:

A. basal Falkirk - a massive, dark gray dolomitic limestone with some eurypterids.

B. Scajaquada - a dark shale and marly limestone (waterlime).

C. Williamsville - a laminated, fine-grained dolomite with a conchoidal fracture.

Thickness: The Falkirk, a resistant unit, varies in thickness from 18-25 feet; the Scajaquada is only 4-10 feet thick, and the Williamsville averages 5-8 feet in thickness. The entire Bertie Formation in west-central N.Y. is then between 40-50 feet.

Contact Relations: The lower contact with the Camillus Shale may be disconformable; the upper contact with the Akron Formation is gradational.

Age: Cayugan

Environment: The units represent intertidal and supratidal mud flats along the edge of the Salina Basin.

AKRON FORMATION

Lithology: The Akron is a fine-grained, gray-buff mottled dolomite with cavities from the solution of calcareous corals.

Thickness: In this region the Akron Dolomite is about 6-8 feet thick.

Contact Relations: The lower contact into the Falkirk Dolomite is gradational; the upper contact is disconformable with the lower Devonian Bois Blanc Limestone.

Age: Cayugan Series

Environment: The sequence represents a low energy intertidal and subtidal or lagoonal environment with an erosional surface.

Lower Devonian Stratigraphy

BOIS BLANC FORMATION (Ehlers, 1945)

Lithology: The unit is thin and is usually a sandy calcareous quartz arenite at its base, grading into a dark gray, fine-grained limestone. It is dominated by brachiopods, rugose corals and conodonts.

Thickness: It is discontinuous, lens-like, and varies from 2-3 inches up to 4 feet. Its presence, or the contact relations (see below) provide an important horizon marker in well cores.

Contact Relations: Though six possible contact relations with the overlying coarsely-crystalline crinoidal Onondaga Limestone are possible (Oliver, 1966, p.35) the lower contact forms a disconformity of a few inches and exhibits sandy quartz grains mixing with a fine-grained limestone. The upper contact may again be sandy (another disconformity) and forms a sharp contact with the crystalline Onondaga.

Age: Ulsterian Series - correlates eastward with the Schoharie Formation.

Environment: The irregular appearance of the limestone across a sandy regional disconformity associated with the older Oriskany Sandstone represents the periodic reworking of these sand deposits along the migrating uppermost Lower Devonian strandline with local lagoonal carbonate deposition.

Middle Devonian Stratigraphy

This interval exhibits the last major limestone unit in this part of the Appalachian Basin and demonstrates the effect of basin subsidence in controlling the clastic distal marine sediments of the evolving Catskill Delta Complex.

ONONDAGA FORMATION (Vanuxem, 1843)

Lithology: The Onondaga has been subdivided into 4 members by Oliver (1954). Beginning as a crystalline crinoidal-coralline limestone with numerous reef and reef shoals, it becomes finer-grained and more chert-rich upward. The units are laterally persistent and quite extensively developed.

A. basal Edgecliff Member - a light gray, coarse, coralline limestone with reef-like characteristics that grades vertically into a fine-grained gray limestone with irregular light gray chert nodules (50 feet). Gas is associated with the porous bioherms and reefs in this unit.

B. Clarence Member - a fine-grained gray limestone with abundant dark gray chert nodules - the unit grades eastward (Central N.Y.) into the thin-bedded argillaceous Nedrow Member. The abundance of chert (up to 75%) in this unit has been noted as the "flint bed" in many drillers logs. In this area it grades eastward into the coarse gray reef flank deposits of the Edgecliff bioherms (40 feet).

C. Moorehouse Member - a medium-grained, gray, massive limestone with both light and dark chert nodules, and an abundance of corals and brachiopods (55 feet).

D. Seneca Member - contains the basal "Tioga bentonite" soft, white clay bed of volcanic origin (4-10 inches) that is an important datum marker and key bed in this region. The limestone is a dark gray, massive unit that becomes darker and more argillaceous upward (40 feet).

Thickness: The formation attains a thickness of about 180 feet across the region.

Contact Relations: The lower contact forms a sharp disconformity with the Bois Blanc Formation; the individual members have gradational contacts; the upper contact is gradational into the black calcareous shales of the Marcellus Formation.

Age: Erian Series

Environment: The various carbonate facies with rugose and tabulate corals, crinoid banks, brachiopods, horizontal laminations, and planar cross-beds, all indicate features associated with reef, reef flank, carbonate shoals, and subtidal carbonate environments.

Middle Devonian - Hamilton Group

MARCELLUS FORMATION (Oatka Creek)

Lithology: The Oatka Creek Member consists of thin bedded black and blue-black fissle, bitumenous shale with some brachiopods.

Thickness: The entire Marcellus Formation consists of about 50 feet of the Oatka Creek Shale.

Contact Relations: The lower contact is gradational with the Onondage Limestone, and contains abundant pyrite nodules; the upper contact is sharp and conformable with the Stafford Limestone.

Age: Erian Series

Environment: Restricted marine environment because of effect of basin subsidence; sediments indicate offshore clays coming from the Catskill delta complex.

SKANEATELES FORMATION

Lithology: The Skaneateles exhibits an upward coarsening sequence beginning with a fine-grained black limestone (Stafford Member) that grades upward into blue-black and gray shales of the Levanna Member. The Levanna becomes coarser and medium-bedded higher in the section.

Thickness: The Stafford is 3-4 feet thick and is overlain by about 50 feet of gray Levanna Shale.

Contact relations: There is a sharp contact with the massive, dark gray Stafford Limestone and a similar contact with the overlying thick bedded, medium-grained Centerfield Limestone.

Age: Erian Series

Environment: Transgressive-regressive succession across the distal prograding slope of the Catskill Delta.

LUDLOWVILLE FORMATION

Lithology: The Ludlowville consists of four members that represent clearwater carbonate sections bounding a thick interval of gray shales.

A. The basal Centerfield limestone is a gray, thick-bedded medium-coarse-grained fossiliferous limestone. (6-8 feet)
B. The Ledyard Member is a thin bedded or knobby gray-black shale with fossiliferous pyrite concretions (26-30 feet).
C. The Wanakah Member is a calcareous, concretionary, blue-gray shale with abundant tabulate corals along its base (35 feet).
D. The Tichenor Member is a light gray, crinoidal limestone (8-10 feet).

Thickness: The entire Ludlowville Formation is about 75-80 feet through this region.

Contact relations: The Centerfield Limestone produces a sharp but conformable contact with the black shales of Levanna Member, and the Tichenor Limestone forms a sharp, slightly disconformable contact with black shales of the Kashong Member.

Age: Erian Series

Environment: The Ludlowville represents a periodic clearing of the seas during the otherwise continuing progradation of the marine slope of the Catskill Delta.

MOSCOW FORMATION

Lithology: The Moscow consists of thin-bedded black shales of the Kashong Member (6 feet) grading up into the knobby gray shales of the Windom Member (30 feet).

Thickness: The Moscow here forms a thickness of about 35 feet.

Contact relations: The lower contact is sharp and disconformable with the coarse-grained crinoidal Tichenor Limestone, while the upper contact is also disconformable with lenses of the Leicester Pyrite and dark gray Penn Yan Shale.

Age: Erian Series

Environment: Moderate basin subsidence gradually produces a more restrictive marine environment.

TULLY FORMATION

Lithology: The Tully Formation is represented by the thin-bedded black irregular lenses of the Leicester Pyrite and, while present eastward, is represented by a disconformity across this region.

Thickness: Interval is represented by a few inches of pyrite nodules in black shale.

Contact relations: Irregular lenses of pyrite occur between the knobby dark gray calcareous shales of the Windom and the black fissle shales of the Penn Yan Formations.

Age: Erian Series

Environment: Restricted reducing environment associated with limestone deposition along the delta front shoals and non-deposition in the deeper part of the basin.

Upper Devonian Stratigraphy

The Upper Devonian section in this region, a frequent reservoir for local oil and gas, consists of nearly 2500 feet of interbedded marine sandstones and shales that form an upward coarsening sequences through the Senecan and Chautauquan Series to form the Genesee, Sonyea, West Falls, Canadaway, Conneaut, and Conewango Groups.

The subdivisions are based on the recognition of rhythmic units of black and gray shales and thin interbedded turbidite siltstones that thicken eastward to form the lower slope of the Catskill deltaic complex. The base of the Genesee, Sonyea and Westfalls Formations are each marked by basal black shales associated with regional basin subsidence (Geneseo, Middlesex and Rhinestreet) that are then overlain by prograding gray shales and siltstones (Penn Yan - West River, Cashaqua, and Angola) of the Catskill Delta. The best key horizons as marker beds include the base of some of these shales or the associated siltstones.

GENESEE GROUP: This sequence consists of three formations:

Lithology:
A. basal Geneseo and Penn Yan black and dark gray shales that are sparsely fossiliferous and grade laterally into each other (2 inches to 3 feet).
B. Genundawa Formation - a gray, thin bedded, fine-grained limestone (2 inches to 2 feet - a marker horizon).
C. West River - a gray to dark gray fissle shale (15 feet).

Thickness: The Genesee Group varies in thickness from 10 to 20 feet.

Lower contact: Defined by the presence of the irregular Leicester Pyrite lenses that form a non-depositional disconformity during the development of the Tully Formation elsewhere (a key bed).

Upper contact: The dark gray shales of the West River grade vertically into the black shales of the Middlesex Formation.

Age: Senecan Series

Environment: The sequence repeats the offshore and distal delta slope environments of the Marcellus Formation, and represents the second major stage of progradation for the Catskill delta.

SONYEA GROUP: (Colten and deWitt, 1958) This sequence consists of two shale formations.

Lithology: A. the basal Middlesex Formation is a black fissle shale (6 to 8 feet) that grades upward into B. the gray Cashaqua Shale which varies in thickness from 45 to 75 feet.

Thickness: The Sonyea Group varies in thickness from about 50 to 80 feet across this region.

Contact relations: The lower contact is gradational with the dark gray shales of the West River Formation, while the upper is gradational with the black fissle shales of the Rhinestreet Formation.

Age: Senecan Series

Environment: The sequence repeats the offshore and distal delta slope environments of the Genesee Group and reflects the third major period of progradation for the Catskill Delta.

WEST FALLS GROUP: (Pepper deWitt and Colten, 1956) This sequence now consists of three shale formations.

Lithology:

A. the basal Rhinestreet Formation is a black fissle shale that is 150-200 feet thick.
B. Angola-Nunda Formations are a series of gray shales and siltstones with interbedded thin limestones and calcareous siltstones. The Angola is the more western equivalent of the Nunda. The units vary in thickness from 220 to 365 feet.
C. Hanover Formation (Java Formation of Pepper and deWitt, 1950) consists of the Pipe Creek Member - a unit only 1-2 feet in thickness with conodonts and carbonized plants; and the Hanover Shale - a unit of gray shales with interbedded thin limestones that is 85-95 feet thick and a continuation of the Angola Shale lithology.

Thickness: The West Falls Group varies in thickness from 450 to nearly 650 feet across the area.

Contact relations: Along the lower contact the black shales of the Rhinestreet are gradational into the gray shales of the Cashequa Formation; at the upper contact the gray medstones and siltstones of the Hanover Formation are gradational with the black fissle Dunkirk Shale.

Age: Senecan Series

Environment: The sequence illustrates the distal and proximal delta slope environments of the Catskill Delta, and represents the fourth major interval of clastic progradation into the subsiding basin.

CANADAWAY GROUP (Caster, 1934) This sequence consists of six formations and represents the final pulse of deltaic progradation during the Devonian Period.

Lithology:

A. basal Dunkirk Formation - a black fissle shale with carbonized plant remains and conodonts, about 120 feet in thickness. Its eastern equivalent is the gray shales and massive siltstones of the Caneadea Formation.
B. Gowanda Formation - a succession of gray and black shales with interbedded massive but thin gray siltstones and concretions, varying in thickness from 120 to 230 feet.

C. Laona Formation - a gray massive siltstone with a unique assemblege of brachiopods and pelecypods. It varies from 3-25 feet in thickness and forms a series of elongate lenses over the area.
D. Westfield Formation - a series of gray shales with a few thin-bedded siltstones and a sparse marine fauna that varies in thickness from 100 to 220 feet.
E. Shumla Formation - a gray massive siltstone similar in texture and form to the Laona Formation but without the distinctive fauna. It attains a thickness up to 35 feet.
F. Northeast Formation - a thick series of gray shales and interbedded siltstones in which the siltstones become thicker and more numerous higher in the section. The unit varies in thickness from 400 to 600 feet.

In Cattaraugus County units B-F cannot be separated, and this thick 1000-foot sequence of undifferentiated gray shale and interbedded siltstones is referred to as Forty Bridge Formation (Rickard, 1975) in the Salamanca region.

Thickness: The Canadaway Group nearly reaches a thickness of 1000 feet across this region.

Contact relations: The black shales of the Dunkirk form a gradational lower contact with the gray shales of the Hanover Formation; along the upper contact the gray shales and siltstones of the Northeast Formation now grade into the massive fine grained sandstones and brown-gray knobby shales of the Dexterville Formation.

Age: Chautauquan Series

Environment: The sequence reflects the proximal delta slope environments of the Catskill Delta and represents the fifth and final pulse of progradation into western New York.

CONNEAUT GROUP (Rickard, 1975) This sequence represents the first appearance of the proximal delta slope of the Catskill Delta in this region, and consists of two formations.

Lithology:
A. basal Dexterville Member consists of a series of gray knobby shales and interbedded gray siltstones with the brachiopod *Pugnoides duplicatus* as an index fossil (100 feet).
B. Ellicot Member is very similar in lithology but without the characteristic brachiopod (150 feet).

In Cattaraugus County where *Pugnoides* is absent the members are not differentiated and the lateral equivalent is the Chadakoin Formation (250 feet).

Contact relations: Since the lithologies are similar, the lower contacts are gradational; the upper contact occurs at the base of the Panama and Wolf Creek Conglomerate.

Age: Chautauquan Series

Environment: As noted, the presence of shales and interbedded fine-grained sandstones indicates the initial presence of a coarser sequence representing the proximal delta slope environments.

CONNEWANGO GROUP (Butts, 1908) This sequence represents the marine and non-marine environments of the Catskill Delta.

Lithology: Gray shales and siltstones with beds and lenses of tan sandstone and conglomerate, red mudstones, gray crossbedded sandstones. (Cattaraugus Facies of Fisher & Rickard, 1975)

A. basal conglomerates (Wolf Creek and Panama) occur as layers and lenses in crossbedded tan sandstones (6 feet)

B. Venango Member - consists of knobby gray shales interbedded with gray-dark gray siltstones with marine fossils (pelecypod Ptychopteria)

C. Cattaraugus Member - contains a varying lithology of shales, siltstones and sandstones (gray, green and red) as layers and lenses with Calcareous nodules, dessication cracks, and plant fossils.

Thickness: The group thins westward, but averages 650 feet in thickness.

Contact relations: Because of the varying lithology, the contact relations also become more variable and the units are separated by several attributes (i.e. facies) as defined by Rickard (1975).

Age: Chatauquan Series

Environment: Marine-non marine fluctuations from changes in position of deltaic lobes and differential basin subsidence to produce delta platform sands, river mouth base, tidal and fluvial channels, mud and overbank deposits.

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GEOLOGIC AND ENGINEERING HISTORY OF PRESQUE ISLE PENINSULA, PA

Joan Pope and Richard J. Gorecki - Coastal Engineering Section,
Buffalo District, U.S. Army Corps of Engineers

INTRODUCTION

Presque Isle is a unique and significant coastal feature on the south shore of Lake Erie at Erie, Pennsylvania. It is a compound, recurved sandspit that arches lakeward about two and one-half miles from an otherwise straight shore (Figure 1). The peninsula has a lake shoreline of about six and one-quarter miles from its narrow connection with the mainland to its distal end where it turns sharply shoreward. It is the only major positive depositional feature along the generally sand-starved south shore of Lake Erie. Presque Isle Peninsula is an old-age geomorphic feature which is migrating eastward into deeper water, thereby resulting in a net annual loss to the sand body. The processes responsible for the geological evolution of this feature will also be responsible for its eventual destruction unless attempts are undertaken to permanently stagnate its migration. The history of coastal engineering measures for shore protection has been played out on the peninsula beaches as man has employed a myriad of engineering efforts dating back to the early 1800's for the purpose of preservation of this migrating and diminishing feature. The peninsula is truly a rare ecological laboratory that allows the process of primary plant and animal succession to be studied in habitat diversity ranging from pioneer vegetation on newly formed shore zones to climax woodland communities on old beach ridges, all within a distance of about three miles. The peninsula is developed as a State park and is a popular recreational area which provides facilities for bathing, boating, hiking, fishing, bird watching, picnicking, and other recreational opportunities. The public has free and unrestricted access to the park and approximately 3,800,000 persons have visited the park annually for the past 10 years.

In 1922, Presque Isle Peninsula was conveyed from the Federal Government to the Commonwealth of Pennsylvania for park purposes, and the care and protection of the peninsula was shifted from prevention of breaches through the peninsula for the purpose of preserving Erie Harbor to the purpose of providing recreational beaches. In 1956, the Federal Government, in cooperation with the Commonwealth of Pennsylvania, completed an erosion control project on Presque Isle Peninsula. Since that time, the project has proven to be inadequate, and sand replenishment measures have been required in order to protect the Federal structures and State's park facilities. The Commonwealth of Pennsylvania, in 1968, requested the Corps of Engineers to make a complete restudy of the Presque Isle beach erosion control project in order to develop a more effective and permanent solution to the erosion problem. This field trip manuscript contains a small portion of the geological and engineering background which was developed by the authors in support of the

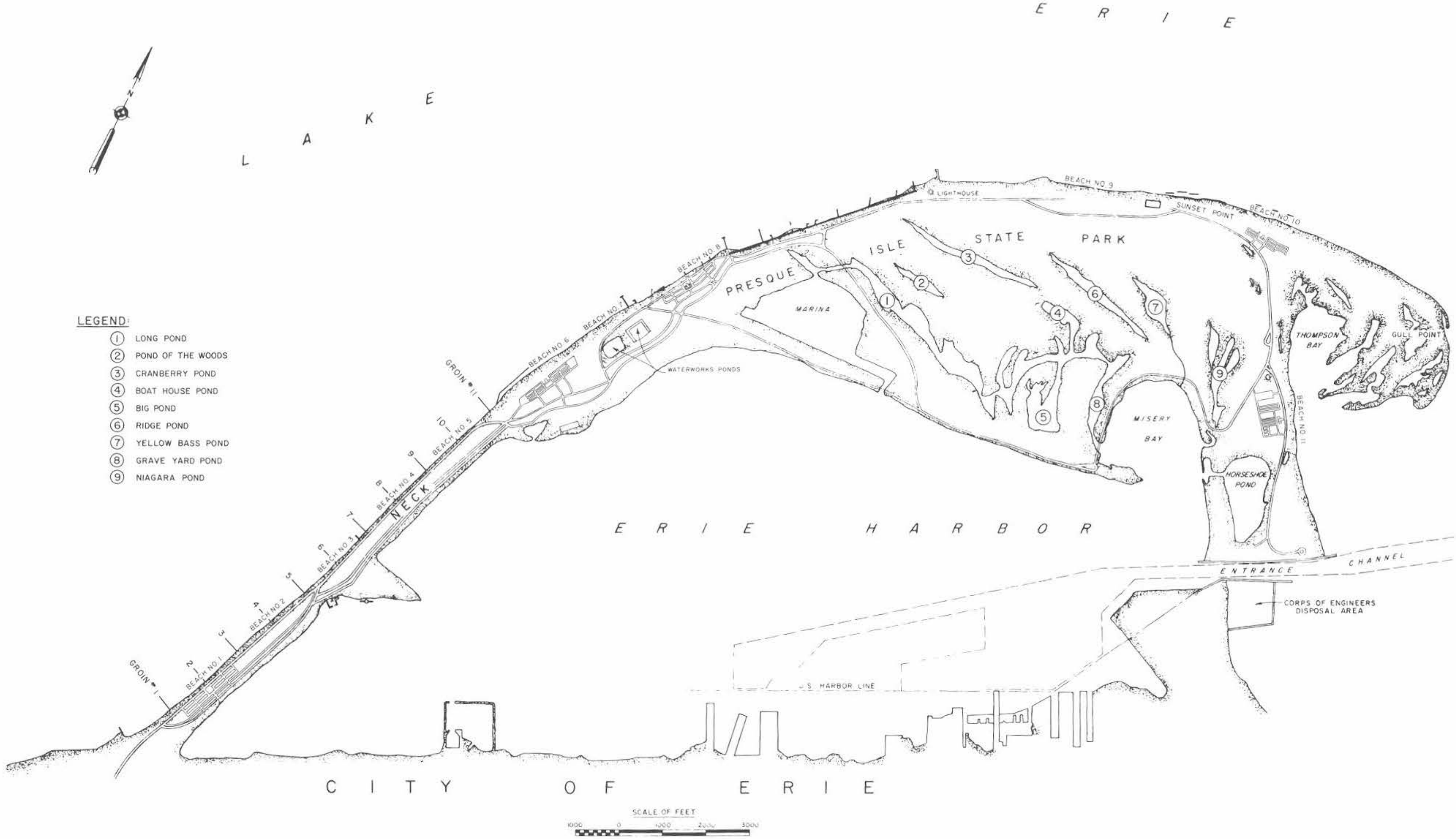

FIGURE I. LOCATION OF BEACHES AND PONDS ON PRESQUE ISLE PENINSULA

1980 General Design Memorandum, which the Corps prepared for Congressional review and approval. In the interest of brevity, many of the complex geologic environmental, engineering, and socioeconomic issues which were part of the Corps study, could not be reproduced herein. This paper is taken from the context of the official study and its purpose is purely academic and is designed to enlighten the reader by providing an understanding of the fascinating geologic evolution of Presque Isle Peninsula and the history of man's attempts at stabilization. Should the reader desire additional insight into the Corps study, he/she is directed to the 1980 General Design Memorandum.

SITE DESCRIPTION

Presque Isle Peninsula, from its mainland root to its distal end where it turns sharply shoreward, is about six and one-quarter miles long. The eastern end of the peninsula terminates in several low, flat, recurving longshore bars. For a distance of about two miles from the westerly root, the peninsula is narrow and has an average width of generally less than 800 feet. This narrow section of the peninsula is called the neck. East of this narrow neck, the peninsula widens abruptly to a width of over one mile. Presque Isle Peninsula consists almost entirely of fine sand reworked from glacial deposits. The general ground elevation of the peninsula is relatively low, averaging about seven or eight feet above low water datum (LWD) which for Lake Erie is elevation 568.6 feet above mean water level at Father Point, Quebec, International Great Lakes Datum (IGLD 1955). There are four major and several minor beach ridges which extend across the peninsula, generally in an east-west direction and rise to a maximum elevation of about 20 feet above low water datum. The higher ground on the peninsula sustains a thick growth of a wide variety of trees and shrubs. The low areas between the beach ridges are comprised of several elongated lagoons and marshes.

The lakeward perimeter of Presque Isle is about nine miles. The lakeward shoreline has been segmented into 11 bathing beaches by the Pennsylvania State Park Service. These beaches vary in width and, with the exception of Beach No. 11, have had a history of serious erosion for at least 150 years. The bathing beaches are backed by picnic areas, and four major beach areas are provided with bathhouse and parking facilities. Roadside parking provides easy access to intervening beach and picnic areas. Numerous protective works consisting of groins, revetments, bulkheads, and offshore breakwaters have been constructed to halt erosion. The bay shoreline is characterized by numerous small bays, coves, and inlets. Encircled between the peninsula and the mainland is Presque Isle Bay, the easterly part of which has been improved as Erie Harbor. The north jetty for the Erie Harbor entrance channel is joined to the distal east end of Presque Isle Peninsula.

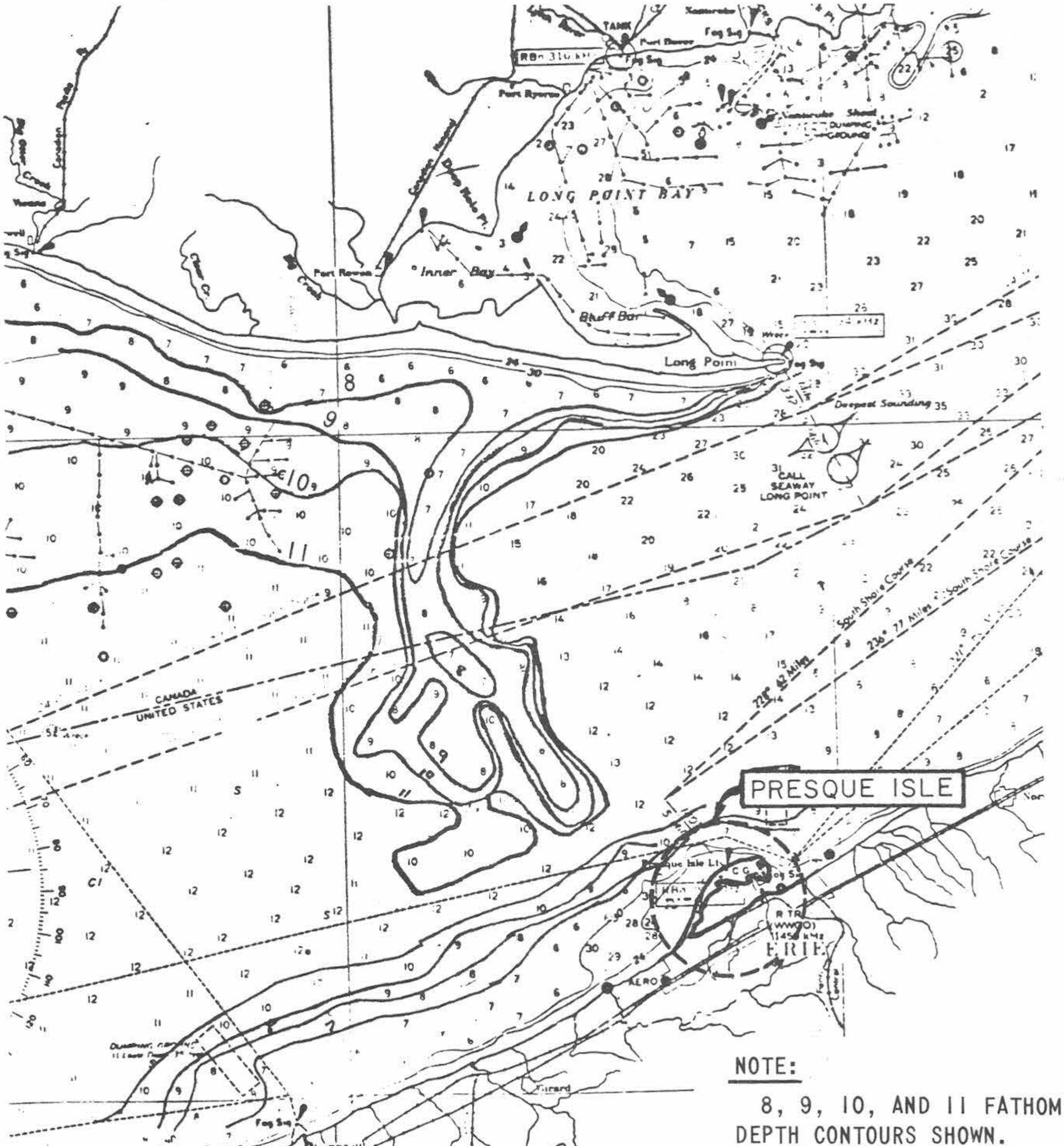

FIGURE 2. LAKE SURVEY CHART NO. 3 ILLUSTRATING THE LONG POINT - ERIE MORAINAL RIDGE.

GEOLOGIC SETTING

Physiography

The major physiographic divisions in northwest Pennsylvania are the eastern lake section of the Central Lowland Province and the glaciated section of the Appalachian Plateaus Province. The eastern lake section is a two to five-mile wide plain bordering Lake Erie. Bluffs along the Lake Erie shore in Pennsylvania are greater than 80 feet in height and are composed of glacial and lacustrine deposits. Bedrock is often found at the base of the bluffs. Sandy beach ridges, representing post-glacial lake strands, cross the lake plain on top of the bluffs. The topography of the glaciated section of the Appalachian Plateaus Province is that of an eroded plateau with gently rolling hills.

Bedrock

Bedrock exposed in Erie County, Pennsylvania, is predominantly Upper Devonian shales and siltstones of the Conneaut and Canadaway Groups. At Presque Isle, there is a lakeward slope of the rock surface with contours parallel to the mainland. At the junction of the neck of the peninsula with the general shore, the bedrock surface is only two feet below low water datum. A gas well drilled near the northeast corner of the Waterworks ponds shows rock to be about 112 feet deep. Borings taken in 1965 by a consulting firm for the State of Pennsylvania, extended in a line across the harbor entrance channel and showed that the rock surface sloped lakeward with a 1 on 125 slope with a depth of approximately 60 feet below LWD near Beach No. 11 (Rummel, Klepper & Kahl-Fertig Engineering Comany, 1968). The bedrock here is likely to be the gray shale of the Portage Formation.

Lake Erie Basin Deposits

Lake Erie can be divided into three separate subbasins. Presque Isle is located at the eastern end of the central basin. The bathymetry of the lake is mostly controlled by lithology and dip of bedrock. Superimposed on the bedrock are Pleistocene and recent deposits. The most prominent glacial features in the lake, are three ridges which traverse the lake between Pelee Point - Lorain, Erieau - Cleveland, and Long Point - Erie. These are thought to be end moraines and are composed of clay till veneered with sand or gravel (Lewis, 1966). The Long Point-Erie Moraine, largest of the three, is broad, flat-topped, and about 40 Km (25 miles) wide (Figure 2). Coring studies, conducted in 1978 by the Coastal Engineering Research Center of the Corps of Engineers, indicate that the sand and gravel overlying the moraine on the United States side is as much as 12.7 feet thick and averages about 7.4 feet. Seismic profiling shows the sand to be 15 to 20 feet thick along the ridge surface. Recent soft, gray mud covers most of the rest of the central basin. In some areas, the mud is 60 to 80 feet thick (Lewis, 1966).

Surficial Deposits

The surficial deposits of northwest Pennsylvania are dominated by the glacial history of this area. During the Pleistocene Epoch, a series of glacial advances and retreats modified the landscape and deposited material. Glacial deposits on the mainland consist of till and stratified drift. The till units are variable in texture and found in hilly end moraines and as ground moraine blanketing much of the area. The stratified deposits are in the form of kames and outwash. Petrographic analysis of the stratified deposits show them to be composed of hard and tough sandstone, siltstone, limestone, dolomite, quartz, and quartzite particles. Strand deposits of Glacial Lakes Whittlesey and Warren also consist of sand and pebble gravel. These deposits, formed about 12,800 years ago (Schooler, 1974), have not been found to be suitable for use as beachfill because of a predominance of shale and siltstone fragments.

Glacial History

The Late Wisconsin stage left the greatest impacts on the topography and the deposits of this region and starts the evolutional trail toward the existance of modern Presque Isle Peninsula. The earliest event of the late Wisconsin significantly affecting the project area occurred about 20,000 years B.P. during the Kent Phase. Deposits of Kent drift include till and stratified drift in the form of kames, crevasse fillings, and outwash. The main characteristic of the Kent Advance is extensive kame deposition. These are found on valley bottoms or perched on valley walls. Most of the sand which has been used in recent years for beach replenishment at Presque Isle is derived from these deposits. During the next event, the Lavery Phase, a glacier advanced to a location marked by the Lavery End Moraine. This occurred about 17,000 years B.P. The surface expression of this deposit varies from smooth hills and swales to moderately hummocky topography. Shepps and others (1959) have mapped morainal kames in locations where the Lavery moraine crosses valleys. Kames and outwash, deposited in valleys, supply some of the sand used for beach replenishment at Presque Isle.

After the Lavery advance, Fullerton (1971) believes that the ice margin retreated as far northeast as Toronto, Ontario, and he refers to this period as the Lake Erie Interval (approximately 15,500 years B.P.) during which both Lakes Erie and Ontario drained eastward through the Mohawk Lowland in New York. A glacial readvance in Port Stanley time (15,000 years B.P.) resulted in the deposition of the Hiram Moraine. Kames were not as well-developed as during the preceding Kent and Lavery advances. Outwash deposits also are not as extensive.

The last glacial advance into northwestern Pennsylvania, according to Shepps and others (1959), and White and others (1969) was the Ashtabula Advance. Fullerton (1971) shows this to have begun 14,100 years B.P. Its limit is marked by a series of end moraines exhibiting knob and kettle topography. Kames are more common in the eastern portion of the moraine than in the western portion. Outwash occurs between the ridges.

The next major event of the Pleistocene is known as the Cary-Port Huron Interval when the ice margin was north of the Lake Erie Basin. At this time, a series of glacial Great Lakes developed in the Erie Basin. Strand lines of Lakes Maumee I, II, III, and Arkona were fairly well-developed in the western portion of the basin but are faint or absent in the eastern part (Leverett and Taylor, 1915). These lakes drained westward, outletting at Ft. Wayne, IN, through the Wabash River and also through the Huron Basin (Hough, 1958).

At 12,900 years B.P., a major glacial readvance, known as the Port Huron Advance, took place resulting in a rise of water in the Erie Basin to form Glacial Lake Whittlesey (Calkin, 1970). The Long Point-Erie Moraine of Lake Erie has been correlated with the deposits of the Port Huron Advance by Lewis (1966), Wall (1968), and Fullerton (1971).

Features of Lake Whittlesey can be found in the vicinity of Presque Isle at an elevation of about 735 feet above mean sea level (MSL). The Whittlesey strand occurs as a 10-foot high wave-cut cliff near the Pennsylvania-Ohio State line. About a mile east, it becomes a 15-foot high, gravelly ridge and then changes to a series of sand dunes south of West Springfield, PA. Across the rest of Erie County, PA, it is a well-defined ridge 15-20 feet high with a steep north slope and gentle south slope. East of Erie, the ridge is replaced by two low, wave-cut cliffs cut in glacial material and bedrock (Schooler, 1974).

Further retreat of the Port Huron glacier resulted in a series of lower lakes. The most important of these is Lake Warren which is evidenced as two ridges occurring at elevations of 725 to 735 feet and 715 to 725 feet (Schooler, 1974).

After the ice had retreated north of the Niagara Escarpment, water in the Erie Basin was allowed to drain into the Ontario Basin. Due to crustal depression caused by the weight of glaciers, the outlet at the escarpment was relatively much lower than the present outlet at Niagara Falls. The lake occupying the Erie Basin at this time was at an elevation of 470 feet MSL, approximately 100 feet lower than today. This stage known as Early Lake Erie existed between 12,370 and 12,790 years B.P. (Lewis and others, 1966). It was during this time that Lewis (1966) and Lewis and others (1966), believe that the sand and gravel overlying the Long Point-Erie Moraine developed.

As the outlet of Early Lake Erie was uplifted by crustal rebound, the elevation of the water surface was raised to its present level. Wave erosion of bluffs along the present shore and streams in addition to the Long Point-Erie Moraine, contributed sand and gravel for the development of beaches and the original Presque Isle sand body.

Modern Lake Erie

The water levels in the Lake Erie Basin have changed much in post-glacial times. This is due to crustal uplift, climatic changes, and diversion of water. The present outlet, the Niagara River, is controlled by a bedrock threshold at Buffalo, NY. During glacial times, this was blocked by ice, and lake water was diverted through higher outlets such as the Wabash, Grand, and Mohawk Rivers. After glacial retreat, the Niagara outlet was opened, but due to crustal downwarping caused by the weight of glaciers, this outlet was more than 100 feet lower than today.

Early investigators (Leverett and Taylor, 1915, and others) determined the differential uplift in the region by comparing the elevations of southern beaches with northern beaches of the glacial Great Lakes. They found that the beaches are horizontal to a point, known as a hinge line, from which the beaches rise vertically to the north. For example, Lake Whittlesey beaches are at an elevation of 735 feet (MSL) throughout most of Ohio and Pennsylvania, but starting at a point east of Erie, PA, they begin to rise up to an elevation of 910 feet (MSL) in New York State (Leverett and Taylor, 1912).

In another study of water levels, Lewis (1969) compared radiocarbon dates with known lake levels and developed the diagram shown as Figure 3. This shows the rate of change in water level in the Erie Basin during post-glacial time. Lewis prefers to use the curve near the upper envelope. If the lower curve is adopted, it would mean that levels in the eastern basin of the lake would have been lower than the channel along the southern margin of the Long Point-Erie Moraine for more than 1,500 years. Lewis' diagram also shows the steep rise of water from 5,000 to 3,800 years B.P. This initial rise corresponds to the abandonment of the North Bay transfering more flow into the lower Great Lakes.

GEOLOGY OF PRESQUE ISLE

INTRODUCTION

The observed sediment transport patterns at Presque Isle are the result of a modern wave climate acting on the glacial and post-glacial deposits of the area. Glacial deposits, some reworked during post-glacial lake level fluctuations, serve as the source for the littoral material. Lake level fluctuation and drainage pattern changes have been frequent in post-glacial time (for the past 12,000 years) and are responsible for denudating the glacial topography and producing many of the present, onshore, offshore, and coastal features including Presque Isle Peninsula. However, Presque Isle is a unique feature. It is the only major positive depositional feature along the southern shore of Lake Erie, and any explanation of its existence must be tied to some specific geologic event.

An understanding of the origin and historical development of Presque Isle Peninsula is necessary in order to understand the processes currently at work and to predict the future condition. Thus, the following discussion concerning the post-glacial development of Presque Isle is presented only as a brief overview in order to provide a better understanding of the observed condition. This discussion is hypothetical and, although it fits with the existing glacial information and theory, has not been rigorously tested.

HISTORICAL ORIGIN

In order for Presque Isle Peninsula to exist prior to recent lake levels, there must have been a substantial source of sand and a reason for that sand to collect in one area. The existence of the platform to the west of Presque Isle may very well be the key which explains how Presque Isle Peninsula evolved (Figure 2). The platform has a total length of 12 miles of which the eastern five miles is currently covered by the peninsula. its average width is about three to three and one-half miles from the average depth is 25 to 30 feet below LWD. Map documentation from the past 150 years shows that the sand of Presque Isle does migrate from west to east across this platform, building a new platform to the east as it moves in that direction. The origin of the platform can be explained as a total sand terrace which has been wave planed by rising lake levels or as a preexisting topographic high rock or glacial till which served as the original base for Presque Isle and was added to as the peninsula grew. Preliminary review of data collected in 1977 and 1978 by the Coastal Engineering Research Center suggests that the western end of the platform is underlain by till (S. J. Williams, personal communication, 1979). If the original platform at the western end is composed of glacial morainal till it is probably the southern end of the Long Point-Erie ridge (Figure 2) which has been traced to the Post Huron-glacial advance (12,800 $\pm$ 250 years B.P.). Hough (1958) describes the moraine as a distinct ridge on the bottom of Lake Erie lying west of the eastern deep basin, emerging on the south side of the lake where it extends eastward into New York as the Lake Escarpment Moraine System (Messenger, 1977). The surface of this moraine, both the ridge and the platform, was probably planed by wave action during lower lake levels, and the silts and clays were carried offshore leaving a lag deposit of sand and gravel. The platform lag deposit was well-sorted by wave action and possibly served as a depositional area for littorally transported material during the Early Lake Erie stage. As lake level rose to approximately 25 feet below today's lake level, about 4,000 years ago, littoral currents transported the sand on the platform toward the east, remolding it into an elongated sand beach. This historical sequence is described in Table 1 and shown on Figure 3.

Migration caused by waves from the west and rising lake levels caused the sand body to move toward the east side of the morainal root. As sand slumped off of the east side of the moraine, a sand platform was

Table 1 - HYPOTHETICAL CHRONOLOGY OF PRESQUE ISLE ORIGIN*

Period (Years B.P.)	Event	Discussion
12,900	Port Huron Advance	Long Point-Erie Moraine formed.
12,500-11,500	Early Lake Erie	Rapidly rising lake level from 120' to 60' below current LWD.
11,500-10,000	Early Lake Erie	Slower rising lake level (from 60' to 50' below current LWD). Crest of Long Point-Erie Moraine planed by rising lake level, beach deposits, and dune field develops from lag deposit.
10,000-4,500		Slowly rising lake level (from 50' to 40' below current LWD). Long Point-Erie Morainal ridge inundated.
4,500-3,500		Rapid rise in lake level (from 40' to 10' below current LWD). Platform of Presque Isle (landward extension of the Long Point-Erie Moraine is subjected to wave attack sand and gravel lag deposit from till released as source material for Presque Isle.
3,500 to present	Modern Lake Erie	Lake level rises at approximate rate of 1 foot per 300 years. Modern Presque Isle evolves as it migrates to the east.

*Based on the historical Lake Erie water levels presented in Lewis (1969) and on a hypothetical development sequence for Presque Isle.

QUATERNARY OF LAKES HURON AND ERIE

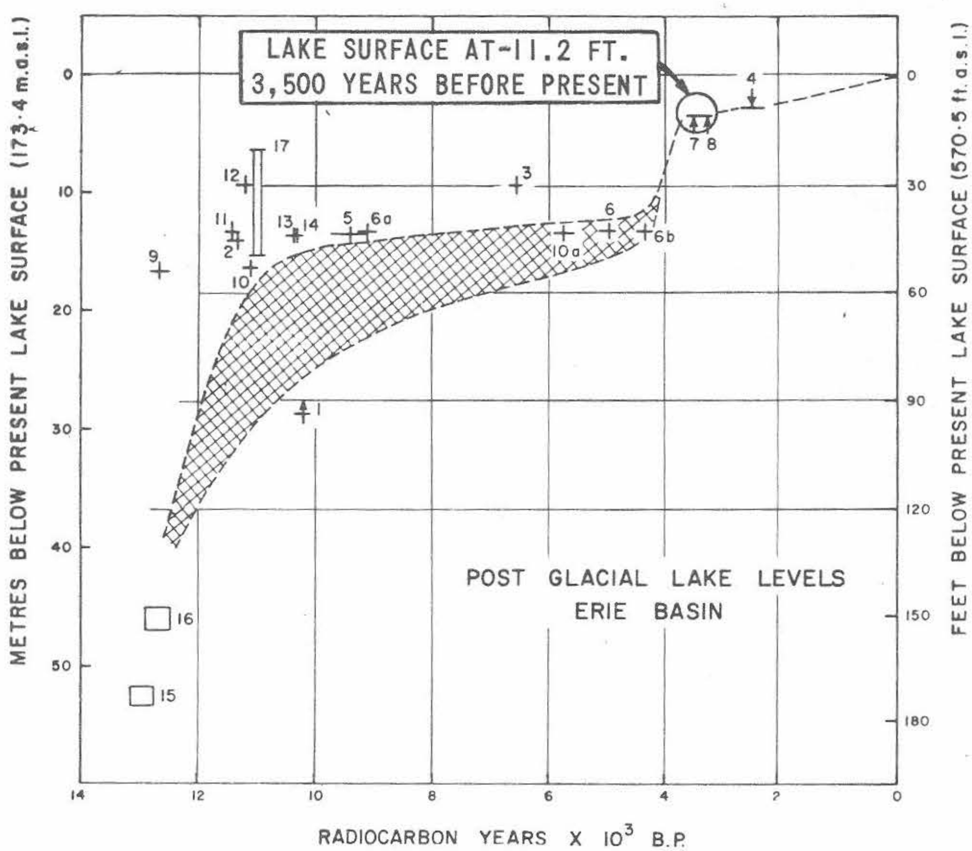

FIG. 12. Post glacial history of Lake Erie water levels. Points are keyed numerically to entries in Table 1.

FIGURE 3. POST GLACIAL HISTORY OF LAKE ERIE WATER LEVELS. LEWIS 1969

built. The feature we recognize as Presque Isle Peninsula evolved as it migrated onto this sand platform. As the sand platform built, the sand volume available for transport diminished. How much of the platform is till and how much is sand are unknown at this time, but subsurface foundation studies planned for the Phase II General Design Memorandum stage of the Corps of Engineers project may provide additional information on the platform and the formation of Presque Isle Peninsula.

Modern Coastal Processes - Migration

The west to east migration of Presque Isle has long been recognized (Figure 4). Presque Isle Peninsula was originally surveyed in 1819. In 1824, the original Erie Harbor project included action as needed to maintain the integrity of Presque Isle Peninsula in order to assure the harbor's future success. Since then, the migratory character of the peninsula has become very evident as erosion and breaching of the neck has demanded continual attention and as accretion at the east end of the peninsula has required jetty extension and dredging to remove shoal buildup in the entrance channel.

Evidence of long-term migration before Federal involvement with Presque Isle is clearly defined by the morphology of the peninsula's internal features, the platform to the west, and the shoreline of the mainland. A comparison of the sheltered shoreline inside Presque Isle Bay to the open shoreline east and west of Presque Isle Peninsula shows no offset. The bay shore should be a positive shoreline and be characterized by a gently sloping shore if it had experienced long-term sheltering by the peninsula. This is not the case. The shoreline is continuous from the west, through the bay, and to the east. The bay shore is characterized by steep, wave-cut bluffs identical to those outside the bay. The sequence of beach ridges, elongated beach ridge ponds, and fingering distal end ponds is repeated and preserved within the interior of the peninsula, documenting previous stages in Presque Isle's migration. The unknown factor is what has been the change in shape and size as Presque Isle has migrated.

The presence of relict features in Presque Isle Peninsula documents the migration from west to east and a continuation of the same general pattern and process of evolution. Presque Isle Peninsula has probably developed in cycles in order for the specific depositional features to be preserved. We can witness the yearly cycle and the long-term cycles of growth related to annual lake level fluctuations, but Presque Isle Peninsula may also be influenced by longer period climatic patterns about which we have no knowledge. High lake levels increase transport rates causing rapid loss of material from the neck area and rapid growth of the distal east end as sand is fed to the growing eastern platform. During lower lake levels, the distal east end matures as the bars are recurved and become subaerial and new material enters the system at the neck, partially healing the eroded areas and widening the neck.

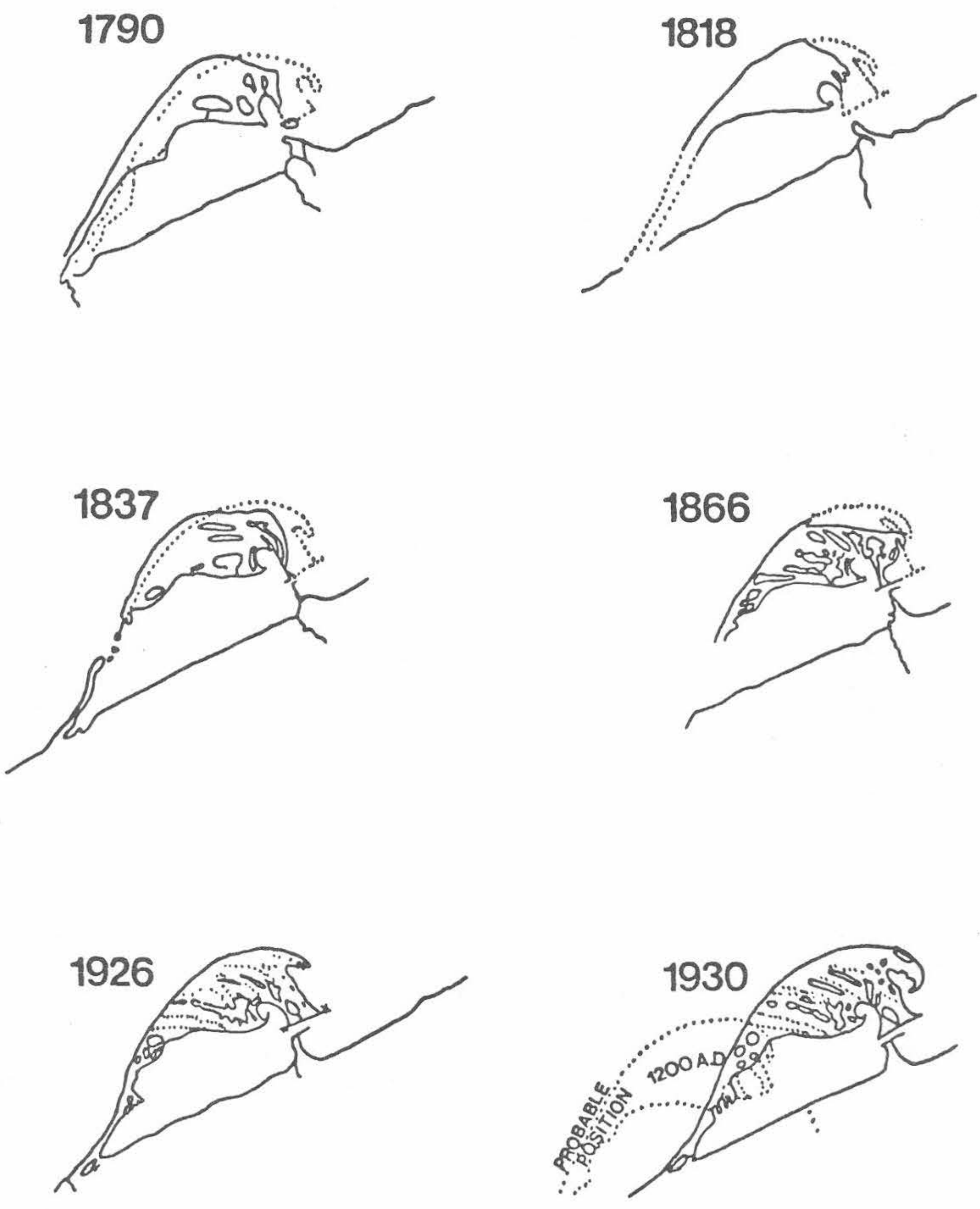

FIGURE 4. GROWTH AND MIGRATION OF PRESQUE ISLE

NOTE: THE DOTTED LINE ON THE 1790 THROUGH 1866 SKETCHES REPRESENTS THE 1930 POSITION OF THE PENINSULA. THE DOTTED LINE SHOWN ON THE 1930 SKETCH DOCUMENTS THE EASTWARD MIGRATION RELATIVE TO A 1200 A.D. POSITION AS HYPOTHESIZED BY JENNINGS, 1930.

The beach ridges evolve as the offshore bars migrate onshore and weld onto the shore as a subaerial bar. They probably build in height as they migrate onshore in response to the steeper waves of the surf zone. Sand is deposited in front of the bar; a lagoon is trapped behind it. Cottonwoods and other vegetation take root on the beach ridge, and dunes build on top of the ridge increasing its height to about 20 feet above low water datum. Low areas behind and between the beach ridges are submerged and appear as a series of elongated ponds oriented WNW-ESE. Examples of these ridge ponds are Long Pond, Cranberry Pond, and Ridge Pond (Figure 1). The recurving offshore bars at the distal east end form a finger shaped array of ponds which are oriented north-south. These distal ponds include Big Pond, Yellow Bass Pond, and Niagara Pond (Figure 1). The Presque Isle system is an eastward migrating system which feeds upon itself as it migrates. Within the system, material is eroded from the neck to the shifting nodal point, which has recently been in the vicinity of Beach 10, and is deposited along the depositional feature which is Gull Point or offshore to create new platform to the east, or landward where it shoals in the harbor entrance channel.

Recent rates for this migration are artificial and directly influenced by the large-scale replenishment operations of the late 1950's through early 1970's. The present migration rate of 289,100 cubic yards per year reflects the replenishment input which has averaged 259,300 cubic yards per year since 1955. Attempts to determine the natural migration rate suffer from a lack of sufficient historical data and the obvious masking influence of the 150-year effort to stabilize the neck. Historical maps extending back to 1819 and aerial photographs extending back to 1939 were used to document the natural drift rate.

Historical maps do suggest that the Gull Point feature is a recent morphological addition to the system. Maps from 1819 through 1907 show a smooth recurved east end to Presque Isle which merges directly with the harbor entrance structures. Since the early 1930's, isolated growth has extended Gull Point as a "Mini Presque Isle" without sufficient recurving to weld this new growth back onto the shore. The original development of Gull Point may be related to a slug of sand which was released to the nearshore processes between 1917 and 1922 by breaching of the neck. The replenishment operations of the 1950's through 1970's continued adding new material to the accretionary end at a rate faster than easterly storms were able to recurve the bars and shoreline onto the Isle.

The incoming quantities of material never really replace the material left behind as the peninsula migrates and as the eastern end of the platform is built up. This continual loss of material plus the effect of a long-term, slowly-rising lake level (post-glacial rise of about one foot every 300 years) has probably caused Presque Isle to shrink. As Presque Isle migrates, it becomes smaller and migrates faster. Any attempt to identify the age and migration rate of Presque Isle must consider a measure of the rate of size change, as well as change in the rate of migration. The background data for this sophisticated analysis does not yet exist.

In summary, a few general statements can be made about Presque Isle's natural development trend:

(1) Presque Isle is an old age feature which is migrating with a net annual loss.

(2) Gull Point is a recent feature which has grown at significant rates because of the effects of artificial nourishment.

(3) Presque Isle Peninsula is a fluid feature; any attempt to permanently stagnate its migration will meet with eventual failure, with respect to geologic time, as all such attempts in the past have. An acceptable beach erosion control alternative will retard migration and/or lengthen the peninsula's life by adding new material to the system to replace that which has been used to build the platform and is a net loss to the littoral system.

Modern Coastal Processes - A Sediment Budget

Gains. Any influx of sediments into the Presque Isle system must either come from the east, from the west, from offshore sources, or from artificial nourishment. Presque Isle Peninsula is probably largely dependent upon influx from the west and artificial nourishment for littoral gains to the system.

Presque Isle is an eastward migrating feature with the Erie Harbor entrance structure and channel blocking any influx of material from the east. In addition, the morphology of Gull Point, plus the known wave energy flux condition for the Erie, PA area (Saville, 1953) further documents the lack of littoral material influx from the east.

Considering the historical development of Presque Isle and the offshore bathymetry, there is little evidence that the offshore is active in supplying a net sediment gain to the Presque Isle system. The platform to the west is below wave base and no longer part of the active Presque Isle system. The offshore is the trailing edge of the migrating feature, and being in deeper water, it does not keep up with the subaerial part of Presque Isle. Thus, there is a continual net offshore loss to the system rather than any gains.

Offshore bars in the nearshore do migrate onshore, but this is simply a redistribution of sand within the system which may result in temporary onshore gains. During lower water periods, the bar system is driven offshore. The importance of the nearshore bar system in influencing the littoral transport patterns of the Presque Isle system has been documented during studies to monitor the shoreline changes to Presque Isle Peninsula and by sand tracer studies.

Thus, all natural influx to the system must come from the west. The approximately 20-mile long shoreline between Conneaut, OH, and the root of the Presque Isle Peninsula is generally unbroken by any dominate stick-out features, headlands, or major shoreline inconsistencies. The Federal harbor structures at Conneaut are a very effective block to any littoral material exchange with shores any f rther to the west. Therefore, this 20-mile section of shore is considered as a single section of shore closed at the west and open at the east where Presque Isle Peninsula serves as the eventual site of deposition for any littoral input. Any littoral sediment input to this section of shore must come from fluvial sources, onshore movement of offshore sands, or bluff recession. The shore to the west is characterized by 20- to 100-feet high eroding till bluffs. The typical section is about 60- to 70-feet high with shale at or just below the waterline, then a coarse-grained till (probably Ashtabula till), followed by a thick clay sequence, and overlain by a thin layer of lacustrine sands (Great Lakes Research Institute, 1975). The recession rate of this sequence ranges from 0.5 ft/year to 2.0 ft/year (Carter, 1977).

Streams in the area, for example Elk Creek and Walnut Creek, flow through steep, shale gorges and have drowned entrance mouths. This combination, plus field data gathered from Elk Creek in support of a proposed Elk Creek Small-Boat Harbor Project, suggest that sand and gravel input from streams is minimal. However, these creeks have such potential for high velocity during periods of discharge (i.e., a steep gradient) that any material which may have collected with the river bed could get washed out into the littoral zone. A field reconnaissance of the upper river basin would be necessary in order to ascertain the presence of any significant fluvial contribution to the littoral zone.

The beaches are generally small, pocket beaches on the updrift side of structures or occur as bay mouth bar complexes at the mouth of each creek. Occasionally, during a period of low water, a narrow beach may collect in front of the bluff areas. The beaches are generally composed of fine to coarse quartz and lithic sands and gravels with shingles of shales and siltstones. Frequently, the beach may appear as a shingle beach.

Little information exists on the offshore area to the west of Presque Isle Peninsula, but it is generally considered to be till or rock surfaced, with little evidence of an offshore sand source except in the area of the Presque Isle platform. The platform area is generally 20 to 30 feet below LWD and, therefore, is considered as below the active wave base. At creek mouths, a delta develops where the bay mouth bars are washed outward during a period of heavy discharge. Some of these delta areas may serve as sites for temporary storage with some minor onshore return from the delta shoals.

In summary, sediment input from the west is dominated by bluff recession rates. There is probably some creek input of a much more minor level, but it is impossible to quantify the level of this contribution at this time. In order to develop a reasonable "ballpark" estimate of littoral transport rates from the west, it is necessary to make the following assumptions:

a. That the drift rate is controlled directly by the amount of material available for transport (This is a high energy shore where the wave energy is capable of transporting all the available littoral material).

b. That the primary source of littoral material is bluff recession.

c. That the major permanent littoral sink for this approximately 20-mile long section of coast is Presque Isle Peninsula. Other losses to the drift regime are limited to temporary storage in fillets associated with stick-out structures and small beaches and to permanent offshore losses. Offshore losses occur, particularly where small creeks divert littorally transported drift offshore into deltas and as material travels around the end of stick-out structures into deeper water. Offshore losses are assumed to be 20 percent.

The annual littoral input due to bluff recession between Conneaut and Presque Isle was calculated from bluff recession rates, bluff heights, reach length, and the stratigraphy presented by Carter (1977). Based on these computations, bluff recession contributes approximately 50,000 cubic yards of sands and gravels per year. Considering that 20 percent of this material is lost to the offshore, only about 40,000 cubic yards of littoral material per year are supplied to Presque Isle from the west (Figure 5).

Artificial nourishment has been a major factor influencing Presque Isle's development for the past 24 years. The need for replenishment reflects the highwater periods of the mid-1950's and the early 1970's which threatened to sever access to the outer peninsula. Over 6,200,000 cubic yards of material have been added to the system since 1955. This input has forestalled breaching of the neck, thus maintaining the neck's position and causing rapid growth at the accretionary east end (Gull Point). Replenishment has caused Presque Isle Peninsula to become elongated and has caused a net gain to the system.

The 4,150,000 cubic yards added in 1955-1956 was fine sand with a median size (50 percent size) of 0.20 mm which was obtained from borrow areas on the bayside of the peninsula. This sand was actually finer than the natural-sized beach material (0.35 mm) and was quickly eroded. The small amount of fill placed in 1965-1966 was medium sand (median size of 0.75 mm) and was considered as successful. As a result of this experience, the sandfill placed in the mid and late 1970's was a medium to coarse sand with a median size to the gradation band of about 1.8 mm. Prior to this period of nourishment, the neck was frequently breached. A major effect of a breach is to cause the neck to migrate eastward through overwash and bayside shoal development. Evaluation of

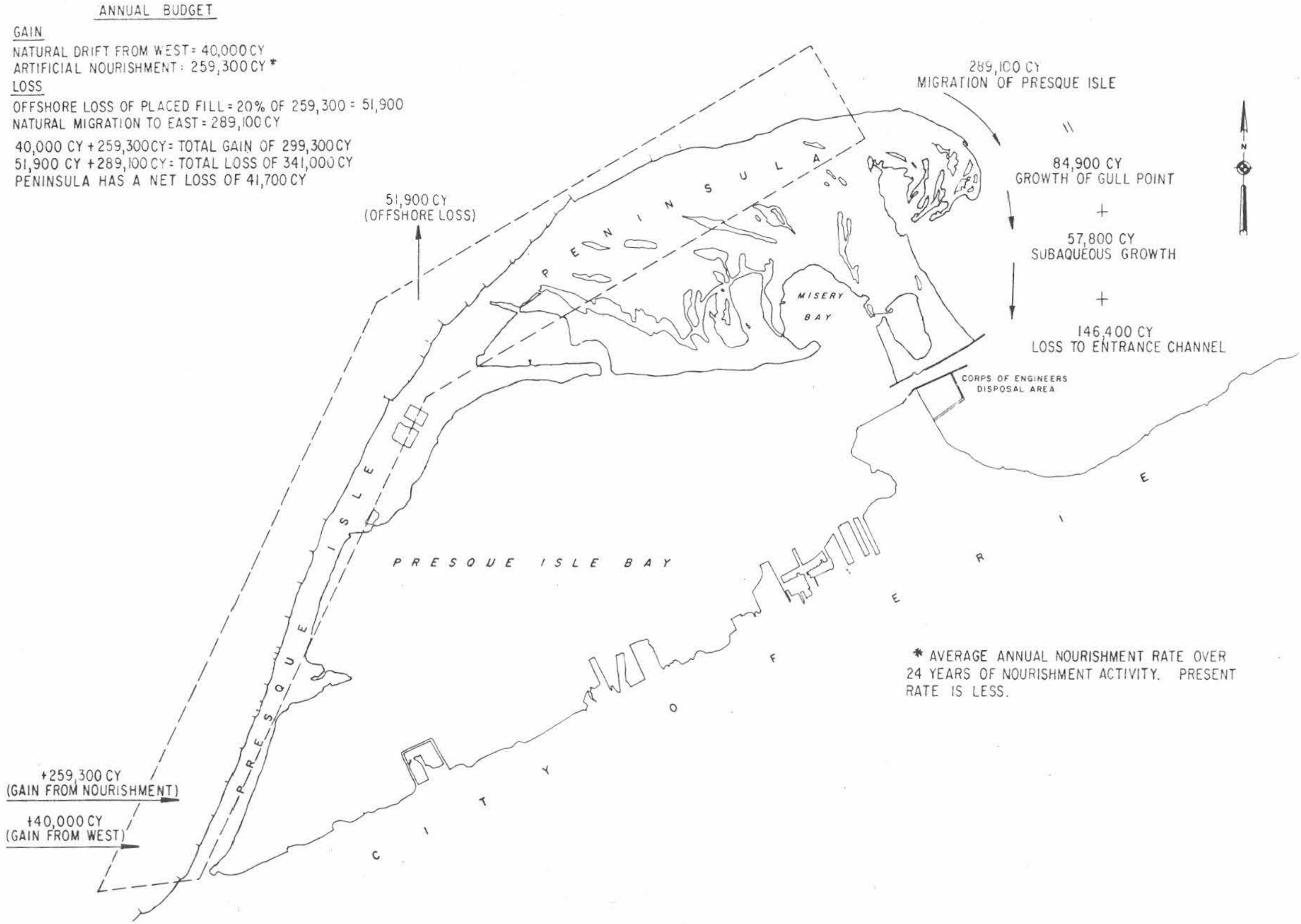

FIGURE 5. PRESENT CONDITION WITH ANNUAL NOURISHMENT SEDIMENT BUDGET

historical maps from the 1800's and early 1900's shows that the accretionary east end (Gull Point) has experienced sporadic growth possibly in response to migration of the neck.

Losses. Although Presque Isle Peninsula is a depositional feature, the dominate present activity is erosion. In 1877, the peninsula was described as eroding along the neck and eastward to a point which was 500' west of the lighthouse. A hundred years later, erosion characterizes the shore as far east as the east end of Beach 10. Thus, the nodal point between erosion and accretion has migrated 9,000 feet to the east in 100 years. Part of this nodal point shift is related to the natural migration of the system, and part is related to a net loss of material. The natural migration has been modified over the past 150 years by the many activities which have anchored and built the neck into a well-defined subaerial isthmus. According to Chief of Engineers reports from the early 1800's, the natural "neck" is a low, nominally vegetated, frequently overwashed, three and one-half mile-long sand spit. Efforts to stabilize the neck have resulted in the whole peninsula system being "stretched." As the distal end migrates, and the neck remains stable, the available littoral load is distributed over a longer shoreline. Thus, the Isle thins, the beaches narrow, and a greater length of shore erodes. This results in an "apparent" loss to the system.

Actual net losses are caused by offshore movement and platform building. Material leaves the system offshore around the total peninsula perimeter and at the distal east end.

Material is lost offshore as a result of bar formation and the migration of the peninsula away from its offshore platform. Typically, the offshore bar system migrates onshore and offshore in response to lake level changes and severe storms. During this cycle, there is a continual net offshore loss. The offshore bars at Presque Isle have been observed to be both complex and dynamic. Nummedal (1979) has identified four different bar forms and believes that substantial amounts of sediment may move along the bar systems. There are also offshore losses associated with the peninsula migrating eastward away from its western platform. That is, Presque Isle migrates east, leaving its platform behind. There is no present knowledge on the offshore losses from the Presque Isle system, but it has been estimated at 20 percent for use in developing a sediment budget (USAED Buffalo (1980)).

The main area of loss to the Presque Isle system is at the distal east end where the drifting sediment not only builds Gull Point, but also spills over the eastern end of the platform, building a new platform, and is recurved shoreward and landward, shoaling across the Erie Harbor entrance channel. Estimates have been made to summarize the losses at the east end based on historical changes at Gull Point, bathymetric charts, and dredging records for the Erie Harbor entrance channel. Based on these figures, the present condition (with replenishment) is that 146,400 cubic yards of littoral material accumulate in the entrance channel per year, 84,900 cubic yards per year are involved in building Gull Point, and 57,800 cubic yards per year build the new platform at the distal end (Figure 5).

From 1960 to the present, the average annual volume dredged from the entrance channel has been about 225,950 cubic yards. Computations presented in USAED, Buffalo 1980, indicate that 146,400 cubic yards of the dredged material per year come from Presque Isle and the rest from the mainland to the east or from siltation of suspended sediments. The 1930 to the present dredging record does not identify the amount dredged each year from the entrance channel, but the bulk of the annual dredging probably is material which originated from Presque Isle Peninsula and was deposited in the entrance channel. The 1930-1977 dredging records show that dredging from 1960 to the present has averaged 95,150 cubic yards per year more than the 1930-1959 period. This probably reflects an increased influx of material as a result of the 1956-1971 beach replenishment operations and suggests that there is about a five to six-year lag between replenishment and increased dredging volumes in the entrance channel.

The annual rate of growth of the distal end (Gull Point) varies from a minimum of 18,400 cubic yards per year with shore protection structures, but no replenishment (1875-1950) to 84,900 cubic yards per year with replenishment (1950-1978). The natural growth rate without structures or replenishment appears to be about 43,600 cubic yards per year (1819-1875).

Therefore, the natural balance for Presque Isle without replenishment is summarized as a 40,000 cubic yard gain from the west, 51,300 cubic yard permanent loss to the entrance channel, 17,400 cubic yards used to build up the new eastern platform, and 18,400 cubic yards to develop Gull Point. The resultant system, therefore, has a migration rate of 87,100 cubic yards per year. Presently, the volume of Gull Point growth and the net loss to the entrance channel are higher (Figure 5), reflecting the additional available sediment load introduced by the replenishment activities.

ENGINEERING HISTORY OF PRESQUE ISLE

The Problem

The geological forces which have created Presque Isle are also gradually destroying it. The natural processes of erosion and deposition continue as Presque Isle continues to migrate. Destructive natural processes, although necessary in a migrating coastal feature, are considered as unacceptable on Presque Isle by the general human factor. Erosion of the lakeshore beaches and breaching of the neck have been counteracted by public and private efforts for over 160 years. A history of the human efforts to retard erosion of the peninsula is lengthy and complex.

History of Shore Protection

When the Federal project for Erie Harbor was first initiated back in the early 1800's, in addition to the work at the entrance, the project required protection of the shore at the neck of the peninsula of Presque Isle, which by its position, forms the harbor of Erie. The preservation of the peninsula is of vital importance to Erie Harbor, and it is for the purpose of preserving the harbor that protection of the long, narrow neck at the western end of the peninsula was originally deemed necessary. The protective works to date have been constructed to prevent breaching through the narrow neck during severe storms from the west. Such a breach would compromise the effectiveness of the harbor. A literature survey of the Chief of Engineers Reports (1867-1978) was undertaken, and the following paragraphs present a documentation on protective works which were implemented for preservation of Presque Isle Peninsula.

The attention of the United States Government was directed to Erie Harbor after the close of the War of 1812 from the fact that it was in Erie that Commodore Perry anchored his fleet after his memorable battle. In 1823, the Board of Engineers presented an elaborate report with a plan for the improvement of the entrance to Erie Harbor. Subsequently, the River and Harbor Act of 26 May 1824 authorized improvement of Erie Harbor and protection of Presque Isle Peninsula.

The first breach recorded appears to have taken place during the Winter of 1828-1829. Its location and extent were not reported, but the entire appropriation of $7,390.25 provided by the River and Harbor Act of 3 March 1829 was used in closing it. During the Winter of 1832-1833, another breach occurred. Nothing was done to close it, and in 1835, it was reported to be nearly one-mile wide. Plans were developed which provided for partially closing the breach with cribwork and to make a 400-foot wide western entrance to the bay. In 1836, work commenced and 420 feet of cribwork breakwater was completed, strengthened by piling, and partially filled with stone. This cribwork breakwater was extended 1,920 feet in 1837 for an aggregate length of 2,340 feet. It was reported that the progress in partially closing the breach was very satisfactory, and in 1838 an additional 1,035 feet of cribwork was built. Work continued in 1839 when 990 feet of cribwork was built. There were no appropriations nor work done during the years 1840 through 1843. In 1844, the breach was reported to be about 3,000-feet wide, and the erosion was such that 470 feet of cribwork was built to protect the barracks built for workmen in 1836. Nothing further was done and in 1852, the breach was reported as still existing, and the cribwork protection built in previous years had been almost destroyed. In 1853, efforts were made to prevent further erosion by protecting the shore with brush weighted with stone. The results were very satisfactory, and this mode of closing the breach was continued in 1854 through 1856. Work was suspended in 1857 due to lack of funds, and no further work was done until 1864. In 1864, it was reported that the breach at the west end of the harbor was entirely closed, although about 500 feet of the peninsula was so low that waves would break clear across during high water and heavy gales. This low portion of the peninsula was strengthened in 1865 by placing old tree trunks, brush, saplings, etc., parallel to the shore, making a layer 30-feet wide.

During the years 1871 and 1872, 51,300 young trees, roots, and slips of silver poplar, American poplar, and willow were planted as an experiment on the west side of the peninsula for protection of the neck. Also, the beach at two exposed points was further protected by anchoring and picketing brush laid in rows and weighted with heavy stone. The Fall and Winter gales of 1873-1874 made alarming attacks on the shore of the peninsula, and in November 1874, the peninsula was once more breached. The breach was closed in 1875 with 400 feet of six-foot high pile and plank fence riprapped on both sides with stone. The protection proved to be successful, and an additional 1,080 feet of pile and plank fence was built at other weak points on the peninsula in 1875. This pile and plank fence was extended 3,056 feet in 1876, another 1,461 feet in 1877, and 550 feet in 1878, making a total length of 6,547 feet. In 1879, the protection fence was badly damaged at various points with the stone washed away, piles broken off, and planks destroyed.

In 1880, eight jetties 200 feet apart were built by driving lines of close piling out to a depth of 6 feet in the lake. A ninth jetty was built about two miles from the neck of the peninsula. In addition, about 2,000 feet of brush and stone protection was built along the lakefront to repair the protective fences which had been destroyed during the previous winter. Violent gales during the Winters of 1880-1881 and 1881-1882 destroyed several portions of the protective fencing built during the period from 1875 to 1878. In 1882, three additional piles were driven between every two old piles from the original protective fencing. About 1,000 feet of this type of protection was built to provide a nearly closed continuous row at a cost of nearly $2,500. This brought the total expenditures for work accomplished on Presque Isle during the period from 1829 through 1883 to approximately $220,000.

There was no work done for protection of Presque Isle Peninsula during the period from 1883 through 1887, and in 1887, it was reported that all the protection fences and pile jetties built in the previous years were so broken down and rotten that they were considered useless. The River and Harbor Act of 11 August 1888 authorized protection of the neck of the peninsula by construction of a 6,000-foot long timber pile and sheet pile breakwater located about 100 feet offshore. About 4,500 feet of breakwater was built by September 1889 at a cost of about $60,000 when a moderate storm badly wrecked all but 1,300 feet of the structure and work was ordered stopped since it was evident that the protection constructed was not going to prove serviceable. The remaining sheet piling and walings were washed away during a severe storm in October 1892.

No further work was done on protection of the peninsula during the period of 1890 through 1895. Several severe storms occurred during this period whereby waves would wash over the peninsula and into the bay, causing severe erosion along the western portion of the peninsula. In 1896, another experimental tree planting project was undertaken whereby 1,000 Carolina poplars, 200 Wisconsin willows, 200 yellow locusts, 200 Scotch pines, three bushels of blue grass, two bushels of orchard grass, one bushel of crimson clover, 300 willow cuttings, and about 60 native poplar trees were planted on the neck of the peninsula at a cost of $360. The purpose of the plantings was to make a growth that would catch drifting sand and increase the height and width of the neck, increase the resistance of the neck to erosion, and lessen the liability of a possible breach from waves washing over the neck of the peninsula. The trees planted in 1896 grew vigorously during the year and therefore, in 1897, about 2,400 yellow locust trees and two bushels of seeds of native shrubs were planted on the neck of the peninsula at a cost of $376. At that time, the plantings were regarded as an important part of the harbor works and further plant growth encouraged since those planted in previous years had thriven very well. Therefore, an additional 2,000 honey-locust trees and 200 willow cuttings were planted in 1898 at a cost of $210.

The River and Harbor Act of 3 March 1899 authorized construction of four protection jetties along the outer edge of Presque Isle Peninsula. The first jetty was built in 1900 and located 5,200 feet west of the Presque Isle Light. The structure cost about $5,390 and was of timber crib construction filled with stone and had a "T" across the outer end. The cribbing was 12-feet wide, 11-1/2-feet deep, and 290-feet long; the "T" was 10-feet wide, 11-1/2-feet deep, and 32-feet long. The second protection jetty was built in 1903 at a cost of $8,560 and located 7,800 feet west of the Presque Isle Light. In 1906, it was determined that the jetties built in 1900 and 1903 were not correcting the beach erosion along the peninsula and therefore, the remaining two jetties authorized by the River and Harbor Act of 1899 were never constructed.

There was no work done for protection of Presque Isle Peninsula during the period from 1904 through 1915. However, in 1916, about $316 was expended for planting 5,000 poplar trees and 2,725 linear feet of willow hedge on the neck of the peninsula to reinforce the existing growth. These trees and hedge grew well during the year and in 1917, an additional 2,310 poplar trees and 2,280 willow cuttings were planted to reinforce the existing growth at a cost of $195.

A severe storm occurred late in October 1917, causing waves to break over the neck of the peninsula and creating a breach about 150-feet wide. Work on closing the breach with a 300-foot timber bulkhead was initiated in mid-November and continued until early December with 270 feet being completed at a cost of $7,000 when another severe storm occurred, uprooting large trees, washing out small growth, destroying the completed portion of the timber bulkhead, and widening the breach to

479 feet. There were no further attempts made to close the breach during 1918, and storms during the Winter of 1918-1919 increased the width of the breach to 1,160 feet. Closure of the breach with sandfill protection was begun in the Fall of 1919 when a 500-foot section of fill protection at the east end of the breach was placed before operations were halted for the winter. When operations resumed in April 1920, the breach was 1,470-feet wide. During 1920, about 3,000 feet of sandfill protection and 1,700 feet of rubblemound protection were placed, and 4,800 small poplar trees were planted on the sandfill protection. In addition, 310 feet of riprap wall was placed on the lakeside of the sandfill protection. The sandfill protection was completed during 1921 with 1,500 feet being placed, and the riprap wall on the lakeside of the sandfill protection was extended 1,465 feet. During the period from October 1920 through November 1921, about 22,700 small poplar and 1,900 small willow trees were planted and 49 bushels of rye and six bushels of cowpeas sown to protect the sandfill. In 1922, the riprap stone wall on the lakeside of the sandfill protection was reinforced and extended 1,160 feet, thus completing the work in closing the breach. Approximately $282,000 was expended on work to close the breach.

The River and Harbor Act of 28 November 1922 reconveyed Presque Isle Peninsula to the State of Pennsylvania for park purposes, and its care and protection were no longer to be considered by the United States as part of the project for improvement of Erie Harbor. The State of Pennsylvania built six sand traps in 1927, a series of seven steel sheet pile groins during 1928 and 1929, and about 5,300 feet of steel sheet pile bulkhead in 1929 on the lakeside of the peninsula at various locations from the neck to the light-house.

The United States Government again became involved with Presque Isle Peninsula for the protection of Erie Harbor in 1930 and 1931 when 5,646 feet of steel sheet pile bulkhead (including shore returns) with 5,052 feet of stone facing, was constructed along the neck of the peninsula at a cost of about $165,400. The State of Pennsylvania extended this protection along the neck of the peninsula an additional 1,230 feet in 1931 and also built a steel sheet pile groin. In 1932, the State built two more steel sheet pile groins and extended the steel sheet pile bulkhead which they built in 1929 an additional 1,500 feet. This bulkhead was again extended 850 feet by the State in 1937.

In 1943 and 1944, the United States Government repaired shore protection works constructed in previous years and further protected the steel sheet pile bulkheads by construction of a rubblemound facing on the lakeside. In addition, 2,750 feet of rubblemound protection was constructed at the root of the peninsula, and two experimental 300-foot long rubblemound groins were built. The work undertaken in 1943 and 1944 was accomplished at a cost of about $1,041,700. Further repairs to the protection works along Presque Isle Peninsula were undertaken by the United States Government during the period from 1947 through 1952 at a total cost of $443,100. During the period from 1924 through 1948, it was estimated that the Commonwealth of Pennsylvania had spent approximately $3,500,000 on maintenance of the peninsula.

Severe storms during the early 1950's led to the establishment of the cooperative beach protection program between the Federal Government and the Commonwealth of Pennsylvania as authorized by the River and Harbor Act of 3 September 1954. Work commenced in the Fall of 1955 and was completed in the Summer of 1956, during which time 4,150,000 cubic yards of sand were pumped on the beaches, ten new steel sheet pile groins constructed, two existing groins altered, and a badly damaged bulkhead section near the lighthouse groin was removed. The total cost of the cooperative project was $2,451,270, which includes a stone seawall 3,000-feet long built in 1952 on the neck of the peninsula.

An emergency sand replenishment was accomplished by the Commonwealth of Pennsylvania in the Winter of 1959-1960 at the cost of about $24,000. The cooperative beach protection program between the Federal Government and the Commonwealth of Pennsylvania was modifed by the River and Harbor Act of 14 July 1960 to include participation in periodic nourishment for a period of 10 years following the first major replenishment operation. The emergency protection in 1959-1960 prevented further damage to the project up to the time of the first major replenishment authorized by the 1960 River and Harbor Act. The first major replenishment was undertaken in 1960-1961 during which approximately 681,500 cubic yards of sand were pumped onto the beaches at a cost of $500,000. In 1963-1964, the Commonwealth of Pennsylvania repaired two groins which were built in 1956 by placing heavy stone at a cost of about $54,000. A second major replenishment authorized by the 1960 River and Harbor Act was required in 1964-1965, at which time approximately 402,300 cubic yards of sand were pumped on the beaches at a cost of $355,000. In 1965-1966, a third replenishment was undertaken whereby 45,000 tons of coarse-grained sandfill were placed, and six of the groins built in 1956 were modified by addition of a stone facing. The total cost for accomplishing the work undertaken in the third replenishment was about $166,000. A fourth major beach replenishment was undertaken in 1968-1969, with 102,700 tons of coarse sandfill being placed on the beaches at a cost of $348,000. The fifth and final beach replenishment operation under authorization of the 1960 River and Harbor Act was accomplished in 1971 when a 1,200-foot long barrier consisting of nylon bags filled with sand and grout was built at Beach No. 6, and 152,500 tons of sand were placed on the beaches at a total cost of $535,000.

In 1973, an emergency sand replenishment was undertaken by the Federal Government, whereby 100,000 tons of sand were placed along the neck of the peninsula at a cost of $240,000. Due to the severe erosion problem which still existed, the cooperative beach protection program between the Federal Government and the Commonwealth of Pennsylvania was again modified. The Water Resources Development Act of 1974 authorized the Federal Government to participate in beach nourishment for a five-year period. Actual work under the program was initiated in 1975 with the placement of approximately 187,000 tons of sand and a total expenditure of $1,097,000. A second nourishment was completed in 1976 at a cost of about $1,097,000 for placement of 183,000 tons of sand. In 1977, sand from land sources was used instead of from an offshore borrow area as in the previous two years, and 287,000 tons of sand were placed at a cost of about $1,089,000. The fourth beach nourishment project was completed in 1978 at a cost of $1,074,000 and included construction of three experimental prototype breakwaters offshore from Budny Beach (Beach No. 10) and placement of 173,000 tons of sand. A fifth and sixth beach nourishment projects were completed in 1979 and 1980 at a cost of $1,061,000 and $1,082,000, respectively, for placement of 216,000 tons of sand on beaches along the lake shoreline each year. A seventh beach nourishment project requiring the placement of 236,000 tons of sand, was completed in 1981 at a cost of $1,213,000. An eighth replenishment operation, requiring the placement of 284,000 tons of sand, was completed in 1982 at a cost of $1,430,000.

During the past 25 years, the Commonwealth of Pennsylvania, in addition to contributing approximately $5.2 million to the cooperative beach nourishment program, has expended several million dollars for performing emergency repairs to roadways on the peninsula which were damaged during storms, for undertaking sand replenishment operations, for placement of stone protection at critical locations on the lakeside, as well as the bayside of the peninsula, and for grout-filled nylon bag barriers.

Existing Structures

The structures built for preservation of Presque Isle Peninsula during the 1800's and early 1900's were mainly of timber construction. These structures had a useful life of only a few years before being destroyed. During the period from 1920 through 1978, rubblemound and steel sheet pile construction methods were implemented. These types of construction are more durable and longer lasting. Structures built of these types of construction make up the majority of the protective structures presently in existence along the peninsula. The locations of protective structures presently in existence along Presque Isle Peninsula, the type of construction utilized, the date the structures were built, and who built them are presented on Figure 6. The types of some of the existing structures and experimental projects implemented at Presque Isle will be observed during the field trips.

PROPOSED PROJECT

Presque Isle is a migrating feature with a continual loss of material. Any project which is designed to stabilize Presque Isle must consider the system as a whole. It is impractical to protect only one portion of the system, as the system will continue to migrate and the bars will continue to carry sediment. Thus, the Corps of Engineers (USAED, Buffalo, 1980) designed a plan which is intended to protect the whole peninsula unit, from serious erosion by utilizing the offshore bar system, to stabilize the shoreline.

The plan consists of construction of a system of 58 rubblemound breakwaters located about 400 feet offshore along the lakeward length of the peninsula, parallel to the shoreline, and positioned in the trough between the first and second offshore sand bars. Each structure would be 150 feet in length with a 350-foot gap between structures. In addition, 500,000 cubic yards of sandfill would be placed along the shoreline in the lee of the breakwaters to provide a recreational beach berm. The breakwaters are intended to attenuate the wave action to such a degree as to reduce littoral drift by approximately 75 percent, thus slowing the migration of the feature, reducing erosion and helping to maintain the beach area in the lee of the breakwaters.

SUMMARY

The Commonwealth of Pennsylvania summarizes the natural wonder which is Presque Isle in a sign which greets all visitors:

> "Welcome to Presque Isle Peninsula. Enjoy its unique botanical and geological evolution, historical background, bird and animal life."

ROAD LOG AND STOP DESCRIPTIONS

	Miles	Time
Buffalo Marriott Inn to entrance of Presque Isle Peninsula	115	2 Hr, 30 Min
Presque Isle entrance to Beach 10 (Bay Route)	8.2	14 Min
Beach 10 to Beach 6 (Lake Route)	3.6	7 Min
Beach 6 to Entrance	2.3	5 Min
Entrance of Presque Isle to Buffalo Marriott Inn	115	2 Hr, 30 Min

Stops Planned (Reference Figure 1)

Groins 1 and 2
Walk Groin 5 to Beach 6
Beach 10
Walk out to Gull Point
Lighthouse
Walk Beach, 8 and 7

● INDICATES STRUCTURE CONSTRUCTED AS PART OF THE COOPERATIVE BEACH EROSION CONTROL PROJECT BETWEEN THE FEDERAL GOVERNMENT AND COMMONWEALTH OF PENNSYLVANIA

FIGURE 6. EXISTING PROTECTION STRUCTURES

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STRATIGRAPHY AND FACIES VARIATION OF THE ROCHESTER SHALE (SILURIAN: CLINTON GROUP) ALONG NIAGARA GORGE

BY

CARLTON E. BRETT, Department of Geological Sciences,
University of Rochester, Rochester, NY 14627

INTRODUCTION

The linear, east-west configuration of Paleozoic outcrop belts in western New York and Ontario imposes a strong two-dimensionality on our knowledge of the stratigraphy and facies relationships of these rocks. It is frequently assumed that the outcrop sections provide a reasonable transect from near shore clastic-dominated facies in the east to offshore carbonates and marine muds in the west. This view is overly simplistic and it ignores the fact that the Paleozoic epeiric seas had a northern as well as an eastern shoreline (directions based on modern geography). Furthermore, the idea of a simple east-west transect is contradicted by detailed examination of stratigraphy in western New York. Stratigraphic sequences such as the Silurian Clinton, Lockport, or Salina Groups exhibit vertical facies changes, manifest in the gradational contacts between beds, members, or formations. Walther's law predicts that such vertical sequences of litho- and biofacies should mirror arrays of laterally adjacent, contemporaneous facies belts. However, lateral gradation of lithologies is rarely observed along the outcrop belt of western New York. Rather, one observes that thin stratigraphic units representing distinctive facies, are traceable along the strike of the outcrop belt for tens to hundreds of miles (see, for example, Belak, 1980).

This "layer cake" aspect of New York stratigraphy has facilitated local correlation. However, it has also led to a common impression that the strata represent vertical stacking of time parallel facies units in a three dimensional sense, implying basinwide environmental changes. This non-Waltherian view of western New York stratigraphy is probably erroneous and it has greatly hampered attempts to decipher facies relationships and depositional environments. I contend that it is largely an artifact resulting from the east-west trend of the modern outcrop belts.

Niagara Gorge provides the longest continuous north-south section of Silurian rocks in western New York. As such, this exposure permits detailed examination of strata along a 10 km section normal to the dominantly east-west trending Niagara escarpment. Curiously, few, if any, previous workers have examined this transect in any detail, perhaps because of the general inaccessibility of exposures in the southern end of Niagara Gorge. Recent detailed examination of Silurian rocks in the gorge (Brett, 1978; in press) provide important new insights into understanding facies relationships and interpreting paleoenvironments. Most importantly, these studies indicate that facies belts within the Rochester Shale, and probably other units, are elongate east-west and subparallel to the present outcrop belt in western New York and the Ontario Peninsula. This single factor provides an important key for the development of a depositional model for the Rochester Shale, and perhaps other units with apparent "layer cake" stratigraphy. The present field trip examines various aspects of the stratigraphy and facies relationships of the Rochester Shale along Niagara Gorge (Figs. 1,7).

GEOLOGIC SETTING OF THE ROCHESTER SHALE AND ASSOCIATED FACIES

The medial Silurian Rochester Shale (Wenlockian; Clinton Group) is a classic unit in American stratigraphy, being among the first formally designated formations in North America (Hall, 1839, p. 20). This formation has been widely correlated in the northern and central Appalachian region, and serves as an important stratigraphic marker in subsurface studies (Schuchert, 1914; Chadwick, 1918; Caley 1940; Gillette, 1940, 1947; Berry and Boucot, 1970). The Rochester Shale is also noted as an important source of fossils; over 200 species of invertebrates have been reported from the unit, including some of the best preserved Silurian fossils in North America (Hall, 1852; Ringueberg, 1888; Grabau, 1901; Sarle, 1901; Bassler, 1906; Springer, 1920, 1926; Brett, 1978).

Yet, despite its historical, stratigraphic and paleontologic significance, the Rochester Shale has received little recent study. Preliminary studies of Rochester depositional environments and paleoecology were undertaken by Thusu (1972) and Narbonne (1977); however these papers are restricted in geographic scope, and their conclusions necessarily generalized and tentative.

In New York State and Ontario, the Rochester Shale constitutes the middle unit of three formations of the upper Clinton Group (Bolton, 1957), (Figs. 2,3). Throughout much of its extent, the Rochester is underlain conformably by the upper member of the Irondequoit Limestone - a light gray to pinkish gray crinoidal biomicrite or biosparite.

From Hamilton, Ontario eastward to Rochester, New York, the Rochester Shale is overlain by fine-grained, buff-colored DeCew Dolostone. Early workers, (Ulrich, 1911; Schuchert, 1914; Chadwick, 1918) postulated a major disconformity between the Rochester and DeCew but the completely gradational nature of the contact in nearly all localities argues against this. For this reason, the boundary between the Clinton and Lockport groups is now generally drawn at the sharp upper contact of the DeCew with the overlying Gasport Limestone in western New York (Gillette, 1947; Bolton, 1957). Crowley (1973) and Nairn (1971) have demonstrated that Gasport and DeCew are also gene-

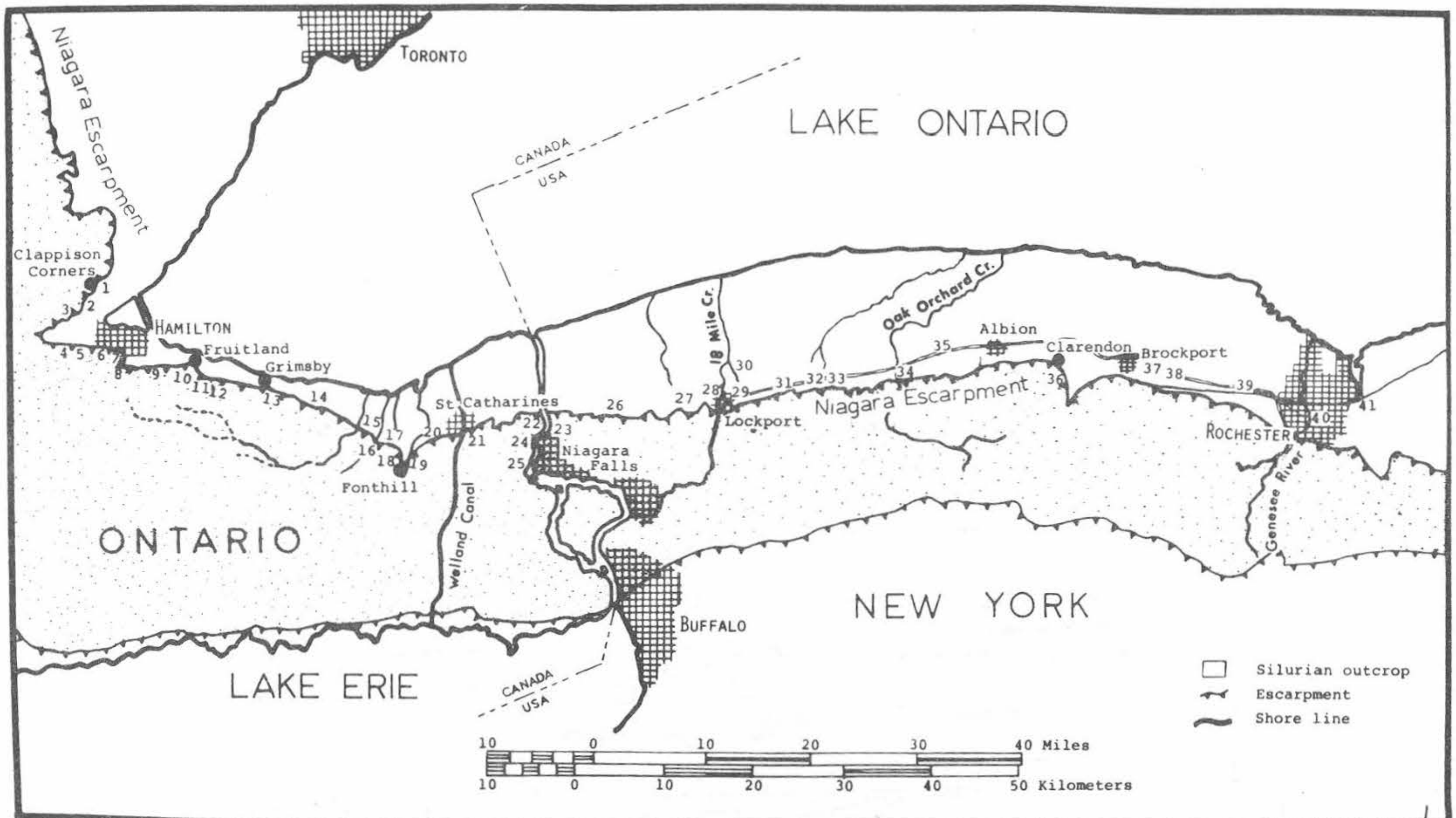

FIG. 1.--Location map for stratigraphic sections of the Rochester Shale in western New York and Ontario. Numbers refer to localities discussed in the text (Localities described in Appendix A of Brett, 1978).

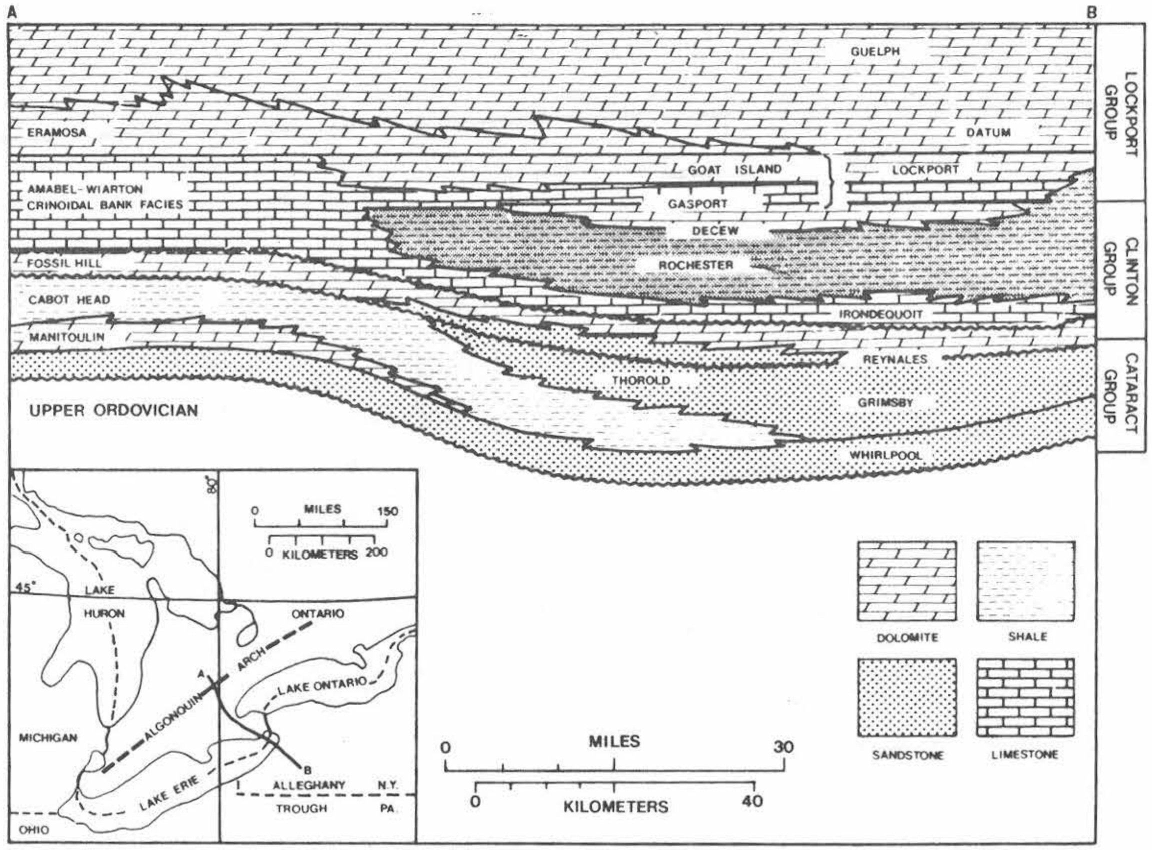

FIG. 2.--Cross-sectional diagram of the upper Clinton-lower Lockport Groups in Ontario and western New York; orientation of section line is indicated on the inset. Modified from Sanford (1969).

tically related facies so that the differentiation of Clinton and Lockport Groups is arbitrary.

The Rochester Shale consists of medium to dark-gray calcareous shale and thin limestones including micrites, pelmicrites, biomicrites, and intramicrites (Thusu, 1972). Along the outcrop belt in New York and Ontario, the Rochester thickens southeastward from a minimum of 0.6 m (2 ft) at Clappison Corners, Ontario, to a maximum of about 37 m (122 ft) near Walcott, New York. Eastward, it thins again slightly to about 30 m (100 ft) near Lakeport, New York, before grading into the Herkimer Sandstone (Gillette, 1947). Subsurface data reveal that the Rochester is thickest in an elongate northeast-southwest trending region from North Victory (Cayuga County), New York to Geneva, New York (Ontario County), where it attains a maximum thickness of 44.8 m (147 ft) (Gillette, 1940; Fig.2). Well logs indicate a southward thickening in western New York from 15-21 m (57-70 ft) in Niagara-Orleans counties to 30-35 m (100-115 ft) in southern Erie, Genesee and Livingston counties (Van Tyne, 1975).

The Rochester Shale maintains a thickness of 12-15 m (40-50 ft) in the subsurface along the southernside of the Ontario peninsula westward to Windsor, Ontario. However, well logs also reveal relatively rapid thinning and pinch-out of the Rochester into Wiarton crinoidal dolostone within 24-32 km northeast of this belt (Caley, 1940; Fig.2).

The northward thinning is apparently due to facies change with only minor, if any, erosive overstep of the Rochester (Bolton, 1957; Sanford, 1969). The Wiarton crinoidal facies is thus the northern equivalent of the entire Irondequoit-Gasport sequence. The line of abrupt facies change approximately coincides with Algonquin Axis, an area of relative uplift during the Paleozoic.

In north-south cross sections (Fig. 2), the Irondequoit and Gasport limestones appear as tongues of the crinoidal shoal facies which extend southeastward into the Appalachian Basin. These tongues are mutually separated by mudstone and argillaceous dolostone of the Rochester and DeCew formations.

STRATIGRAPHY OF THE ROCHESTER SHALE IN ONTARIO AND WESTERN NEW YORK

General Stratigraphic Subdivisions

Local subdivisions of the Rochester Shale were recognized by early workers (Ringueberg, 1888; Grabau, 1901), but subsequent authors tended to treat the Rochester Shale as a homogenous lithologic unit, lacking mappable subdivisions (Gillette, 1940; Bolton, 1957; Thusu, 1972). Detailed stratigraphic study of the Rochester Shale indicates the existence of several widely traceable units within the formation, and four new members have recently been proposed (Brett in press).

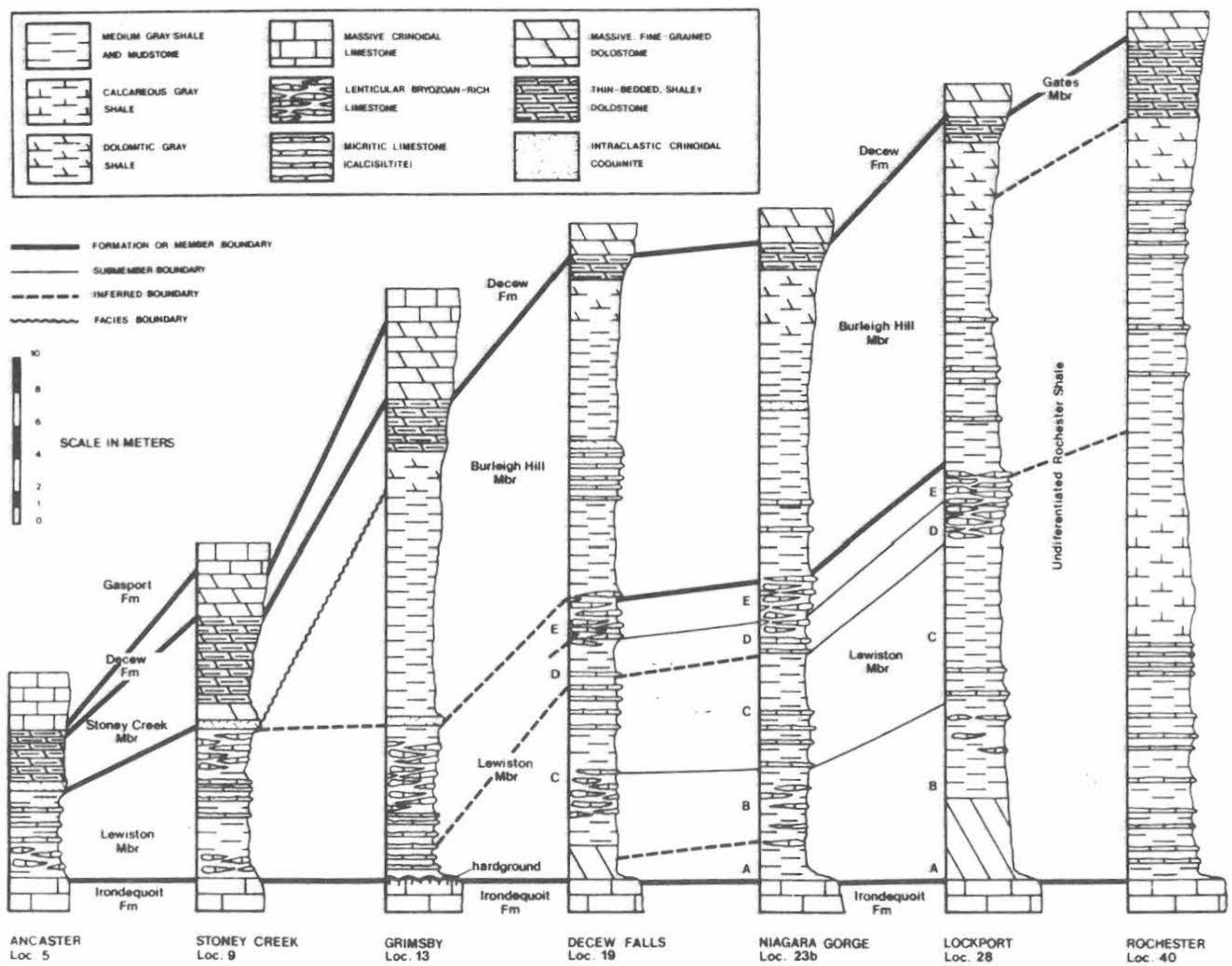

FIG. 3.--Stratigraphic sections of Rochester Shale at seven localities in Ontario and western New York. Locations of numbered sections are shown in Figure 1.

A twofold subdivision of the Rochester Shale is recognizable in outcrops along the Niagara Escarpment from Hamilton, Ontario (west) to Brockport, New York (east) (Fig. 3). The Rochester Shale can be differentiated into units of roughly equal thickness in nearly all exposures in this area.

The lower (= "lower shales" of Grabau, 1901) - previously termed the Lewiston Member (Brett, in press), consists of interbedded mudstone, lenticular fossiliferous limestones and barren calcisiltites. At its type section in Niagara Gorge, the Lewiston Member is 8.7 m (28.6 ft) thick and is divisible into five submembers, as follows (Fig. 3).

A) a basal transition zone, up to a meter thick, of brachiopod-crinoid-bearing silty, argillaceous biomicrite and biosparite lenses in shale.

B) a one-to-three meter interval of fossiliferous brownish-gray, slightly calcareous mudstone with abundant patches of ramose bryozoans and bryozoan-rich biomicrites.

C) a middle unit of sparsely fossiliferous shale and interbedded dark gray laminated calcisiltites (pelmicrites).

D) an interval, up to two meters thick, resembling submember B, and gradational into Unit E.

E) an upper unit two meters or less in thickness, of closely-spaced lenses of bryozoan-rich biomicrites and biosparite with thin interbedded shales. The uppermost bed of this interval, up to 30 cm thick, is, locally, glauconitic and contains shale clasts and abraded fossil debris. Submember E, which corresponds to the "bryozoa beds" of Grabau (1901), forms a readily recognizable upper unit of the Lewiston Member.

All of the Lewiston submembers can be recognized from St. Catharines, Ontario to near Brockport, New York, and their thicknesses remain proportionally constant, despite thinning of the Lewiston Member, as a whole, from over 12.2 m to 7.9 m (40-26 ft) (Fig. 3).

Although the Lewiston Member is recognizable to Hamilton, Ontario, it becomes highly condensed and contains an increased abundance of calcisiltite, calcarenite and intraclastic limestone. Differentiation of submembers (B-E) becomes obscure in this region.

The upper portion of the Rochester Shale is consistently distinct from the Lewiston Member, and in western New York comprises sparsely fossiliferous platy and highly calcareous or dolomitic shale, the Burleigh Hill Member. At most localities, this member is gradational upwards through about 1 to 2 m of dolomitic shale into the overlying DeCew dolostone. The contact between the Burleigh Hill and the underlying Lewiston Member is generally sharp and is placed at the top of bryozoan-rich limestones, (submember E) the "bryozoa beds" of Grabau (1901). Bryozoan biomicrites are absent from the Burleigh Hill Member.

Bryozoan-rich limestones and bryozoan clusters are also lacking from the lower Rochester east of Brockport, New York, where the entire formation thickens considerably and consists of a more homogeneous sequence of shales and barren calcisiltites. This undifferentiated and less fossiliferous Rochester Shale of Monroe County closely resembles the Burleigh Hill Member (or Lewiston-C submember) throughout.

A 3-5 m (10 to 15 ft) interval, near the middle of Rochester Shale at Genesee Gorge which exhibits a concentration of barren calcisiltites and calcareous shales, may correspond to the bryozoa beds farther west. Again, the upper Rochester Shale grades upward into an interval of about 20 m of argillaceous dolostone and dolomitic shale, which Chadwick (1918) designated the Gates Member.

North-South Facies Variations in the Rochester Shale

Detailed studies of Rochester Shale stratigraphy in Niagara Gorge demonstrate very rapid north-south facies changes, contrasting to the continuity of facies along the east-west oriented outcrop belt (Brett, in press, Fig. 4, herein). Indeed, the section at Niagara Falls more closely resembles the undifferentiated Rochester Shale at Genesee Gorge, some 130 km (70 mi) to the east, than it does sections in the north end of Niagara Gorge, only 10 km (7 mi) to the north (see Fig. 1).

Units B and D of the Lewiston Member diverge from one another southward as the barren middle shales (Unit C) thicken rapidly toward the south end of Niagara Gorge. Unit C changes lithologically somewhat, containing thicker and more closely-spaced calcisiltite bands in the northern end of Niagara Gorge. Near Niagara Falls, the unit is more than doubled in thickness and consists almost entirely of barren shale.

The bryozoan-rich intervals of the Lewiston Member (units B,D, and E) die out southward and are replaced, laterally, by calcisiltite-shale units (Fig. 5). This is most notable in the case of Unit E, the "bryozoan beds" of Grabau (1901). Nearly two meters thick near Lewiston (loc. 23), this bryozoan biomicrite-shale interval thins to less than 0.5 m at Whirlpool State Park, 2 km to the south. At this point, the lower bands of bryozoan-rich limestone grade into dolomitic calcisiltites with only scattered bryozoans and other fossils. Farther south, at the Niagara Sewage Plant exposure (Stop 5) bryozoans are completely absent from all but one uppermost band, the remainder of Unit E consisting

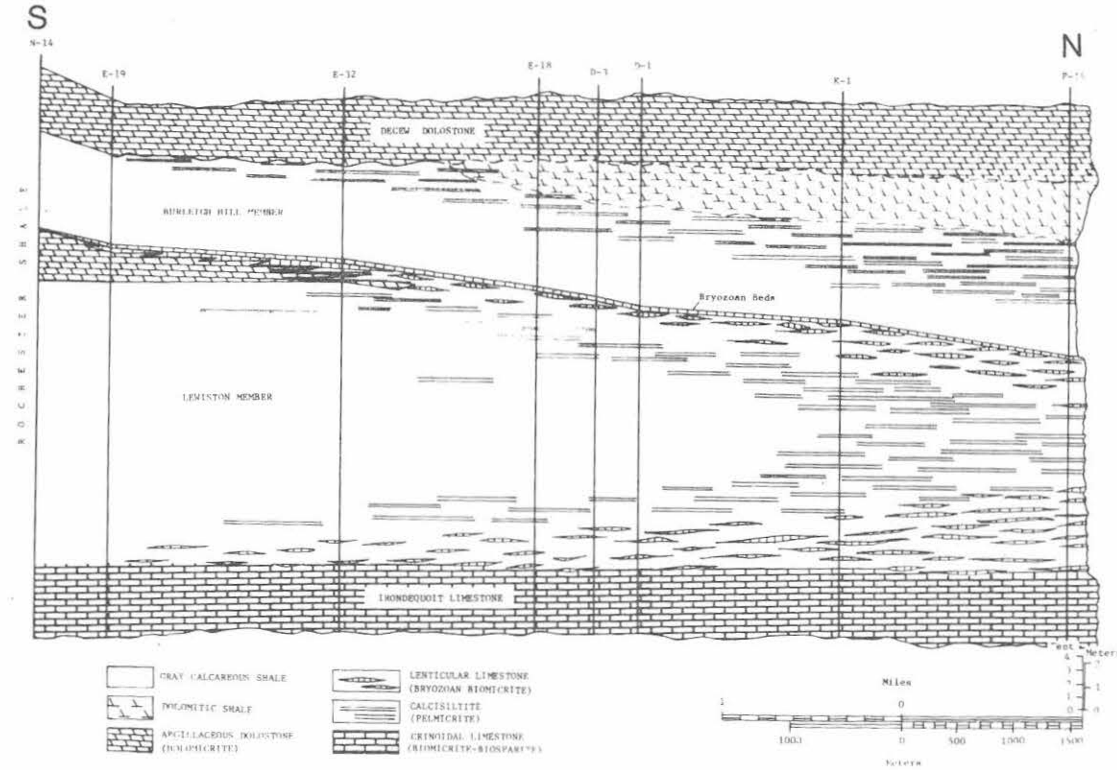

FIG. 4.-- North-south stratigraphic relationships of the Rochester Shale in Niagara Gorge; vertical lines indicate positions of measured sections from drill cores logged by Bolton, (1957; appendix B, p. 107-141); numbers correspond to codes assigned by Bolton to each drill core.

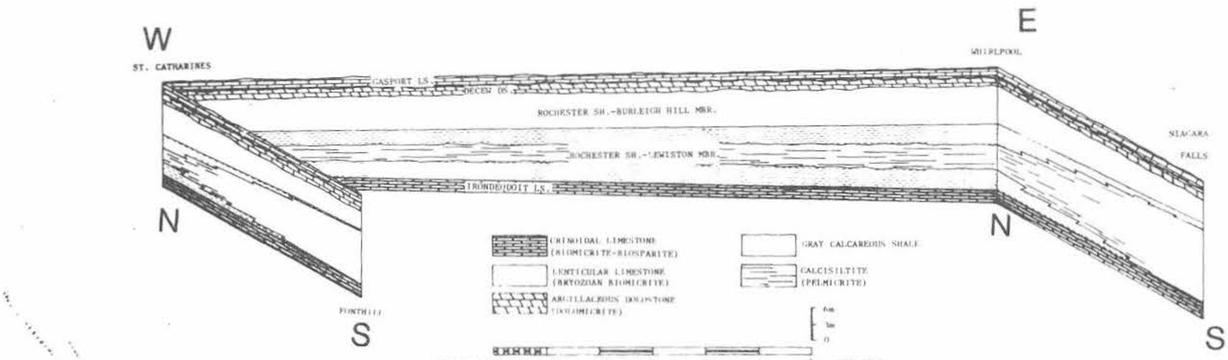

FIG. 5.-- Stratigraphic relationships of the upper Clinton and lower Lockport Groups from Niagara Gorge (east section) to the Fonthill Reentrant (west section). Note consistent north-south facies changes in each area, contrasting with facies continuity in the east-west oriented section.

of five or six 10-30 cm thick bands of barren, laminated dolomitic calcisiltites, closely resembling the higher DeCew Formation. At the site of the old Schoellkopf Hydroelectric Plant in the city of Niagara Falls, the bands merge into a single blocky dolomitic band about a meter thick; this unit persists to the southern end of Niagara Gorge. The only vestige of the highly fossiliferous limestones of unit E remaining at Niagara Falls is a thin (5-10 cm) uppermost bed of biosparite containing abundant crinoid ossicles and scattered brachiopod and bryozoan fragments.

Lithological changes in the upper Burleigh Hill Member in Niagara Gorge are less pronounced than those of the Lewiston. However, there is a general southward loss of calcisiltite beds in the upper portion of the unit. The lower contact remains sharply defined at the top of the upper calcarenite bed of the Lewiston Member. In contrast, the upper contact with the DeCew Dolostone, which is completely gradational at the north end of the Gorge, becomes sharp and undulatory south of the Whirlpool. The DeCew locally contains clasts of Rochester Shale, suggesting erosional rip-up.

The Burleigh Hill thins to the south along Niagara Gorge, as the Lewiston Member thickens, maintaining an approximately uniform thickness of the entire Rochester Shale. This is anomalous with respect to the general southward thickening of the Rochester (see above), and is attributed to two factors. First, the sharp erosional upper contact of the Rochester Shale, south of Whirlpool, suggests truncation of upper Rochester beds (possibly by downslope erosion). Second, south of the Whirlpool, the DeCew Dolostone exhibits a 1-2 m interbed of calcareous shale, closely resembling the Stoney Creek Member. This indicates interfingering of DeCew and upper Rochester facies.

Rapid southward facies change is also corroborated by field data from stream exposures of the Rochester Shale in the Fonthill Reentrant, an 8 km (5 mi) southward embayment in the Niagara Escarpment just west of St. Catharines. Here again, the Lewiston Member thickens and grades southward into a more monotonous, sparsely fossiliferous shale interval. Bryozoan-rich beds of unit E, well represented at DeCew Falls (loc. 19), again diminish and are replaced by calcisiltite beds capped by one or two thin beds of crinoid-brachiopod-rich calcarenite. A three-dimensional fence diagram (Fig. 5) demonstrates consistent southward changes in the Niagara Gorge and Fonthill sections with relatively little facies change along the main (east-west) escarpment outcrop belt.

A third area where abrupt north-south changes are observable in the Rochester Shale is the Burlington Bay Reentrant, at the western end of Lake Ontario. Here the Niagara Escarpment makes an abrupt bend from east-west to north-south; exposures of Rochester Shale occur in both north and south sides of the 8 km (5 mi) wide reentrant (Fig. 1). At Route 403 (loc. 5), on the south side of the reentrant, the Rochester

Shale is 4.3 m (14 ft) thick and is subdivisible into Lewiston and Stoney Creek members; 8 km (5 mi) to the north of this area at Clappison Corners (Rt. 6; loc. 1) the Rochester consists of 0.6 m (2 ft) of intraclastic sandy limestones and calcareous shales. North of this locality, the Rochester is represented by merely a shaley parting or is absent altogether. As noted previously, this rapid northward thinning and pinch-out of the Rochester Shale can be recognized in the subsurface as far west as the Windsor area of Ontario.

The rapid and substantial north-south facies changes of the Rochester Shale across the Burlington and Fonthill Reentrants and in the length of Niagara Gorge contrast with the east-west persistence of facies. This indicates that, in western New York and adjacent Ontario, Rochester facies belts are elongate east to west, parallel to a northern paleoshoreline and subparallel to the modern Niagara Escarpment. The rare north-south outcrop sections are approximately perpendicular to the strike of facies belts and thus show rapid internal changes. Finally, the stratigraphic cross sections of Niagara Gorge and the Fonthill Reentrant closely resemble hypothetical transgressive regressive cycles (cf. Raup and Stanley, 1978, p. 215).

DISCUSSION

The general depositional environment of the Rochester Shale in Ontario and western New York is envisaged as a shallow (less than 50 m) gently south-eastward sloping, muddy shelf (Fig. 6). This region was bordered on the southeast by a sandy shoreline and on the northwest by carbonate shoals. A hypothetical northwest to southeast transect would include the following facies: A) crinoidal bank-biohermal facies (Wiarton), coinciding with the Algonquin axis; B) either i, an argillaceous-carbonate facies reflecting mixed terrigenous and detrital carbonate sedimentation (Stoney Creek and upper Burleigh Hill-Gates Members), or ii, a bryozoan belt facies consisting of abundant patches of ramose bryozoan, and associated diverse brachiopods, echinoderms and other fossils, formed during times of low sedimentation rates (Lewiston B, D, and E); C) a calcisiltite facies comprising mudstone and interbedded barren, carbonate, storm-silt layers (Lewiston C, lower Burleigh Hill); D) a sparsely fossiliferous mudstone facies (undifferentiated lower Rochester Shale, lower Burleigh Hill in part).

Vertical facies changes in the Rochester Shale at a given section are attributed to lateral (north or south) shifts of environmental (facies) belts due to migration of a northern paleoshoreline. The Irondequoit to Gasport sequence constitutes a major transgressive-regressive cycle (Dennison, 1970; Dennison and Head, 1975), with two superimposed subcycles in the Rochester Shale.

As noted earlier, the Irondequoit and Gasport limestones appear in north-south cross sections as two basinward extensions of the Wiarton crinoid bank facies. From the preceding discussion, it is evident that these

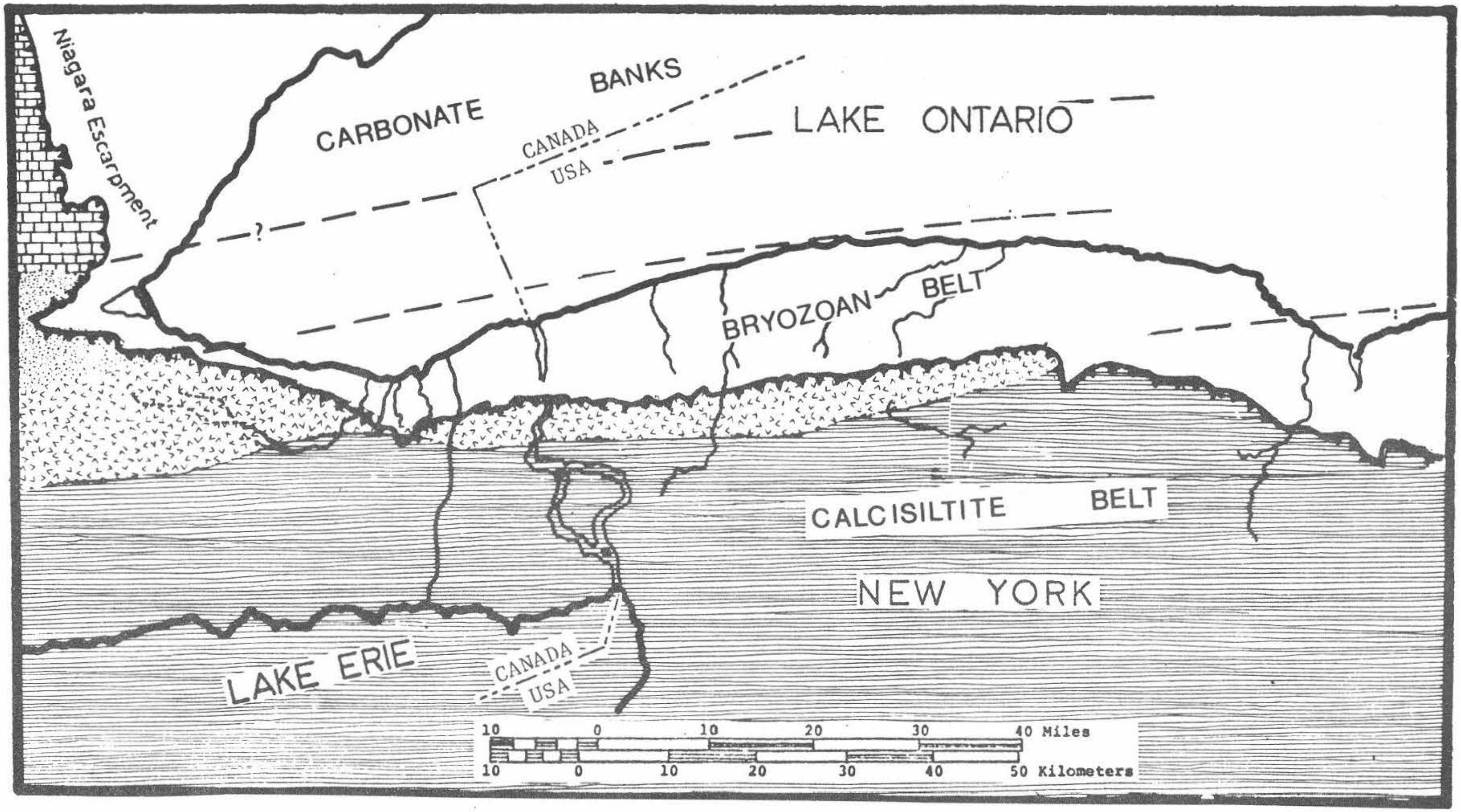

FIG. 6.--Facies map for Rochester Shale in Ontario and western New York State. Reconstruction shows approximate geographic position of facies belts during deposition of the upper portion of the Lewiston Member (submember E). Note parallelism between facies belts and modern Niagara Escarpment.

tongues define a large-scale transgressive-regressive cycle comprising Irondequoit-Rochester-DeCew-Gasport. Previous authors (Dennison and Head, 1979) suggested that the Rochester Shale was deposited during a single rise of sea level. However, the preceding discussion indicates that this is not entirely correct. Detailed stratigraphy of the Rochester reveals evidence for a minor regressive event of lesser magnitude than that which produced the Irondequoit or Gasport limestones during the middle of Rochester Shale deposition.

The Lewiston Member records a subsymmetrical cycle of deepening (units submembers A-C) and shallowing (submembers C-E). In contrast the upper unit (Burleigh Hill or Stoney Creek member) reflects aspects of an asymmetrical shallowing-upward hemicycle. The transgressive portion of this sequence is poorly preserved or absent and is usually represented by a slight disconformity: the contact between calcarenites of Lewiston E submember and upper barren shales. Greatly increased detrital carbonate sedimentation during deposition of the upper Burleigh Hill-DeCew interval inhibited the growth of bryozoans in shallow water and prevented the development of bryozoan facies analogous to those observed in the Lewiston Member. Predictably, in outcrops toward the center of the basin, e.g., in the Genesee River Gorge, these smaller scale cycles are less obvious, and the entire section takes on a more monotonous, homogeneous aspect.

ACKNOWLEDGMENTS

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ROAD LOG FOR SILURIAN ROCHESTER SHALE FACIES IN WESTERN NEW YORK

CUMULATIVE MILES	MILES FROM LAST POINT	ROUTE DESCRIPTION
	0.0	Begin trip at Buffalo Marriott Inn. At entrance turn left and proceed north on Millersport Highway (Route 263).
0.9	0.9	Coventry entrance to SUNY at Buffalo Amherst Campus.
1.8	0.9	Junction North Forest Road.
2.5	0.7	Junction Campbell Blvd. (270). Turn left (north) on Route 270.
		Junction Route 356.
3.3	0.8	Junction French Road.
5.7	2.4	Bridge over Tonawanda Creek. Enter Niagara County.
7.5	1.8	Junction Bear Ridge Road.
8.7	1.2	Junction Fiegel Road.
10.5	1.8	Junction Lockport Road.

11.7	1.2	Junction Saunders Settlement Road (Route 31).
13.2	1.5	Junction Upper Mountain Road. Route 270 ends, proceed across Upper Mountain Road onto Lockport Junction Road (Route 93 North).
13.5	0.3	Begin road cut in Lockport and Rochester formations.
13.6	0.1	Exposure of upper Rochester/DeCew Dolostone contact. Pull vehicles off in parking area on left side of road.

STOP 1. LOCKPORT JUNCTION ROADCUT. The first two stops of this trip provide an overview of the Rochester Shale in eastern Niagara County, for comparison with Niagara Gorge sections farther west. This exposure illustrates the gradational upper contact of the Rochester Shale with the overlying DeCew Dolostone. The upper 3-4 m of the Burleigh Hill Member of the Rochester Shale are exposed at the base of this cut. Nearly barren, friable, calcareous shale grades upward into a 1-2 m interval of interbedded dolomitic shale and buff weathering, burrow-mottled layers of argillaceous laminated calcisiltite. These beds appear to be gradational into the overlying DeCew Dolostone. This transitional upper Rochester unit thickens eastward to form the Gates Member which is recognized near the type section at Rochester, New York. The upper portion of the roadcut exposes an excellent section of the Gasport Formation (Lockport Group) and the sharp contact between the DeCew Dolostone and the cross-bedded, crinoidal biosparites of the Gasport.

The uppermost Rochester and DeCew are interpreted as rapidly-deposited carbonate silt and clay sediments, winnowed from Gasport crinoidal shoals which lay nearby to the northwest. Cross lamination and abundant soft sediment deformation in the DeCew suggest rapid influx of sediment followed by dewatering. Body fossils are exceedingly rare in the upper Rochester and the DeCew, but burrows, including Chondrites, Planolites and Teichichnus are abundant. This may,again,reflect high turbidity.

13.6		Return to vehicles and reverse directions, proceeding south on Lockport Junction Road.
14.0	0.4	Junction of Upper Mountain Road. Turn right (west).
14.7	0.7	Junction Thrall Road on right. Turn right (northwest).
15.1	0.4	Small exposure of Gasport Limestone.

16.1	1.0	High cutbank in Rochester Shale and Lockport Group, in pasture along south side of Thrall Road. Park vehicles on right hand side of road in front of driveway to farm.

STOP 2. THRALL ROADCUT, CAMBRIA, NEW YORK. Although somewhat weathered, this exposure provides a good reference section for the middle Rochester Shale in eastern Niagara County. Here the upper 4 or 5 m of the Lewiston Member, and its sharp contact with the overlying Burleigh Hill are exposed. Although largely covered by slumped talus, limestone ledges of Lewiston submember E ("bryozoa beds" of Grabau, 1901) protrude from the bank in several locations. This is an excellent locality for collecting the diverse fauna of middle Rochester beds. Weathered talus at the base of the slope has yielded over 20 species of brachiopods, corals, gastropods and complete thecae and crowns of *Stephanocrinus*, *Caryocrinites*, and various crinoids. Weathered slabs of limestone replete with ramose and fenestrate bryozoans are abundant here. Note the sharp contrast with the overlying barren mudstone of the Burleigh Hill Member. Thin calcisiltites comprise the only interbeds in this upper portion of the Rochester. Weathered slabs from the upper bank have yielded abundant trilobite remains (*Dalmanites*, *Trimerus*), ostracodes, *Tentaculites* and crinoid columnals. However, the bulk of the interbedded shale is very sparsely fossiliferous, in marked contrast to the underlying Lewiston beds. The top of this section is again capped by buff weathering DeCew Dolostone and Gasport Limestone.

16.1		Return to vehicles and proceed northwest along Thrall Road.
16.7	0.6	Junction Blackman Road.
16.8	0.1	Y-junction of Thrall Road onto Lower Mountain Road. Bear left (west) onto Lower Mountain Road.
17.6	0.8	Junction Cambria Road.
29.8	12.2	Junction Route 425 on right.
29.9	0.1	Junction Shawnee Road. We are now riding along the base of Niagara Escarpment.
30.6	0.7	Junction Baer Road. Turn left (south) and proceed up low ridge formed by lower platform of Niagara Escarpment on Irondequoit Limestone.
30.72	0.12	Roadcut in Rockway and Irondequoit Limestones. Park near south end of cut.

STOP 3. BAER ROADCUT. There are very few outcrops in Niagara County where the basal beds of the Rochester Shale can be examined. This will be a brief stop to observe the uppermost Irondequoit Limestone which immediately underlies the Lewiston Member. Here the Irondequoit is a crinoidal biosparite, similar to the Gasport Limestone. Well-preserved fossils occur sparingly in the upper Irondequoit, including thecae of Stephanocrinus, Caryocrinites and other echinoderms. Small, white-weathering bioherms composed of fistuliporoid bryozoans and bound biomicrite occur near the top of this exposure. Such small bioherms occasionally protrude upward into the basal beds of the Rochester Shale.

30.72		Continue southward along Baer Road crossing the small plateau on Irondequoit Limestone and then up the main Niagara Escarpment.
31.4	0.7	Small exposure of Gasport Limestone at turn in road.
31.9	0.5	Intersection Upper Mountain Road. Turn right (west).
32.0	0.1	Overpass over Route 429. The classic Pekin roadcut through a Gasport bioherm is exposed along 429 immediately below Upper Mountain Road at this locality.
32.2	0.2	Access road to Route 429 (Grove Street).
32.7	0.5	Junction Bridgeman Road.
33.1	0.4	Meyers Hill Road.
33.6	0.5	Excellent view off Niagara Escarpment onto Ontario Lowlands Plain. On clear days this affords a view of Lake Ontario.
34.7	1.1	Junction Black Nose Spring Road.
35.3	0.6	Junction Walmore Road.
37.5	2.2	Y-intersection with Model City Road. Bear left staying on Upper Mountain Road. View to right out to Ontario Lowlands Plain and Lake Ontario.
38.1	0.6	Western boundary of Tuscarora Indian Reservation.

38.4	0.3	Reservoir for Robert Moses Hydroelectric Plant appears as bank on left.
39.1	1.7	Junction of Upper Mountain Road and Military Road at stop light. Proceed straight across intersection onto access road for Robert Moses Parkway, Route I-190, and Route 104.
39.15	0.05	Exit for Route I-190 north onto Lewiston-Queenston International Bridge.
39.35	0.2	Exit for Route I-190 south.
39.55	0.2	Junction Route 104 east (north) to Lewiston.
39.7	0.15	Junction Route 104 west (south) to Niagara Falls.
39.9	0.2	At Y-intersection merge with southbound lane of Robert Moses Parkway and cautiously cross to right hand lane.
40.0	0.15	Pull off on broad shoulder of Robert Moses Parkway and park at 55 mph sign. Passengers will disembark here, proceed over guard rail and down a slope, bearing to the right or north to cliffs along Niagara Gorge (see Fig. 7 point A).

STOP 4. NORTH END, NIAGARA GORGE NEAR LEWISTON, NEW YORK. Cliffs along the east face of Niagara Gorge between Lewiston Queenston Bridge and the Robert Moses Power Plant provide an excellent section of the entire Irondequoit, Rochester, and lower Lockport units. Here the Rochester Shale is approximately 17 m thick; the lower half comprises the Lewiston Member which consists of about 8.7 m (28.6 ft) of medium to brownish gray calcareous shaley mudstone with interbedded thin carbonates. The member exhibits a gradational lower contact with Irondequoit Limestone, but is sharply differentiated from the overlying Burleigh Hill Member at the top of a 1-2 m interval of bryozoan rich limestones (unit E). As at most localities in western New York and adjacent Ontario the Lewiston Member can be subdivided into six informal units or submembers designated by letters A-E. All of these beds are readily accessible in this cliff section and will be described as follows:

A) The basal transition zone of the Lewiston is up to 1 m thick and consists of dark-brownish to ashy-gray, silty, very calcareous and fossiliferous mudstone with interbeds of argillaceous limestone (biomicrite and/or biosparite). Brachiopods are very abundant and may be packed into layers; characteristic species include: Atrypa reticularis,

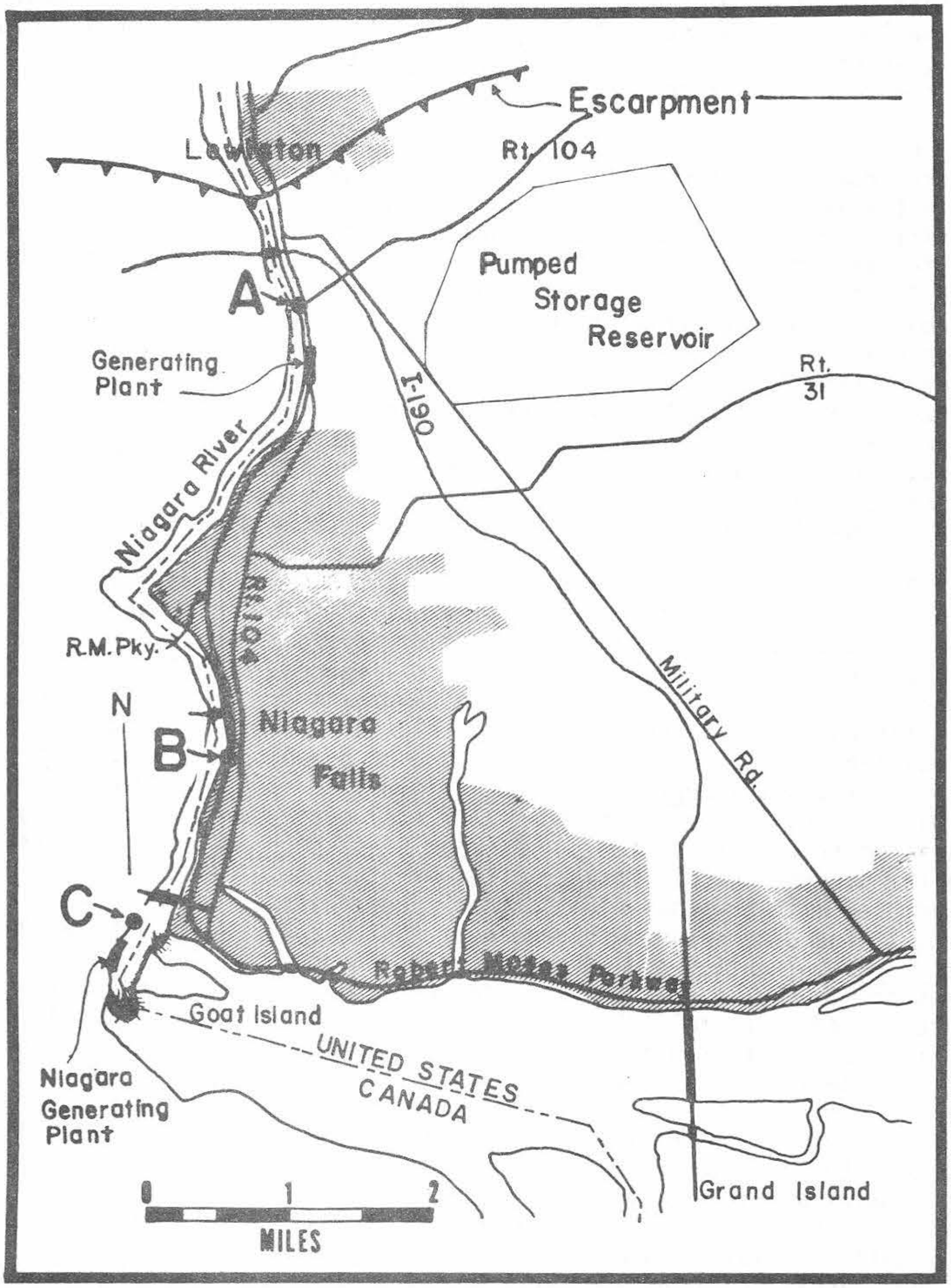

FIG. 7.-- Location map for fieldtrip stops along Niagara Gorge. Locations include: A) Stop 4, Cliffs along east side of Niagara Gorge south of Lewiston-Queenston bridge; B) Stop 5, exposures at Niagara sewage pumping station, Niagara Falls, New York; C) Stop 6, exposures along access road for Ontario Hydro Niagara Generating Plant, Niagara Falls, Ontario.

large Leptaena rhomboidalis, Whitfieldella oblata and Plectodonta transversalis. Bryozoans are less common than in overlying beds, but laminar fistuliporoids are not uncommon. Echinoderm debris is abundant in the lower shales and unit A has yielded certain unique forms including a high, narrow morph of Stephanocrinus angulatus, large Eucalyptocrinites, and the very rare crinoid Paracolocrinus (Brett, 1978).

Locally, small bioherms 1-5 m across and up to 2 m high protrude upward from the Irondequoit Limestone into the base of the Rochester Shale (Sarle, 1901). These mounds are composed of whitish to massive micritic limestone with abundant laminar fistuliporoid bryozoans. A particularly good example of one of these mounds, now inaccessible, but readily visible, occurs along the high wall of Irondequoit Limestone immediately south of this Rochester Shale outcrop. This mound appears to interfinger laterally with shales of unit A; however, it also contains pockets filled with a greenish clay shale, unlike the basal Rochester. Bioherms such as this have yielded a number of unique fossils not seen elsewhere in the Rochester Shale. Large Bumastus and Calymene trilobites, sometimes as coquinites of cephalic and pygidial shields occur in shaley pockets within the bioherms. Unit A clearly exhibits features transitional between typical Irondequoit Limestone and Rochester Shale.

B) The lower beds of the Rochester Shale grade upward into a 2.5-3 m interval of softer, less calcareous, medium to brownish-gray mudstone with numerous lenticular beds of bryozoan-rich biomicrites. As a whole, this interval is richly fossiliferous, although some mudstone beds are nearly barren. Ramose bryozoans are particularly characteristic and they frequently occur in clusters packed in a mudstone matrix. Associated with these clusters are abundant brachiopods such as Atrypa reticularis, small Leptaena, and Whitfieldella. This interval is particularly notable for the abundance of echinoderm remains including thecae of the blastoid-like pelmatozoan Stephanocrinus angulatus, and the rhombiferan cystoid Caryocrinites ornatus, both of which are typically associated with bryozoan-rich patches. Holdfasts of Caryocrinites are commonly found cemented to bryozoans. Gastropods, the trilobites Bumastus, Arctinurus, Dalmanites and others are also abundant in these beds. Certain limestone lenses in unit B appear to represent starved ripples of storm-concentrated, skeletal debris. Rotated geopetal structures and infillings of brachiopod shells with matrix identical to that of the underlying beds suggest derivation of these fossils by reworking during storms.

C) Near the north end of Niagara Gorge the middle interval of the Lewiston Member consists of about 3-4 m of sparsely fossiliferous gray shales and interbedded dark-gray, burrowed, laminated calcisiltites (pelmicrites). Bryozoan clusters and biomicritic limestones are absent from this interval. These shales exhibit a low diversity fauna dominated by the brachiopods Striispirifer, Strophonella and Parmorthis and the trilobites Dalmanites, Trimerus and Arctinurus near the base of unit C occur thin horizons which locally contain exceptionally well-preserved fossils. These layers, in all about 10-20 cm thick, were designated the Homocrinus band at Lockport, New York (Ringueberg, 1888); they contain the characteristic minute inadunate crinoid Homocrinus parvus, as well as completely articulated specimens of Asaphocrinus and Macrostylocrinus, the cystoid Caryocrinites, edrioasteroids and rare starfish. These fossils are associated with dense bedding plane assemblages of Striispirifer and other brachiopods. Such assemblages suggest rapid smothering of the sea floor by muds, producing catastrophic (live) burial of benthic communities by storm-generated mud layers (Brett, 1978; 1980)

Homocrinus-bearing layers have been recognized at Niagara Gorge about 4.6 m (15 ft) above the base of the Lewiston Member. However, no large assemblages have as yet been obtained from this locality. Calcisiltites of Lewiston C also appear to record a type of storm-generated layer. These units form thin (1-2 cm) persistent bands. Weathered surfaces reveal fine, planar- and hummocky- cross lamination, often in alternating sets. Their upper surfaces may be rippled and basal contacts are invariably sharp. The undersurfaces of some calcisiltites yield abundant sole marks including flute and groove molds which suggest current scouring of cohesive muds; occasionally, well-preserved fossils may also occur on the undersides of these beds. Most calcisiltites are barren of body fossils, but they contain abundant trace fossils including large vertical shafts - possibly escape burrows - and post-depositional mining structures such as Chondrites and Planolites.

Rochester calcisiltites closely resemble storm silt layers described by Reineck and Singh (1972), from the modern North Sea and by Kreisa (1981) from the Ordovician Martinsburg Formation. They are interpreted as deposits from storm-generated density currents which transported carbonate silts from shallow platform areas into the Rochester depositional basin.

D) Overlying unit C is an interval, up to 2 m thick, of fossiliferous mudstone with biomicrites which closely resembles unit B, both in lithology and fauna. Again, clusters of ramose bryozoan are abundant and exhibit virtually all of the associated fauna found within the lower unit B. These beds, again, yield abundant atrypids and Whitfieldella, Striispirifer, Leptaena, Bumastus trilobites, dendroid graptolites and the pelmatozoans Stephanocrinus angulatus and Caryocrinites. Lenticular calcarenites resembling those in unit B become increasingly abundant upward in unit D whereas barren calcisiltites are less frequent than in unit C.

E) "Bryozoa beds." The top of the Lewiston Member is marked by a series of thin lenticular limestones (biomicrites and biosparites) interbedded with soft, gray, fossiliferous shale. Grabau (1901) termed this unit the "bryozoa beds," because of the great abundance of ramose and fenestrate bryozoans which constitute a bulk of the thin limestones. Brachiopods and echinoderms are also abundant in these beds although typically disarticulated and fragmented. Notable is the reappearance of a large, slender morphotype of Stephanocrinus seen also in unit A at the base of the Rochester Shale.

The uppermost bed of unit E is generally somewhat thicker and more continuous than the underlying layers and may be richer in brachiopod valves and crinoidal debris, resembling the uppermost layers of the Irondequoit Limestone. This unit marks the top of the Lewiston Member in nearly all localities. Locally, this topmost limestone bed is glauconitic and contains shale intraclasts, evidently derived by erosion from the underlying beds.

At this locality, as in most areas of western New York, the Lewiston Member is abruptly overlain by medium to dark gray platey and calcareous shale of the Burleigh Hill Member, named for exposures along Burleigh Hill Drive in St. Catharines, Ontario (Brett, in press). In the north end of Niagara Gorge the Burleigh Hill is about 10 m thick. However, as will be noted, this unit thins rapidly southward along Niagara Gorge. The basal 4-5 m of the member are sparsely fossiliferous to nearly barren, laminated and platey shales which weather into elongate slabs. Lithologically and faunally this portion of the Burleigh Hill Member most closely resembles unit C of the Lewiston Member, and, as with the latter unit, these shales contain occasional thin, laminated calcisiltites which increase in frequency upward. No bryozoan biomicrites have been observed in the Burleigh Hill Member at any locality in western New York and bryozoans are generally absent. Near the base of the Burleigh Hill are a few thin shell layers composed chiefly of the brachiopods _Fardenia_, _Parmorthis_, and _Strophonella_ as well as trilobite fragments.

The upper half of the Burleigh Hill Member is gradational into the overlying DeCew Dolostone. Bands of calcisiltite become more closely spaced and eventually merge into a silty dolomitic shale near the top. A few thin layers near the top of the Burleigh Hill have yielded local patches of well-preserved crinoids including _Dimerocrinites_ and _Dendrocrinus_ as well as brachiopods and other fossils. Burleigh Hill dolomitic shales appear to pass upward gradationally into the overlying DeCew Dolostone Formation. Unlike the Lewiston Member, which appears to represent a subsymmetrical cycle, Burleigh Hill reflects a shallowing-upward hemicycle.

40.05		Return to vehicle and proceed southward along Robert Moses Parkway.
40.15	0.1	Beginning of penstocks for Robert Moses Power Plant.
40.45	0.3	Overpass of walkway to Robert Moses Power Vista.
41.05	0.6	Pass access road for Adam Beck Plant on Canadian side of Gorge. View of South Haul Road for Robert Moses Power Plant on U.S. side.

41.65	0.6	Devils Hole pulloff.
43.0	1.35	Entrance for Whirlpool State Park.
43.15	0.15	Exit on left for reversing directions on Robert Moses Parkway. Continue south on Robert Moses Parkway.
43.0	0.35	Exit on right for Whirlpool Street and bridge to Canada. Bear right onto exit.
43.75	0.25	Junction Whirlpool Street. Turn right.
43.85	0.1	Underpass for Whirlpool bridge to Canada.
43.9	0.05	Underpass for railroad bridge to Canada.
44.05	0.15	Junction Route 182. Keep right on Whirlpool Street.
44.13	0.8	Fork of Second and Third Streets at the end of Whirlpool Street. Bear right onto Second Street
44.23	0.1	Niagara Aquarium on left.
44.33	0.1	Junction of access road for Niagara sewage pumping station. Make sharp right turn onto dead end access road, and bear right under overpass to Schoellkopf Museum (on left).
44.53-44.63	0.2-0.3	Small exposures of Goat Island Dolostone of the Lockport Group in the roadcut.
44.58-44.68	0.05	Underpass of access road beneath Robert Moses Parkway. Proceed beneath underpass.
44.73-44.83	0.15	Pull off on left and park. Proceed on foot down old access road to Niagara sewage pumping station in the Gorge. Note well-exposed bioherm in the Gasport Limestone along roadcut.

STOP 5. NIAGARA SEWAGE PUMPING STATION. This stop illustrates the contrast between northern and southern sections of Rochester Shale in the Niagara Gorge. A large artificial cut behind the new sewage pumping building exposes the upper portion of the Lewiston and the entire Burleigh Hill Members of the Rochester Shale. At this location in the gorge the Lewiston Member is approximately 12.5 m (41 ft) thick; the majority of this thickness consists of unfossiliferous shales of a greatly thickened submember C, here about 11 m (36 ft) thick in contrast to 3.6 to 4 m (11-14 ft) at Stop 4. The upper portion of this unit, visible at the base of the cut, consists of nearly barren, silty shale with one or two thin calcisiltite bands. The fossil-rich submember D is not recognizable at this locality; its position appears to be occupied by a section of interbedded shales and thin (5-10 cm) calcisiltite bands. Four or five prominent ledge-forming dolomitic calcisiltites occur near the center of the outcrop, apparently representing the lateral equivalent of "bryozoa beds" (submember E). These bands exhibit sedimentary structures including hummocky cross lamination, rippled upper surfaces, and sharp, scoured basal surfaces, suggestive of current deposition of fine carbonate silt. The beds are interpreted as layers derived from winnowing of bryozoan-crinoid shoals which existed to the north (e.g. at the northern of Niagara Gorge). Ripple- and cross-bed orientation suggest southerly current transport of sediments.

Immediately overlying the highest dolomitic band are 1 or 2 thin layers of coarse, rusty weathering biosparite containing an abundance of brachiopods, pelmatozoans and some bryozoan debris, interbedded with slightly fossiliferous shales. These are essentially the only fossil-rich beds in the entire outcrop and they appear to represent a southern vestige of the topmost bed of unit E in the north. Two km to the north the Whirlpool State Park these bands thicken into a 10-20 cm thick interval of fossil debris and the underlying dolomitic calcisiltites interfinger with typical bryozoan-bearing beds indicating a clearcut transition from typical biosparites of unit E, to the dolostones seen at this locality. The upper bed of fossil debris is followed abruptly by barren shales of the Burleigh Hill Member, which closely resemble those of unit C. The Burleigh Hill is about 6m thick and displays a sharp, probably erosional, upper contact with the DeCew Dolostone.

This sparsely fossiliferous section stands in striking contrast to sections just a few miles north. It suggests the presence of a fairly abrupt, southward dipping paleoslope.

44.83		Return to vehicles and reverse route to Second Street.
45.13	0.3	Junction Second Street. Turn right (south).
45.28	0.15	Junction Main Street (Route 104). Turn right (southwest) and proceed straight ahead to entrance for Rainbow International Bridge.

45.53	0.25	U.S. Toll Booth of Rainbow Bridge.
45.83	0.3	Canadian Customs Booth. Those requiring passports should have these ready at this time.
45.88	0.05	Bear right from exit of Customs Booth onto street toward Falls.
45.98	0.1	Turn right onto Hiram Street. Junction River Road. Turn right (south) and proceed toward Falls.
46.28	0.3	Pass circle at Maid of the Mist information building. Pull off bus along the river road here and park. Proceed on foot to access road for Ontario Hydro Niagara Plant, and Maid of the Mist launch area. Note excellent exposures of Lockport Group beginning at the top of the roadcut with Eramosa Dolostone, followed below by the vuggy stromatolitic and stromatoporoid-bearing Goat Island Dolostone and the dolomitic crinoidal limestone of the Gasport Formation; this unit, in turn, is underlain by buff-weathering, sugary-textured, DeCew Dolostone, exhibiting prominent soft sediment deformation ("enterolithic structure"). The access road forks into two branches, one leading to the straight ahead to the Ontario Hydro Plant, and the other, left hand fork leads to the Maid of the Mist launch area. Near the fork are exposures of the upper part of the Rochester Shale.

STOP 6. ONTARIO HYDRO ACCESS ROAD CUT. This roadcut provides one of the southernmost exposures of the Rochester Shale in New York State. Here the upper contact of the formation is well exposed, just north of the fork in the road. The buff-weathering enterolithic DeCew Dolostone rests with sharp, wavy, and apparently channeled contact on the barren shales of the middle Burleigh Hill Member. No transitional interval is observed at this locality and the Burleigh Hill is only about 4.7 m thick. This situation contrasts with the gradational upper contact of the Rochester Shale observed in most western New York localities. It suggests that the upper Burleigh Hill beds have been removed, possibly by downslope erosional truncation, prior to DeCew carbonate silt deposition. The prominent enterolithic structure of the DeCew Dolostone has long been noted and may have resulted from downslope slumping of rapidly deposited carbonate sediments.

The Lewiston Member is exposed to the south of the road fork, and exhibits an upper blocky band of dolomitic mudstone. Note the contrast to the several rippled bands seen at Stop 5. This unit exhibits many features in common with the DeCew Dolostone, including soft sediment deformation structure. Presumably both units record rapid deposition, dewatering and possible slumping of winnowed carbonate silts. The central block (Schoellkopf bed) is followed by interbedded calcareous shale and dolomitic limestones, including one laminated calcisiltite band about 10 cm thick. The uppermost beds include crumbly calcareous shales and lenticular dolomitic limestone, bearing scattered brachiopods and bryozoans, virtually the only fossils found in this outcrop. Again, these beds are thought to represent a remnant of unit E seen farther north. If time and conditions permit we may continue on foot down to exposures in cuts behind the Ontario Hydro Plant. A thick section of nearly barren dolomitic shales with relatively few limestone bands comprises the bulk of the Lewiston Member. Slightly more fossiliferous beds near the base of this section may reflect a southern equivalent of unit B; the basal contact with the underlying Irondequoit Limestone cannot be observed here. The blocky upper beds of the Lewiston Member appear to phase southward into dolomitic shales. Thus, the distinction between Lewiston and Burleigh Hill Members is becoming vague and blurred. Extrapolating trends seen in this outcrop, one might predict that the entire Rochester Formation becomes nearly uniform, calcareous and sparsely fossiliferous shales only a few km to the south.

46.28		Return on foot to vehicles and retrace to Rainbow Bridge.
46.58	0.3	Pay toll at Canadian booth.
46.88	0.3	U.S. Customs. Proceed straight from Customs following signs for Robert Moses Parkway south and Route I-190.
47.08	0.2	Junction Route 384. Turn right (south).
47.18	0.1	Junction of entrance road to Robert Moses Parkway, near old Nabisco Shredded Wheat building. Turn right (south) and continue onto entrance ramp for Robert Moses Parkway southbound, to Route I-190.
48.98	1.8	Intake gates for Niagara Power Project.
50.28	1.3	Exit for Route I-190 south. Bear left onto exit lane and merge onto I-190 southbound lane.

50.7-51.2	0.5-1.0	North Grand Island Bridge over Niagara River just east of the junction of east and west branches of Niagara River around Grand Island. Toll Booth is on Buckhorn Island, a small subsidiary island on the north end of Grand Island. Proceed straight on I-190 southeastward across Grand Island.
56.7-58.8	6.0-7.6	South Grand Island Bridge across east fork of Niagara River.
58.0-60.1	1.3	Junction I-290 (Youngmann Highway). Take exit for I-290 east.
66.5-68.6	8.5	Exit for Millersport Highway (Route 263) northbound. Take exit and proceed north on Millersport Highway.
66.8-68.9	0.3	Entrance to Marriott Inn. End of field trip.

GLACIAL AND ENGINEERING GEOLOGY ASPECTS OF THE NIAGARA FALLS AND GORGE

Parker E. Calkin, Department of Geological Sciences, State University of New York at Buffalo, Buffalo, NY 14214
Thomas A. Wilkinson, U.S. Army Corps of Engineers, Buffalo District, Buffalo, NY 14207

INTRODUCTION

The Niagara River separating Canada and the United States (Frontispiece II) is unusual as compared to other rivers for its (a) short length and (b) discharge stability due to the immense storage capacity of its drainage basin. It is largely supplied by the excess discharge brought into Lake Erie from Lakes Superior, Michigan and Huron. In addition to their stabilizing effect, the Great Lakes trap most of the basin sediment so that the river is essentially sediment-free (Philbrick, 1970). From its mouth at the north end of Lake Erie at 174.4 m (572 ft) the Niagara descends 23.3 m (73 ft) to the brink of the Falls, drops vertically 51 m (167 ft) (Fig. 1), and then descends another 22.6 m (74 ft) in gorge to Lake Ontario at 75 m (246 ft). The mean natural flow is about 5,721 m^3 sec^{-1} (202,000 ft^3 sec^{-1}), and can be increased temporarily by as much as 50 percent due to water surface set up during storms along Lake Erie. However, it is otherwise very stable. Since about 1905, the flow over the Falls itself has been markedly limited by water diversion for hydroelectric power generation. Presently it carries 50 percent of the river's natural flow or about 2,800 m^3 sec^{-1} (100,000 ft^3 sec^{-1}) during tourist hours and only about 25 percent (1,400 m^3 sec^{-1}/50,000 ft^3 sec^{-1}) at other times. About 92 percent of undiverted flow passes over the Horseshoe Falls, 8 percent over the American Falls, and this percentage may have been the average even under precontrol discharges (American Falls International Board, 1974).

The differences in elevation that cause the Falls is related to the presence of the Niagara cuesta (escarpment) that divides the Erie and Ontario basins (Frontispiece I and II). The canyon (gorge) of the Niagara has been cut southward from this escarpment into a section of jointed, gently south-dipping sedimentary rocks capped by the tough Lockport dolostone. The stratigraphy is considered in detail in a subsequent section of this paper and in Brett (1982, this volume).

The modern Niagara River was initiated as a multi-outlet river-lake system following the last retreat of the Last Winconsin ice sheet; the cutting of its deep incised gorge through time is related in various degrees to the changes in drainage controlled by this ice retreat. The ancestral development culminating with a more detailed chronology for development of the Upper Great Gorge is the initial subject of this paper. The second part considers details of the geology in the present area of the Falls and the geologic processes that control stability as well as retreat of the Falls and Gorge as they are revealed in the context of major engineering projects undertaken principally along the east (American) side of the Niagara Gorge. A third section considers the future of the Falls.

DEVELOPMENT OF THE ANCESTRAL NIAGARA

The generalities and many details of the geology and evolution of the Niagara River drainage system have been presented and updated successively by a number of authors since early work by Charles Lyell (1838-1842) and James Hall (1842). Such accounts are found in studies by Gilbert (1891, 1895, 1907), Grabau (1901), Spencer (1907), and particularly Taylor (Kindle and Taylor, 1913; 1933). Much of this section is taken freely from a recent synthesis and study of Calkin and Brett (1978). The Late Wisconsin glacial events leading up to the formation of the modern Falls and Gorge are considered by Calkin (1982, this volume).

The Lake Tonawanda Phase

Initiation of the Niagara River occurred following the Late Wisconsin, Port Huron Stadial as the ice margin retreated from western New York for the last time. In this area, the orientation of striations, bedrock fluting, and drumlins indicate that the ice had moved accross the area from the northwest (see Kindle and Taylor, 1913). Glacial retreat in the Erie Basin like those in the rest of the Great Lakes drainage was down the regional slope and therefore accompanied by a generally lowering succession of large proglacial Great Lakes (Calkin, 1982, this volume). Glacial Lake Dana, the last and lowest of this succession, formed in the northern part of the Erie Basin and southernmost Ontario Basin as the ice margin backed northward from a position near the Albion-Rochester Moraine along the Niagara Escarpment (Frontispiece II).

Lake Dana drained westward through channels at Syracuse and existed only long enough to form weak beach ridges at about 205 m (673 ft) near Buffalo before continuing retreat opened lower outlets to the east. This event led to emergence of the Niagara Escarpment and consequent formation of Early Lake Erie. This discharged via the incipient Niagara River across the emergent escarpment to Glacial Lake Iroquois which formed almost contemporaneously in the Ontario Basin. Lake Iroquois drained into the Mohawk River over a threshold near Rome, New York and thence to the Hudson Valley. Stabilization occurred just before a minor glacial readvance to the Carlton Moraine along the present south coast of Lake Ontario (Muller, 1977a).

The time for inception of these events and the Niagara River is not bracketed closely by reliable radiocarbon dates; however, those available suggest formation about 12,300 yr B.P. (Calkin and Brett, 1978; Karrow, 1981).

Early Lake Erie received the discharge from Glacial Lake Algonquin in the Hudson Basin via Port Huron (through Lake St. Clair and Detroit River) and was very narrow; its outlet near Buffalo was depressed as much as 42 m (138 ft) from that of the present (Lewis, 1969). Former major outlets cut across the Niagara Escarpment west of Niagara Falls in the preglacial ? Erigin Valley (Karrow, 1973), or the St. Davids Gorge (Figs. 1 and 2) were drift-filled. The outflow of Early Lake Erie therefore flooded the lowland between Onondaga and Niagara escarpments forming the shallow, multi-outlet Lake Tonawanda (Frontispiece II).

Lake Tonawanda was really a river-lake extension of the upper Niagara River; it stretched 93 km (58 mi.) eastward at its high strand-producing level of $\sim$178.3 m (585 ft) near Niagara Falls and averaged about 10 m (30 ft) in depth. Sediments are typically mottled, buff, silty fine sands, sparsely fossiliferous (see Calkin and Brett, 1978) and up to 3 m ($\sim$10 ft) thick. Finer sediment was carried from Lake Tonawanda with the general outflow via spillways cut through the Niagara escarpment to Glacial Lake Iroquois at Lewiston, and farther east at Lockport, Gasport, Medina, and Holly, respectively (Kindle and Taylor, 1913; Frontispiece II).

At least one of these bedrock spillways had been cut previous to this outflow. Clayey sand and gravel described from deep borings in the lower portion of the Lockport spillway is interpreted as till (Calkin and Brett, 1978).

During initial stages of the Niagara River, Lake Tonawanda discharged to the low, short-lived Newfane phase of Glacial Lake Iroquois; this built weak beaches 26 m (85 ft) above the present Lake Ontario level at Lewiston (Kindle and Taylor, 1913). Wood from the Lockport site (Frontispiece II) dated at 12,100 $\pm$ 400 yr B.P. may have been washed across the gravelly delta of the Lockport spillway during Newfane time (Miller, 1973). Uplift subsequently raised Lake Iroquois 12 m (39 ft) to its most persistent "Lewiston" phase. The main Iroquois strand is characterized by a very well developed and continuous storm beach ridge about 6 m (20 ft) high.

Lake Iroquois may have drained about 12,000 to 11,000 yr B.P. but its duration is difficult to determine because of uncertainties in the radiocarbon dates from shell materials (Karrow, 1981). Iroquois was succeeded by several post Iroquois proglacial lakes and subsequently by the low, Admiralty phase of Early Lake Ontario as concurrent ice margin retreat allowed the outflow to move respectively from the Mohawk Valley (thence Hudson Valley), to the Covery Hill outlet north of the Adirondacks, and then to the Champlain Sea in the St. Lawrence Valley. The position of the ice front during this period and the timing and verification of outlet routes from the Ontario Basin may have controlled in a more complicated manner by the Valders (Great Lakean or Two Rivers) advance allowing fluctuation of Ontario waters through more than one sequence of >150 m (Gorman and others, 1978; see also Gadd, 1980 and Karrow, 1981).

The main channel of the Lewiston spillway system headed 1,200 to 2,000 m (4,000 to 6,500 ft) north of the American Falls and eventually became the main "Niagara River channel" as differential isostatic uplift toward the northeast caused the eastern outlets to give way successively to those in the west closer to the Lake Erie outflows. Dated wood from the interface of gravelly clay of Lake Iroquois and overlying wood marl depsoits at the lower end of the Lockport spillway indicates that major sediment discharge and most water flow through this spillway had ceased by about 10,920 ± 160 yr B.P. and that Lake Iroquois had drained (Calkin and Brett, 1978).

Fluctuation of the Upper Great Lakes and Niagara River Discharge

As the Lewiston spillway (future Niagara Gorge) was receiving a successively greater share of the Niagara discharge through uplift of the eastern outlets, Lake Erie was also rising, causing an increase in lake size. Likewise, changes in the position of the ice margin to the north and west caused discharges into Lake Erie and hence the Niagara via Port Huron and the Huron Basin to change or cease entirely (see maps in Hough, 1966; Prest, 1970). A reduction of 80 to 90 percent must have occurred as ice retreat opened the Kirkfield Ontario outlet to the Trent River and allowed waters of Glacial Lake Algonquin in the Huron Basin to drain directly into the post-Iroquois lakes or Early Lake Ontario. Complete bypassing by Huron and western waters may have occurred in one or even two episodes about 11,500 yr B.P. (Karrow and others, 1975; Karrow, 1981). A possible second episode may have occurred following a very brief post-Two Creeks - Valderan advance across the Trent River (Hough, 1963; Fullerton, 1980) before isostatic uplift closed the Kirkfield outlet. Continued ice retreat opened direct eastward discharge to the St. Lawrence basin about 10,400 yr B.P. (Karrow and others, 1975) at North Bay, Ontario. The upper lakes then drained directly through the Ottawa Valley into the St. Lawrence Valley until about 5500 yr B.P. when outlets to the south at Chicago and again at Port Huron became active (Lewis, 1969). Between 4700 and 3700 yr B.P. during Lake Nipissing II stage in the Upper Great Lakes, drainage was shared by only the Chicago and Port Huron outlets; full discharge from the upper lakes returned through the Port Huron-Lake Erie-Niagara River route in post-Nipissing time, about 3700 yr B.P.

Timing of Local Events

Discharge variations through the Niagara and fluctuations of base-level in the Ontario Basin must have affected the rate and manner of Niagara Falls recession as did a multiplicity of other factors (see succeeding successions of this paper). Taylor (Kindle and Taylor, 1913; 1933) proposed a correlation of Gorge sections with lake history (Table 1). This and similar correlations based largely on physical dimensions of the gorge (Fig. 1) are yet unproved and speculative but are at least compatible in part with the recent dated chronologies (Calkin and Brett, 1978; Fullerton, 1980). Some recent studies of the Niagara River throw light on the local timing of gorge development.

TABLE 1 - CORRELATIONS OF GORGE SECTIONS BY TAYLOR (1933)

Glacial Lake Event	Niagara Gorge Section of Event *
Early Algonquin to Kirkfield Algonquin	Lewiston Branch Gorge
Kirkfield Algonquin	Old Narrow Gorge
Port Huron Algonquin	Lower Great Gorge
Termination of Iroquois	Falls at Head of Niagara Glen
Nipissing ** (North Bay open)	Whirlpool Rapids Gorge
Post-Nipissing (similar to present drainage)	Upper Great Gorge

* See Fig. 1.

** This is the Early L. Nipissing-Stanley lake stage of Prest (1970).

Niagara River Gravels. As the Gorge was enlarged by recession of the Falls, the river a few hundred meters above respective cataract positions may have resembled its present aspect above Goat Island. The banks of this ancestral Niagara are terraced and outline a channel much broader than the Gorge. Remnant sands and gravels along this channel (Fig. 3) contain a well-preserved mollusk assemblage in part like that of the present (Letson in Grabau, 1901; Calkin and Brett, 1978). The main pattern of development of the Gorge involved minimal vertical cutting with general headward recession of one major cataract line from the Niagara Escarpment at Lewiston.

At the Whirlpool (Figs. 1 and 3b), radiocarbon dates of 9770 $\pm$ 150 yr B.P. were obtained on unionid pelecypod fragments and 9915 $\pm$ 165 yr B.P. on gastropod shells within the Niagara River terrace gravels along the Gorge walls 88 m above the present river. These suggest that these beds were abandoned by the river about 9800 yr B.P. (Calkin and Brett, 1978). The cataract itself may have been about 800 m downstream from the Whirlpool at this time. The overall average rate of recession to this point may have been on the order of 1.6 m yr^{-1} (if the cutting at Lewiston began about 12,400 yr B.P.).

Dates of 9080 ± 130 yr B.P. and 9115 ± 215 yr B.P. obtained on mullosk shells from the upper meter of a Niagara River gravel section capping Goat Island (Fig. 3a) may reflect abandonment of, or shoaling over, this surface resulting from the opening of the Upper Lakes' outlet at North Bay. The North Bay event would have caused a lowering of Lake Erie by 3 to 5 m (10 to 16 ft) and reduction of the Niagara River discharge by as much as 90 percent (Lewis, 1969). The dates above also provide a minimum age for a mastodon tooth recovered from the gravels near Prospect Point by James Hall.

The Lake Tonawanda Deposits. Deposition in the eastern parts of Lake Tonawanda must have ceased during reduction in Lake Erie levels. Wood, dated at 10,450 ± 400 yr B.P. near an outflow area in the southernmost part, overlies scoured clay and underlies sediments containing mastodon remains (Muller, 1977b). However, Lake Tonawanda was high enough in the eastern (Niagara) area to discharge through a spillway channel immediately east of the city to form Devils Hole plungepool (Fig. 1) after the main cataract had receded past the site, perhaps shortly before 9800 yr B.P. (Calkin and Brett, 1978). Furthermore, Lake Tonawanda waters apparently persisted here even until recent centuries (Calkin and Brett, 1978). The return of full discharge of the upper lakes to the Niagara River in Lake Nipissing time is recorded by a mollusk assemblage dated at 3780 ± 90 yr B.P. This overlies a scoured glacial lake clay at the Niagara Falls Sewage Treatment Plant (Frontispiece II; Calkin and Brett, 1978).

RATE AND MANNER OF ENLARGEMENT OF THE UPPER GREAT GORGE

The Upper Great Gorge is 3700 m in length and as suggested by Taylor (1933) appears to be the product of the full, post-Nipissing cutting under Great Lakes outflow similar to that of the present. This may be supported at least in part by comparison of its mean rate of cutting with that of the historic data. The mean recession rate must have been on the order of 0.8 or 1.0 m yr (3.2 ft yr^{-1}) for the historic period 1842 through 1905 (International Joint Commission, 1953, p. 14) before major man-made water diversions. However, a rate of 1.1 m yr^{-1} (3.6 ft yr^{-1}) has been determined for the total years of record (1670 through 1969, American Falls International Board, 1974; Fig. 3). Projection of these rates indicates that recession past the American Falls and separation of flow into the two channels may have occurred about 600 years ago.

Analysis of the historic rates of recession by the International Joint Commission (1953) suggested that the average recession of the Horseshoe Falls had decreased from 1.28 m yr^{-1} (4.2 ft yr^{-1}) between 1842 and 1906, to 3.2 ft yr^{-1} (0.98 m yr^{-1}) between 1906 and 1927, and to 2.2 ft yr^{-1} (0.67 m yr^{-1}) from 1927 to 1950. They initially attributed this to three main causes including: (1) the southerly dip of the Lockport Formation; (2) the southward thickening of the Lockport cap rock from 6 m (20 ft) at Lewiston to 23 m (80 ft) at the Falls; and

(3) diminishing discharge of the river as a result of increased diversion for hydroelectric power. However, more detailed analysis of historic data (Philbrick, 1970) and that from soundings in the Upper Great Gorge indicate that the rate of retreat has been much more variable than this suggests despite nearly uniform flows and continuous bedrock materials. Furthermore, Philbrick (1970, 1974) has argued that the planimetric configuration of the Falls may be as great or of greater control on recession than factors mentioned above. His ideas relative to retreat of the Horseshoe Falls are summarized below.

Philbrick maintains that there are three stages of the horizontal configuration of the Falls. The vertical cross wall is an unstable form which progresses to the horizontal arch, the most stable form (Fig. 4). Recession will be slow during the arch stage and deep plunge pools will be produced (Fig. 1). This stage in turn may be "broken down into" the notched stage (Fig. 4) when the highest stresses and greatest strains are generated. The notch stage then coincides with the highest rates of recession * and with shallow plunge pools. Model studies (Philbrick, 1970) show that stresses during the notch configuration may be three times those that occur during times when the crest is arched. Depths of the plunge pools are therefore inversely related to the rate of retreat.

Comparison of bottom surveys in the Upper Great Gorge between 1842 and 1966 (Fig. 4) confirms this hypothesis and shows that recession rates of the Horseshoe Falls during existence of a horizontal arched crest may reach 19 ft yr^{-1} (5.8 m yr^{-1}). The rate of present arch reaches about 2.5 ft yr^{-1} (0.76 m yr^{-1}) (Philbrick, 1974, p. 94). In addition, instead of having a uniform downstream slope as might be expected if the Horseshoe retreated at a unifrom rate, soundings in the Maid of the Mist Pool (in the Upper Great Gorge) display a series of "basins" or plunge pools separated by highs which are progressively lower in elevation in the upstream direction (Fig. 1). This profile reflects "an intermittent recession at the progressively slower rate" (1974, p. 94). Philbrick (1970, 1974) has correlated the basins with horizontal arches and long stands, the highs with horizontal notches, and relatively high stresses, and faster recession. The joint spacing on the Gorge walls at Prospect Point (Fig. 4) and Goat Island reflect indications on the bottom profile that suggest notch retreat with high stress oposite Prospect Point and broad arch retreat past the upstream side of the American Falls near Goat Island.

* Also recognized by Gilbert (1907) and Kindle and Taylor (1913).

ENGINEERING GEOLOGY OF THE AMERICAN AND HORSESHOE FALLS

Several engineering geology studies have been undertaken in the area of Niagara Falls on both sides of the border. Three of the more recent published reports concerning the American side include those of Dunn (1954), Acres American, Inc. (1972), and the American Falls International Board (1974). The second author directed the geotechnical aspects of the latter study; therefore, much of this section is taken from the 1974 report on the preservation and enhancement of the American Falls and the Terrapin Point flank of the Horseshoe Falls (SI values added for this paper).

Stratigraphy and Structure

The face of the American Falls (Fig. 6) and the Horseshoe Falls is formed in the Lockport and Rochester formations (Table 2) *. The Lockport Formation is mainly a dolomite. The Oak Orchard through the DeCew members are present at both Falls and appear to be fairly consistent although reef structures cause irregularities in the contacts between the Goat Island and Gasport members. The Rochester Formation, mainly a laminated to blocky dolomitic shale, supports the Lockport Dolomite. It has been divided into six zones. Interbedded limestones, dolomites, sandstones, and shale underlie the Rochester Shale. In general, the Irondequoit, Thorold, and Grimsby formations are resistant units at the American Falls and support talus.

Stratigraphic contacts at the Horseshoe Falls are about 3 m (10 ft) lower in elevation than at the American Falls because of the regional dip. For a generalized section of rocks in the vicinity of the Falls see Table 2.

The strata dip approximately 1/2° to the south and are relatively undeformed with only minor folding.

Joints are present in all formations. A preliminary survey (American Falls International Board, 1974) of the near regional area of the Falls indicated joints are oriented N77°E and N7°W as major joints. These were classified as shear joints. Significant horizontal stresses are present in the rock. There are a number of reported cases of load releases due to excavation at major construction projects in western New York (Rose, 1951; Feld, 1966; American Falls International Board, 1974). Movement has also been attributed to rebound from glacial loading and unloading.

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* See Brett (1982, this volume) for brief explanation of varying stratigraphic terminology used in the Niagara Gorge. The term "Formation" is used in the American Falls International Board Report (1974) where traditionally lithologic terms have often been used in standard geologic literature (e.g., Whirlpool Formation vs. Whirlpool Sandstone in the formal sense).

TABLE 2. GENERALIZED SECTION OF PALEOZOIC SEDIMENTARY ROCKS IN THE AMERICAN FALLS VICINITY. FROM AMERICAN FALLS INTERNATIONAL BOARD (1974, TABLE C2)

System	Series	Group, formation, member or zone			Thickness (ft.)	Lithology
Silurian	Niagaran	Lockport Group	Lockport Formation	Oak Orchard member	70±	Dolomite, medium-gray to medium dark-gray; thin- to thick-bedded, numerous irregular shale and stylolitic shale partings, slightly argillaceous; chert nodules and white dolomite crystals are common. The member is finely crystalline and sugary textured. Vugs commonly are filled with calcite, gypsum and sphalerite. Stromatolite domes are present. The rock is moderately hard.
				Eramosa member	14	Dolomite, medium-gray to grayish-brown; thin- to medium-bedded with numerous bituminous and carbonaceous stylolitic shale partings and stylolites. White porous chert and coarsely crystalline dolomite masses are common; gypsum, anhydrite, fluorite and sphalerite occur in lesser amounts. The member is very finely crystalline and sugary textured. The rock is moderately hard. Occasional vugs are filled with calcite and gypsum.
				Goat Island member	26	Dolomite, medium-gray in upper, dark-gray in middle and light-gray to light tannish-gray in the lower part, occasionally mottled; massive in upper and lower part and thin- to medium-bedded in the middle, slightly to very argillaceous with abundant shale and stylolitic shale partings in the middle. The member is finely to medium crystalline and sugary textured. The lower part is coarsely crystalline, pitted and vuggy; the vugs are occasionally filled with secondary dolomite crystals. The rock is moderately hard.
				Gasport member	18	Dolomite to dolomite-limestone, light- to medium-gray; massive with abundant stylolites and discontinuous shale partings throughout; slightly argillaceous in the upper and conglomeratic with pebbles of DeCew lithology in the base. The member is finely to coarsely crystalline and sugary textured. The rock is moderately hard. Vugs and pits are filled occasionally with gypsum.
				DeCew member	10	Dolomite, medium- to dark-gray; thin- to medium-bedded with an occasional thick bed; argillaceous with wavy irregular shale partings that contain well developed slickensides; stylolites and stylolitic shale partings are common; occasionally masses and nodules of gypsum occur. The member is finely crystalline to crystalline with a well-cemented mosaic texture. The rock is moderately hard. Outcrops contain a "flow" or "enterolithic" structure.
		Clinton Group	Rochester Formation	zone 1	1-2	Shale, medium dark-gray; laminated to platy, slightly dolomitic, dense and moderately hard. This zone is a transition from the Rochester below to the DeCew above.
				zone 2	13±	Shale, medium dark-gray to dark-gray; laminated to blocky and contains discontinuous partings and bands of light-gray dolomitic limestone. Clay minerals are illite, chlorite, kaolinite and traces of montmorillinite; the zone also contains scattered pyrite and gypsum masses near the base. Microcrystalline dolomite is interspersed with finely crystalline illitic clay and quartz. The rock is moderately hard.
				zone 3	6±	Dolomitic shale and shaly dolomite, medium dark-gray to dark-gray; thin-bedded in upper and basal, massive in the middle; shale partings occur in the upper and lower parts. Illitic clay, traces of chlorite, kaolinite and silt size quartz grains are dispersed throughout the zone. The rock is dense, contains very few pores, and is moderately hard.
				zone 4	25-29	Shale, medium- to dark-gray; laminated to blocky and contains numerous discontinuous partings and bands of light-gray dolomitic limestone; gypsum partings are abundant. Nodules of calcite and gypsum occur occasionally. The clay minerals are illite, chlorite, kaolinite and mixed layered clay. The zone has a dense, microcrystalline texture. Portions of the zone are fairly well cemented with dolomite; other parts are more shaly and rapidly fracture upon drying. The rock is moderately hard.
				zone 5	6-10	Shale, medium- to dark-gray; laminated, dolomitic and contains occasional gypsum partings, white calcite nodules, numerous discontinuous partings and bands of limestone and thick beds of shaly dolomite. Silt size quartz grains are scattered throughout. Clay minerals are illite, chlorite, kaolinite and mixed layered clay. The texture of the zone is microcrystalline. The rock is moderately hard.
				zone 6	4-5	Shale, medium- to dark-gray; laminated to blocky; contains light-gray laminae and bands of calcite and dolomite. The blocky shale contains gypsum partings. Quartz grains are common; pyrite and marcasite occur with carbonaceous matter as a replacement of organic matter. The clay minerals are illite, chlorite and kaolinite. The rock is moderately hard.
			Irondequoit Formation	Unnamed member	6-9	Limestone, light-gray with pinkish tint, medium-bedded to massive with frequent wavy irregular green or black shale partings near the top. The member is coarsely crystalline, The rock is moderately hard. A few vugs and small pores are present.
				Rockway member	10-11	Dolomite, varies from light- to medium-gray in the upper part to brown in the middle and brownish-gray at the base; thin- to thick-bedded in the upper, massive in the middle and thin-bedded at the base; argillaceous to very argillaceous at the base. Dark-gray shale partings and a few gypsum nodules occur throughout the member. The member varies from dense to finely crystalline and has a sugary texture. The rock is moderately hard.
			Reynales Formation	Hickory Corners member	2-3	Unconformity Limestone, light- to medium-gray; argillaceous, calcitic and highly siliceous; numerous wavy, dark-gray shale partings and bands produce a pseudonodular appearance. The texture of the member is very finely crystalline to dense. The rock is moderately hard.
			Neahga Formation		6	Shale, dark greenish-gray; platy to fissile with a waxy appearance; shaly sandstone at base. Masses of pyrite and gypsum partings occur along the bedding planes; calcite and dolomite occur in small amounts and quartz is the most abundant non-clay mineral. Illite is the dominant clay mineral with lesser amounts of chlorite, kaolinite and mixed layered clay. The rock is soft and flakes readily during wet-dry cycles. Slickensides are present.
		Medina Group	Thorold Formation		9	Sandstone, light-gray to greenish-gray; medium-bedded to massive; irregular green shale partings occur throughout. The sandstone is orthoquartzitic. The texture of the formation is very fine grained. Silt size to fine grained quartz particles are cemented with secondary silica. The rock is hard.
			Grimsby Formation	zone b	8	Sandstone, pink to reddish-brown; thin- to thick-bedded, hematitic, calcareous. The texture varies from fine to medium grain. The rock is moderately hard to hard. A weathered zone frequently occurs at the top of the formation.
				zone a	43	Siltstone and sandstone with interbeds of shale, variegated from red to pale green; pink, white or mottled siltstone or sandstone with red shale and red sandstone interbeds. Gypsum partings occur in shale beds. The sandstone is fine- to medium-grained and well-cemented. The siltstone and shale vary from soft to moderately hard.
			Power Glen Formation		34	Shale with siltstone beds and stringers of silty limestone and dolomite; dark-gray to grayish green shale and siltstone, and light-gray limestone and dolomite; laminated to banded. Quartz is the most abundant non-clay mineral. Clay minerals consist of illite, chlorite and small amounts of montmorillonite and mixed layered clay. The rock is slightly soft to moderately hard.
			Whirlpool Formation		18	Sandstone, light-gray to white; medium-bedded and cross-bedded; fine- to medium grained. The quartz grains are frosted and well rounded, and are well cemented by secondary silica. Feldspar grains altered to kaolinite are abundant. Occasional green shale inclusions and chloritic shale partings occur throughout. The rock is slightly soft to moderately hard.
Ordovician	Cincinnatian	Richmond Group	Queenston Formation		100±	Unconformity Shale (technically classified as a claystone) reddish-brown (ferric) shale with interbeds and nodules of green (ferrous) shale; massive to blocky. The shale is silty and is cemented by dolomite and calcite. Scattered gypsum nodules occur throughout; quartz is a common constituent. Clay minerals are illite, chlorite, kaolinite, montmorillonite and mixed layered clay. The shale is highly compacted and moderately hard. Numerous small, high angle slickensides are stained with iron oxide.
			Total		429±	

Note:

The stratigraphy is compiled from Zenger (1965 and 1966), Fisner (1966) and Kilgour (1966), from stratigraphic studies by U.S. Army Engineer District, Buffalo New York and from petrographic descriptions by U.S. Army Engineer Division, Missouri River. The Rochester Formation zonation was developed for this study by the Buffalo District.

In the Lockport Dolomite of the American Falls the major set of high-angle joints (primary shear set S_1) trends N70°E to N80°E and penetrates into the Goat Island Member. The S_1 joints are generally straight, spaced 3 to 24.4 m (10 to 80 ft) apart, and have traceable lengths up to 39.6 m (130 ft). A variable, complimentary shear set (S_2) has an approximate range of N to N30°E. Bisecting the angle between these sets results in a principal stress direction of about N45°E. This is in general agreement with the principal high horizontal stress shown by overcoring (N34°E ± 30° and N54°E ± 12°). There is a minor concentration of joints between N40°E to N50°E which is parallel to the maximum stress direction; this set has been classified as tension set (extension) T_1. There is a weak T_2 set (release) perpendicular to the maximum stress direction oriented N40°W to N60°W. In addition to high-angle joints, several bedding plane joints were found that are generally open across the face. For the general location of joints in the Lockport Dolomite on the dewatered riverbed see Figure 5.

The joints in the Rochester Shale at the American Falls vary from N10°W to N80°E, generally dip toward the Gorge (approximately northwest), and have spacings from 0.06 m (0.2 ft) to several tens of ft. The joints on the face have variable vertical penetration and quite often terminate or become offset at the more massive dolomitic Zone 3. However, some joints extend through the entire Rochester Shale. Joint openings vary from closed to open more than .3 m (1 ft). The exposed joint planes are usually curvilinear. The frequency of joints decreases with distance from the face (Fig. 6). In some areas joints are present at least 34.1 m (112 ft) back from the face. The joints generally parallel the trends of the American Falls face and are classified as stress release joints originating from tensile stresses directed into the Gorge. Jointing in underlying formations at the American Falls is known from borings and a few outcrops; however, there are not enough data to determine joint sets.

The prominent joints in the Rochester Shale at the Horeshoe Falls trend N50°E to N60°E, N70°E to N80°E, and N40°W to N50°W and probably reflect the change in direction of recession of the Horseshoe Falls. Frequency decreases with distance from the face. The joints appear to be stress release joints and in some areas are present 27.4 m (90 ft) back of the cliff face.

Two thrust faults were found on the face of the American Falls and are located entirely within the Rochester Shale. One fault is near the flank of the 1931 rockfall area and the other is located just downstream of Luna Island. Both faults have a very small vertical displacement.

Engineering Properties

Consideration of repair for the preservation of the American Falls required a thorough study of the physical strength characteristics of the rocks that make up the Falls escarpment. Testing of the rocks was performed by the Missouri River Division Laboratory (1971) to establish criterion upon which the repair could be based. The conclusions from the report are summarized below.

The five members of the Lockport Formation are dense, hard, low absorptive rock with a unit weight of 26.2 kN/m^3 to 27.2 kN/m^3 (167 to 173 pcf). They are highly durable and, as shown by the following test results, have adequate strength for most engineering consruction: tensile strength 2,756 to 9,646 KPa (400 to 1,400 psi); unconfined compressive strength 117,130 to 192,920 KPa (17,000 to 28,000 psi); Tangent Young's modulus 41.3 to 75.8 X 10^6 KPa (6 to 11 X 10^6 psi); and Poisson's Ratio 0.25 to 0.37. Angle of internal friction ranges from 59 to 68 degrees and cohesion intercept values vary from 9,646 to 17,914 KPa (1,400 to 2,600 psi). Frictional resistance of smooth sawed joints averages 0.60 and is 0.73 for the moderately rough natural joints. Cohension intercept values of 344.5 to 2,756 KPa (50 to 400 psi) were obtained for the natural joints; the value was largely dependent upon the surface roughness.

The Rochester Formation, comprised of six zones, is quite variable in composition and structure. Carbonate content, mostly as dolomite ranges from 34 to 67 percent by weight of the rock. High unit weights of 26.5 kN/m^3 to 27.1 kN/m^3 (168.5 to 172.6 pcf) and low absorption, 1.5 to 2.3 percent, indicate a highly compacted, well-cemented shale. All of the zones are very susceptible to deterioration in freezing-and-thawing. Additionally, all but Zone 3, a shaly dolomite, also break down almost completely in the wetting-and-drying test.

The Rochester Shale is very weak in tension normal to the bedding (34.5 to 1,309.1 KPa) (5 to 190 psi); however, it is 10 to 20 times as strong in a direction parallel to it. Unconfined compressive strength is variable ranging from a low of 19,774.3 KPa (2,870 psi) to a high of 88,329.8 KPa (12,820 psi). Average strength for the more shaly beds is about 31,349.5 KPa (4,550 psi) with Zone 6 being the weakest at about 2/3 of this value. The shale has a tangent Young's Modulus of 2.1 to 25.5 X 10^6 KPa (0.3 to 3.7 X 10^6 psi) and a Poisson's Ratio of 0.30 to 0.50 for most of the zones. In the range of 385.8 to 1,378 KPa (56 to 200 psi) confining pressure, the Mohr strength envelopes of the shale zones have slopes of 60 to 73 degrees and cohesion intercepts of 344.5 to 3,927.3 KPa (50 to 570 psi). The smooth sawed joints of the shale have greater frictional resistance, averaging about 0.75, than the moderately rough natural joints, 0.49. Values of cohesion intercept range from 172.3 to 2,411.5 KPa (25 to 350 psi) for the natural joints. Direct sheer tests on bedding planes gave an average sliding friction angle of 26 degrees. Illite and chlorite with a small amount of mixed layered clay and kaolinite are the principle clay minerals present in all of the shale zones.

Both the limestone and sandstone formations of the underlying beds have similar strength characteristics as the Lockport Dolomite. The Neahga Shale is the weakest shale with an unconfined compressive strength of 5,167.5 KPa (750 psi). It is a highly fissile, illitic rock with a frictional resistance of about 0.31 to 0.60 derived by direct shear on bedding planes. These friction values reduce to 0.09 at 1,378 KPa (200 psi) normal stress indicating a low shear strength material.

Contemporary Processes

Ground Water. Open fractures in the Lockport Dolomite provide lateral and vertical seepage paths to the face and to underlying formations. There is a continuous recharge from the river to the permeable zones. The Lockport Dolomite generally has low permeability except in areas where structural defects occur. Several boreholes accepted 18.9 to 53 cubic decimeters/min (5 to 14 gal/min) at member contacts or at partially detached (translated) zones near the face. Seepage occurs in the cliff walls adjacent to the Falls; the seepage occurs along horizontal open bedding planes intersected by major joints in the Lockport Dolomite. Piezometer studies indicated that seasonal fluctuations generally range from 0.3 to 1.8 m (1 to 6 ft), piezometric levels are influenced by the level of flow in the river, and winter blockages of drainage are possible.

Pressure tests and dye tracer tests indicated that the Rochester Shale generally is tight except near the face where there are concentrations of high-angle stress release joints. Very small quantities of water were observed seeping along the Lockport-Rochester contact at Prospect Point.

Pressure tests indicated that underlying units contain some zones with moderate permeability.

Weathering and Erosion. The rock at the Falls weathers and erodes because of effects from flowing and falling water, from seasonal ice buildup, from wetting and drying (in some areas), and from stresses within the rock. The Lockport Dolomite is more resistant to erosion and weathering than the Rochester Shale. Solution action of water was evident from cores in which there were weathered, stained, widened, and sometimes mineralized (gypsum and calcite) fractures and joints. The Rochester Shale, except for Zone 3, cannot stand repeated wet-dry cycles. After 10 cycles the losses during wet-dry testing ranged from 85 to 100 percent (except for Zone 3). Ice-jacking and freeze-thawing are significant processes which contribute to erosion. Accumulations of ice have varied from local accumulations to complete freezing of the American Falls. Freeze-thaw tests showed losses as a result of fragmentation ranging from 0.5 percent to 20 percent for the Lockport Dolomite and from 32 percent to 75 percent in the Rochester Shale.

Undermining. Undermining is the result of progressive surface deterioration from various modes of weathering, collapse of weakly supported masses, and the buckling of vertically jointed masses of rock from loading.

At the American Falls, undermining is most pronounced in the 1931 rockfall area (Fig. 7). There, a segment of the face, about 36.6 m (120 ft) wide, is recessed a maximum of 9.1 m (30 ft). In profile, undermining actually begins in the lower part of the Goat Island Member, but is deepest in the Rochester Zone 2.

At the Horseshoe Falls (Terrapin Point) the Lockport Dolomite generally is undermined; the maximum undermining which is commonly found, is about 4.6 m (15 ft).

Talus Accumulation. Rockfalls have resulted in huge piles of talus at the American Falls. The volume of talus is estimated to be 214,200 m^3 (280,000 yd^3). The resistant Irondequoit, Thorold, and Grimsby formations form ledges that underlie the talus accumulation at the American Falls. Most of the talus consists of blocks of Lockport Dolomite. The sizes of 130 of the largest accessible blocks varied from slightly less than 90,718.5 kg (100 tons) to greater than 2,177,243.3 kg (2,400 tons). Boulders about 0.9 m (3 ft) or smaller in diameter can be moved and reduced by the tumbling action (abrasion) of falling water at the American Falls.

The talus beneath the Horseshoe Falls at the central apex area is ground up and dispersed by the flow; there is no visible accumulation comparable to that at the American Falls. The talus mound from the 1934 rockfall adjacent to Terrapin Point is the only significant accumulation.

Rockfalls, Failure Mechanisms, Stability. The earliest reference to rockfalls found is a passage from Lyell (1845) quoted in Dow (1921) which reported the sudden descent of huge rock fragments of undermined limestone at the Horseshoe Falls in 1828 and another at the American Falls in 1818. At the American Falls other rockfalls of significant size occurred in 1931, July 1954, and December 1954 (Fig. 7). Rockfalls of less significant size occurred in 1907, 1920, 1967, and 1974. At the Horseshoe Falls other rockfalls of significant size occurred in 1823, 1846, 1850, 1852, 1882, 1889, 1905, 1934, 1936, 1937, 1963 and 1981. Numerous intermittent and sometimes unnoticed rockfalls of generally lesser significance have also occurred at both Falls and flank areas.

Since the separation of the American and Horseshoe Falls about 600 yr ago, the American Falls has retreated on the order of 61 m (200 ft), while the Horeshoe Falls has retreated about 762 m (2,500 ft). The results of surveys since 1842 show that the American Falls has undergone drastic changes attributable to massive rockfalls beginning in

1931. The recession has been sporadic. The general rate of recession of the American Falls in 600 years is about 0.09 m yr^{-1} (0.3 ft yr^{-1}). This is considerably slower than the 3.6 ft yr^{-1} (1.1 m yr^{-1}) rate of recession obtained for the Horseshoe Falls between 1678 and 1969 mentioned in a previous section of this paper (see Fig. 4).

A computerized two-dimensional analysis of the Niagara Falls area using the finite element method was performed (Fairhurst, 1969) in order to determine the influence of the direction of the known regional stress field in western New York on the stability of the Falls. The analysis indicated that a lateral compressive stress parallel to the Niagara Gorge tends to develop a concentrated compression behind the arcuate crestline of the Horseshoe Falls. Those compressive stresses tend to close the joints in the Lockport Dolomite stabilizing the center portion and thereby preventing the fall of large rock fragments. However, in the region of the American Falls, the analysis indicated that at promontories in the crestline, such as at Prospect Point, there would be a tendency for tensile stresses to develop somewhat behind the crest acting to open vertical fractures parallel to it. Compression parallel to the Gorge would thus tend to contribute to failure of rock-mass in promontories along the Gorge, particularly where water under pressure could enter the fractures.

At the American Falls measurements of the horizontal stresses were made in two shallow vertical holes located 54.9 and 176.8 m (180 and 580 ft) back from the crestline (Fig. 5). Results indicted that significant horizontal stresses exist in the Lockport Dolomite and act in a generally NE-SW direction (parallel to the Gorge). At the 176.8 m (580 ft) location, the horizontal stresses had an average magnitude of +6,890 KPa (+1,000 psi) (compression) and -68.9 KPa (-10 psi) (tensile) while at the 54.9 m (180 ft) location the stresses had an average magnitude of +5,994.3 KPa (+870 psi) (compression) and -2,273.7 KPa (-330 psi) (tensile). The high tensile stress near the Gorge wall is particularly significant since the average tensile strength of the Rochester Shale parallel to the bedding is only 1,446.9 KPa (210 psi). These high tensile stresses probably extend down into the underlying Rochester Shale and explain the development of stress release joints in the shale.

There are at least two modes of failure at the American Falls (Fig. 8). The classical mode of failure occurred in January 1931, when erosion of the Rochester Shale undermined and caused failure of the overlying Lockport Dolomite along existing high-angle joint planes. The jointed cap rock could no longer support its own weight and failed.

Another mode of failure occurred in July 1954. That failure occurred at the wetwall-drywall contact at Prospect Point. Prior to the rockfall, photographs indicate that there was some undermining of the cap rock. Since the landward depth of the failed mess 39.6 m (130 ft) far exceeded the depth of undermining, the mode of failure was something other than ordinary gravity collapse. Close evaluation of photography taken during the rockfall indicate a slight down-dropping of the top of rock prior to its toppling. The down-dropping probably was the result of shearing within the Rochester Shale. The main features contributing to this rockfall appear to have been: (1) the open joints in the Lockport cap rock and possibly in the Rochester Shale, which created the weak zone for the back limit of the rockfall; (2) the unbuttressed cliff face; (3) the failure plane at the bottom of the block (assumed to be the Rochester Shale); and (4) the hydrostatic pressure.

Large rockfalls are likely to occur in the future. One of the most significant areas with questionable stability at the American Falls is Indian Head Point (Fig. 7). It is a detached block of rock about 53.3 m (175 ft) long by 30.5 m (100 ft) wide with open vertical fractures extending into the shale below. The block is considered to be just barely stable. If it collapsed, it could drag down large amounts of adjacent rock. At Terrapin Point, a significant mass of rock adjacent to the river flow and in front of the viewing area is failing. The structural conditions were identified in 1972 and the outer portion of the Terrapin Point viewing area was fenced off by Park authorities. An earthquake induced failure here could involve a significant portion of Terrapin Point.

Preservation, Enhancement, and Remedial Work

The most recent comprehensive study of the Falls (American Falls International Board, 1974) considered the measures necessary to preserve or enhance the beauty of the American Falls at Niagara as well as public safety at the flanks of the American Falls and at the Terrapin Point flank adjacent to the Horseshoe Falls.

The Niagara Frontier State Park and Recreation Commission completed a contract recently for instrumentation and drainage of the viewing area at Terrapin Point and also completed a study of possible rock removal and restoration of the viewing area at Terrapin Point. The Commission removed overhangs at Luna Island in 1955 and performed additional remedial work there (installation of rock bolts, tendons, and drain holes) in 1972.

OTHER MAJOR ENGINEERING PROJECTS ALONG U.S. SIDE OF THE NIAGARA GORGE

The Niagara River has been used for power purposes for over 200 years. Hydroelectric power was produced at Niagara Falls as early as 1880. The first large scale output of commercial power at Niagara Falls began in 1895 (Adams Station). Around the turn of the century advances in electric power development made practical the construction of power plants at the bottom of the Gorge (Schoellkopf Stations). The Robert Moses Niagara Power Plant was completed in 1963 and at the time was one of the world's largest with an installed capacity of 1,950,000 kilowatts.

Schoellkopf Power Plant Failure

On June 7, 1956, a massive rockfall destroyed Schoellkopf Powerhouse Stations 3B and 3C and damaged 3A near the current site of the Schoellkopf Geological Museum. The powerhouse stations had been excavated in the Grimsby Sandstone at the base of the Gorge wall. The rockfall was progressive; it started at the south end and worked northward. Prior to the rockfall, large leaks of water emerged from the cliff face. Apparently no report which included an analysis of the failure was ever published. From the limited information available, it appears that removal of rock at the base of the cliff, possible weak shale zones at the base of the failure, open joints and fractures (some blasting induced?) back of the cliff face which permitted high joint water pressures to develop, and a source of recharge of water (an unlined canal) were the major reasons for the failure. The triggering cause for the failure may have been the blockage of drainage as a result of grouting programs performed just prior to the failure. The grouting was intended to reduce leakages from penstocks.

Robert Moses Niagara Power Plant

This project consists of an intake section, two parallel conduits about 6.4 km (4 mi) long, an open canal about 1,219.2 m (4,000 ft) long, a main generating plant, and a pump-generating plant (Power Authority of the State of New York, 1965).

Large quantities of rock excavation were required for construction of the intake, conduits, canal, and plants. At the start of the project different methods of establishing the breakline in rock were tried: line drilling on close centers, line drilling with light explosive charges placed at intervals in every second or third hole, slashing, and modified cushion blasting. However, none of those methods resulted in suitably clean walls for the placement of concrete. After considerable experimentation, presplitting (the establishment of a free surface or shear plane in the solid by the controlled usage of explosives and blasting accessories in appropriately aligned and spaced drill holes) was developed and superior results were obtained (Paine, Holmes, and Clark, 1961). Apparently this was among the first large scale application of presplitting.

More than 5,355,000 m^3 (7,000,000 yd^3) of excavation, mostly rock from the Lockport through the Queenston formations were removed to accommodate the massive power plant (Fig. 9). Another ~1,912,500 m^3 (2,500,000 yd^3) were removed for the forebay.

Feld (1966) reported that rock movement from load release occurred during the excavation for the two parallel conduits. The trenches in rock were up to 50.3 m (165 ft) in depth. According to Feld with the excavation to subgrade, the sides at 11.6 m (38 ft) above subgrade moved inward 1.3 cm (1/2 in) and the subgrade developed a longitudinal crack and about a 7.6 cm (3 in) heave at the center of the trench. One month later the heave had increased to 21.6 cm (8-1/2 in) at the center and was 6.4 cm (2-1/2 in) at each side wall. Those movements were accompanied by loosening of the exposed rock sheets at subgrade and with ravelling of the rock at the side walls. With the completion of the concrete conduits, backfilling, and filling the conduits, apparently normal stable conditions were restored. The concrete conduits were designed with articulated arch roofs and longitudinal center joint in the floor to provide flexibility should internal rock movements occur.

FUTURE OF THE FALLS

In his "Principles of Geology," Lyell (1860, p. 181) suggested that on the basis of contemporary rates of recession, the Falls would eventually reach Lake Erie in $\sim$30,000 yr. Projections of the twentieth century rate of $\sim$2 ft yr^{-1} (0.6 m yr^{-1}) via the shortest (west or Chippewa) channel (Frontispiece II and Fig. 10) yields 48,000 yr for the 29 km (18 mi) distance. Such projections are based on fallacious assumptions but the results may be as good as any number that could be generated considering the complicated scenario of retreat (Philbrick, 1974). Philbrick (1974) suggests the following sequence of events during progressive recession of the Falls and Gorge to Lake Erie (see Fig. 10):

(1) Retreat at 2 ft yr^{-1} (0.6 m yr^{-1}) causes capture of the American channel above Goat Island about 2000 yr AP and within 7000 yr AP, a lowering of the Chippewa-Grass Island Pool above the Falls with consequent slight lowering of Lake Erie levels.

(2) Splitting of the Horseshoe Falls into first, two falls, then three, and subsequently two falls again as recession causes the channel to split at Navy Island and again around the north end of Grand Island.

(3) Headward migration of Chippewa Horseshoe Falls down south dip of the Lockport Formation to a height of 50 ft (15 m) at which point recession is so retarded as to make it a quasi-stationary waterfall (similar to present American Falls).

(4) Rapid erosion to form a broad, gentle gorge and stepped rapids in the Salina Group rocks. Eventual capture of slower-eroding Tonawanda (east) Channel by Chippewa Channel near upstream end of Grand Island.

(5) Marked slowing of recession to Lake Erie and Niagara River mouth as gorge develops in Bertie dolomite and single Falls on overlying tough Onondaga Limestone.

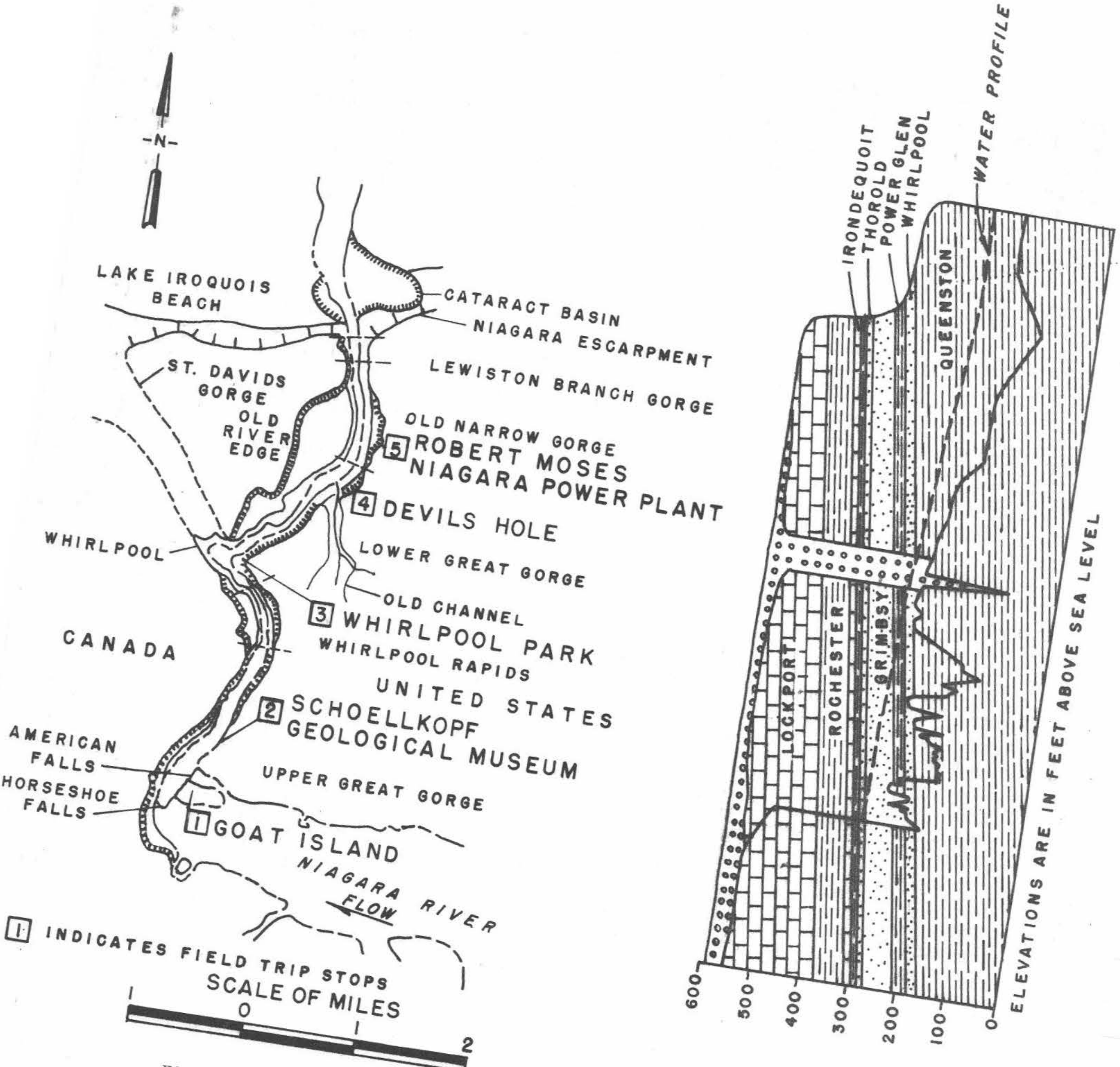


Figure 1. Location map of the Niagara Gorge showing field trip stops, Gorge sections (see Table 1), and profile modified from American Falls International Board (1974).

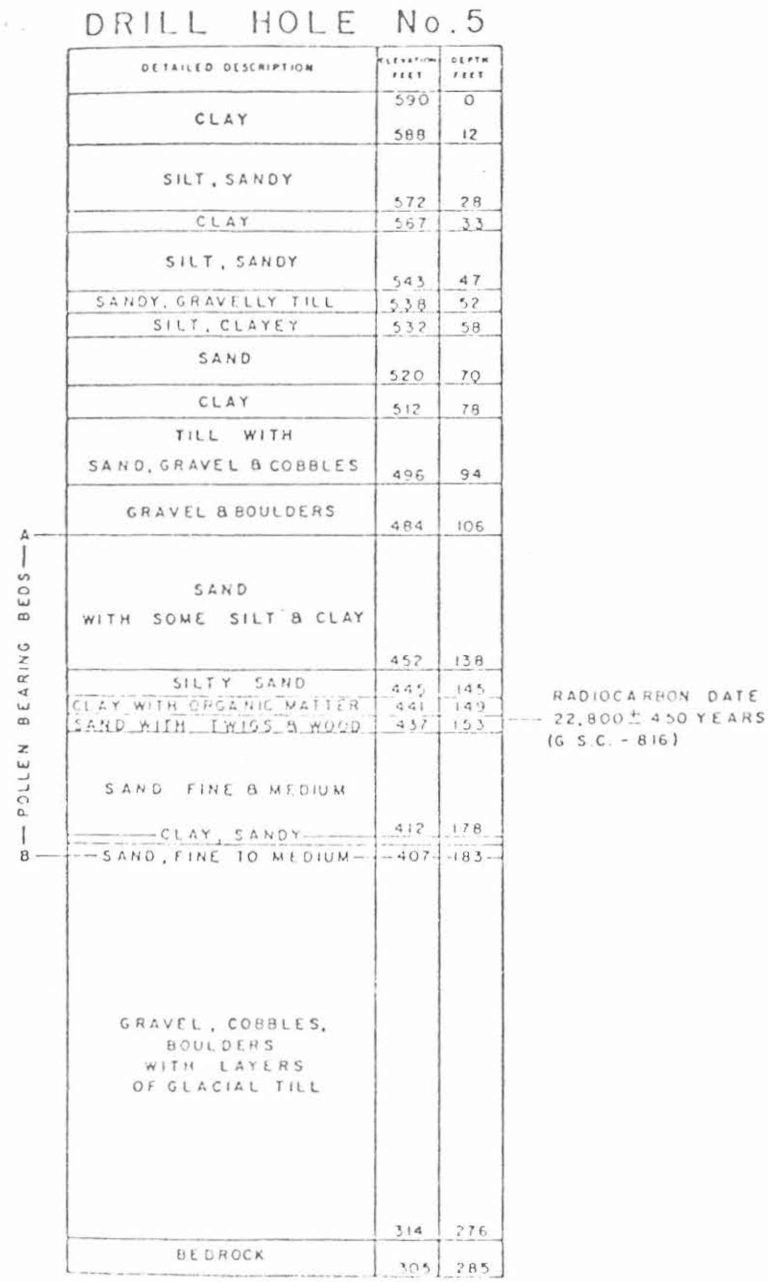
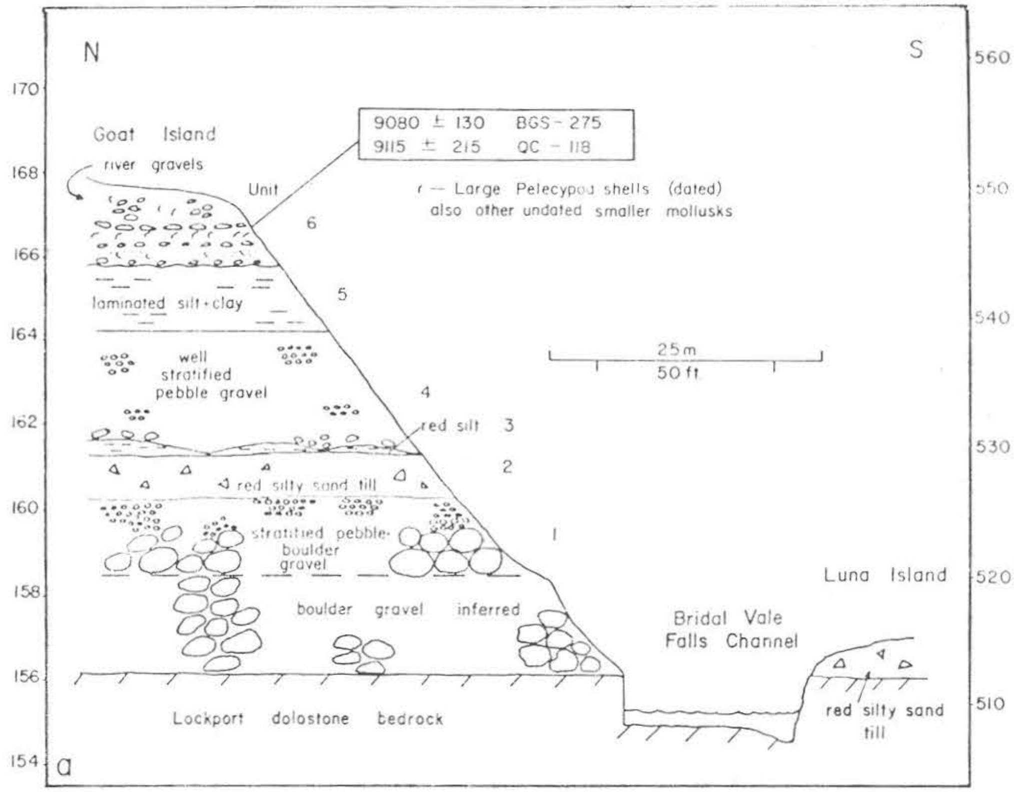


Figure 2. Generalized drill log no. 5 from the upper end of St. Davids Gorge. The radiocarbon date may provide a maximum age for advance of Late Wisconsin ice across the area. The organic matter is assumed to be Middle Wisconsin based on this date and the pollen assemblage; the lower tills may be Middle and/or Early Wisconsin (or older).



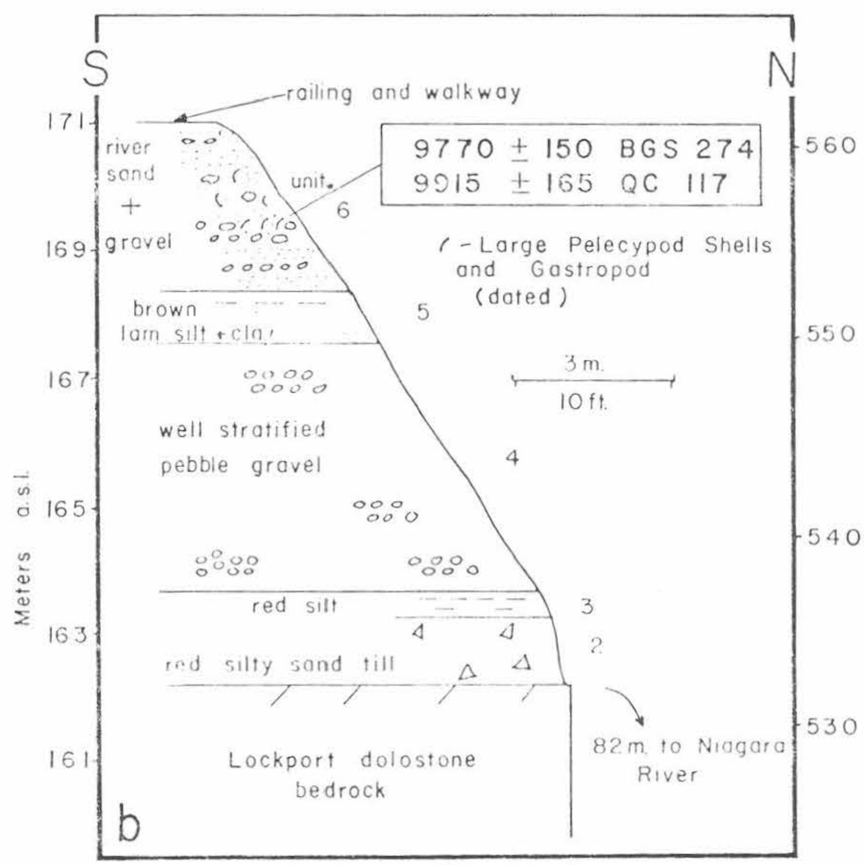


Figure 3. Stratigraphic sections through the surficial deposits at Goat Island (3a) and Whirlpool Park (3b). From Calkin and Brett (1978).

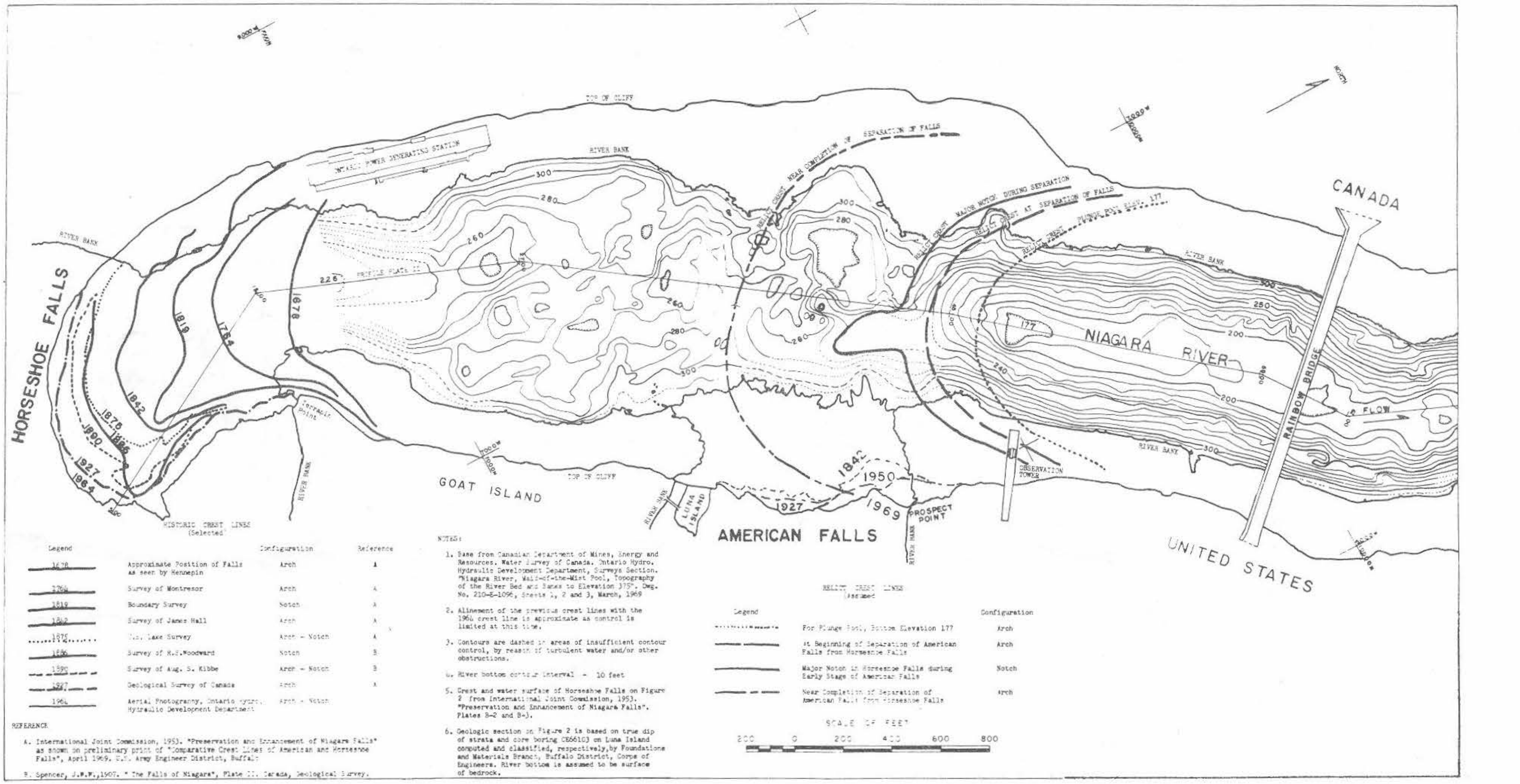


Figure 4. Former crestlines of the American and Horseshoe Falls. Modified after Philbrick (1970, Figure 1); former crestlines at the American Falls added from American Falls International Board (1974, Plate C29).

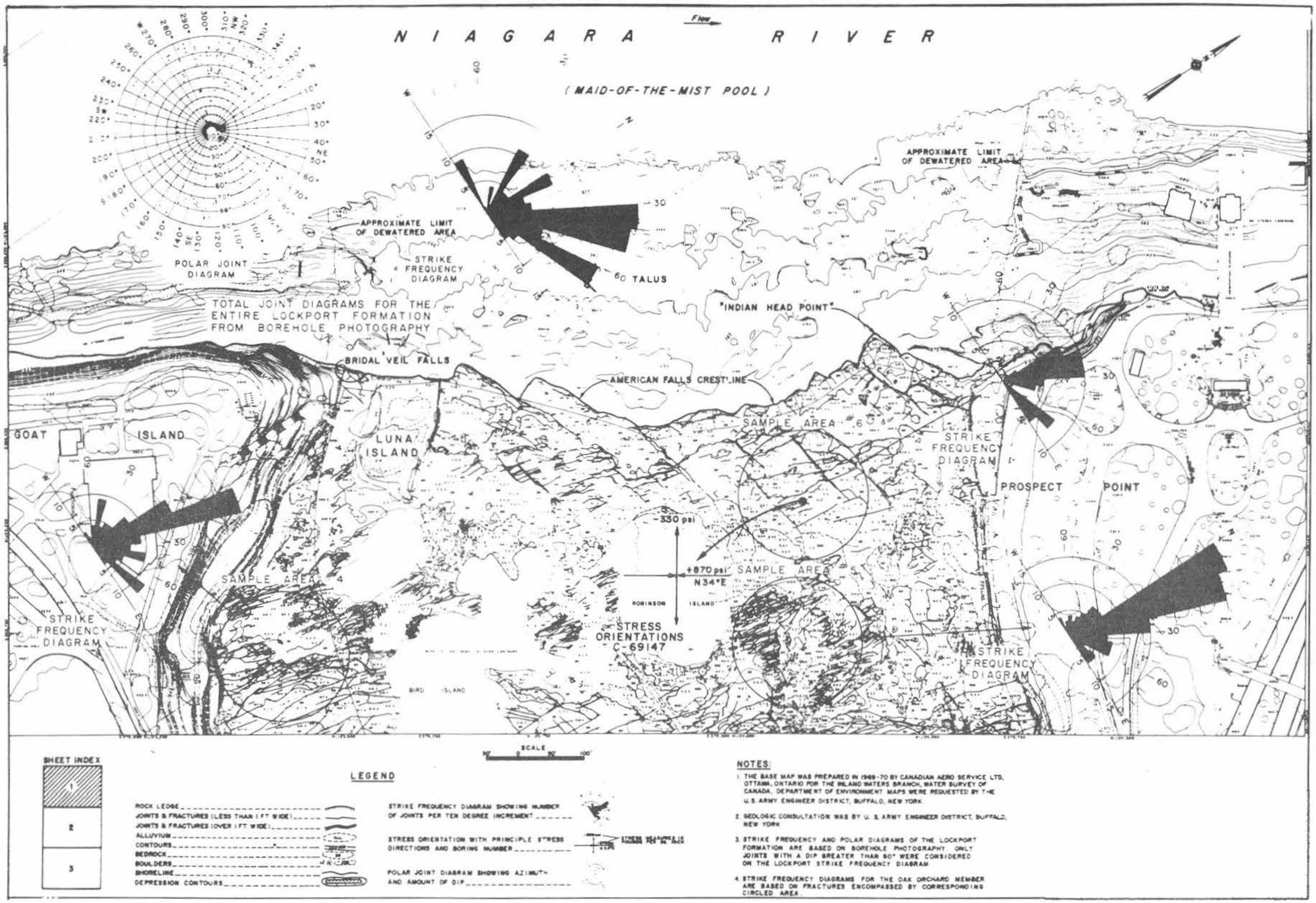


Figure 5. Joints in Lockport Formation of dewatered river bed. From American Falls International Board (1974, Plate C11).

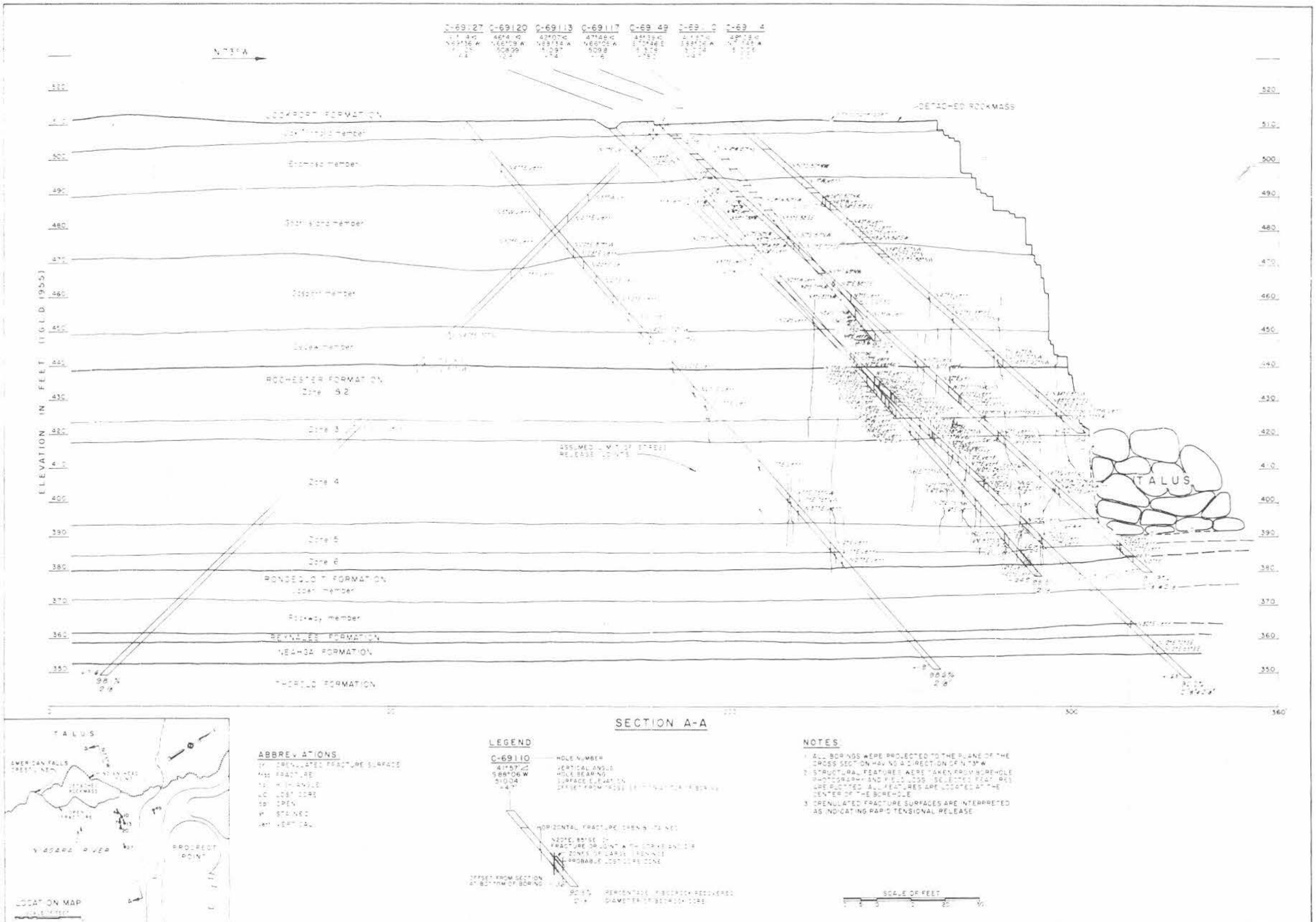


Figure 6. Geologic cross section, Prospect Point flank. From American Falls International Board (1974, Plate C16).

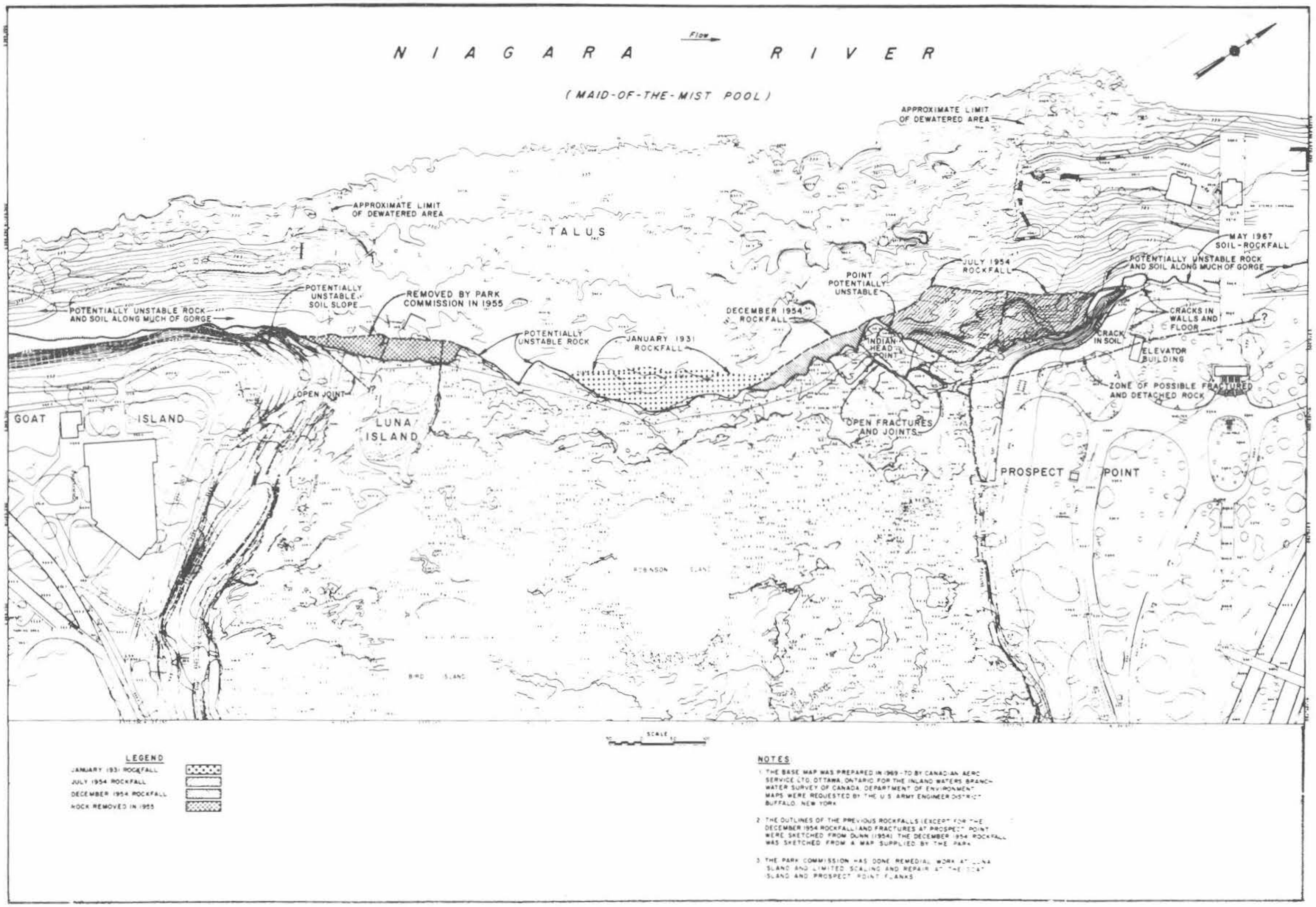


Figure 7. Rockfalls and unstable areas along American Falls crest and flanks. From American Falls International Board (1974, Plate C30).

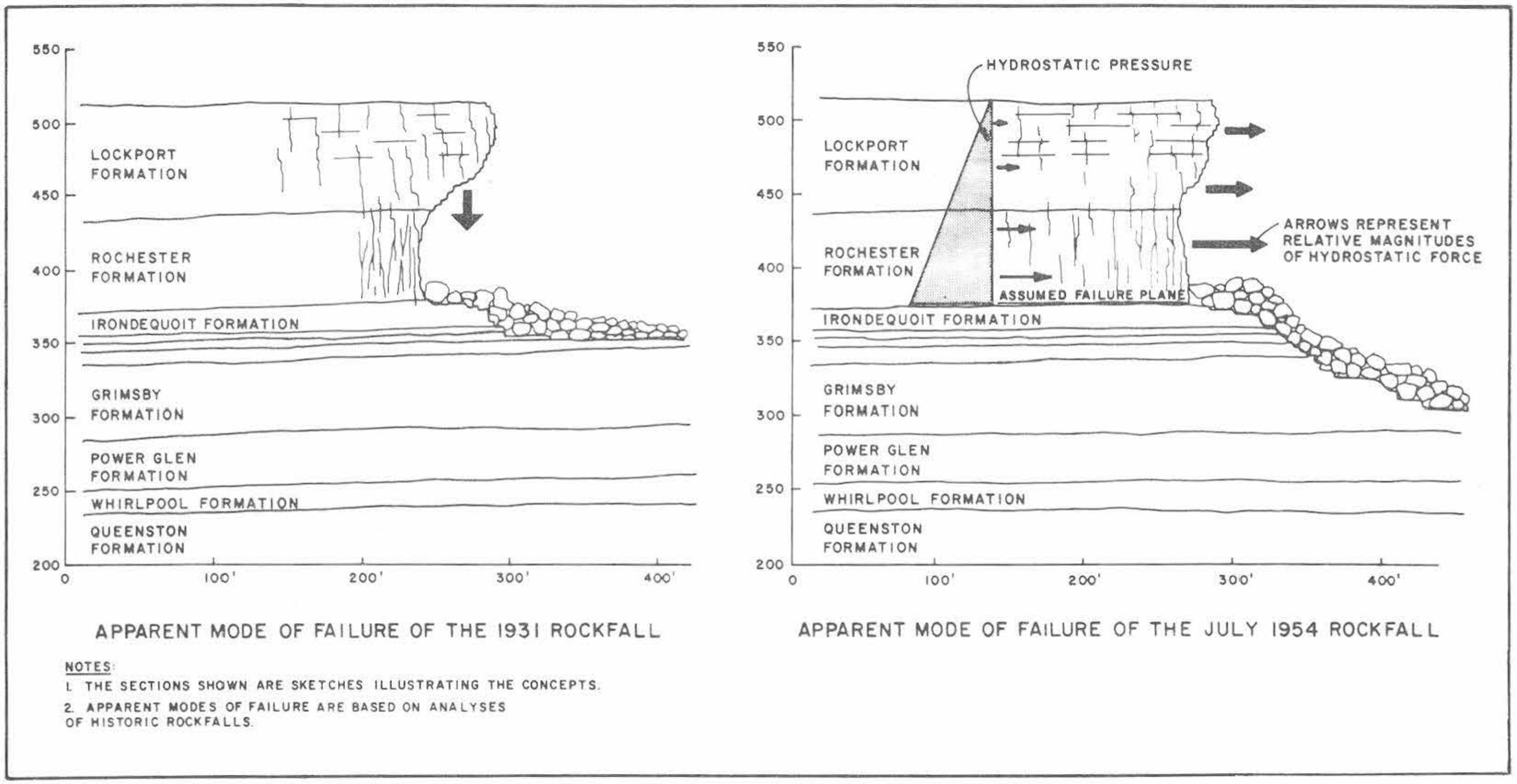


Figure 8. Modes of failure, American Falls. From American Falls International Board (1974, Plate C45).

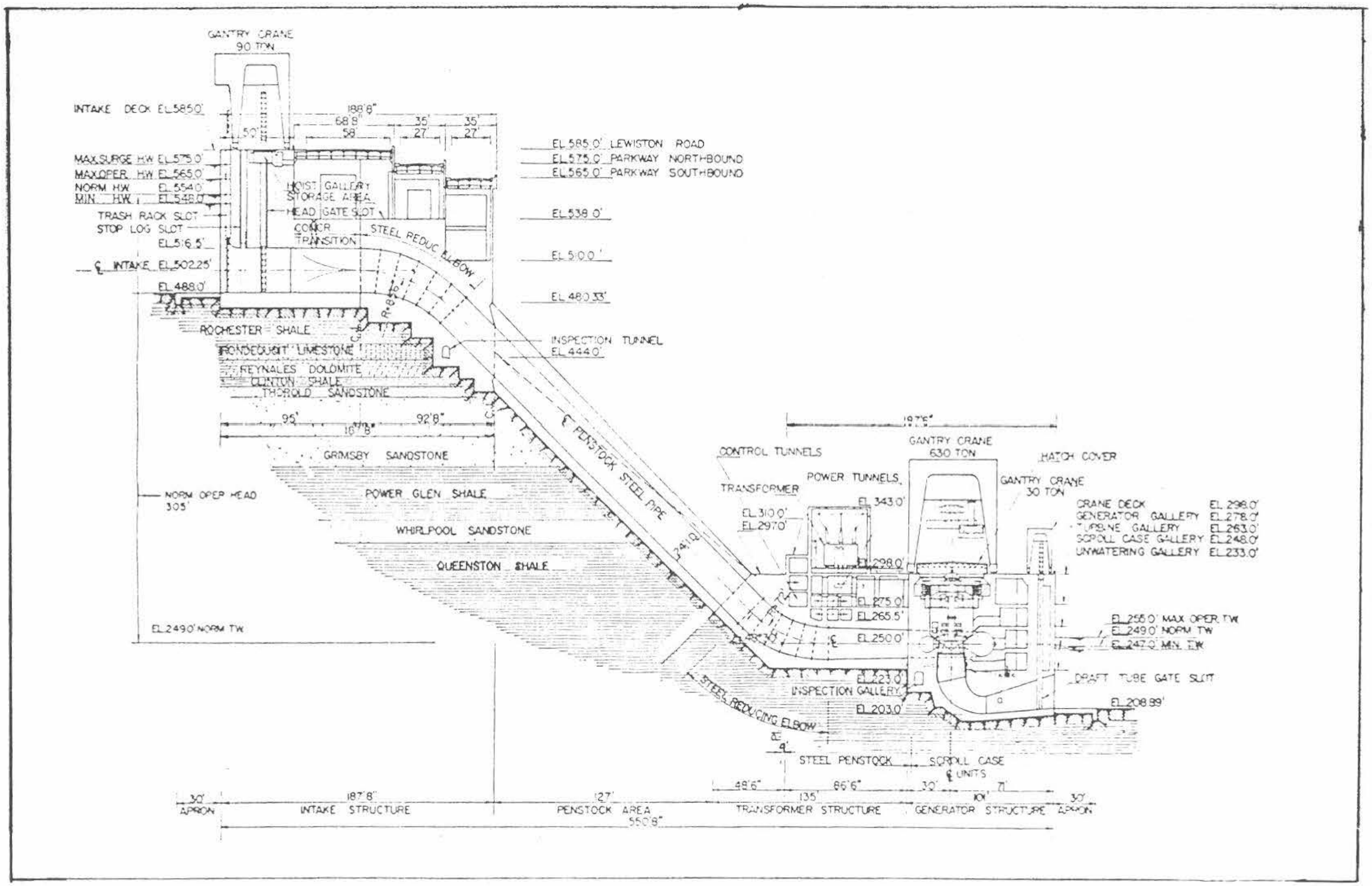


Figure 9. Cross section, Robert Moses Niagara Power Plant. From Power Authority of the State of New York (1965, p. 24).

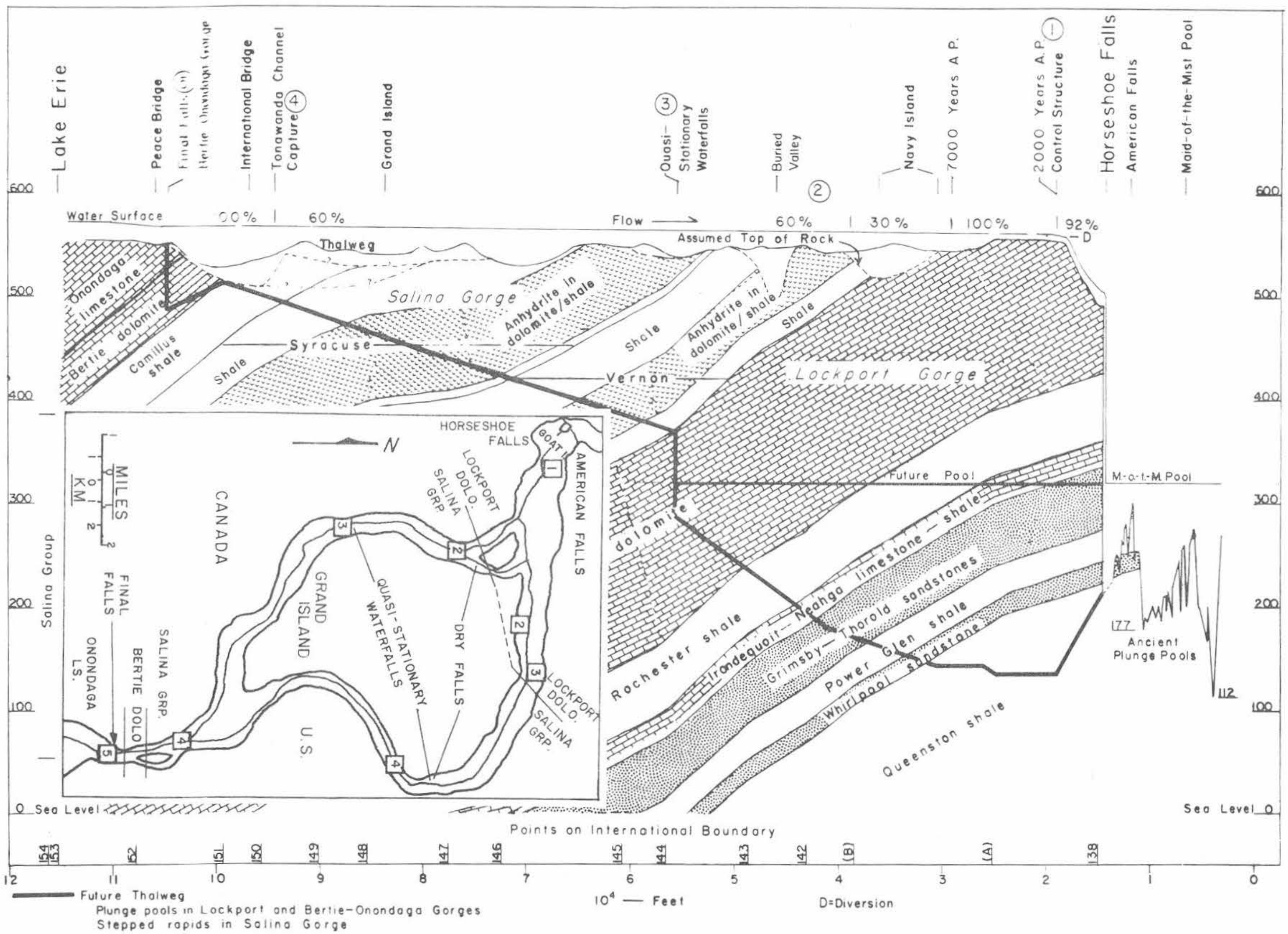


Figure 10. Geologic cross section along the thalweg of the Chippewa Channel (west branch) Niagara River, Lake Erie to Horseshoe Falls. Numbers on top of profile correspond to those on inset map (with equivalent numbers having equivalent dates on both channels) and events in text. Percentages refer to the share of undiverted flow carried during recession of the Falls through the Chippewa Channel. Modified from Philbrick (1974).

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ROAD LOG AND STOP DESCRIPTIONS FOR GLACIAL AND ENGINEERING GEOLOGY ASPECTS OF THE NIAGARA FALLS AND GORGE

(Route to Goat Island duplicates that of Drexhage and Calkin (1982, this volume, Fig. 1). Route and stops along the Gorge are shown in Fig. 1. Figure 11 shows the Lewiston - Lockport Spillway Stop 6 to Marriott Inn route.)

CUMULATIVE MILEAGE	MILES FROM LAST POINT	
0.0	0.0	From Buffalo Marriott Inn turn right at signal light onto Millersport Hwy. and immediately right onto Youngman Expressway (I-290) west (Fig. 11).

		Travel over glacial Lake Warren plain with red clay colored by Queenston Shale. We are also at south margin of Lake Tonawanda plain.
6.0	6.0	Exit right to Rt. I-190 N to Niagara Falls.
7.3	1.3	Toll barrier, South Grand Island Bridge. Cross Tonawanda (east) Channel of Niagara River onto Grand Island.
		Subdued topography due to subaqueous deposition in Lake Warren and washing by initial stages of Niagara River System.
10.7	3.4	Crest of Niagara Falls Moraine.
13.0	2.3	North Grand Island Bridge across Tonawanda Channel of the Niagara. Note spray of Falls at Left.
14.0	1.0	Exit right off I-190 to Robert Moses Parkway to Falls. Niagara Falls Water Treatment Plant and Hooker Chemical Plant on right west of bridge.
15.4	1.4	Structures at left house gates for water intakes to Robert Moses Niagara Power Plant.
		Pass Carborundum Company and Niagara Falls Sewage Treatment Plant on right, Grass Island Pool and control structure (mostly on Canadian side) on distant left.
18.0	2.6	Road bears right under Goat Island Bridge to stop sign.
18.4	0.4	Turn right onto Falls Street and right again onto Rainbow Blvd.
18.7	0.3	Turn right onto First St; proceed through intersection at light over bridge and American Rapids onto Goat Island. Circle island past Horseshoe Rapids.
19.2	0.5	Turn into parking lot after passing Terrapin Point.

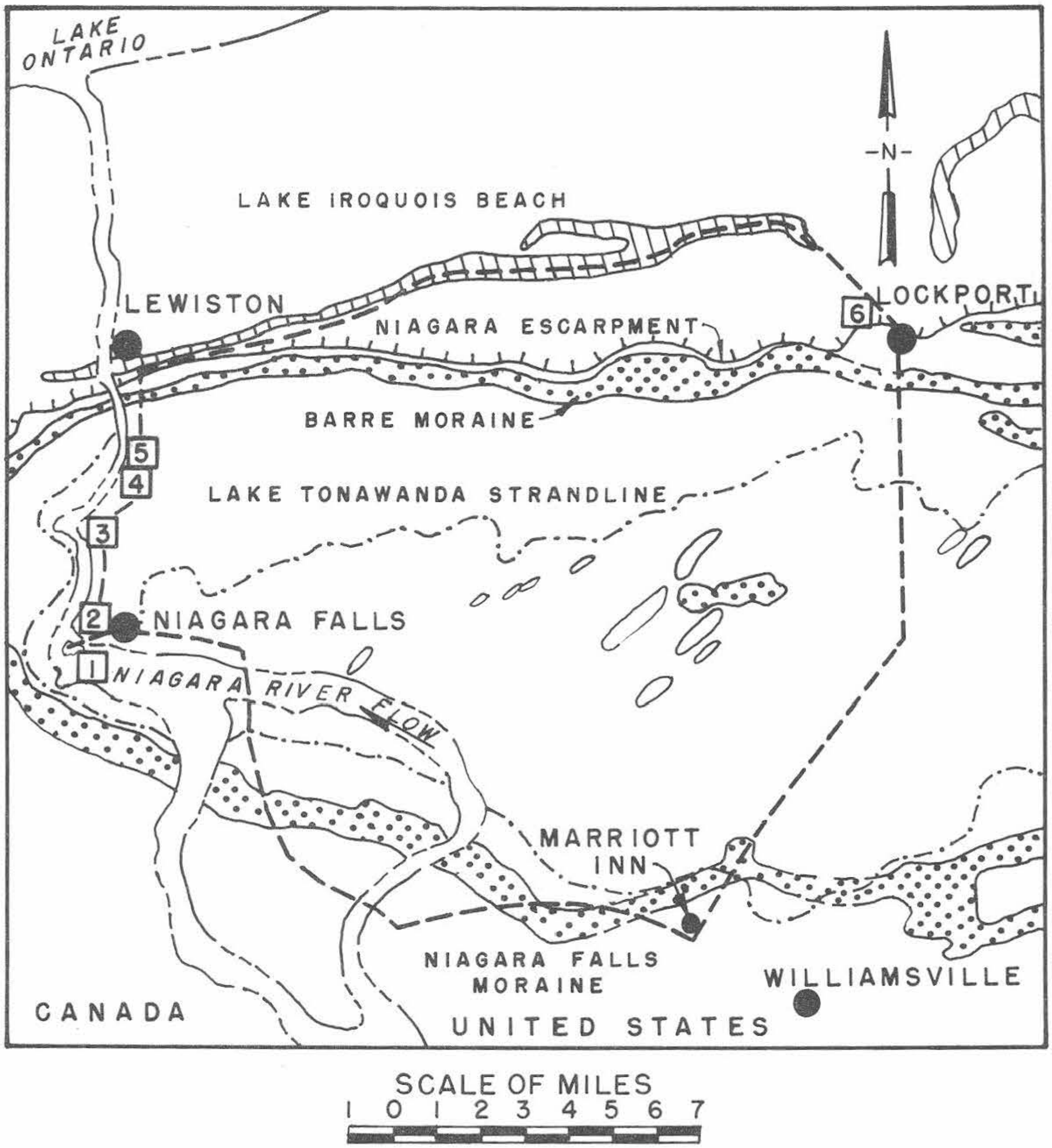

Figure 11. Map of field trip route. Base modified after Muller (1977b).

STOP 1. GOAT ISLAND

The Horseshoe Falls is seen between emerged bedrock terraces of the ancestral river. The Niagara Falls Moraine forms the bluff behind the Falls on the Canadian side. The bedrock slopes which form the Horseshoe and American Rapids are connected beneath Goat Island, at this site, and represent the eastern bank of Spencer's (1907) "pre-glacial Falls-Chippewa Valley." This was believed to trend southward from headwater c. 2 Km (1.2 mi) to the north (near head of north-trending St. Davids Gorge). Figure 3a shows the surficial deposits on Goat Island.

The stratigraphy, structure, talus accumulation, and general location of previous rockfalls at the American Falls will be observed at this stop.

Leave parking lot and take exit back to bridge.

20.5	1.3	Leave Goat Island across bridge over American Rapids; proceed straight past Hotel Niagara along First St. (4 lights) to intersection with Main St.
21.2	0.7	Turn left onto Main St. and immediately take first right following signs to Schoellkopf Geological Museum.
22.7	1.5	Park at Museum.

STOP 2. SCHOELLKOPF GEOLOGICAL MUSEUM

This is the site of the massive rockfall that destroyed the Schoellkopf Powerhouse Stations 3B and 3C and damaged 3A on June 7, 1956. This area of failure will be observed and described.

Leave Museum and retrace route back to Main St.

23.0	0.3	Turn left on Main St.
23.1	0.1	Turn left immediately at light onto Robert Moses Parkway to Whirlpool.
25.4	2.3	Turn left and immediately right around traffic oval into Whirlpool State Park.

STOP 3. WHIRLPOOL PARK

Descend from the parking lot a series of bedrock or gravel river terraces to Gorge margin and view of Whirlpool. The Whirlpool formed soon after 9800 yr BP when the retreating Falls intersected the buried St. Davids Gorge (see Figs. 1, 2, 3b). Known since the time of James

Holland and Charles Lyell, this gorge is as deep and nearly as wide as the Upper Great Gorge (Hobson and Terasmae, 1969), and extends northward across the Escarpment at the community of St. Davids, Ontario, to the mouth of the Niagara at Niagara-On-The-Lake. Taylor (1933) showed that it probably reached as far south (upstream) as the Railroad Bridge where till was encountered in the Gorge bottom and probably up to the head of the Whirlpool Rapids section (Fig. 1). The age and nature of the drift fill (Fig. 2) suggest cutting occurred during or more probably before Middle Wisconsin time under nonglacial conditions when river discharges and baselevel conditions were similar to those of the present.

The Gorge walls expose Lockport Dolostone down through Lower Silurian, Whirlpool Sandstone (type area).

Leave Whirlpool with right and immediate left turns half around traffic oval to Parkway going north.

27.2	1.8	Turn right (at first opportunity) into Devils Hole parking lot.

STOP 4. DEVILS HOLE

This is a plunge pool formed by drainage from Lake Tonawanda via a shallow channel east of the main Niagara channel (Fig. 1). The cave has enlarged through solution localized along a master joint. This site is a good one for viewing the stratigraphy and structure of the Lockport Formation.

Bloody Run, the stream now occupying the lower end of this channel, is named after an Indian massacre of settlers which is reported to have occurred here in 1763.

Leave Devils Hole parking onto Rt. 104 heading north. Pass Niagara University on right and nearby Hyde Park Landfill site of Hooker Corporation.

27.9	0.7	Turn right into parking area for public Power Vista of Robert Moses Niagara Power Plant.

STOP 5. ROBERT MOSES NIAGARA POWER PLANT

A brief stop at this plant will be made to observe the limited exposures of rock excavation surface. The Power Vista itself displays models of Niagara Falls and Gorge, the power generating plant setup and in addition, views of the Gorge stratigraphy. Elevator service to the base of the Power Plant Access Road (traversing good exposures of the Queenston through the Lockport formations) may be obtained for geology field trips with special permission.

Leave Power Vista road right (north) onto Rt. 104E.

29.5	1.6	Cross Barre Moraine at crest of Niagara Escarpment with view of Lake Iroquois plain to north.
		Descend Escarpment; note meandering Lower Niagara River with drop of less than 1 m in 6 mi (9 Km) distance to Lake Ontario.
30.5	1.0	Turn right (east) of underpass following Rt. 104 E (Ridge Rd.) with left turn at stop sign.
		Rt. 104 E here follows crest of wave-cut bluff of Lake Iroquois with thin lacustrine silts on till. Niagara Escarpment at right.
35.7	5.2	At Dickerson Rd., Rt. 104 starts to follow crest of Lake Iroquois beach ridge.
41.3	5.6	Cross Rt. 425 (Cambria Rd.). A second ridge occurs to North (North Ridge spit) probably derived from the Lockport spillway to the east.
45.1	3.8	Turn off Rt. 104 onto Rt.93 at fork and proceed straight (not on Rt. 93) onto Stone Rd.
48.1	3.0	Turn right after crossing Eighteenmile Creek onto Plank (= Purdy) Rd. Enter bedrock embayment of Lockport spillway system for Lake Tonawanda.
48.8	0.7	Turn left up Escarpment on East Jackson St. The Lockport Wastewater Treatment Plant on right is Lockport Gulf site of Calkin and Brett (1978) and Miller and Morgan (1982). Subsurface exposure of the Lockport Gulf site have revealed the following stratigraphy:

Top 0.5 Fill
2.5 m Massive silty clay (post glacial)
2.0 Gray organic silt, clay, and marl with pollen dominated by spruce; basal wood date of 10,920$\pm$ 160 yr BP (I-5841)

3.5 Laminated red and gray sand, silt, and clay (Lake Iroquois?)
Base 9.0 Red clayey sand and gravel interpreted to be till (only in borings).

East Jackson St. follows one of the three major spillway channels at Lockport.

49.4	0.6	Turn back, sharp right onto Glenwood Ave. past St. Patrick's Cem. Kame deposits are related to Albion-Rochester Moraine.
49.7	0.3	Turn left onto North Transit Street up escarpment.
49.8	0.1	Turn right at top onto Outwater Drive.
50.3	0.5	Turn right into short loop road to escarpment edge just before reaching water tank.

STOP 6. OUTWATER PARK, LOCKPORT

View northward from Lockport Escarpment of Lake Iroquois plain and Lake Ontario (if clear) 11 mi. distant. The Lockport Gulf (W. Jackson St.) spillway circles just below. Good exposures of the bedrock surface and northeast-southwest oriented striations occur in the park dump area to the west of the water tank.

50.4	0.1	Turn left back eastward on Outwater Drive.
51.0	0.6	Turn right (south) onto North Transit.
51.7	0.7	Cross Erie/New York State Barge Canal proceeding south on Transit Road (Route 78).
51.9	0.2	Cross High Street on crest of Barre Moraine ridge.
52.6	0.7	Cross Summit Street at light, another of four Barre Moraine ridges mapped by Gilbert (Kindle and Taylor, 1913).
57.5	4.9	Cross Tonawanda Creek and bear right onto Route 263 South (Millersport Highway). You are in middle of the former River-Lake Tonawanda plain. A few low NE-SW-oriented drumlins can be seen to the northeast.
66.2	8.7	Turn right at light into Buffalo Marriott Inn.

QUATERNARY STRATIGRAPHY AND BLUFF EROSION WESTERN LAKE ONTARIO, NEW YORK

PARKER E. CALKIN, Department of Geological Sciences, State University of New York at Buffalo, Buffalo, New York 14226

ERNEST H. MULLER,* Department of Geology, Syracuse University, Syracuse, New York 13210

THOMAS F. DREXHAGE,* Acres American Inc., Liberty Bank Building, Buffalo, New York 14202

INTRODUCTION

Wave-steepened bluffs along the south coast of Lake Ontario display the most continuous and some of the most interesting exposures of glacial drift in New York. Study of these bluffs provides detail of local and regional Pleistocene history, as well as insight into mechanisms of glacio-lacustrine sedimentation. In addition, knowledge of the composition, structure and correlation of bluff materials is fundamental to interpretation of rates and modes of coastal recession and sediment contribution to the lake basin. Wave and subaerial erosion of these bluffs has been particularly severe because bedrock comprises a very small part of the bluff sections.

Record, or near-record high lake levels between 1972 and 1978 greatly accelerated bluff recession and provided the stimulus for a Sea Grant study at SUNY Buffalo of more than 400 km (250 mi) of Lake Ontario coast in New York (Drexhage and Calkin, 1981).

In this guide book we focus on 1) the geologic setting, composition and stratigraphy of the bluffs west of Rochester (Fig. 1); 2) the historic rates and distribution of bluff recession and sediment loading to the lake along this same reach of coast; and 3) some of the factors that influence local differences in recession rates.

Much of the data on bluff recession in this report is from the comprehensive study of historic recession along the whole coast in New York (Drexhage and Calkin, 1981), as well as more local studies (Fortune and Calkin, 1981; Brennan and Calkin, in press). A comprehensive account of the Lake Ontario bluff stratigraphy in New York will be published elsewhere (P.E.Calkin and E.H.Muller, in prep.), but preliminary accounts have been presented by Calkin and Brennan (1976) and Calkin and others (1978).

Basic field measurements and sampling for the stratigraphic study were undertaken during the summers of 1974 and 1975 when Lake Ontario levels were, at times, as much as 1.5 ft (0.45 m) above the 100-year monthly averages which range from 245 to 246.6 ft (74.68 to 74.86 m), respectively. Bluff sections were measured at least every kilometer. Attempts were made to trace stratigraphic units continuously along the coast. In this guide, the same station numbers are used for location on Figures 5 through 25 and are referred to in the text without figure numbers and usually within parentheses.

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* Field trip leaders

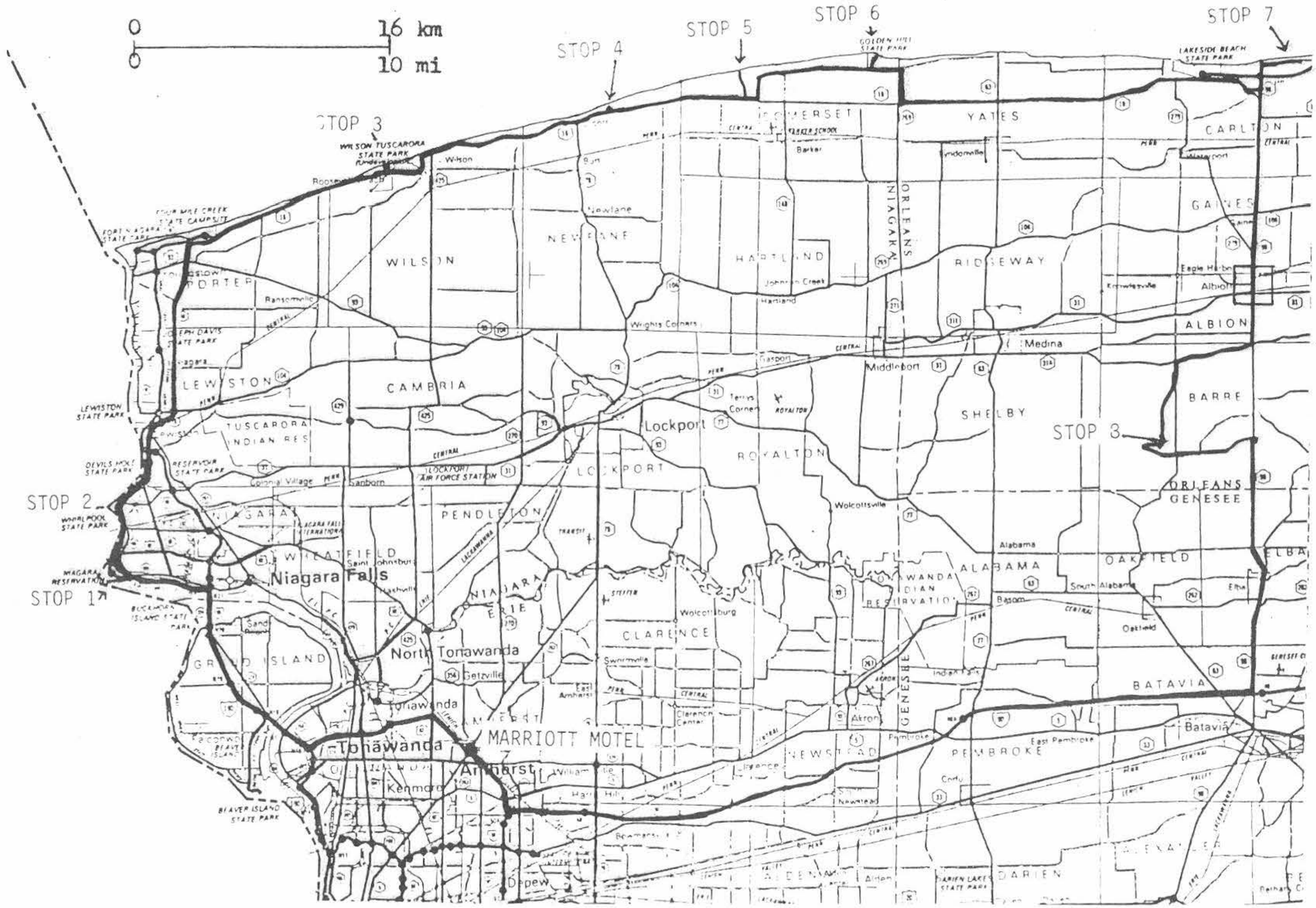

Figure 1A. Field trip route and stops for Lake Ontario shore bluffs field trip. Figures 1A and 1B are the same scale; e.g. Golden Hill State Park (Fig. 1B is at Thirty Mile Point (Fig. 1A).

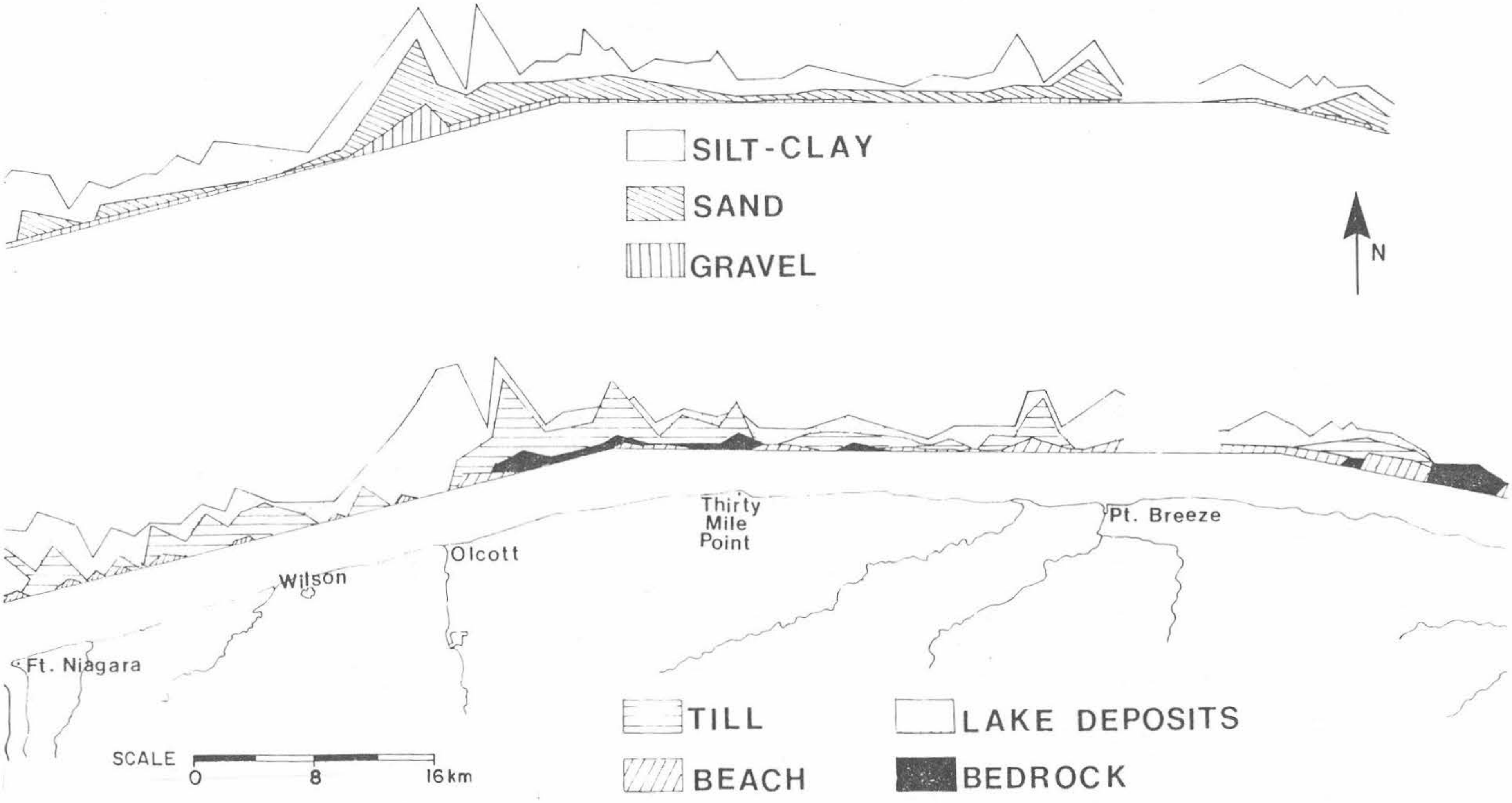

Figure 12. Areal distribution of sediments in bluff exposures along the south shore of Lake Ontario, by textures (above) and genesis (below) (Brennan and Calkin, in press).

The glacial geology of the south coast of Lake Ontario has been studied by many workers, particularly between 1880 and 1930 (Muller, 1965). The most pertinent works are those of Kindle and Taylor (1913), Dames and Moore (1974), Salomon (1976) and Muller (1977a, 1977b). Adjacent portions of the Niagara Peninsula in Ontario have been mapped by Feenstra (1972a, 1972b, 1975, 1981).

LATE QUATERNARY HISTORY

Glacial deposits in western and central New York originated during at least one pre-Wisconsin glaciation and five successively less extensive Wisconsin stadials (Calkin, 1982; Calkin and Wilkinson, 1982; Calkin et al., 1982). During the last, or Port Huron Stadial, starting about 13,000 yr BP (Dreimanis and Goldthwait, 1973), ice advanced through the Ontario Basin to the margins of the Allegheny Plateau. Glacial lobation into the Erie Basin formed the Hamburg Moraine (Frontispiece II) southwest of Buffalo while the ice margin reached a position at or south of the Waterloo-Auburn Moraine in central New York (Connally and Sirkin, 1973; Muller, 1977b). At the same time, glacial advance in Michigan formed the Port Huron Moraine and initiated the westward-draining glacial Lake Whittlesey in the Erie-Huron Basins. Oscillatory retreat and moraine formation in the Erie Basin is correlated successively with glacial Lakes Warren I and II, Wayne, Warren III and possibly lower, short-lived Lakes Grassmere, Lundy and Early Algonquin before initiation of Early Lake Erie and Lake Iroquois about 12,400 yr BP (Calkin, 1970; Muller, 1977b). During this interval, ice movement was dominantly southwestward in the field area as shown by alignment of drumlin forms, bedrock fluting and striation.

Impounding of water in the Ontario Basin below the Niagara Escarpment began during retreat from the Batavia, Barre and Albion Moraines in the Erie-Ontario Lowlands. Eastward drainage to the Mohawk was initiated by ice marginal withdrawal from the Onondaga bench north of Skaneateles. Fairchild (1928) mapped shore features of Lake Dawson at 470 to 480 ft (143 to 146 m) locally in the Rochester area, but the extent of sediment contribution by this lake and its predecessors was limited along the present Ontario bluffs.

Renewed glacial recession in central New York allowed expansion into central New York by glacial Lake Iroquois, controlled by the col at about 440 ft (134 m) near Rome. Muller (1977b) indicates that inception of Lake Iroquois preceded the main readvance of the ice front to the Carlton Moraine (Frontispiece II).

Work of Gilbert and Taylor (Kindle and Taylor, 1913) suggests that early Lake Erie drained first to a short-lived low stage of Lake Iroquois called Newfane, whose beaches are now at about 360 ft (109 m) near Lewiston. Subsequently, uplift of its outlet raised the Iroquois waters 25 to 40 ft (7.6 to 12 m) to the most persistent beach-forming stage. The prominent Iroquois beach ridge (Frontispiece II) has been strongly tilted by glacial rebound. The strand line now rises from 360 ft (109 m) to 545 ft

(165 m) between the Niagara River and the east end of Lake Ontario (Fairchild, 1916). Rebound apparently continues even to the present time. As indicated by long term records at gauging stations, the outlet of Lake Ontario is rising faster than the southwest end by about .23 m/century (Clark and Persoage, 1970; Kite, 1972), thus continuing gradually to drown river valleys and raise lake level along the south coast.

An analysis of dates and events in the Ontario - St. Lawrence area led Karrow and others (1975) to suggest that Lake Iroquois drained to much lower "post-Iroquois" levels (Prest, 1970) shortly after 12,000 yr BP, but at least before 11,000 yr BP. Evidence of post-Iroquois stages, including those of early Lake Ontario are not known in exposure in western New York.

The present coastline is cut into a surface of low relief, broken by very subtle northeast-southwest trending ridges or "giant fluting" (Kindle and Taylor, 1913) paralleling the direction of former glacier flow. Low hummocks of the Carlton Moraine are also distributed in a narrow generally east-west band (Muller, 1977a, 1977b).

BEDROCK

Lake Ontario is underlain by Middle and Late Ordovician sedimentary rocks with a regional east-west strike and dip of about 40 ft/mi (7.5 m/km) south or slightly west of south. This sequence as measured in borings along the southern Lake Ontario Lowland, includes from youngest to oldest: 200 to 600 ft (61 to 183 m) of sandstone and shale of the Queenston Formation; 950 ft (290 m) of Lorraine Group rocks including the Oswego and Pulaski sandstones and the Whetstone Gulf siltstone and shale; 150 to 250 ft (46 to 76 m) of Utica Shale; and 450 ft (137 m) of limestone of the Trenton Group (Fisher, 1977). The Late Ordovician red Queenston rocks underlie bluffs along the south coast of Lake Ontario.

Figure 1B shows the linear distribution and vertical relief of bedrock and unconsolidated units in bluff exposure. Variable elevation of the bedrock surface is primarily a consequence of glacial erosion. Outcrops of Queenston bedrock form many of the "points" or lakeward projections of land west of Sodus Bay. The main exposures are south and east of Thirty Mile Point (75) where the Queenston rises to near its maximum height of about 4 m above mean lake level (Stop 6).

Queenston rocks exposed in bluffs between the Niagara River and Rochester include laminated to thick-bedded, red to dark red (purplish or cherry red) fine- to very fine-grained calcareous argillaceous sandstone and red calcareous shale or siltstone. Beds of green or gray sandstone and shale make up less than 10 per cent of the exposed rock. Red shale is most common in exposures near the Niagara River with more siliceous sandstones to the east along the coast.

The attitude of beds in the regional homocline is relatively uniform but dips as steep as 4^{0} occur 20 km west of Rochester (Dames and Moore,

1965) and locally near some of the many points of land where secondary fold axes occur oblique or normal to the regional east-west strike. Small, but prominent tight anticlinal structures have been described at or near the shoreline east and west of Thirty Mile Point (Stop 6)(Gilbert, 1899; Dames and Moore, 1965) and a few kilometers west of Olcott (near sta. 54)(Kindle and Taylor, 1913).

Wave-cut platforms in the field area display prominent and consistent joint patterns with a predominant set striking between 65° and 75°, dipping at 80° to 90°. A second set strikes between 330° and 350° with 60° to 90° dips. The spacing is generally 0.5 to 1.5 m (Dames and Moore, 1965, 1974; Fakundiny and others, 1978). A number of small faults occur along the coast (Kindle and Taylor, 1913; Dames and Moore, 1974), but the only major fault known to transect the area is the north-south Clarendon-Linden normal fault system, expressed as a drift-filled bay on the Ontario coast just east of the field trip area near Troutburg (Van Tyne, 1975; Hutchinson and others, 1979).

QUATERNARY STRATIGRAPHY

A maximum of six stratigraphically superimposed lithostratigraphic units may be distinguished within the drift between the Niagara River and Rochester. Figure 2 shows a composite stratigraphic section. Only very locally are all of these units exposed together in one section. Two tills with interbedded glaciolacustrine units are distinguished and correlated laterally on the basis of moist field color, relative stratigraphic position, and associated primary structures as well as texture (in order of decreasing reliability) (Brennan and Calkin, in press). However units may also be locally distinguished and correlated using gross lithology and compaction due to minor differences in ice-flow trajectories (Salomon, 1976).

Lower Glaciolacustrine Deposits

The oldest unit is a red, or less commonly, gray glaciolacustrine clay and silt up to one meter thick. This unit crops out in the Somerset bluff area (61) during times of strong wave erosion and is distinguished in test boring in the region (Dames and Moore, 1974; Pendleton, personal communication, 1975) beneath a lower red stony till and above the Queenston bedrock. It is locally stony and/or well-laminated in upper parts and absent or unrecognized over most of the bluff exposure where the lower red stony till either lies directly on bedrock or is itself not distinguished from the upper gray silty till.

Red Stony Till

A very stony, weak red (2.4YR 4/2), compact basal till overlies the lower glaciolacustrine unit or in other locations lies directly on the Queenston bedrock with gradational contacts (Fig. 2). At its base this till displays evidence of minimal reworking or glacial transport but locally includes small, well-striated and polished stones. Textural maturity is proportional to thickness and height above bedrock. This unit is reported to reach a thickness of 2 m in the field trip area (Dames and Moore, 1974). Borings in the Somerset area (59 to 71) reveal thicknesses of less than one meter.

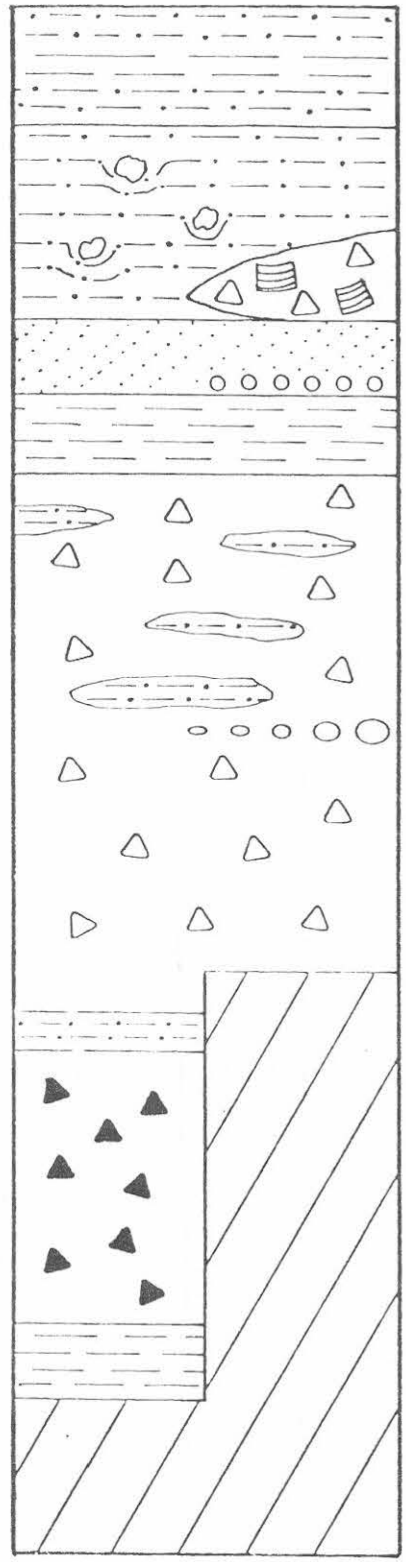

Figure 2. Composite stratigraphic section of the bluff stratigraphy of the south coast of Lake Ontario, Niagara River to Rochester, New York.

Till matrices are poorly sorted and composed of silty fine sand, sand silt clay, or less commonly sandy or clayey silt where shaly or weathered. Salomon (1976) showed that percentages of silt and clay are closely correlated with those of red (Queenston) clasts in the till. Stones scattered irregularly in the sandy matrix comprise up to 80 per cent by weight and consist predominantly of fine to coarse red brown, micaceous Queenston sandstone. Westward, near the Niagara, the percentage of Queenston clasts may be equaled by that of fine to medium-grained micaceous Oswego Sandstone. The stones are angular to subangular and tabular in all but small-pebble sizes. Locally, reworking is indicated by discontinuous lensoid silt and fine sand at the upper contact.

Exposures of the red stony till and any underlying glaciolacustrine unit are uncommon in the coastal reach west of Rochester, partly because they are discontinuously preserved and partly because they are at the base of the bluff and readily obscured by slumping. Exposures have been reported most frequently near Somerset (61), and from the Wilson area (34 and 48) by Salomon (1976). Some exposures of stony red till lying directly on Queenston bedrock may display characteristics misleadingly similar to those indicated above because of assimilation of underlying rock material, yet may be shown to be a facies of the younger purplish-gray till described below.

Intertill Glaciolacustrine Deposits

Glaciolacustrine deposits or till overlie red stony till. Where superjacent till is exposed in the bluff, as near stations 48 and 61, it may be separated from the underlying red stony till by discontinuous gray, or less frequently, red glaciolacustrine clay to fine sand units up to 0.5 m thick. Near Roosevelt Beach, Gilbert (1898) described a boulder pavement which may represent this interval between deposition of two till units.

Silty Purplish-Gray Till

The predominant till of this reach of coast is a silty gray or purplish deposit (5 YR 5/1 or 10 YR 6/2) which weathers to a red brown (Fig. 2). This unit is here referred to as the "Somerset Till" after the town in Niagara County, New York where it is particularly well exposed. This till has been locally observed directly on the red Queenston Formation in a few stations where it typically shows sharp color and textural contacts with the bedrock (e.g. near 59 and 83). However, in most areas it lies over fine glaciolacustrine deposits and/or lower red stony till (61). The stratified deposits were the sources of its finer texture and gray color (Pendleton, personal communication, 1975; Salomon 1976).

The "Somerset Till" consists of a lower, compact, texturally and lithologically homogenous silty till unit. Preliminary study suggests that it is largely a basal lodgement till but that some stratified subaqueous till is typical as well. The compact till facies generally grades upward to a subaquatic allo-till facies with similar matrix but enclosing much stratified material. We interpret it to have been deposited near the glacial margin by secondary processes much like the well-published Catfish Creek Till of Ontario, Canada (Dreimanis, 1976; Evenson and others, 1977; Gibbard, 1980; Dreimanis, 1982). The Somerset subaquatic till facies has indistinct upper contacts with overlying glaciolacustrine sediments which blanket most of the bluff top. One or both till facies may be present in any one exposure and maximum thicknesses reach about 3 m.

The"Somerset Till"is generally distinctly siltier and less stony than the underlying red till. Sand-silt-clay to clayey silt matrices are typical. Stones average about 4 percent by weight. Slightly higher percentages are characteristic in the lower than in the upper facies. Larger stones show strong evidence of glacial handling, particularly in the compact till facies. The subangular red Queenston clasts and gray to green Oswego sandstones are subequal in number with subordinate content of well-rounded dark fossiliferous limestones. The lithologic assemblage is much like that of the red stony till below except that the ratio of greenish-gray to red sandstone or siltstone is slightly higher (Salomon, 1976). Clasts of lacustrine clay and of the older red stony till are also reported (Dames and Moore, 1974; Pendleton, personal communication, 1974).

The upper, subaquatic allo-till facies is up to 3 m thick and composed of massive to laminated till enclosing subhorizontal, blue, green, gray, yellow, red, or brown lenses, stringers, or wisp-like beds of very fine to coarse sand, silt or clay. These may make up from 20 to 80 percent of the unit. Clay is less common than silt or sand lenses. The individual sorted and bedded units are from 1 to 50 cm thick, locally contain small-scale current structures and drop stones, and are rarely continuous laterally for more than a few meters. The beds may also be wrinkled, or grossly tilted. Isoclinal-flow folds (Evenson and others, 1977) occur on a scale of less than a meter within some individual till units.

Stones associated with the subaqueous facies in some places occur along distinct horizons in the waterlaid units as lag deposits or pavement. For example, on either side of the Wilson Inlet (34 and 38; Stop 3) a stone lag deposit bearing parallel striations occurs between the contact of the compact till facies and the upper weathered subaqueous till.

The subaqueous allo-till deposits have remained unweathered and at a distance are not readily distinguishable from the compact lower facies where overlain by more than 3 or 4 meters of relatively impermeable lake silt and clay. However, they appear to be more readily oxidized and display colors of higher chroma than the underlying compact till when nearer to the ground surface. For example, purplish-gray (2.5 YR 4/1) has been oxidized to red (2.5 YR 4/2) and except for the paucity of stones, resembles the red stony till unit. The contacts between purplish-gray (unaltered) and overlying red (oxidized) tills is typically sharp. Horizontal and vertical partings which develop in the weathered till allow it to maintain nearly vertical slopes emphasizing the apparent differences in the two facies.

A model suggesting possible mixed subaqueous origins of the purplish gray "Somerset Till" is shown in Figure 3 with inset. As the ice margin retreated across the current coastal position it was fronted by Lake Iroquois waters more than 70 m deep. While grounded, the temperate-based ice had deposited compact lodgement or meltout till. However, the thinning margin became increasingly buoyant, shedding debris by meltout, marginal or submarginal flowage and settling.

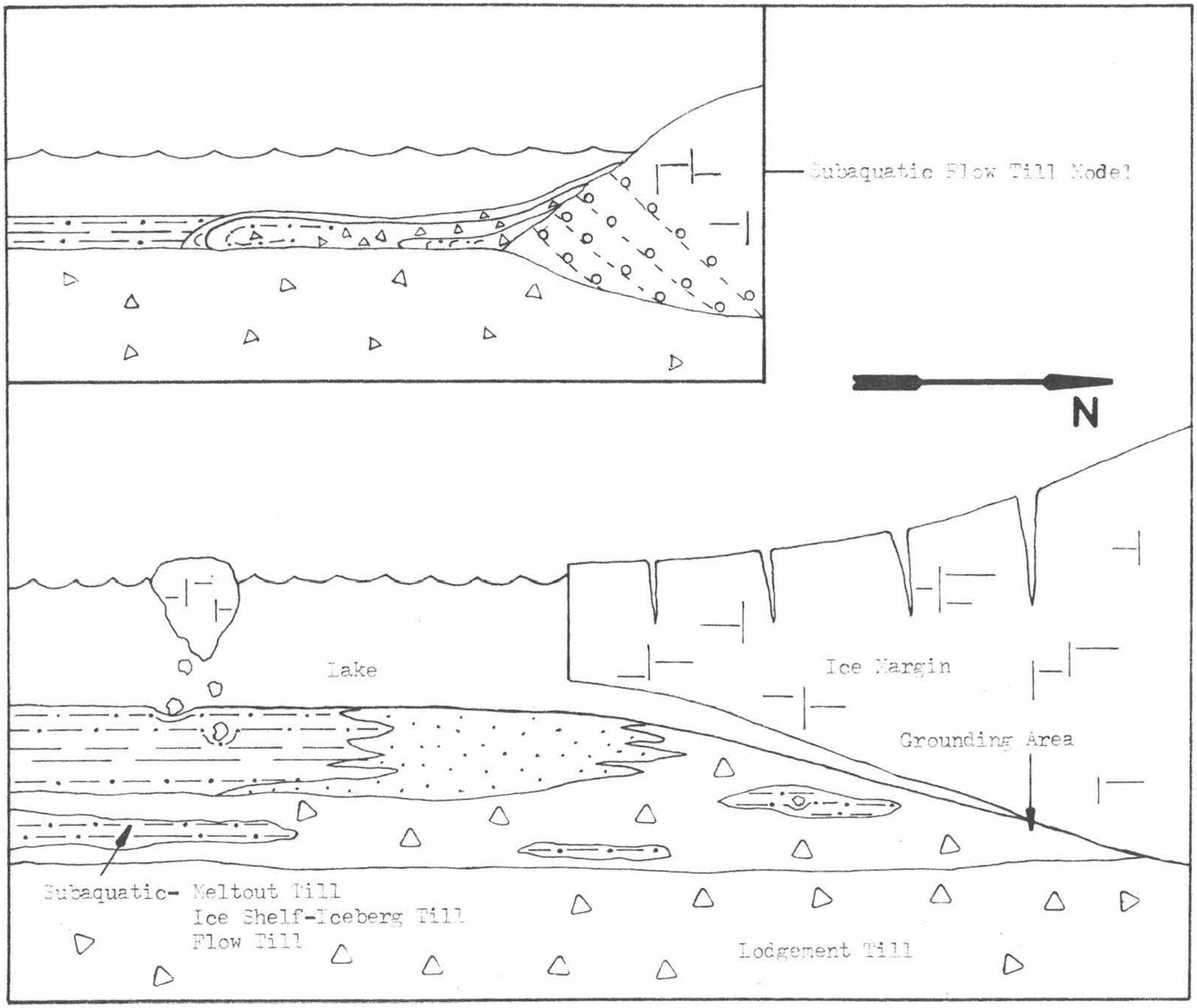

Figure 3. Models for till deposition along the south coast of Lake Ontario, Western New York.

Upper Glaciolacustrine Deposits of Lake Iroquois

Glaciolacustrine deposits averaging 3 to 4 m, but as thick as a maximum of 10 to 15 m (57), overlie the "Somerset Till" or older deposits (Fig. 2). These are considered to be largely Lake Iroquois deposits; however southwestern parts of the Ontario shore bluffs may include Lake Warren III through Hyper-Iroquois deposits and northeastern parts may include some post-Iroquois lake sediments. Red or red-brown, oxidized, massive to thin-bedded silt layers predominate, but rhythmic gray clay and silt units with couplets up to 8 cm thick lie directly on "Somerset Till". Fine yellow brown, oxidized, sandy beds are also locally an important component in upper parts of the glaciolacustrine unit.

Drop-stones up to cobble size are common locally, but the fine lacustrine beds also enclose ice-marginal deposits in several localities as discussed below. Lower contacts may be diffuse, or, in more than 10 per cent of the reach, may display a pebble or boulder lag on the till/lake clay interface.

Ice-Marginal Deposits

Gravelly silt, coarse sand and gravel, and pebbly lag deposits and/or intercalated silty till lenses interpreted as subaquatic flow till (Evenson and others, 1977) occur within the upper glaciolacustrine unit at lateral intervals generally less than 2 to 3 mi (3 to 5 km)(Fig. 2). Because of their composition, form and location adjacent to remnants of the Carlton Moraine (Frontispiece II), they are interpreted as near-ice or ice-marginal deposits related to an active glacial stand or short glacial readvance with subsequent oscillatory retreat soon after initiation of Lake Iroquois (Muller, 1977a). They are typically separated from underlying purplish-gray "Somerset Till" by fine lacustrine deposits and/or by an erosional unconformity with this lag concentration of clasts. These units are particularly developed in at least two locations in the field trip area (Stops 4 and 7). Just east of the field trip area, near Hamlin Beach State Park, flow till lenses occur within the cross-bedded, lacustrine sand unit (Drexhage and Calkin, 1981, p. 55).

CORRELATION OF TILL UNITS

The lower red stony till is tentatively correlated eastward with thicker but similar tills through the Sodus Bay area and with some tills of the inner cores of the drumlins beyond to Oswego. The purplish-gray "Somerset Till" can be traced eastward into the gray clayey tills of the Rochester-Sodus Bay area and perhaps the pink and gray upper drumlin tills. Properties of the tills, but particularly the lower one, mirror changes of regional bedrock geology and minor differences in ice-flow trajectory along the coast (Salomon, 1976; Calkin and Brennan, in press).

Southward from the Lake Ontario bluffs, the till units are generally buried beneath lake deposits. Borings in the Somerset area (Dames and Moore, 1974) showed both tills separated by lake clay; however in this area the purplish-gray till has not been traced more than 1600 to 8400 ft (488 to 2500 m) south of the shoreline. Elsewhere, abundant subsurface data (unpublished) verify the occurrence of the two contrasting till units, at least southward to about the Iroquois Beach. Systematic tracing and correlation of these units southward across the Niagara Escarpment has not been established in New York.

Detailed work in the Niagara Peninsula of Ontario may provide a valid working hypothesis for correlation in New York. Gray to brownish-gray silty or clayey till exposed directly beneath postglacial lake sediments in the bluffs west of the Niagara River and underlain by, or interbedded with, red sandy silt till are correlated with the Halton Till (Feenstra, 1972a, 1972b, 1981). This is, in turn, traced southward in the Niagara-Welland areas by buried drift valleys across the Escarpment and represents the last advance of the Ontario glacial lobe "across the Niagara Peninsula into the Lake Erie Basin during the Port Huron Stadial" (Feenstra, 1981, p. 87), Karrow (1963, p. 44) notes that in the Hamilton-Galt area of Ontario, the "Halton Till is predominantly a silt till that is red brown when oxidized and dark purple when fresh". He further indicates that at the type locality it contains an upper waterlaid zone that is stratified.

Borings along the eastern half of the Lake Ontario coast in the Canadian Niagara Peninsula reveal a red sandy silt till. This is traced southward beneath the fine textured gray till and is correlated with Late Wisconsin glacial advances of Nissouri-Port Bruce Stadials (Calkin, 1982, Fig. 2).

SHORE EROSION

Shore erosion rates may be variously measured, for instance, in terms of bluff recession and of sediment loading on the lake system. Bluff recession rates are of immediate consequence to coastal structures built atop of the receding bluffs. Sediment loading rates are of concern relative to beach maintenance and shoaling of navigation channels.

Shore erosion rates are related to a variety of factors, among them, bluff stratigraphy, geotechnical properties of bluff sediments, groundwater conditions, beach characteristics, offshore gradients and wave climate. Variations in Quaternary stratigraphy and associated properties of materials in shore bluffs of Lake Ontario lead to corresponding variability of coastal morphology and shore erosion rates.

Among the factors mentioned above, bluff materials and morphology, beach characteristics and offshore gradients are both affected by and themselves affect rates of shore erosion. They are dependent variables in the erosion system. Lake stage, and to a lesser degree, wave climate, are external or independent variables imposed upon the system. Higher lake stage and greater storm frequency, by steepening offshore gradients and undercutting shore bluffs initiate new cycles of accelerated shoreline erosion (Emery and Kuhn, 1982). The succession of high-water years through the mid-seventies initiated just such a cycle of accelerated erosion. For this reason, bluff recession rates averaged over a 13 to 18 year period prior to 1955 were compared with rates averaged over 99 years prior to 1974. As would be expected from the episodic nature of accelerated erosion, greater extremes in bluff erosion rates are recorded in short term data (Drexhage and Calkin, 1981).

"Classification of sea cliffs in just a small region ... involves more generalizations than can be accepted for engineering purposes." (Emery and Kuhn, 1981, p. 652). With full recognition of this fact, the following generalizations are offered relative to shore erosion in the field trip area.

A general decrease in bluff recession rates is found from west to east across the field trip area. Most of the bluffs in Niagara County (1 thru 75) face north-northwest. Such bluffs are exposed to wave action induced, for instance by strong westerly and northwesterly winds which follow frontal passages and to waves generated by northeasterly winds with maximum fetch. Bedrock elevations are higher in eastern Niagara County, reaching a maximum in the vicinity of Thirty Mile Point. Approximately 2×10^5 m^3/yr average erosional loading occurs from bluffs in that portion of the field area included in Niagara County (Fig. 1B). However, due to generally steep nearshore slopes, typically 10 to 27.5 ft/mi (5 to 14 m/km), and the large amount of fine material present in the bluffs, little sediment remains on the shoreline to protect the bluffs from wave attack. The lack of well developed beaches permits wave onslaught against glacial till and associated glaciolacustrine deposits. This occurs particularly during intervals of high lake levels and severe storms, resulting in high recession losses in a short time. For Niagara County, the mean recession rate between 1875 and 1934 was 46 cm/yr, whereas between 1938 and 1954 the rate was 79 cm/yr .

The portion of Orleans County included in the field trip area (75 to 99) has generally experienced less recession than Niagara County. Bluffs face more northerly (Fig. 1A) and are less subject to direct wave attack. At Stop 6 (75 and 76) less than 30 cm/yr recession was evidenced between 1875 and 1974. Recession in this area has been fairly uniform, reflecting the relative regularity of the coast and similarity of the bluffs. Annual erosional loading for Orleans County was computed at an average of 2.5×10^4 m^3/yr, an order of magnitude lower than for Niagara County. Low bluffs, thin drift and relatively slow bluff recession all contribute to poor development of beaches along much of this coast. An apparent paradox lies in the fact that greatest long-term bluff recession rates in Orleans County have been observed where beaches are relatively well developed, east of Brighton.

The greatest erosion rates in the field trip area are recorded in the vicinity of Roosevelt Beach (Fig. 4). This area experienced 2.37 m/yr average recession from 1938 to 1951 and 1.28 m/yr over the longer term from 1875 to 1974. Recession has been so severe that Twelvemile Creek which in 1874 entered Lake Ontario through Tuscarora Bay has been truncated and now enters separately more than a mile to the west. Numerous residences and several streets in Roosevelt Beach have been lost to bluff erosion. Bluffs in the vicinity have steep faces, without bedrock exposure above lake level. Beaches are generally less than 10 ft (3 m) wide and in present lake stage, essentially nonexistent. They are largely composed of cobble-sized material (Corps of Engineers, 1945, 1973) affording little protection against wave action (Fortune and Calkin, 1981). Tributaries which might nourish the beach are insignificant and long-shore drifting is minimal.

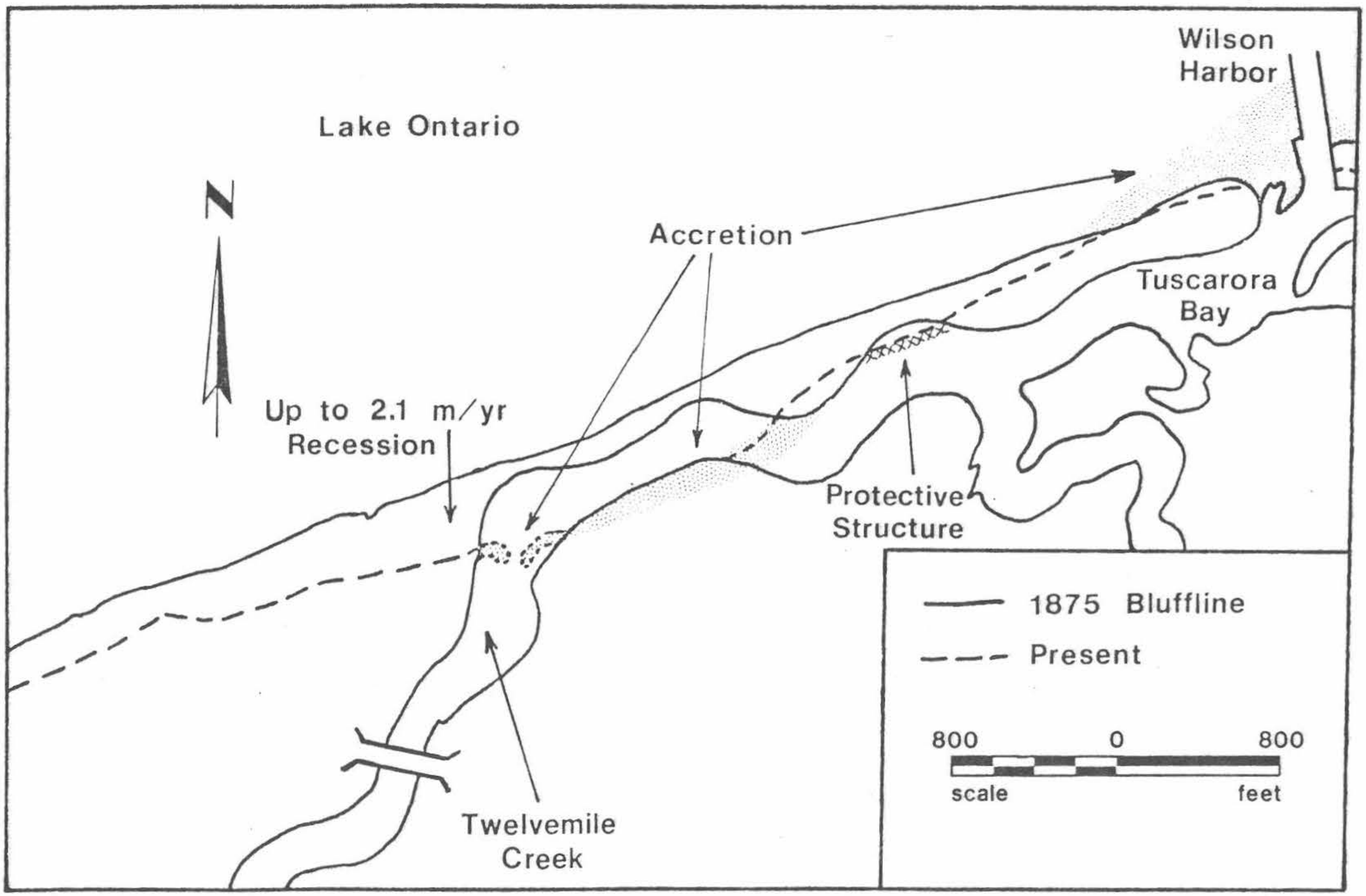

FIGURE 4 : Map of Twelvemile Creek - Wilson, N.Y. area, showing the 1875 and 1974 shoreline positions. Note the change in location ofthe mouth of Twelvemile Creek during the 99-yr period (after Fortune and Calkin, 1981).

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ROAD LOG
QUATERNARY STRATIGRAPHY AND BLUFF EROSION
WESTERN LAKE ONTARIO, NEW YORK

(Route and location of stops are shown on Figure 1a)

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	From Buffalo Marriott Inn turn right at light onto Millersport Highway and immediately right onto Expressway (I-290) west.
		On Lake Warren plain (note red clay exposures at right). We are just south of the edge of Lake Tonawanda plain.
6.0	6.0	Exit right to Route I-190 N to Niagara Falls
7.3	1.3	Toll barrier, South Grand Island Bridge. Cross Tonawanda (east) Channel of Niagara onto Grand Island (Erie County, N.Y.).
		Subdued topography due to subsequent deposition tion in Lake Warren and washing by initial stages of Niagara River system
10.7	3.4	Crest of Niagara Falls Moraine
13.0	2.3	North Grand Island Bridge across Tonawanda Channel of the Niagara. Note spray of Falls at left.
14.0	1.0	Exit right off I-190 to Robert Moses Parkway to Falls. Niagara Falls Water Treatment Plant and Hooker Chemical Plant on right, west of bridge.
15.4	1.4	Structures at left house the gates for water intakes to Robert Moses Power Plant.
		Pass Carborundum Co. and Niagara Falls Sewage Treatment Plant on right; Grass Island Pool and control weir (Canadian side) on distant left.
16.1	2.6	Road bears right under Goat Island Bridge to stop sign.
18.4	0.4	Turn right at first opportunity onto Niagara Street and right again onto Rainbow Blvd.

18.7	0.3	Turn right onto First Street; proceed through intersection with light over American Rapids onto Goat Island. Pass Horseshoe Rapids to main parking for Falls and Terrapin Point.

STOP 1. TERRAPIN POINT, GOAT ISLAND, NIAGARA FALLS. We will take time only for photographs. See Calkin and Wilkinson (this volume, Stop 1) for details.

20.5	0.6	Leave Goat Island across bridge over American Rapids; proceed straight past Hotel Niagara along First Street to 4th traffic light.
21.2	0.7	Turn right from middle lane onto Main Street at 4th light; turn left almost immediately at blinking light onto Robert Moses Parkway north toward Whirlpool. Pass Schoelkopf Geological Museum along Gorge on left (Calkin and Wilkinson, this volume, Stop 2).
23.4	2.2	Turn left and back to Whirlpool State Park, then right into parking lot.

STOP 2. WHIRLPOOL STATE PARK. This again is only a photostop. See Calkin and Wilkinson (this volume, Stop 3) for details.

		Exit right onto Robert Moses Parkway and immediately turn back left going northward to first right turnoff.
24.7	1.3	Turn right off Robert Moses Parkway through parking lot for Devils Hole. See Calkin and Wilkinson (this volume, Stop 4) for details. Proceed left onto Route 104 north past Niagara University.
25.7	0.7	Pass under walkway to Robert Moses Power Plant and Power Vista. See Calkin and Wilkinson (this volume, Stop 5) for details.
26.9	1.5	Cross Barre Moraine at Country Club atop the Niagara Escarpment. View north of Lake Ontario (if clear) across Lake Iroquois plain in foreground below Escarpment.
27.1	0.2	Turn right off Route 104 onto Route 18E (Creek Road) and descend Escarpment.
28.2	1.1	Pass under Route 104 across wave-formed bluff and strand of Lake Iroquois. Proceed north onto Iroquois lake plain on Route 18.
32.1	3.9	Cross Fourmile Creek; in Blairsville, bear right at fork, staying on Route 18. Follow field trip route, referring to Figures 5 through 25.

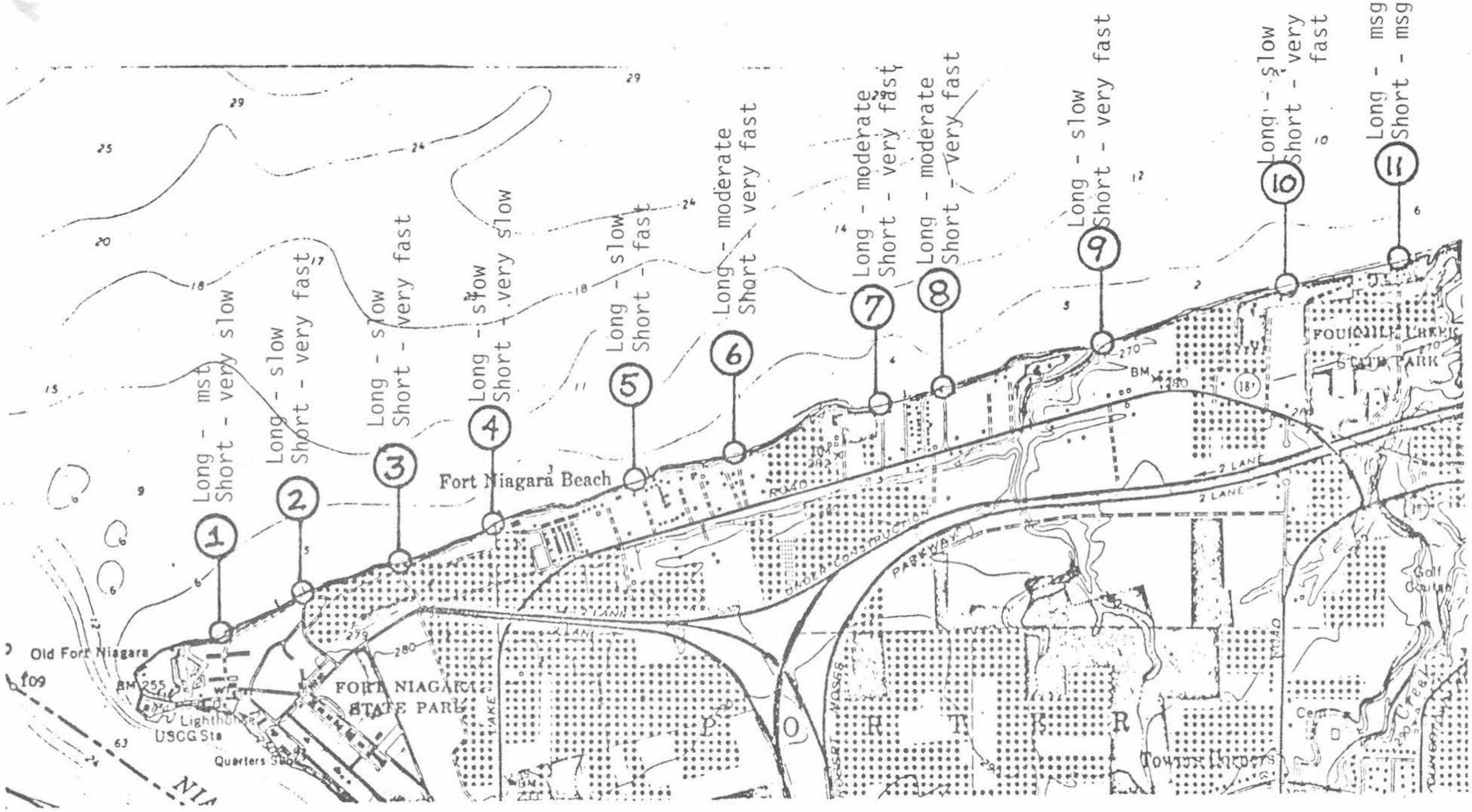


FIGURE 5
Bluff recession rates after Drexhage and Calkin, 1981. Very slow = less than 30 cm/yr; slow 30-60 cm/yr; moderate 60-90 cm/yr; fast 90-120 cm/yr; very fast = more than 120 cm/yr.

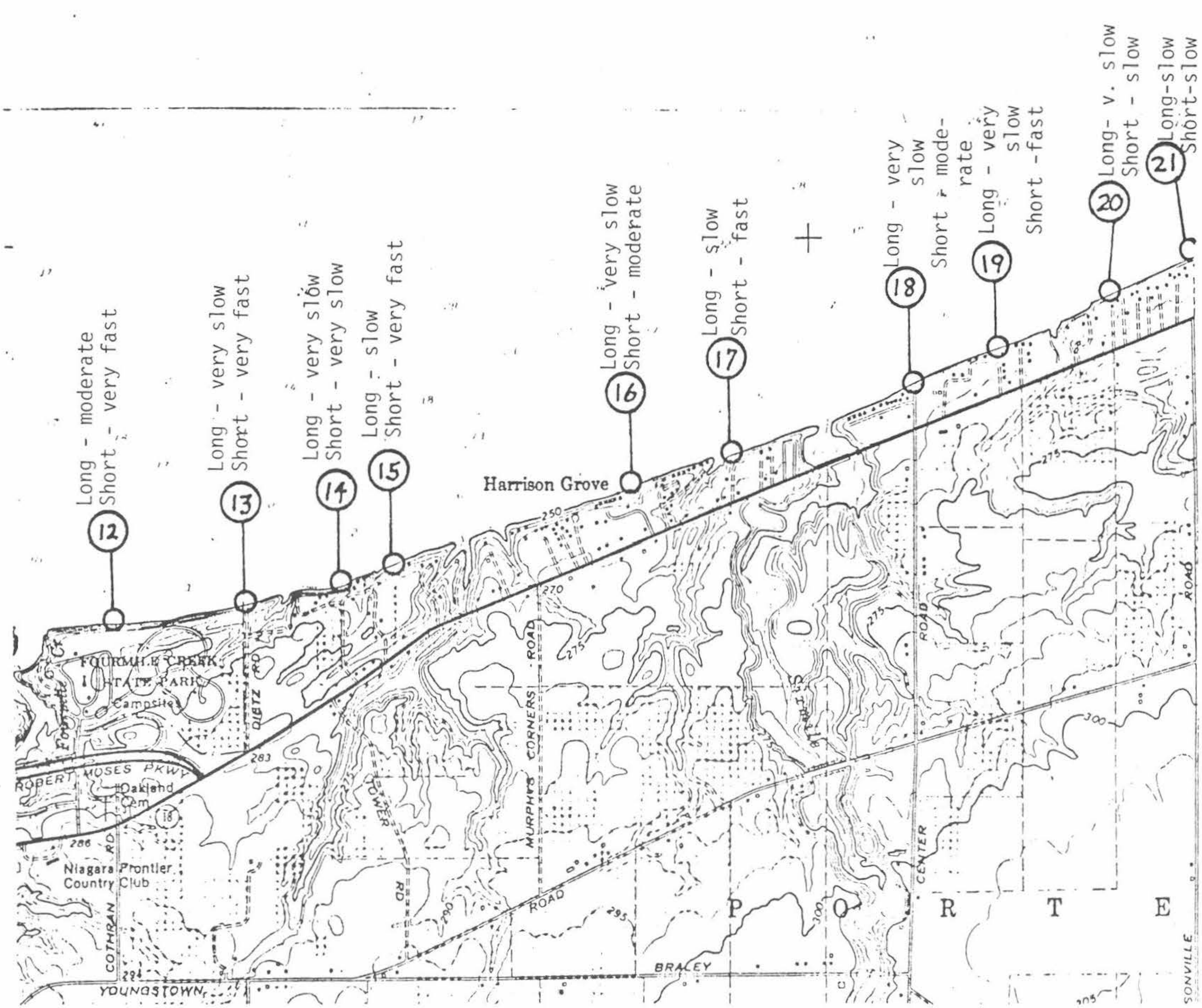


FIGURE 6
Bluff recession rates after Drexhage and Calkin. See Figure 5 for explanation of terms

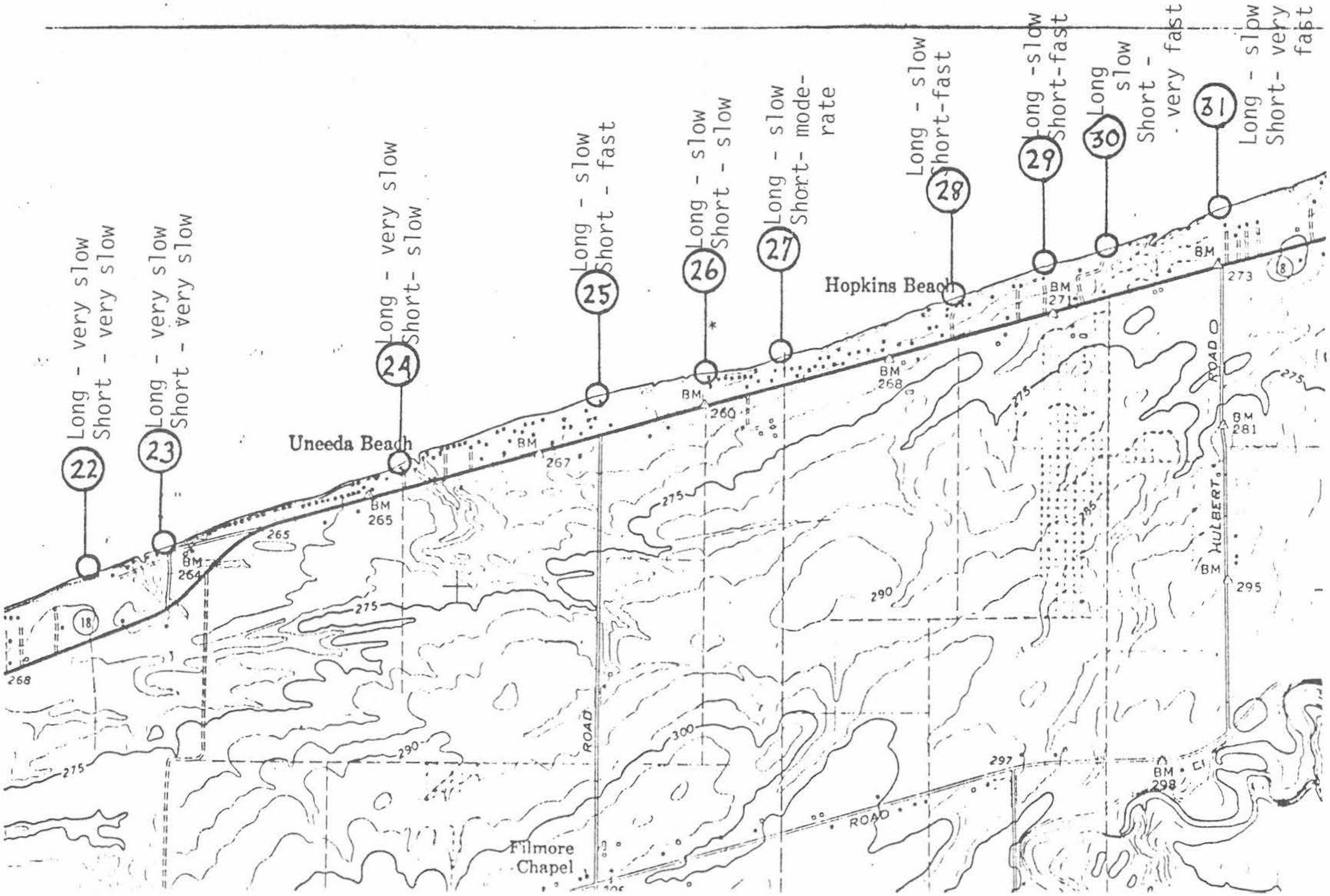


FIGURE 7

.Bluff recession after Drexhage and Calkin (1981). See Figure 5 for explanation of terms.

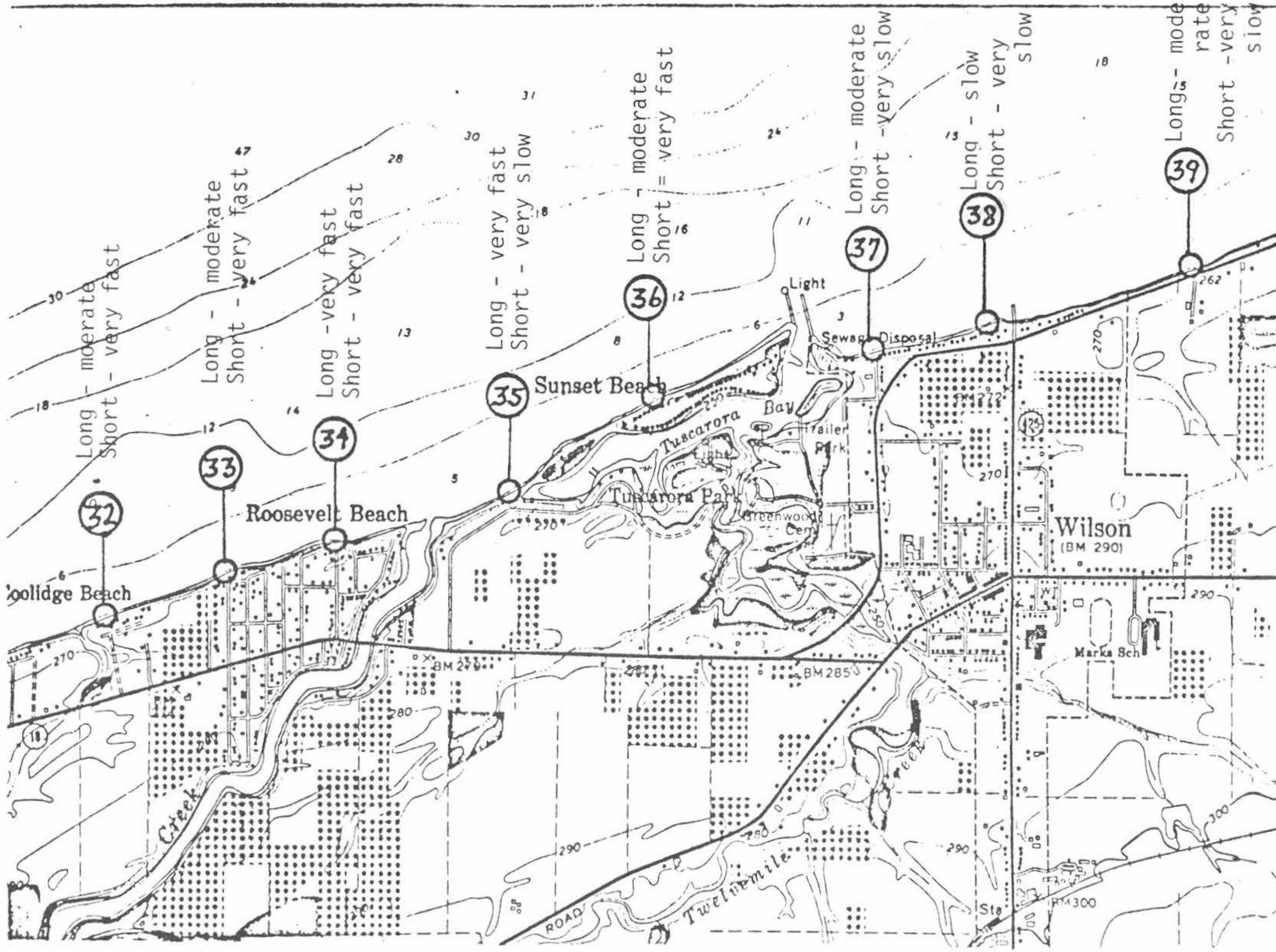


FIGURE 8

Bluff recession after Drexhage and Calkin (1981). See Figure 5 for explanation of terms.

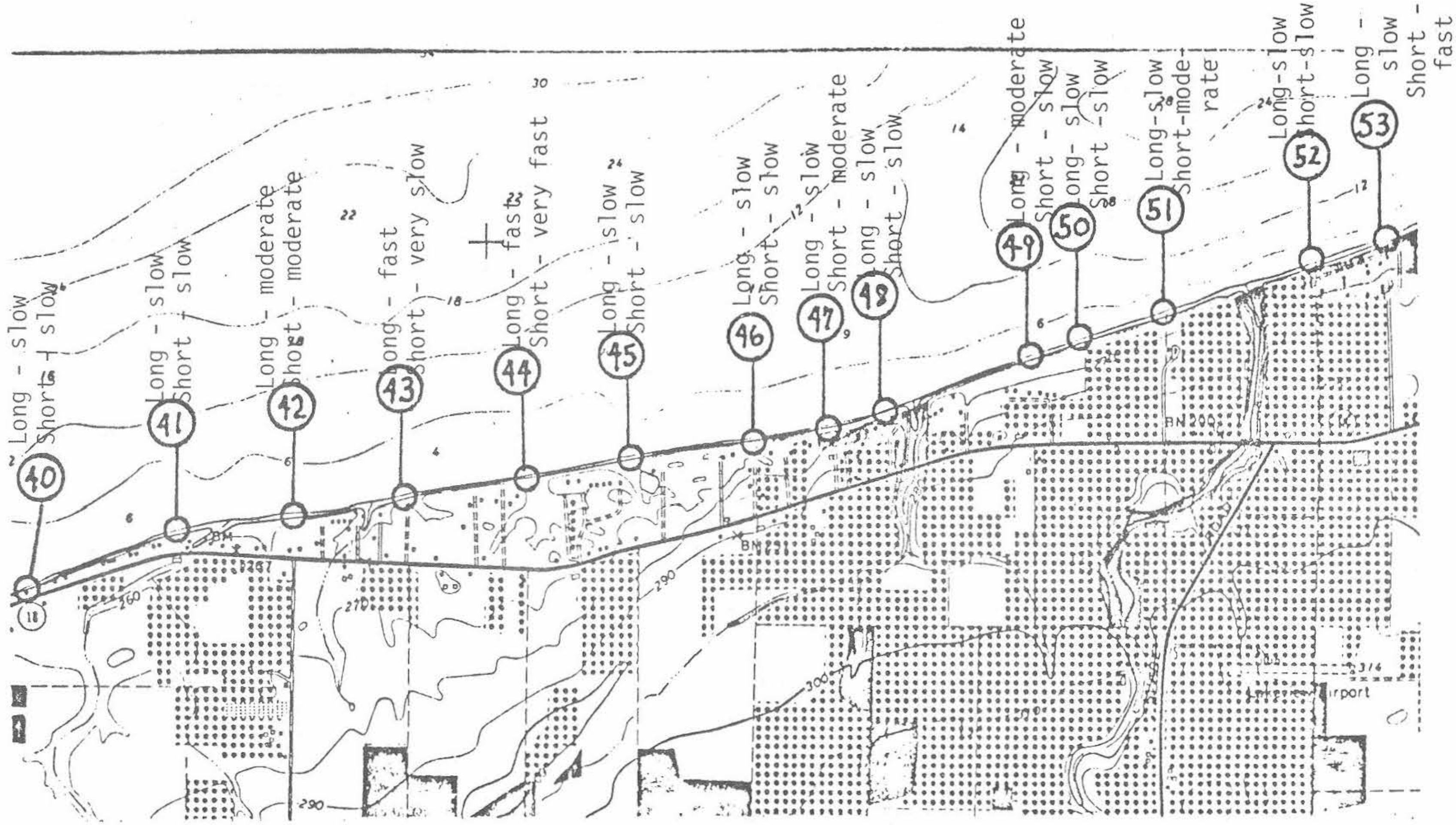


FIGURE 9.

Bluff recession after Drexhage and Calkin (1981). See Figure 5 for explanation of terms.

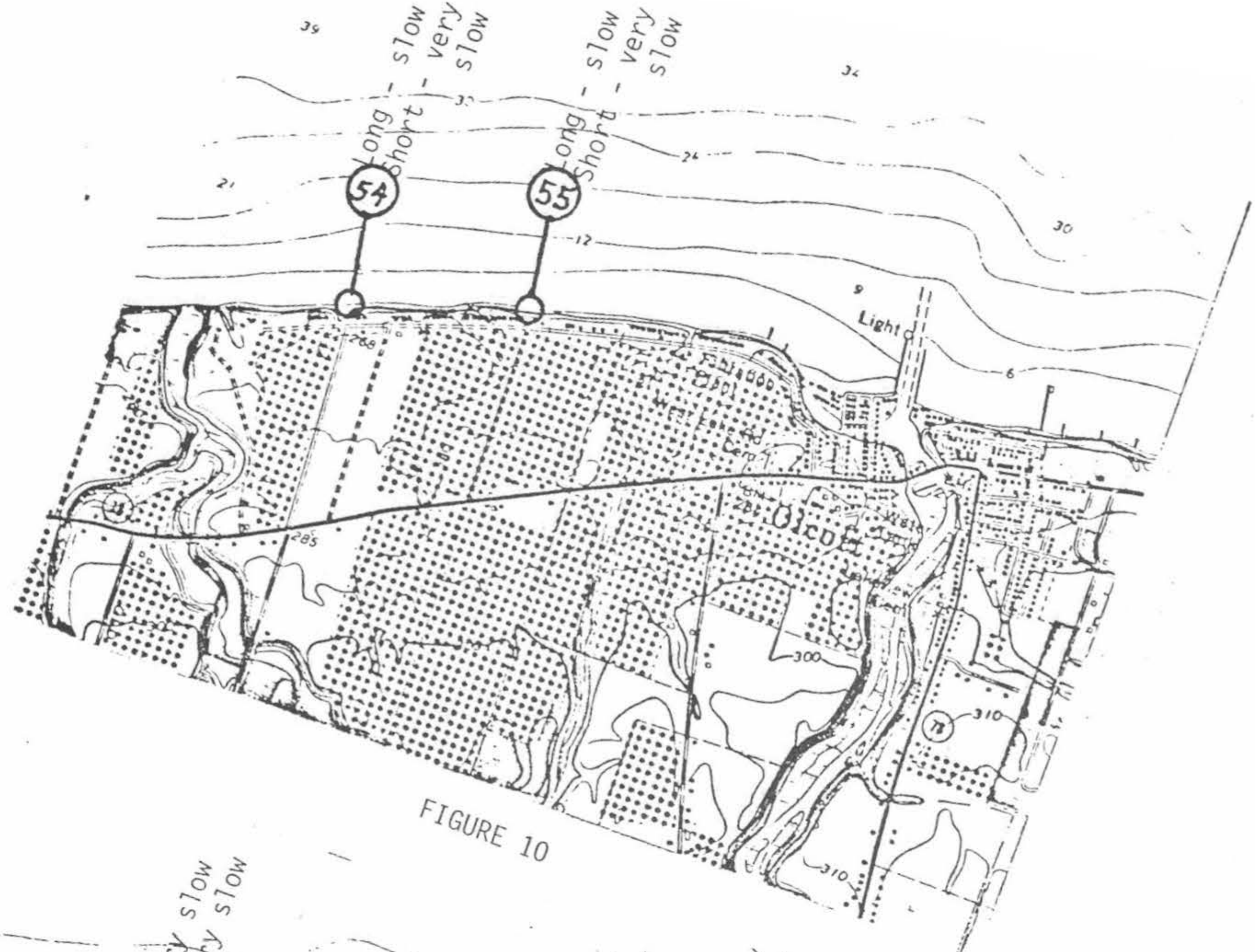


FIGURE 10

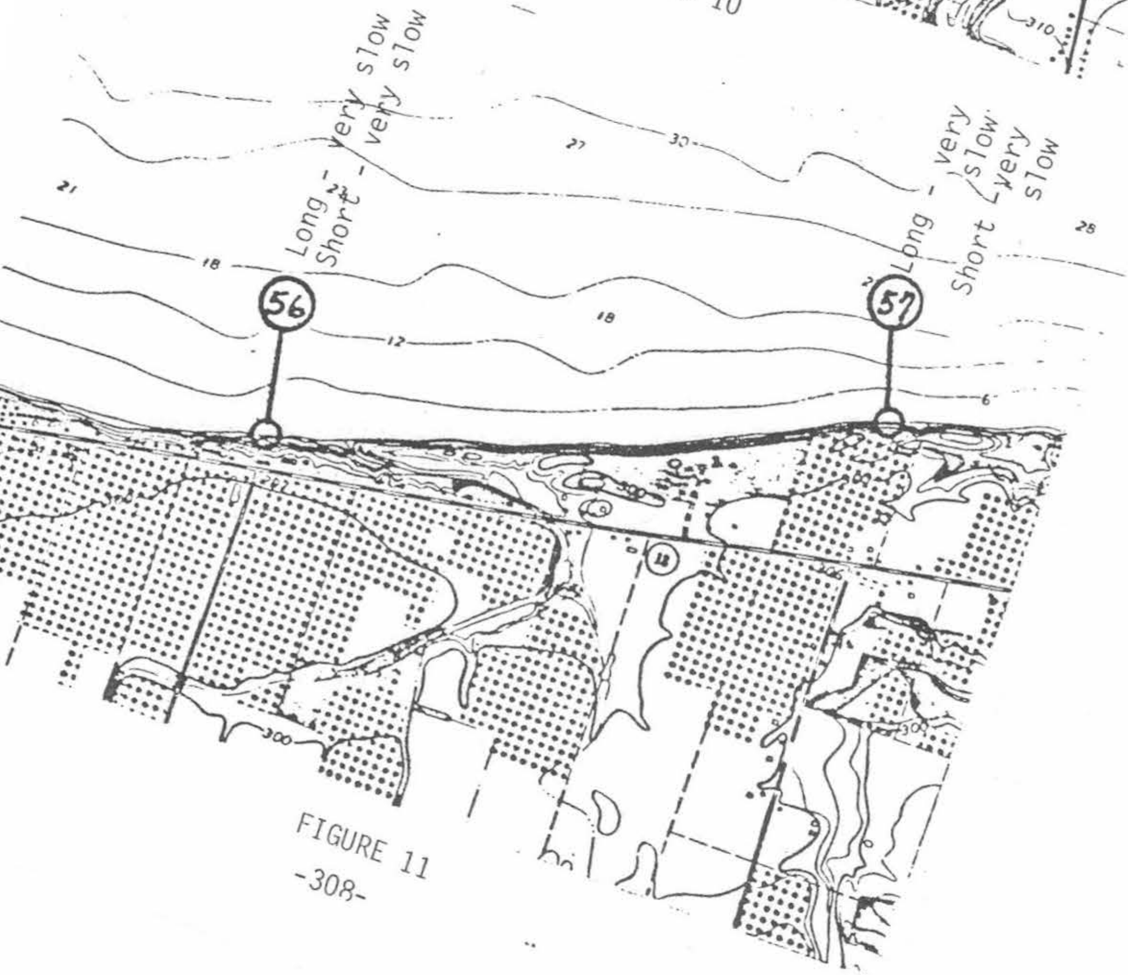
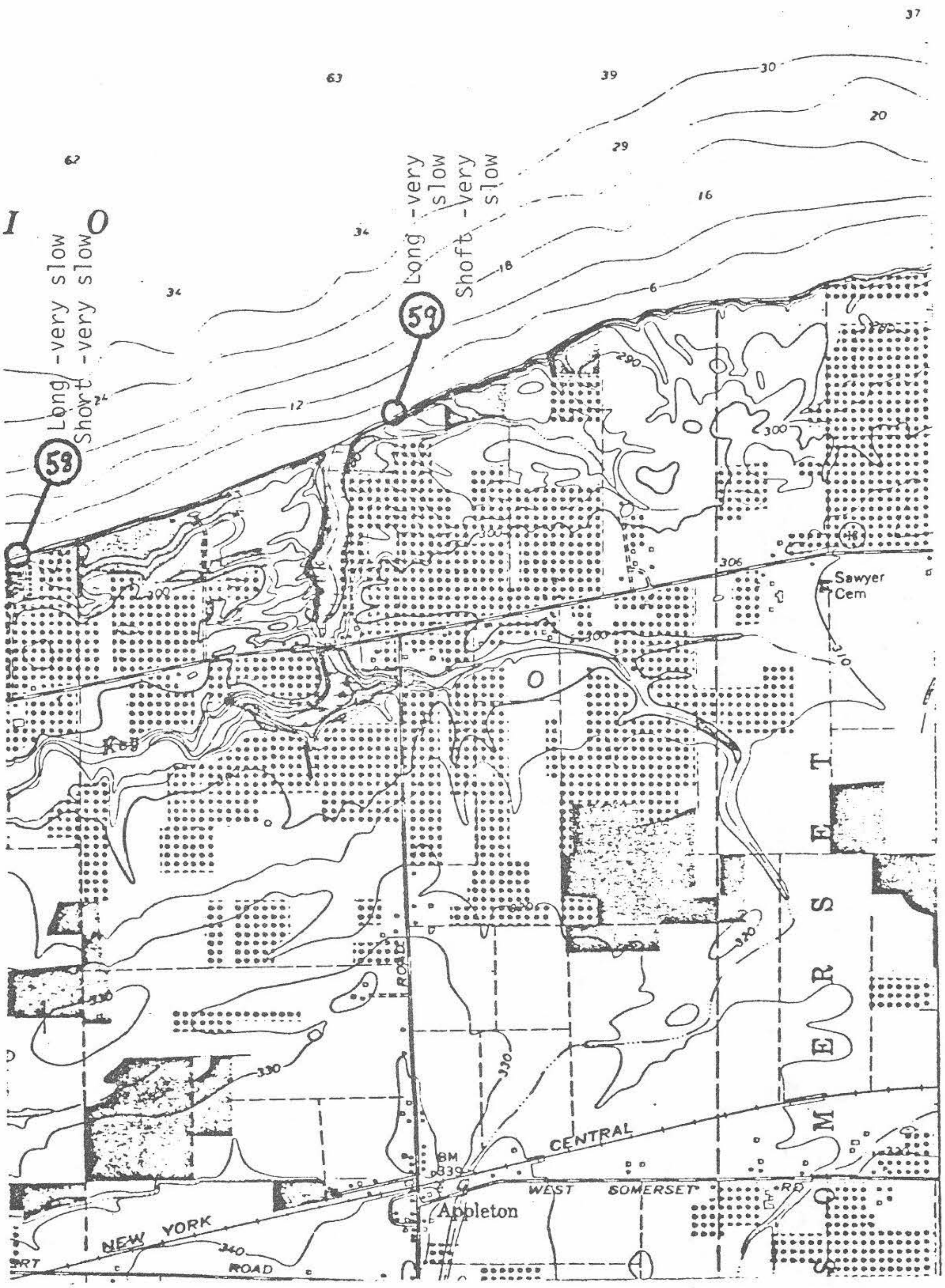


FIGURE 11



Long -very slow
Short -very slow
58
Long -very slow
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Sawyer Cem
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NEW YORK
ROAD
WEST SOMERSET RD
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BM 339
SOMERSET

FIGURE 12

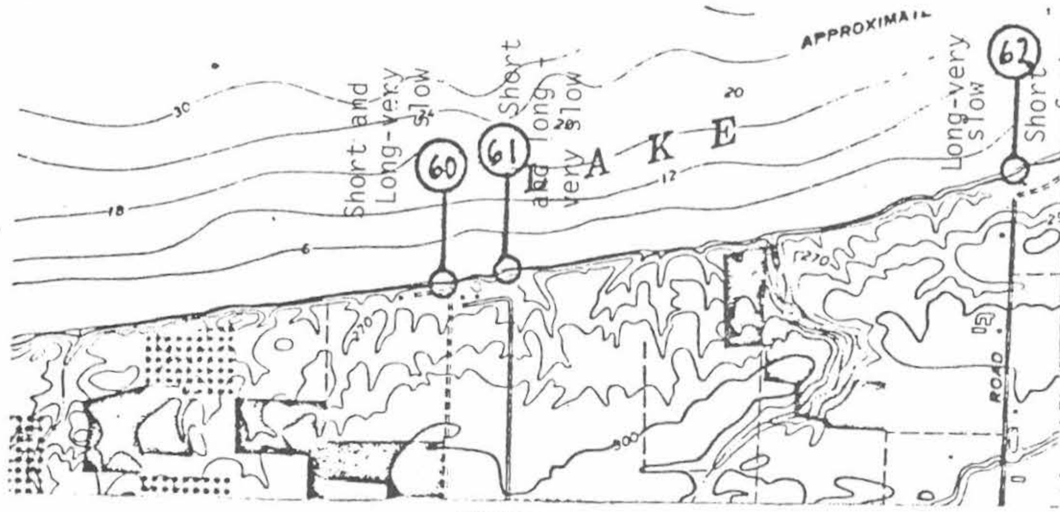

FIGURE 13

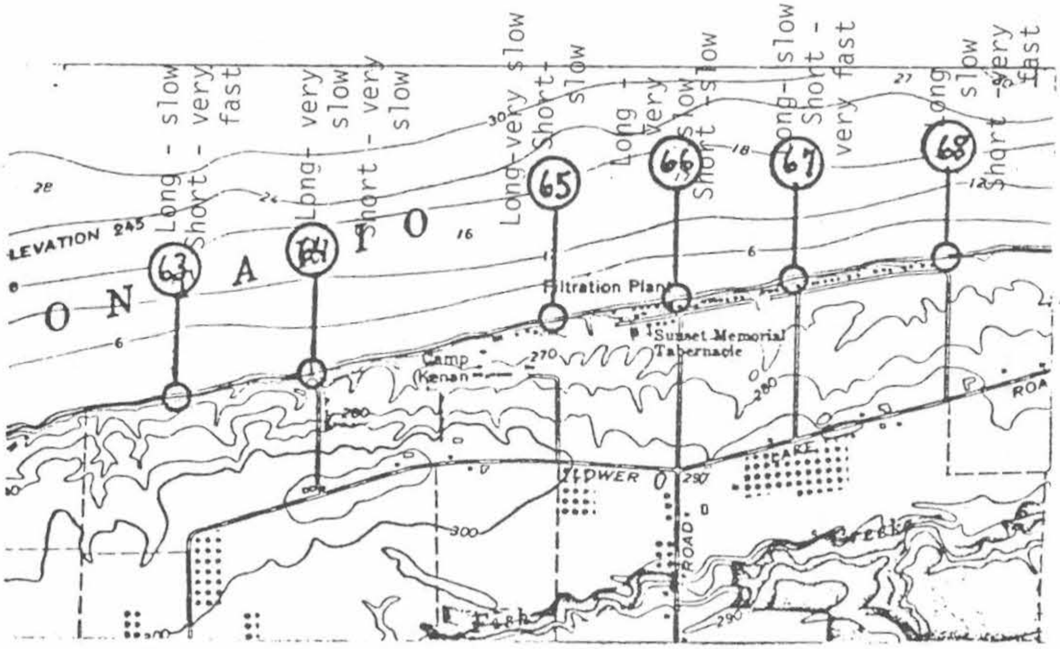

FIGURE 14

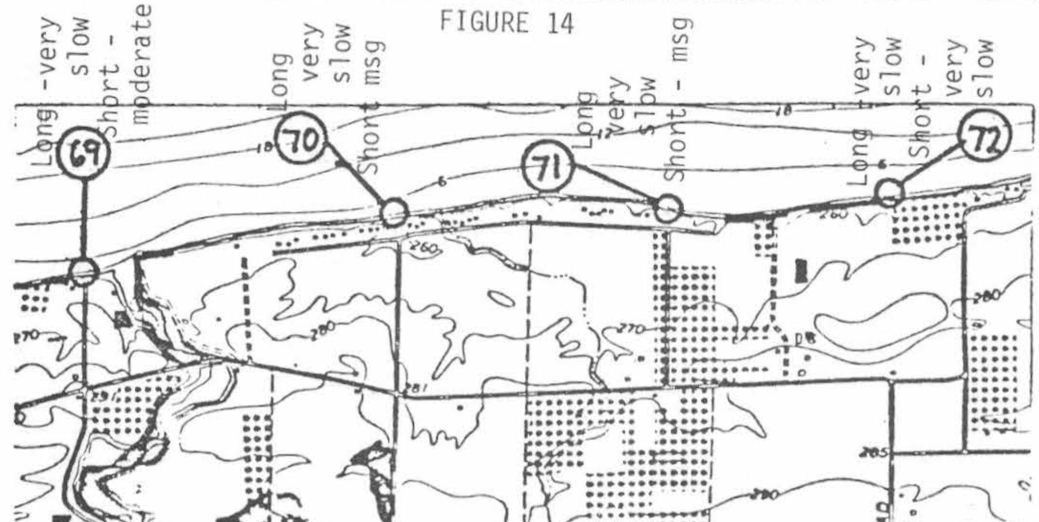

FIGURE 15

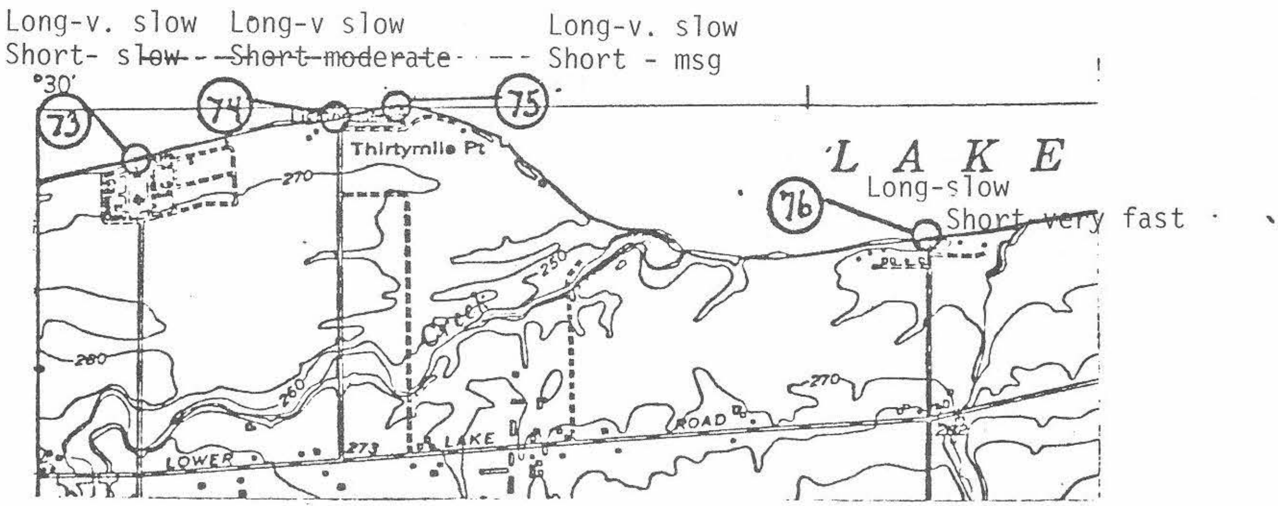

FIGURE 16

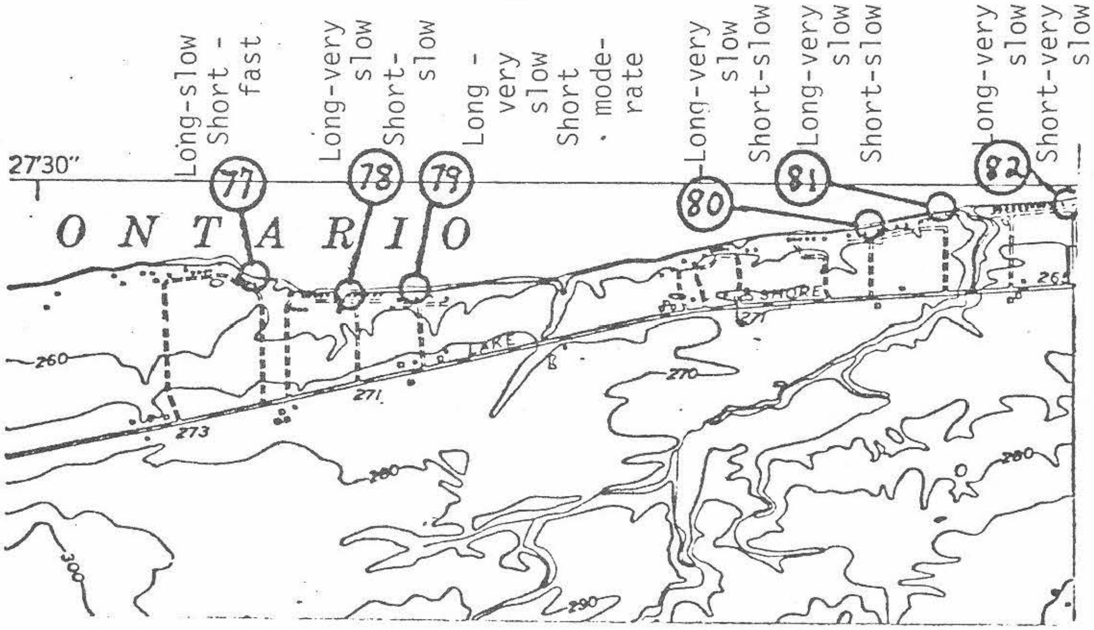

FIGURE 17

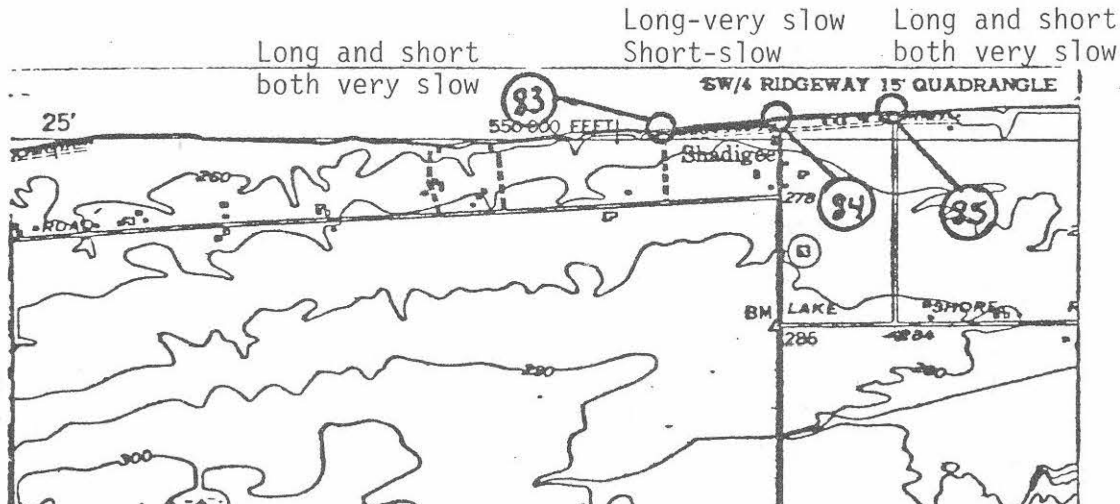

FIGURE 18

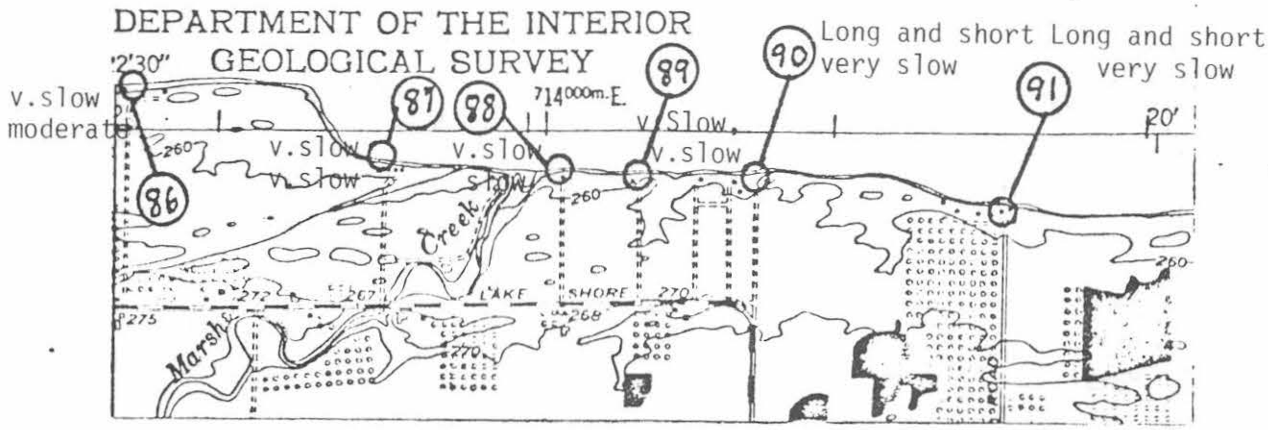

FIGURE 19

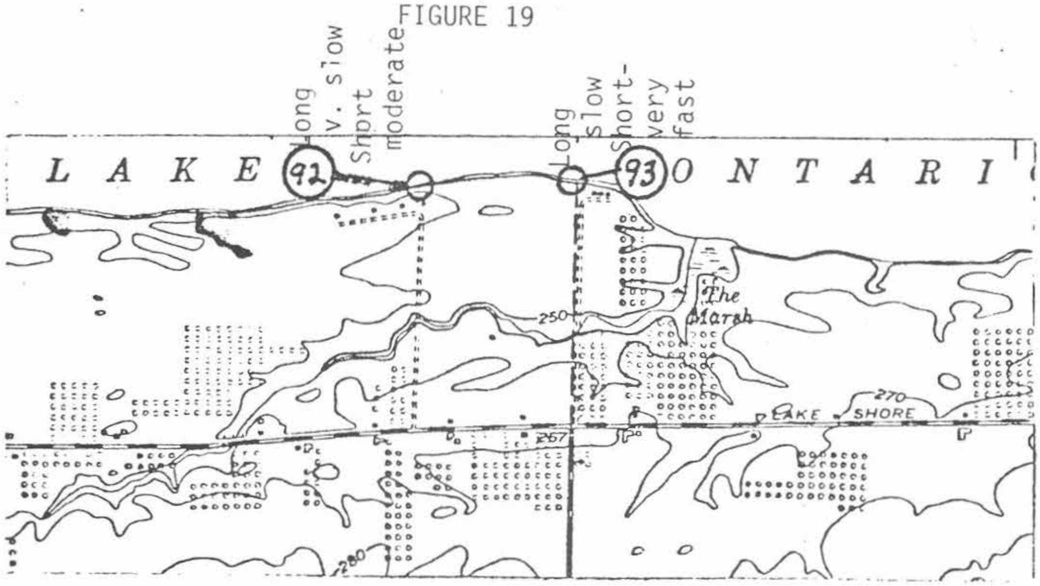

FIGURE 20

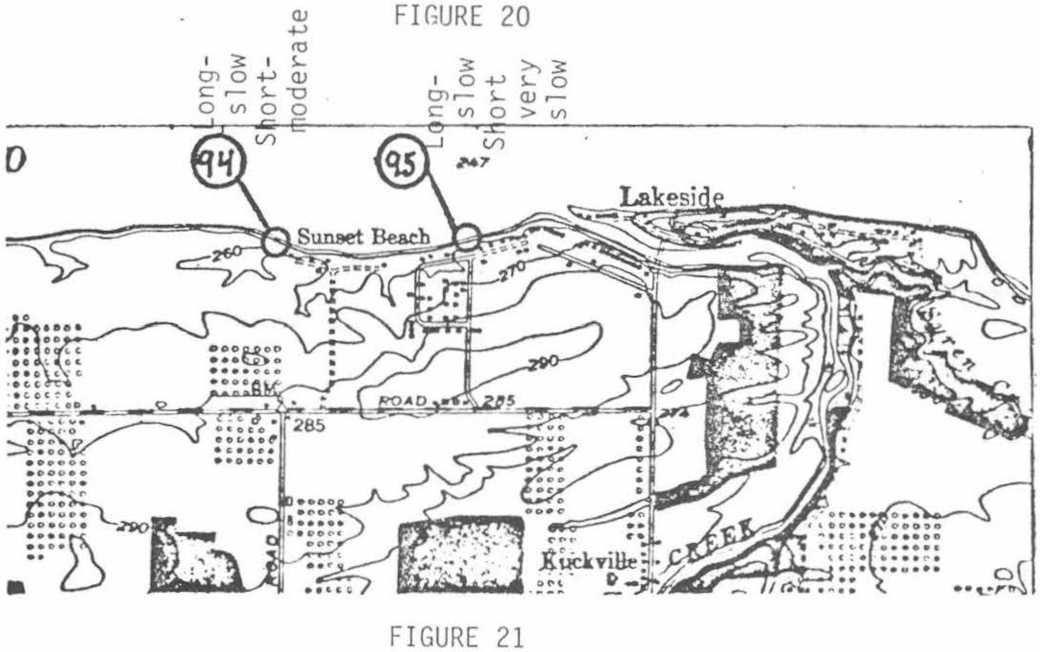

FIGURE 21

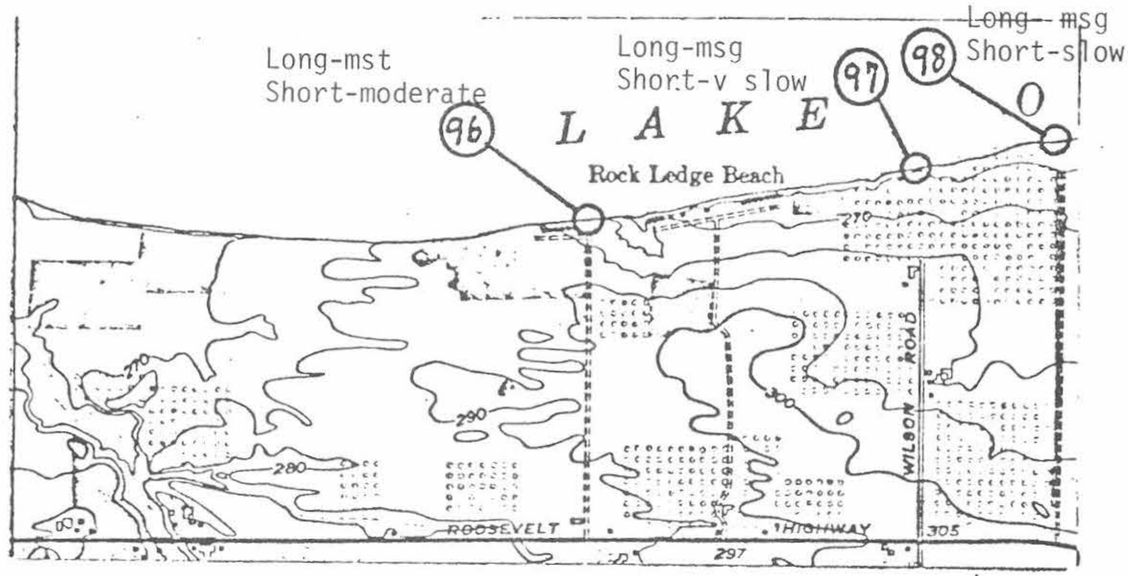

FIGURE 22

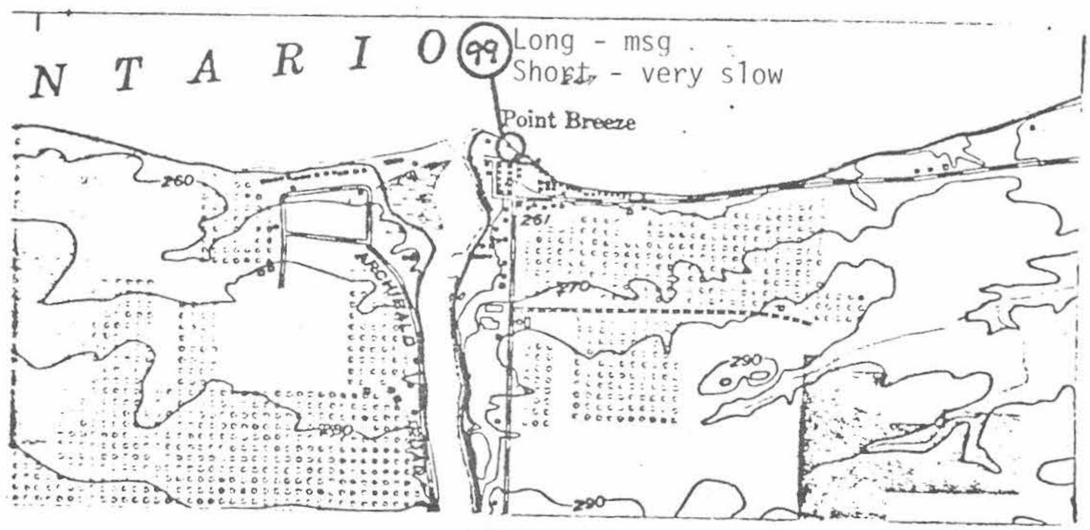

FIGURE 23

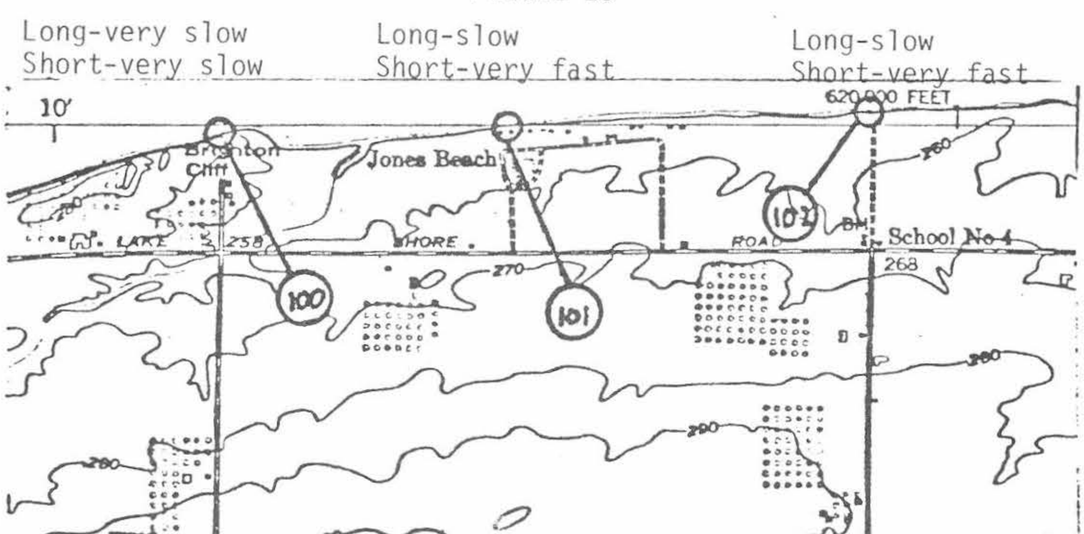

FIGURE 24

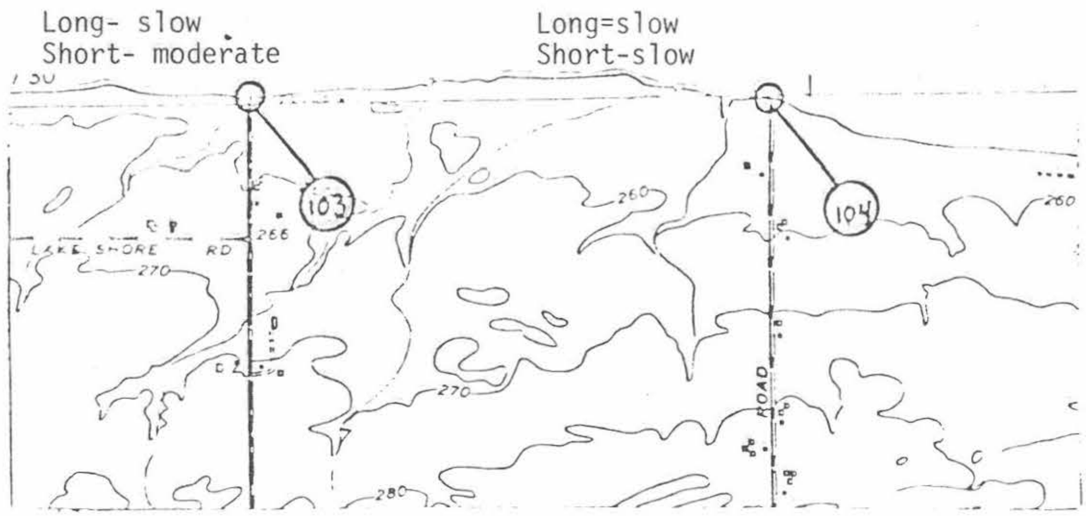

FIGURE 25

Bluff recession rates after Drexhage and Calkin, 1981.

For Figures 5 through 25

Very slow = less than 30 cm/yr
Slow = 30 to 60 cm/yr
Moderate = 60 to 90 cm/yr
Fast = 90 to 120 cm/yr
Very fast = more than 120 cm/yr

Long term rate is the average rate from 1875-1974
Short term rate is the average rate from 1938-1951 for Niagara County and 1938-1954 for Orleans County.

34.0	1.9	Pass on Route 18 east across Fourmile Creek and entry to Fourmile Creek State Campground
37.5	3.5	Cross Sixmile Creek.
38.8	1.3	Turn left and drive on loop bypass to view Lake Ontario and bluff at 23 (Fig. 7). Near-shore gradient is about 8.6 m/km; both short- and long-term recession average under 30 cm/yr. (Drexhage and Calkin, 1981). Return in 0.7 mi turning left east onto Route 18.
44.7	5.9	Turn left at bridge before crossing Twelvemile Creek onto Riverview Road; take second left onto Lakeview Road.
45.1	0.4	Park at boat ramp site enclosed by storm fence and gate. You are 100 m east of Station 34.

STOP 3. ROOSEVELT BEACH. The bluff stratigraphy as exposed in 1975 was:

Top	0.5 m	Lake clay, red, massive, scattered pebbles (Lake Iroquois). Till pockets locally in base.
	2.0 m	Till, red (weathered); sand:silt:clay per cents 41:35:24; incorporates lenses and discontinuous beds of yellow silt or fine sand; cobbly at base.
	2.0	Till, purplish to red (5 YR 4/1); sand:silt:clay per cents 22:57:21; compact (denser than overlying till); clast content 5%.
Water	2.0	Rip rap protecting base of bluff

Approximately 1200 m to the west, the lower compact till is separated from the overlying till by a stone pavement. Gilbert (1898) described a well developed stone pavement exposed in shore bluffs for a half mile east of the mouth of Tuscarora Bay. Marked parallelism of striae on the flat upper surface of 10 of these boulders indicated deposition of the upper till to have been by ice flowing S50W. He interpreted the stone pavement as distinguishing two tills. An alternative explanation recognizes the upper till as a subaqueous allo-till facies; the lower till as lodgement facies representing the same glaciation.

The highest bluff recession rates in Niagara County are measured at Roosevelt Beach (Drexhage and Calkin, 1981). Between 1938 and 1951 the bluff receded at an average rate of 2.1 ft/yr. Prior to 1902, Twelvemile Creek entered Lake Ontario through Tuscarora Bay. In 1902, the receding shore bluffs breached the left bank of Twelvemile Creek. Subsequent sedimentation has built a barrier across the former channel, isolating the mouth of Twelvemile Creek from that of its East Branch and tying the "Island" east of the State Park to the mainland. (See also Brennan, 1979, Fic. 13, Stn. 11, pp 36-37, and Fortune and Calkin, 1981.)

		Turn left onto Hamilton Street, right on Riverview Road and back to Route 18; turn left to the east.
45.3	0.2	Cross West Branch Twelvemile Creek; continue east on Route 18, pass Wilson-Tuscarora State Park entrance.
46.4	1.1	Bear left at fork, continuing on Route 18 through village of Wilson toward Olcott.
48.8	2.4	Note elongate, northeast-trending ridge at right. This is the general trend of ground moraine ridges in this part of the lake plain. Though very subdued, this grain controls orientation of minor drainage lines. In part, at least, it is considered to reflect fluting of underlying bedrock by glacial erosion.
53.1	4.3	Cross Eighteenmile Creek at Village of Olcott at left; cross Route 78 and continue east on Route 18.
54.8	1.7	Turn left off Route 18 opposite creek and park beside Newfane Wastewater Treatment Plant.

STOP 4. OLCOTT BLUFFS. Located between Stations 56 and 57 (Fig. 11). This is the eastern edge of a somewhat uneven silt-topped sand plain that stretches along the coast from Olcott and inland at least one mile to a low ridge, 10 to 20 ft. high and correlable with the Carlton Moraine. Ice-contact features have not been recognized in the thick cross-bedded, subaqueous sands, but their occurrence above abnormally thick clay-silt rhythmites and below fine silt suggests ice marginal oscillation. Figure 26 shows the section as measured 1 km to the west. Seeps near base of the stratified drift induce mass wasting which now largely obscures stratigraphic relationships at this site. Queenston bedrock crops out at bases of bluffs on headlands adjoining this stretch of shoreline. Bluff recession less than 30 cm/yr.

		Proceed east on Route 18
59.1	4.3	Turn left into New York State Electric and Gas Fossil Fuel Generating Plant, Somerset. Permission to enter must be obtained at gate.

STOP 5. SOMERSET BLUFFS. Located between Stations 60 and 61. In 1974, the wave-eroded bluffs here recorded two tills, a red stony till overlain by purplish-gray silty till. This section is also revealed in borings for the facility. The following section was measured in 1975:

Top	0.4 m	Lake clay, silty, oxidized
	0.55 m	Lake silt, yellow; contains convoluted beds and dropstones.

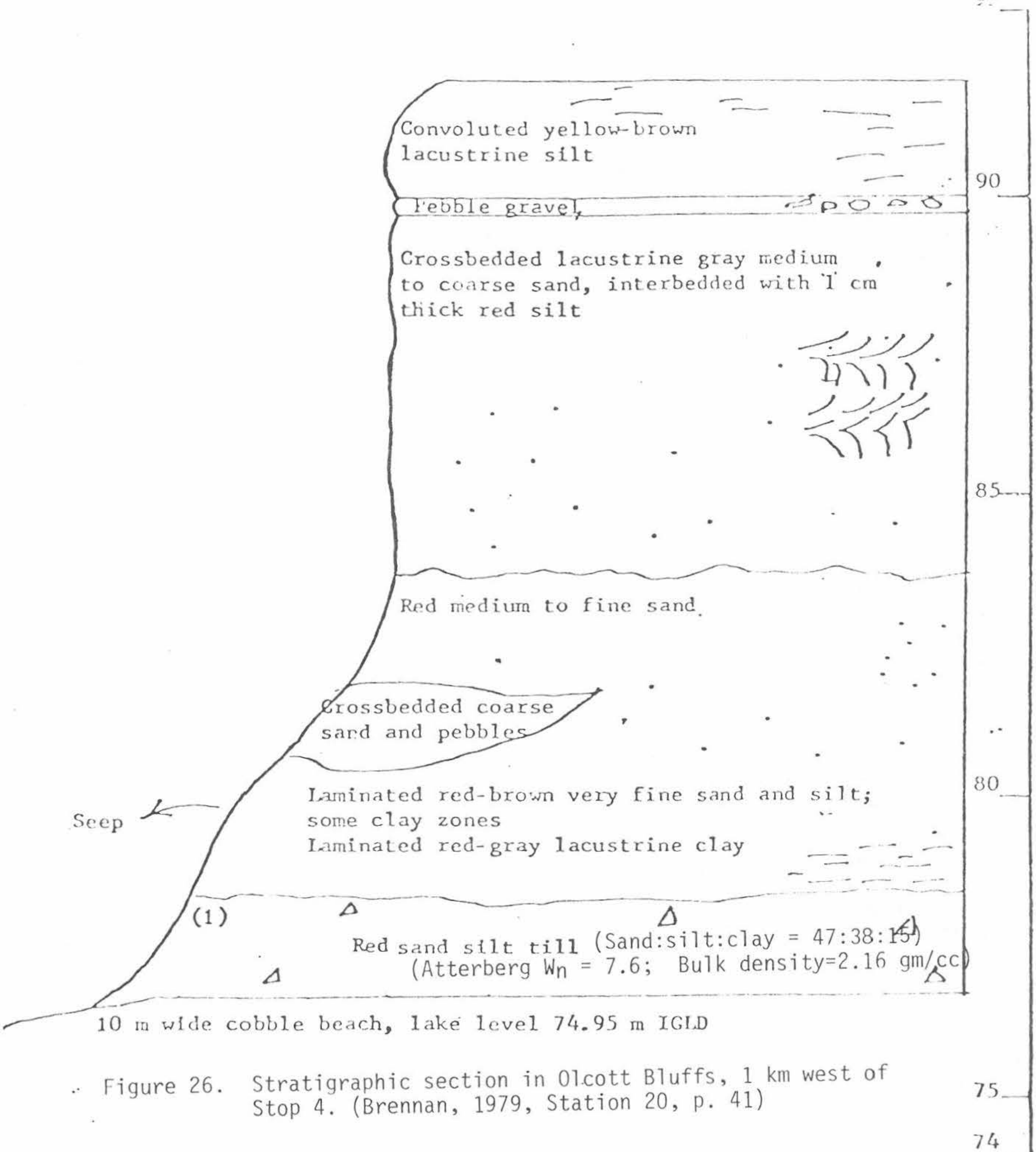

Figure 26. Stratigraphic section in Olcott Bluffs, 1 km west of Stop 4. (Brennan, 1979, Station 20, p. 41)

	1.1 m	Till, interbedded yellow sand, locally gravelly; subaqueous allo-till (?)
	2.1 m	Till, purple-red, 10% stones, compact; lodgement and subaqueous till.
	0.5 m	Queenston bedrock

As at Stop 3, the question is raised whether the two till units can be assigned to a single glaciation as different depositional facies of the same till, or whether two glaciations are represented. Salomon (1976) at a nearby exposure reported 4 ft of silt, clay and sand between the two tills. Both short- and long-term bluff recession rates are "very slow", i.e. less than 30 cm/yr (Drexhage and Calkin, 1981).

		Continue east on Route 18
60.4	1.3	Turn left (north) on Hartland Road and follow around bend east onto Lower Lake Road.
65.8	0.9	Turn left into Golden Hill State Camp Site and Thirtymile Point. Proceed to parking location immediately east of lighthouse.

STOP 8. THIRTY MILE POINT. This is Station 75 (Fig. 16). Bedrock reaches its greatest elevation above lake level in shorebluff of the south coast of Lake Ontario at Thirty Mile Point. Lake silts and sandy silt till overlie some 3 to 4 m of Queenston bedrock.

Of particular interest is the small, fractured bedrock anticline with northwest-southeast trend paralleling the coastline. As described by Gilbert (1899), the crest was broken by a steeply northeasterly dipping fracture, the northeast side displaced upward 2m with overturning of fold axis to the southwest, enclosing till. Although continuing below lake level, the deformation seems to die out downward. Evidence suggests glaciotectonic thrusting by southwesterly moving ice following initial till deposition (Kindle and Taylor, 1913).

Gilbert wrote "Should this description rouse enough interest to induce others to examine the locality, their visits should not be long delayed. This part of the coast is specially exposed to the attack of storm waves and is rapidly beaten back." Eighty years later, however, the exposure appears much as illustrated by Gilbert with apparent bluff recession of no more than 2-3 m in that interval.

67.3	1.5	Leave Park and proceed east on Lower Lake Road.
68.5	1.2	Turn right (south) at first road, County Line Road.
69.8	1.3	Turn left (east) onto Route 18, Roosevelt Highway .
73.6	3.8	Cross Route 63, Lyndonville Road.

77.7	4.1	Note very gently hummocky topography of Carlton Moraine trending parallel to route.
83.6	5.9	Turn sharp left off Route 18 at three-pronged intersection and cross Orchard Creek on Marsh Road. Take first left onto Route 98N; pass under Lake Ontario State Parkway to Point Breeze. Orchard Creek has been incised deeply into bedrock by early postglacial stream erosion. Like other creeks on this south coast of Lake Ontario, its gradient has been diminished and its debouchment into Lake Ontario drowned by postglacial tilting. Shore bluff retreat partly counters the estuary development that tilting has produced.
85.0	1.4	Turn right (east) onto Lake Shore Road at Point Breeze.
86.1	1.1	Park along road adjacent to Station 99.

STOP 7. BRIGHTON CLIFF. Bouldery to sandy supraglacial drift correlated with ridges of the Carlton Moraine 1 to 2 mi south. Underlying weakly bedded till overlies compact till similar to the main till unit of the Niagara River to Rochester reach of the Ontario coastline. The stratigraphy in the bluff is depicted in Figure 27 from Brennan (1979)."At the top of the till, a deformed boulder layer lies on a distinct erosion surface; above the boulder layer pebbly sand grades upwards to fine sand and silt". (Brennan, 1979, p. 47). Pods of till within the sand and small-scale normal faults further suggest ice contact deposition. Features like the fault in Figure 27b may represent deformation during the Carlton glaciation.

		Retrace route westward along Lake Shore Road to Point Breeze and follow Route 98 south.
87.7	1.6	Pass southward under Lake Ontario Parkway.
88.5	0.8	Cross Orchard Creek on Route 98 and proceed through type area of Carlton Moraine at the railroad tracks.
95.9	4.5	Cross Lake Iroquois beach ridge at Route 104 intersection.
98.0	2.1	Albion Village limit; continue south, crossing New York Barge Canal.
99.1	1.1	Cross Route 31 and pass over Albion Moraine
101.0	1.9	Turn right (west) off Route 98 onto Route 31A (West Lee Street). Cross esker.
106.3	5.3	Turn left (south) onto Eagle Harbor Road. Note esker ridge angling obliquely toward road.

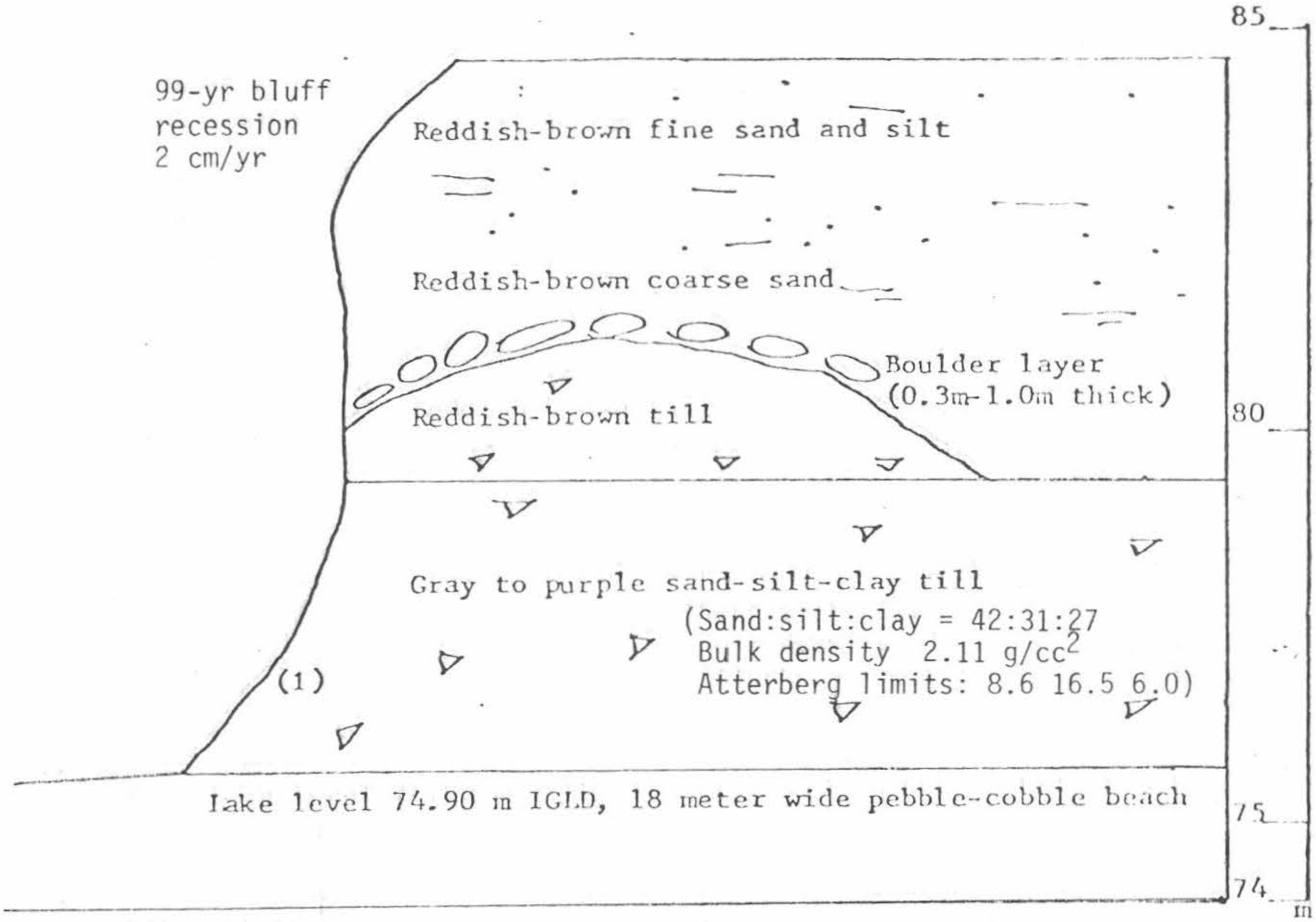

Fig. 27A. Bluff section near Brighton (after Brennan, 1979).

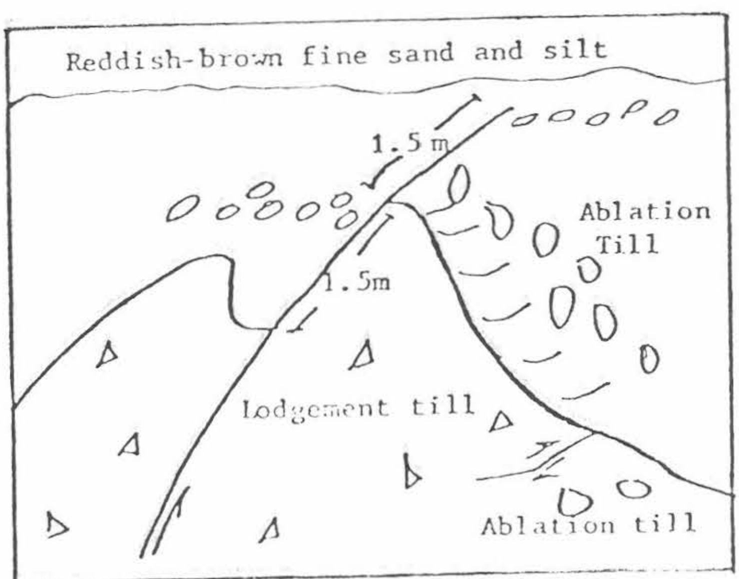

Fig. 27B. Near Brighton. Structure in till showing lodgement and ablation facies. (after Brennan, 1979).

106.8	0.5	Bear right at fork onto Kams Road and along esker ridge.
107.9	0.8	Turn right onto Maple Street across the Burma Woods kames.
108.0	0.4	Turn left onto Pine H ill Road and proceed south over Burma Woods kames toward intersection with esker.
109.4	1.4	Turn off road into gravel pit.

STOP 8. BURMA WOODS ESKER/KAME COMPLEX. The extensive area of glacier disintegration topography traversed for the past three miles developed during wastage of the glacier front from the Barre Moraine, about .5 miles south of this point. The stagnating ice margin apparently fronted on Lake Lundy (Lake Dana (?)) at a stage when it was too shallow to float large ice bergs. The Burma Woods Esker can be readily traced for 4 mi (6.4 km) and conceivably extends all the way to the Albion Moraine, making it perhaps the longest esker in western New York. It consists of a number of discrete segments, the last to be deposited being at the north end. One of the segments terminates in a kame delta at about 735 ft. suggesting the level of Lake Lundy into which it was deposited. Distinct deltaic foreset bedding can be observed in this pit. See Figure 28 for location.

		Leave pit, continuing south
109.6	0.2	Turn right on Gray Road
110.3	0.7	Turn left onto Hemlock Ridge Road. This is the frontal ridge of the Barre Moraine. Proceed east through West Barre on Root Road (continuation of Hemlock Ridge Road) and cross lake flats with several closely spaced drumlin ridges oriented northeast-southwest.
115.0	4.7	Turn right (south) onto Quaker Hill Road which leads to Route 98 southbound. Cross eastern end of Oak Orchard Creek headwaters and surrounding swamp, as well as several southwest-oriented drumlins.
119.4	5.4	Elba Town Line -- the "onion capital of N.Y.S."
122.3	2.9	Cross Onondaga Escarpment and rise onto Onondaga Bench. The Batavia Moraine crosses Route 98 at top of the scarp.
123.9	1.6	Turn left onto New York Thruway and west towards Buffalo.

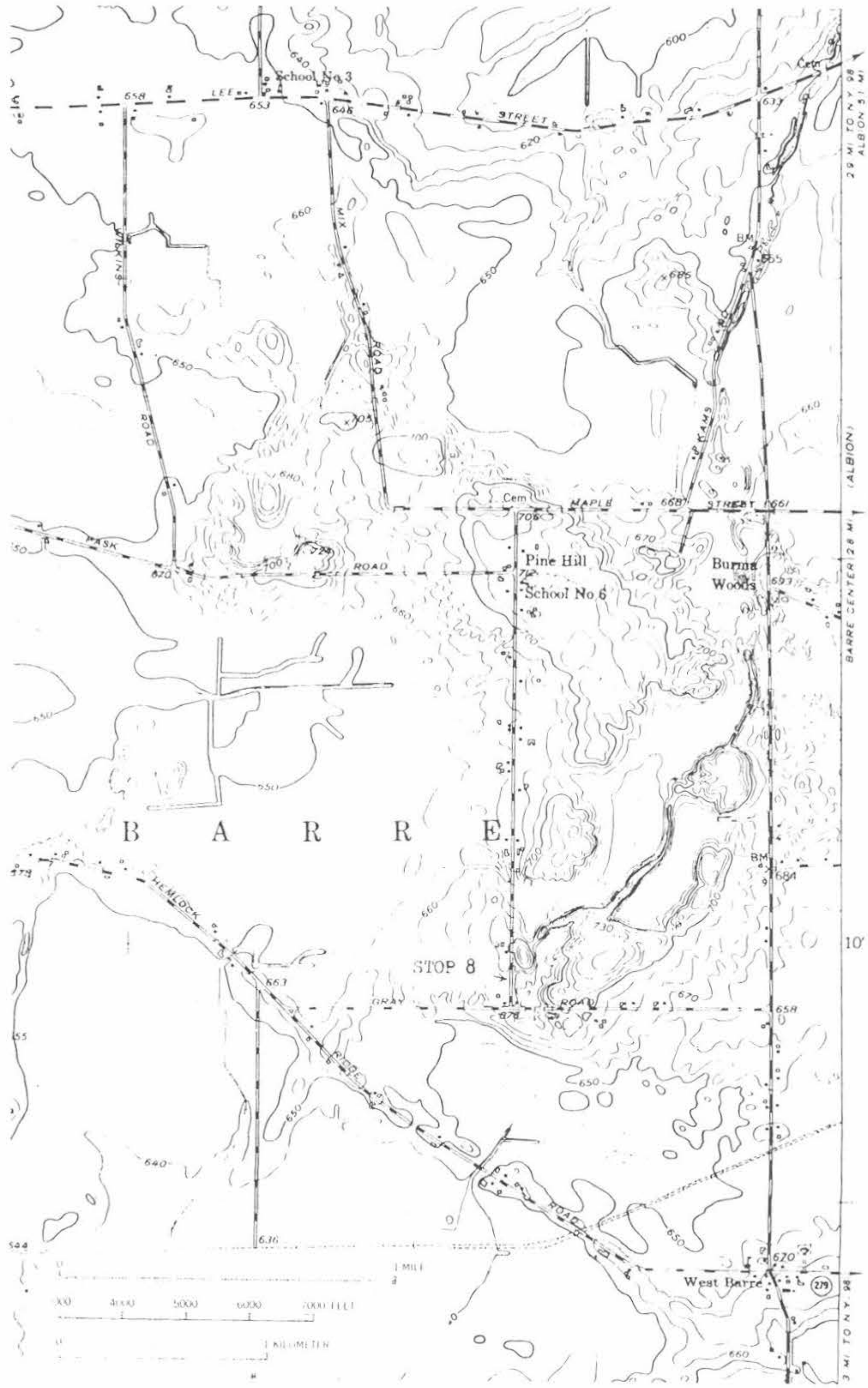

Figure 28. A portion of the Knowlesville, N.Y. 75' Quadrangle showing the Burma Woods esker/kame complex and the Barre Moraine (Hemlock Ridge Road).

124.2	0.3	Pass west on New York Thruway back under Route 98. Gravel kame topography correlated with the Buffalo Moraine trends northwest away from Thruway on right.
129.4	5.2	Cross through partly quarried-out kames related to the Buffalo Moraine.
143.1	13.7	Pass gravel pits in thick gravel associated with Buffalo Moraine (Calkin, this volume, Figure 2).
147.9	4.7	Quarry on right next mile in Onondaga Limestone
153.1	5.2	N.Y.S. Thruway Toll barrier. Take first exit right onto Route I-290 after leaving barrier. On I-290 move promptly into middle lane.
154.1	1.0	Cross Onondaga Limestone Escarpment and under Route 5 (Main Street).
156.3	2.2	Exit right (north) onto Millersport Highway
157.0	0.6	Turn left into Marriott Hotel parking lot.
		HAVE A SAFE TRIP HOME !!!

SEDIMENTOLOGIC AND GEOMORPHIC PROCESSES AND EVOLUTION OF BUTTERMILK VALLEY, WEST VALLEY, NY

JON C. BOOTHROYD Department of Geology, University of Rhode Island. Kingston, RI 02881
BARRY S. TIMSON Earth Surface Research, Inc., 114 State St., Augusta, ME 04330
LORIE A. DUNNE New York State Geological Survey, Albany, NY 12230

INTRODUCTION

The purpose of this trip is to investigate the sedimentologic and geomorphic processes active in a small non-glacial gravel stream and on adjacent valley walls and tributaries. The work that led to this field trip is part of a larger geologic and hydrologic study by the New York State Geological Survey of the low-level nuclear waste-disposal site and other use areas of the West Valley Nuclear Service Center. The geomorphic study is being done to determine, as accurately as possible, the denudation rate in the Buttermilk drainage basin, and to estimate rate and magnitude of morphologic changes to the waste-burial site.

Buttermilk Creek, located in southwestern New York (Fig. 1), has a drainage basin area of 78.4 km^2 and is a tributary of Cattaraugus Creek. Figure 2 shows detail of that part of the drainage basin adjacent to the West Valley Nuclear Service Center (WVNSC).

GENERAL GEOLOGY

Surficial Geology

The area of interest is in the Ashford Hollow and West Valley 7½ minute quadrangles. The regional surficial geology has been mapped by Muller (1977) and discussed by Coates (1976). Wisconsinan glacial features and later Holocene modifications of the 2 quadrangles have been mapped in great detail by Lafleur (1979) and discussed by him on the 1980 Friends of the Pleistocene field trip (Lafleur, 1980). An upper plateau containing the waste-burial trenches, is underlain by till capped in places by a thin layer of fluvial gravel that marks the original fluvial surface immediately following the late Woodfordian deglaciation. Movement of groundwater in the vicinity of the waste-burial trenches has been investigated by Prudic (1979) and Prudic and Randall (1979). Mapping by Boothroyd et al. (1979, 1981) has further refined Lafleur's units in the vicinity of the WVNSC. There are eight till units, identified by age and present steepness of slope, and ten fluvial/alluvial fan systems ranging from the presently-active bar and channel system to late Woodfordian proglacial channels.

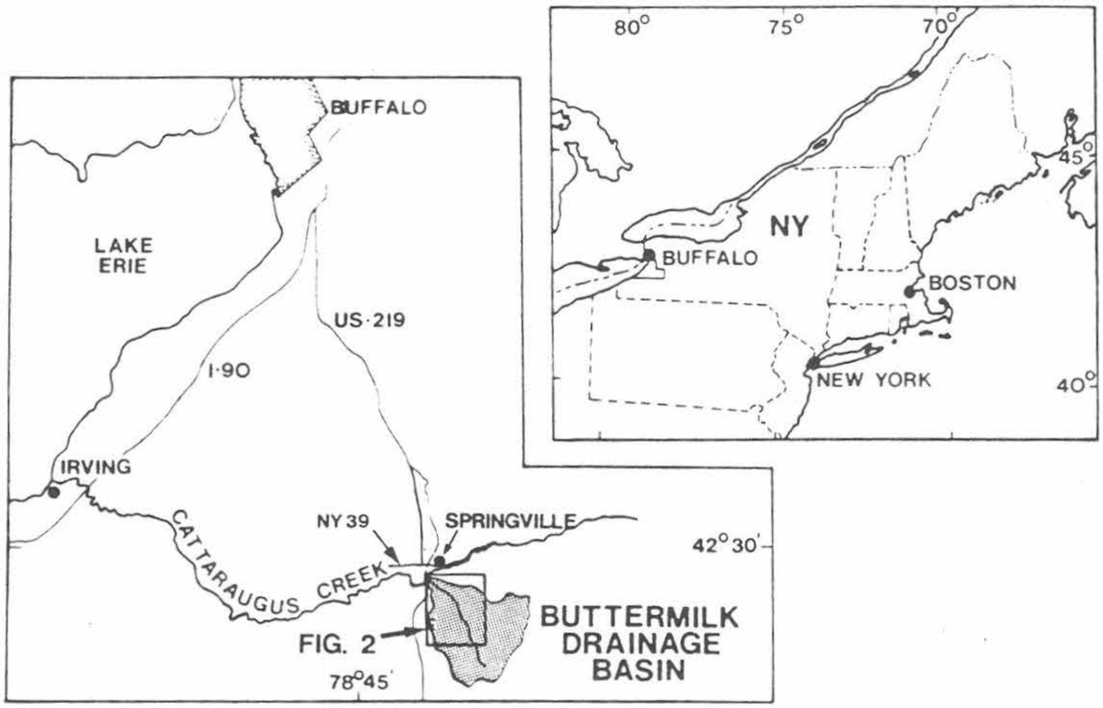

Figure 1. Location map.

Bar and Channel Pattern

At present, Buttermilk Creek and tributaries are incised up to 50 m into compact Lavery till. The basic channel pattern of Buttermilk Creek is an entrenched meander system when the active, unvegetated bars and the low-stage thalweg are considered together. However, most of the meandering appearance is a valley-wall feature with a secondary low-stage thalweg pattern that is usually in phase with the valley wall meanders. The meander wavelength of the valley wall (500+ meters) is many times longer than the wavelength of the low-stage thalweg as indicated by shorter-term fluctuations (years) on channel-sweep diagrams constructed from sequential aerial photographs. The low-stage thalweg is bent around or cuts through bars, and thus is controlled in part by bar movement. Many inactive channels that have been cut off by sudden channel switching or by having chute heads filled by bar gravel.

Gradient and Clast Size

The mean gradient of Buttermilk Creek is 6.7 m km^{-1}, as measured along the low-stage thalweg. The longitudinal profile is segmented of varying spacing or wavelength (Fig. 3). The shortest segments, 50 to 100 m in length, are pool and riffle sequences developed over

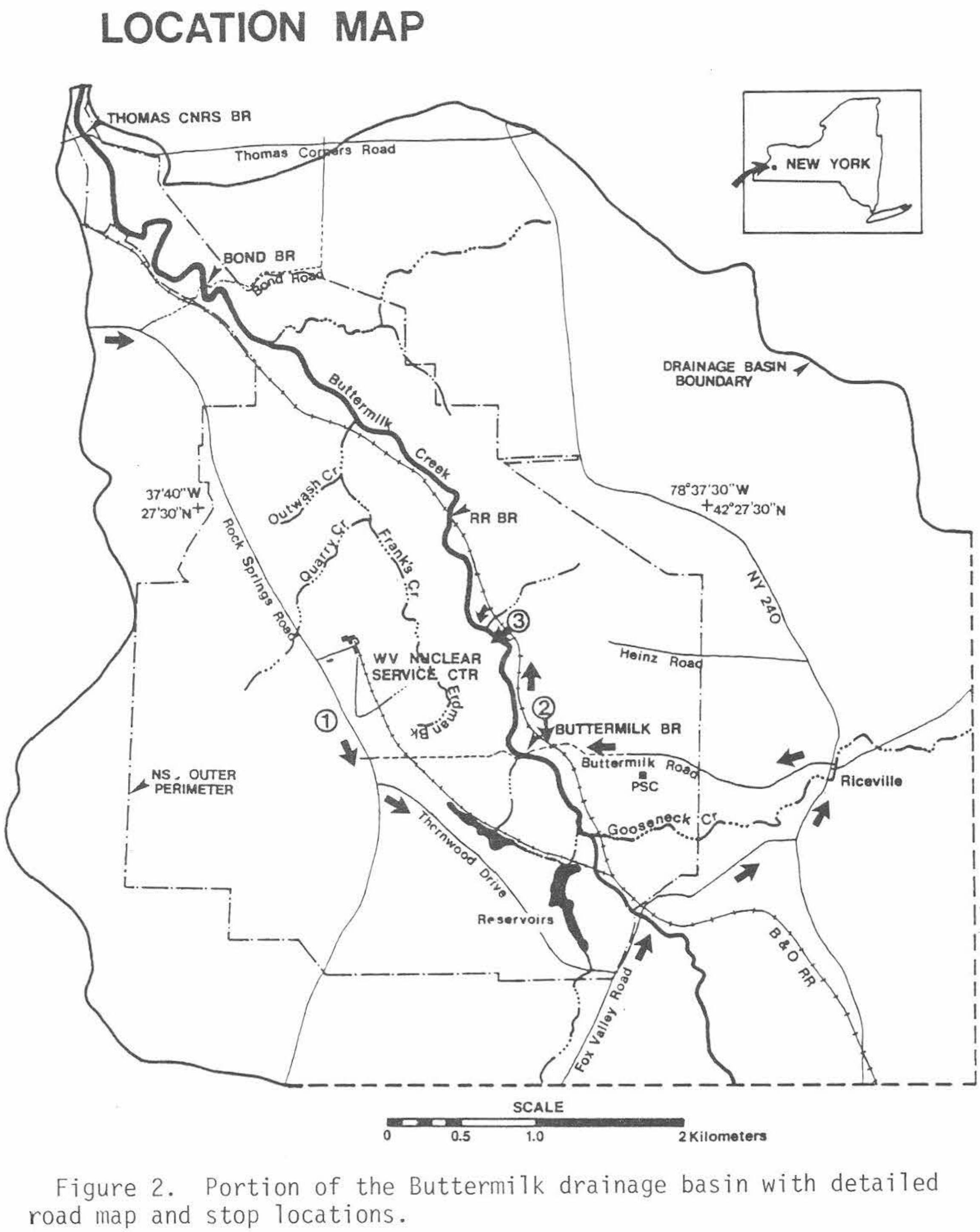

Figure 2. Portion of the Buttermilk drainage basin with detailed road map and stop locations.

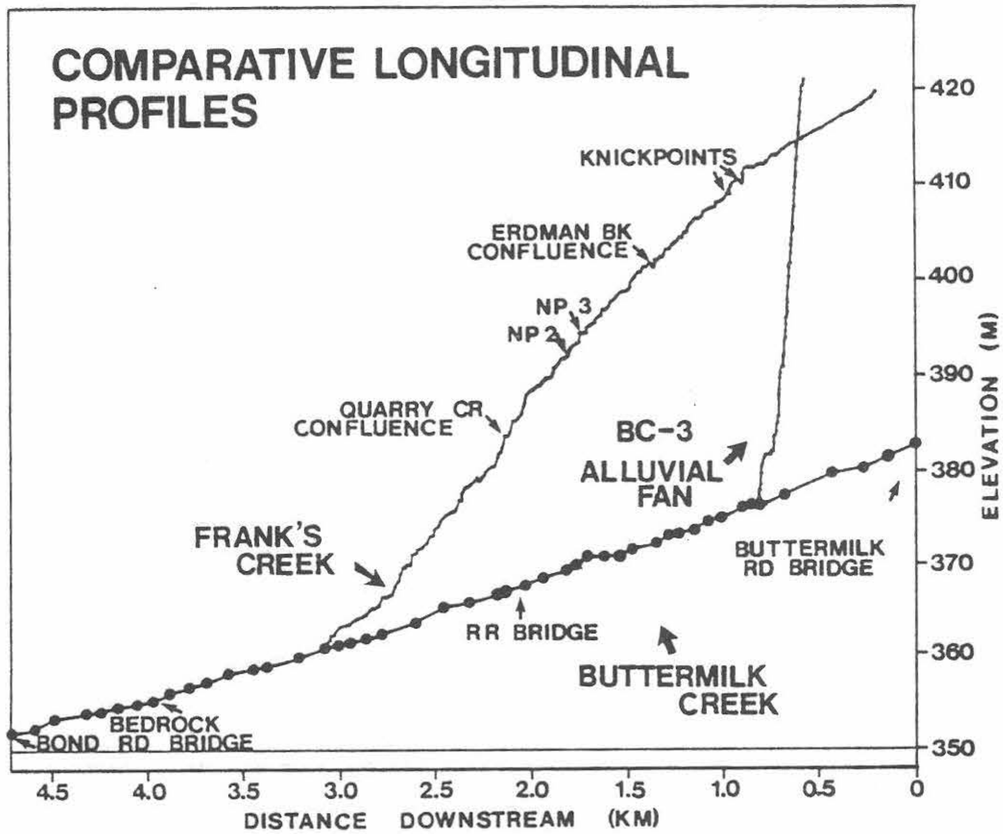

Figure 3. Longitudinal profiles of Buttermilk Creek; Frank's Creek, an important tributary; and a small valley-wall alluvial fan.

and around bars. Longer-spaced segments, on the order of 500 m to 1 km spacing, are due to the sediment load of inflowing tributaries that tends to flatten the gradient near the confluences. Several 200-300 m long flat spots in the profile are flood-bar surfaces and represent temporary storage of gravel in the bars. Average maximum clast size (long axes) is 25 cm and shows no systematic variation in a downstream direction. Clast size does not vary downstream because clasts are continuously supplied to the bars and channels along the reach by lateral erosion of valley walls and previously deposited terraces. The clasts are well imbricated and occur in groups or concentrations on the surfaces of bar complexes and in chutes. Clasts of all sizes form an armor on the bar, channel, and chute surfaces; about 2 percent of these clasts are much larger than the average maximum size; about 5 to 10 percent are of average maximum size, and the remainder are smaller. The clasts are very bladed to very platy in shape, and consist mostly of Devonian sandstones, the most resistant bedrock of the area.

Grain Size

Large (150-225 kg) sediment samples were collected from the valley-wall till, gravel bars, and low terraces. Figure 4 illustrates the textural classes of those samples.

Till. The till samples are silt-rich, with 80-85% silt and clay that constitutes the suspended-sediment load of the fluvial system. Visual inspection of till cropping out at landslide localities and at the base of slope along scarps cut by the Buttermilk channel, reveals that few large clasts are contained in the till and that a low overall gravel percentage is in line with our two analyses. Lafleur (1979) and Dana et al. (1979, 1980) report similar findings at other outcrop localities and in research trenches cut in Lavery till on the plateau adjacent to the low-level, waste-burial site.

Bar Gravel. All samples contained 75-95% gravel with little sand matrix and very little silt and clay. Some sand, and all silt and clay, occurred as a falling-stage drape over the gravel with some infiltration downwards into available pore space.

Terrace Samples. Two samples are similar in gravel percent and overall grain-size distribution to the bar samples. These two samples represent previously deposited bar complexes resulting from the cross-valley channel sweep. The third sample, GS-3, is a fine-grained sandy silt with essentially no gravel. It was obtained from the topmost unit in the stratigraphic section upstream of transect 1, opposite bar complex 3. This sediment was deposited in a small depression (pond) on the gravel terrace at the base of a small alluvial fan.

GEOMORPHIC AND SEDIMENTARY PROCESSES AND LANDFORMS

Bar and Channel Geometry

Geometry. The bars in Buttermilk Creek are 100 to 200 m in length and exist as complex, elongate, erosional and depositional features flanked, and often cut, by the present low-stage channel (Fig. 5A, 5B). Riffles are present where the channel crosses over a large bar and in the vicinity of the downstream edges of the bars. Pools are present as deeper areas in the narrow thalweg and also as very deep scour holes between advancing bars and eroding banks. Dry chutes exist along the margins of many bars. The overall bar and channel pattern is similar to that of small, gravelly, braided streams (Boothroyd and Ashley, 1975; Smith, 1970). Some bar complexes have well-developed slip faces or depositional edges on their downstream margins. Depositional edges, gravel or sand, record the downstream growth of bars. The fact that the bar complexes may be treated as braided-stream features at high-flow stage means that observations and conclusions regarding braided-stream gravel transport in other areas apply to Buttermilk Creek. Masses of gravel may move by growth of longitudinal bars, formation of unit bars,

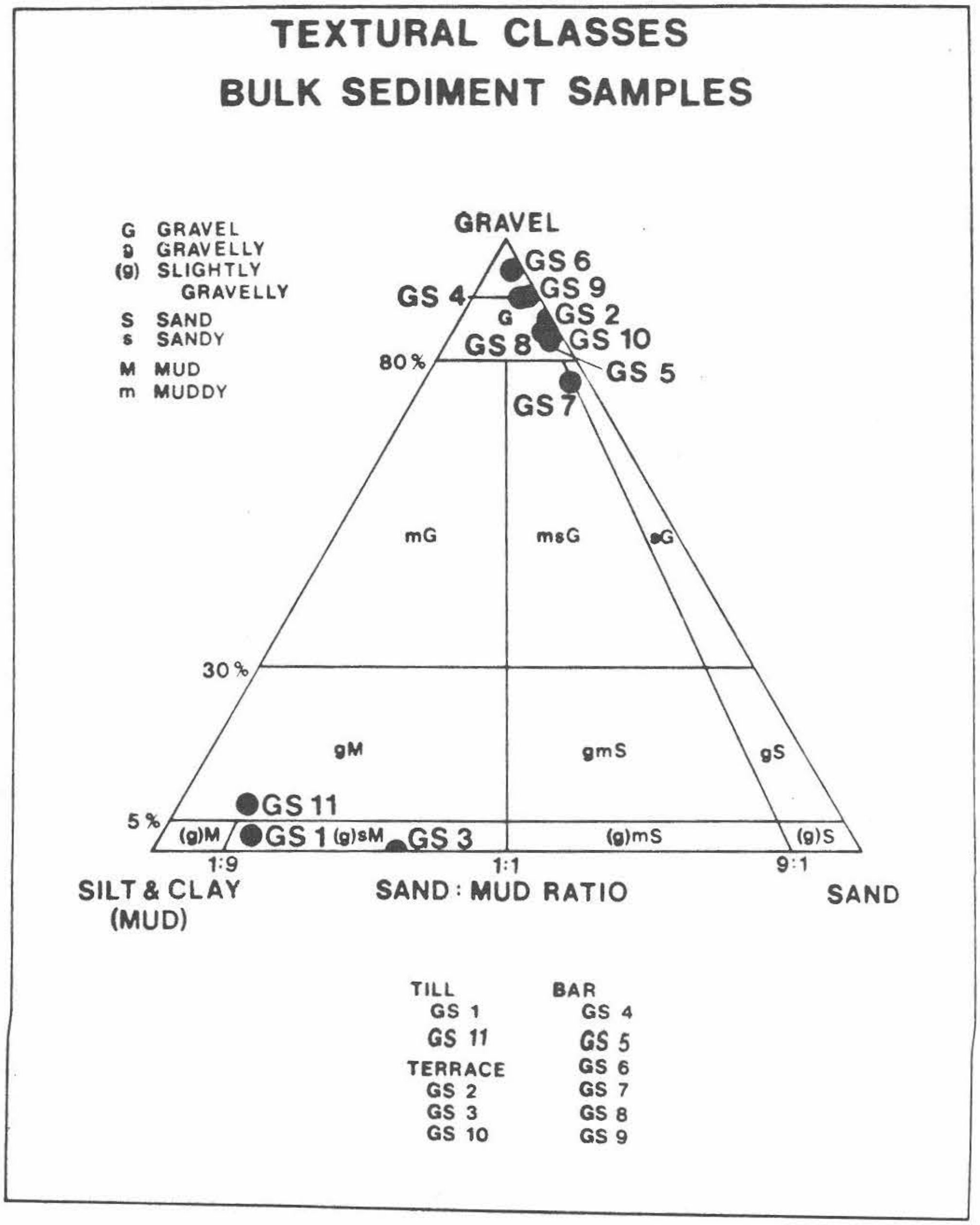

Figure 4. Textural classes of sediment from basal till, fluvial bar, and fluvial terrace depositional environments.

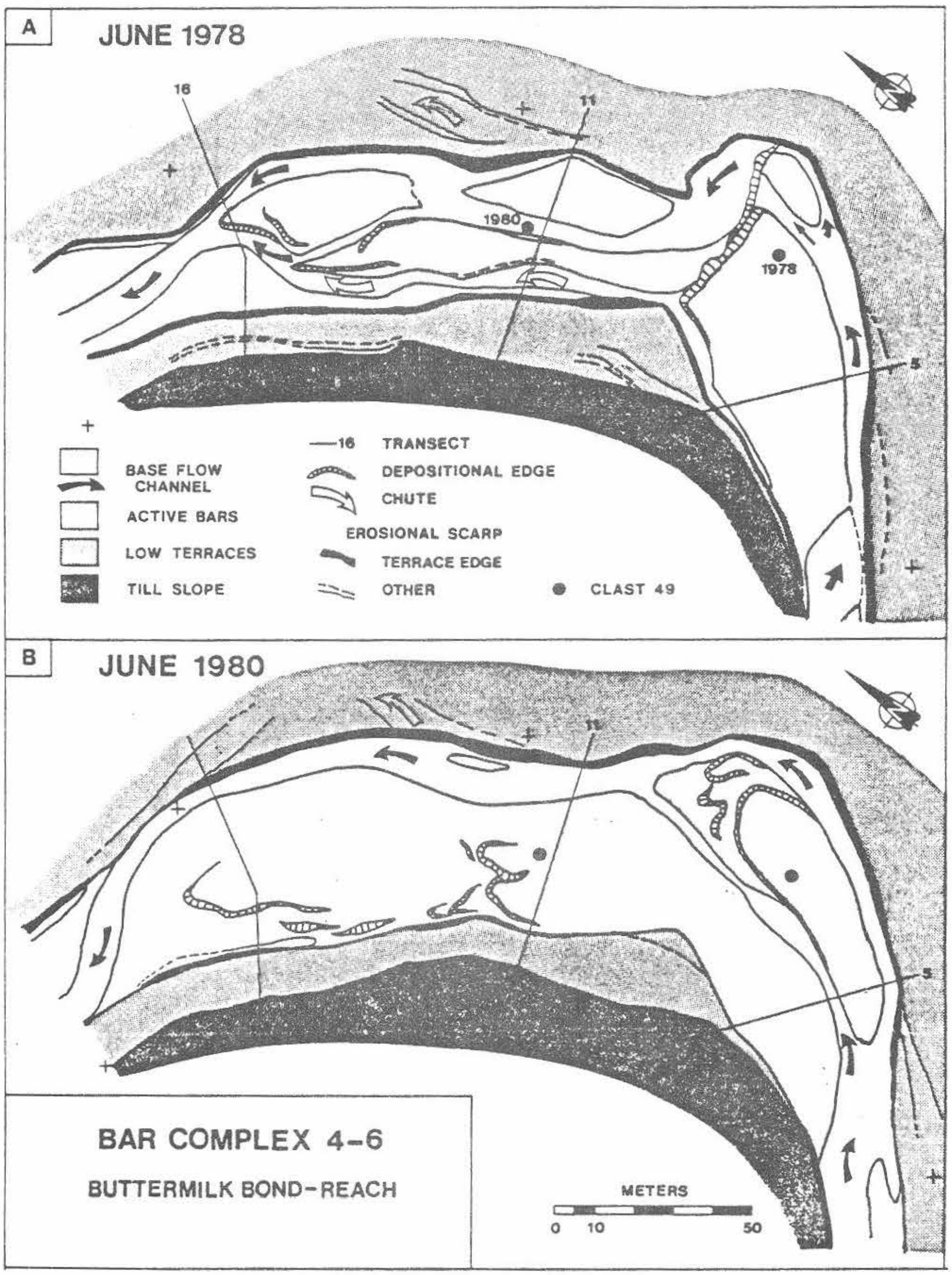

Figure 5. Bar complex 4-6. A. June 1978. B. June 1980.

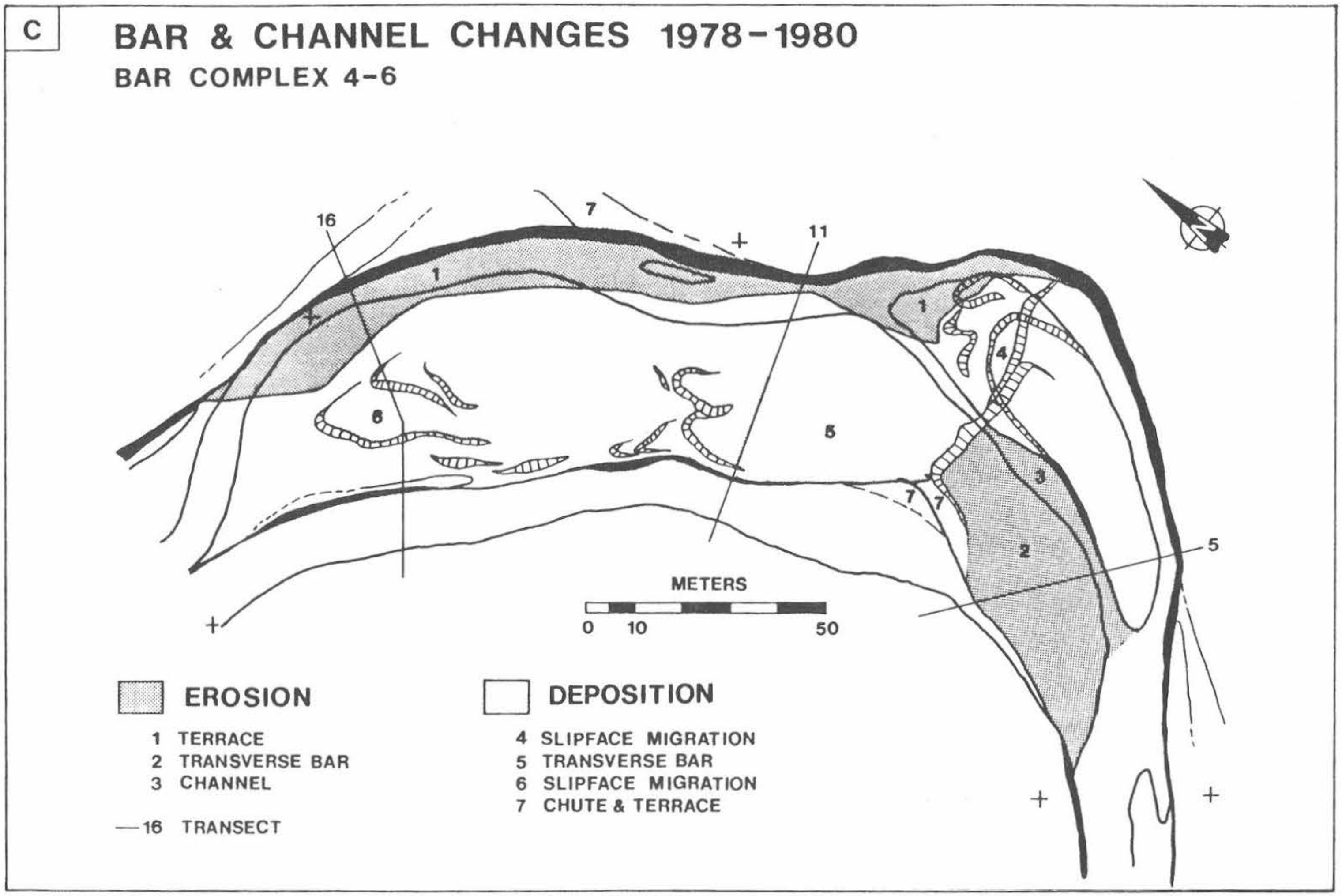

Figure 5C. Bar complex 4-6. Note changes between 1978 and 1980 (due to Hurricane Fredric flooding).

or by movement of diffuse gravel sheets. All of these features are well-displayed on many of the bar complexes, particularly on bar complex 4-6 (Fig. 5).

Bar Migration. Bar complex 4-6 (Fig. 5) was vastly changed by the flood event accompanying Hurricane Fredric on Sept. 8-9, 1979, mostly by the migration of large transverse bars in equilibrium with the high discharge depth and velocity conditions. The upper transverse bar (BC-4) attempted to migrate forward but the east side slipface encountered the debris pile and terrace at the kink in the channel (Fig. 5A). The intense turbulence created by this situation caused rapid erosion and removal of terrace gravel that resulted in a wider bend in the channel (Fig. 5B). The gravel was redistributed onto bars farther downstream, effectively recycling the low-active terrace material. The complete bar form migrated downstream approximately 60 m by a combination of stoss-side accumulation and bar slipface migration. Additions of east side terrace gravel resulted in the vertical accretion and slipface migration of smaller transverse bars on the downstream part of the complex (BC-5). In addition to removal of terrace material and recycling it back to active bars, bar gravel was deposited up on the low-active terraces as longitudinal bars during Hurricane Fredric. At bar complex 4-6, unvegetated chutes adjacent to active bars were filled and excess gravel deposited on the west side terrace. Higher elevation chutes on the terraces were activated during peak flooding, and gravel longitudinal bars accumulated in them.

Alluvial Fans

Alluvial fans along Buttermilk Creek can be classified into 3 groups: 1) short, steep active fans, measuring 100-200 m long; 2) larger fans with both inactive and active segments; and 3) large fans at the junciton of tributary streams with Buttermilk Creek. All are heavily vegetated. The short fans begin part-way up the valley wall, with an entrenched stream extending to the top of the wall. The fanhead may or may not contain an incised stream with the fan commonly having a single active lobe. An example is the small fan at bar complex 3, on the west wall, adjacent to the waste-burial site (Fig. 6). The larger fans also head in the valley wall, but the entrenched streams above the fanheads are incised into the upland surface. These fans contain both inactive and active segments with inactive lobes existing as terraces above incised, active streams that feed distal lobes. The fans may have multiple active lobes. The largest fans occur at the confluence of small tributary streams with Buttermilk Creek. They are similar to the medium-sized fans in appearance, except for larger size and deeper incision of the active stream.

The small fans are active only during rainfall events and have dry channels the rest of the time. The larger fans remain active during lower-flow stages, even though runoff is very low. The largest tributary fans have an active low-flow channel even during base-flow conditions. During flood stage, overbank flow can reactivate fan lobes.

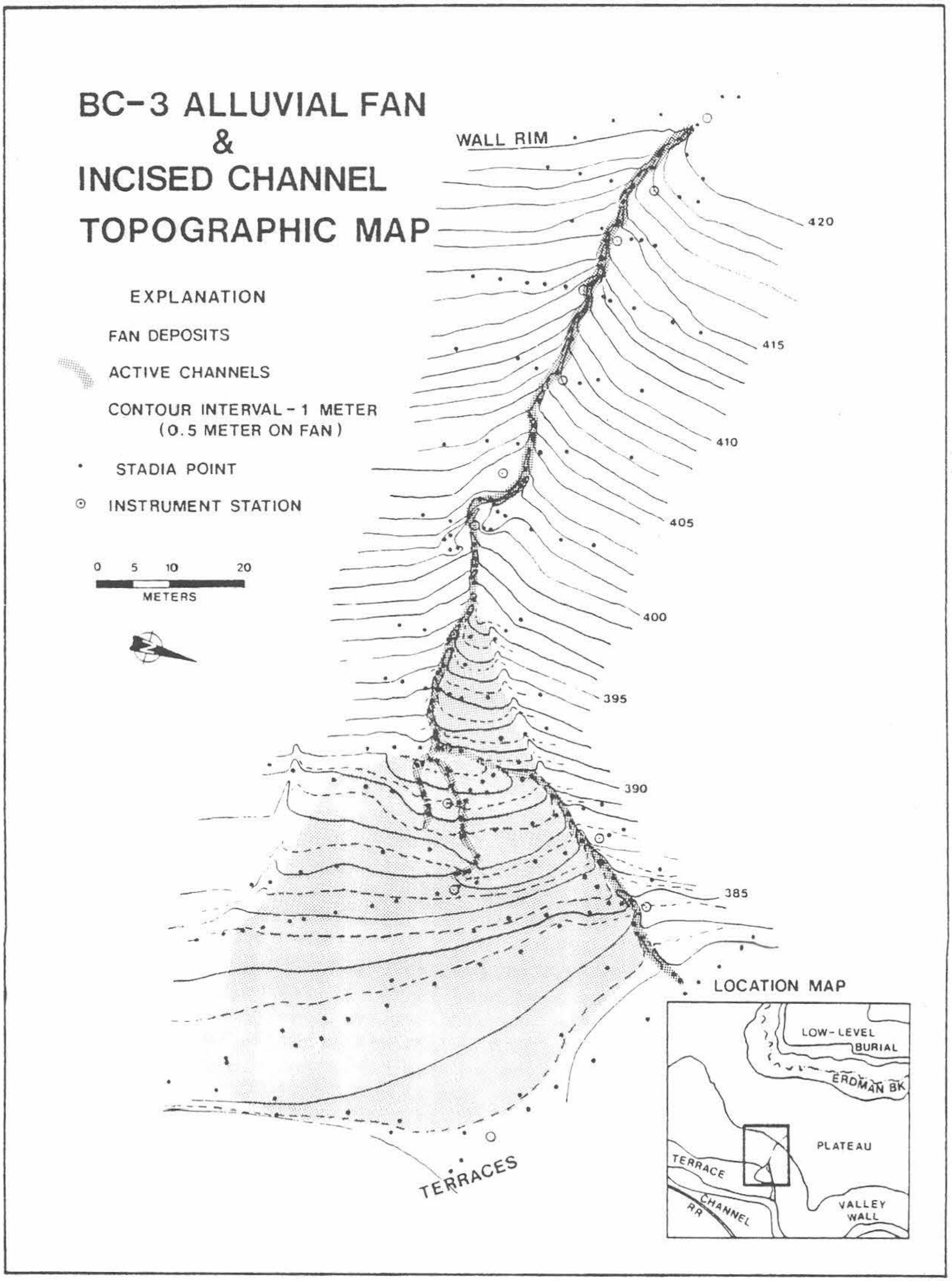

Figure 6. BC-3 alluvial fan.

The processes associated with alluvial-fan development are important agents in the widening of Buttermilk Creek and its tributaries. Gravel, sand, and some silt and clay eroded from the upper incised channels is deposited on the fans, whereas other silt and clay may collect in ponded depressions on the terraces. An unknown percentage of fine-grained sediment bypasses the fan and is fed directly as suspended-material load into Buttermilk Creek, or into tributaries and then into Buttermilk.

Landslides

Active landslides occur in areas where the channel impinges on, and cuts, the valley wall over a period of at least tens of years. Sequential aerial photographs showed that the Buttermilk channel was at, or near, the large landslide area on the west valley wall at bar complex 6, from 1939 to 1977. Monitoring of the BC-6 landslide (Fig. 7) showed downslope movement of 8 to 32 meters horizontally and 0.8 to 10 meters vertically during the time period 1978-1980. The movement actually occurs as a series of coherent slumps, 20-50 m wide at the top of the slide, changing to a hummocky, tension-cracked, earthflow mass at the toe of the slide. Downslope trajectories of the upper slide slumps are steeper than the lower earthflow, contributing to a pile-up of material at the base of the slide. This material can rapidly flow out onto Buttermilk bar and channel areas. The earthflow accumulation of material had been removed by April, 1980; the last was probably eroded by Hurricane Fredric flooding.

Buttermilk Terraces

153 separate fluvial terraces mapped within the Buttermilk-Bond reach of Buttermilk Creek are shown on Figure 8. Terraces have been grouped into categories according to their general elevation above active bar surfaces as follows: 1) low active (0-3 m); 2) low inactive (3-8 m); 3) middle (8-35 m); and 4) high (35 m). Arrays of terraces also can be grouped according to events that generated them, or allowed their preservation after they were formed.

The low-active terraces are associated with the present processes of Buttermilk Creek and its tributaries. It is evident from examination of historical aerial photographs and field mapping that the active bar-and-channel lateral movement incorporates the lower terraces and reactivates them as active bars. Catastrophic floods may serve to devegetate the low terraces and convert them to active surfaces. Large uprooted trees are common as flood debris on bar surfaces and in channels. The heavy vegetation cover of these terraces is confusing and serves to mask their degree of activity. Some terraces may be preferentially preserved as discussed below.

The largest number of terraces that are higher in elevation than the low-active level are associated with the confluence of tributaries with Buttermilk Creek. Gravel transported down the tributaries is

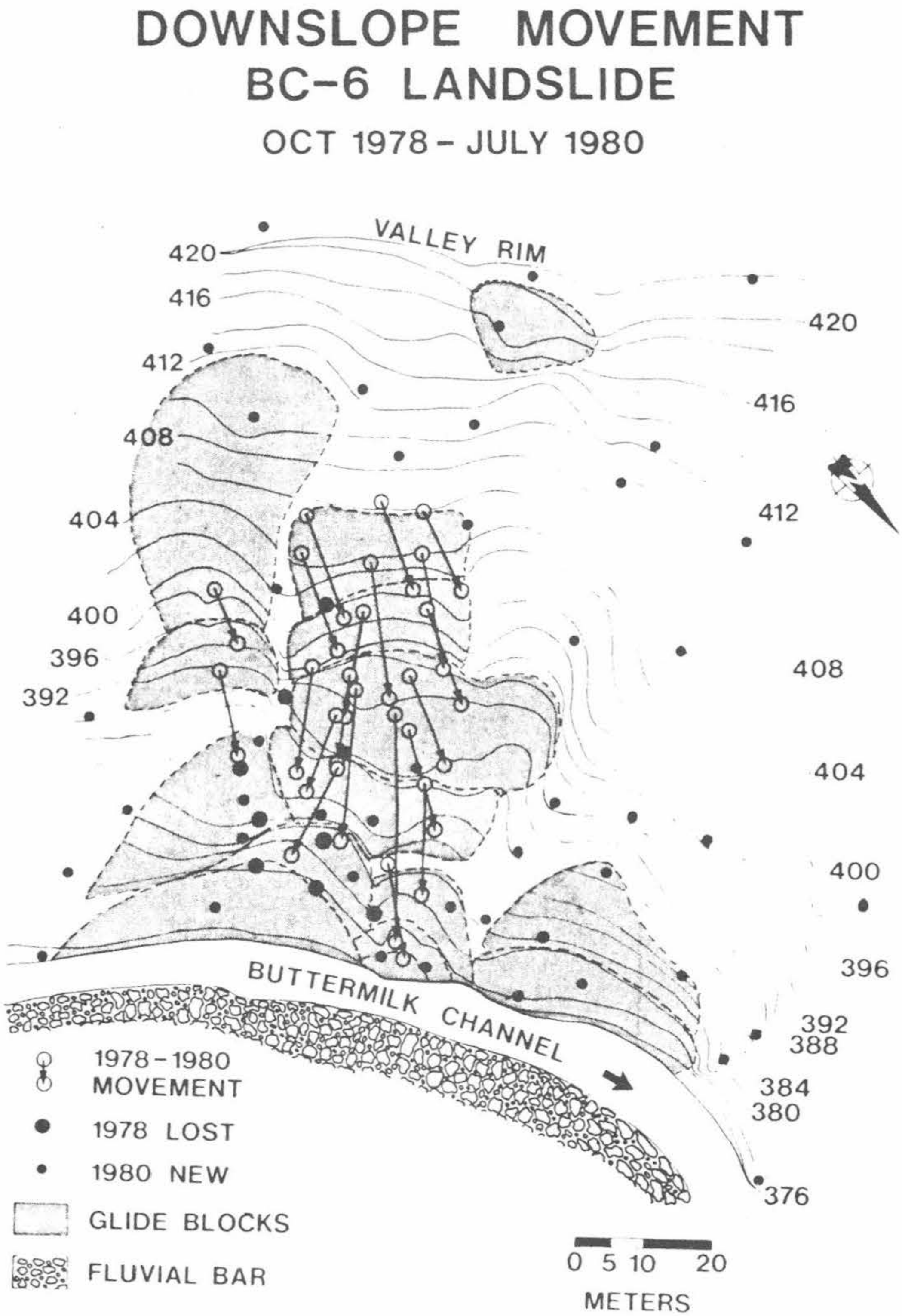

Figure 7. BC-6 landslide. Note location and amount of downslope movement of glide blocks.

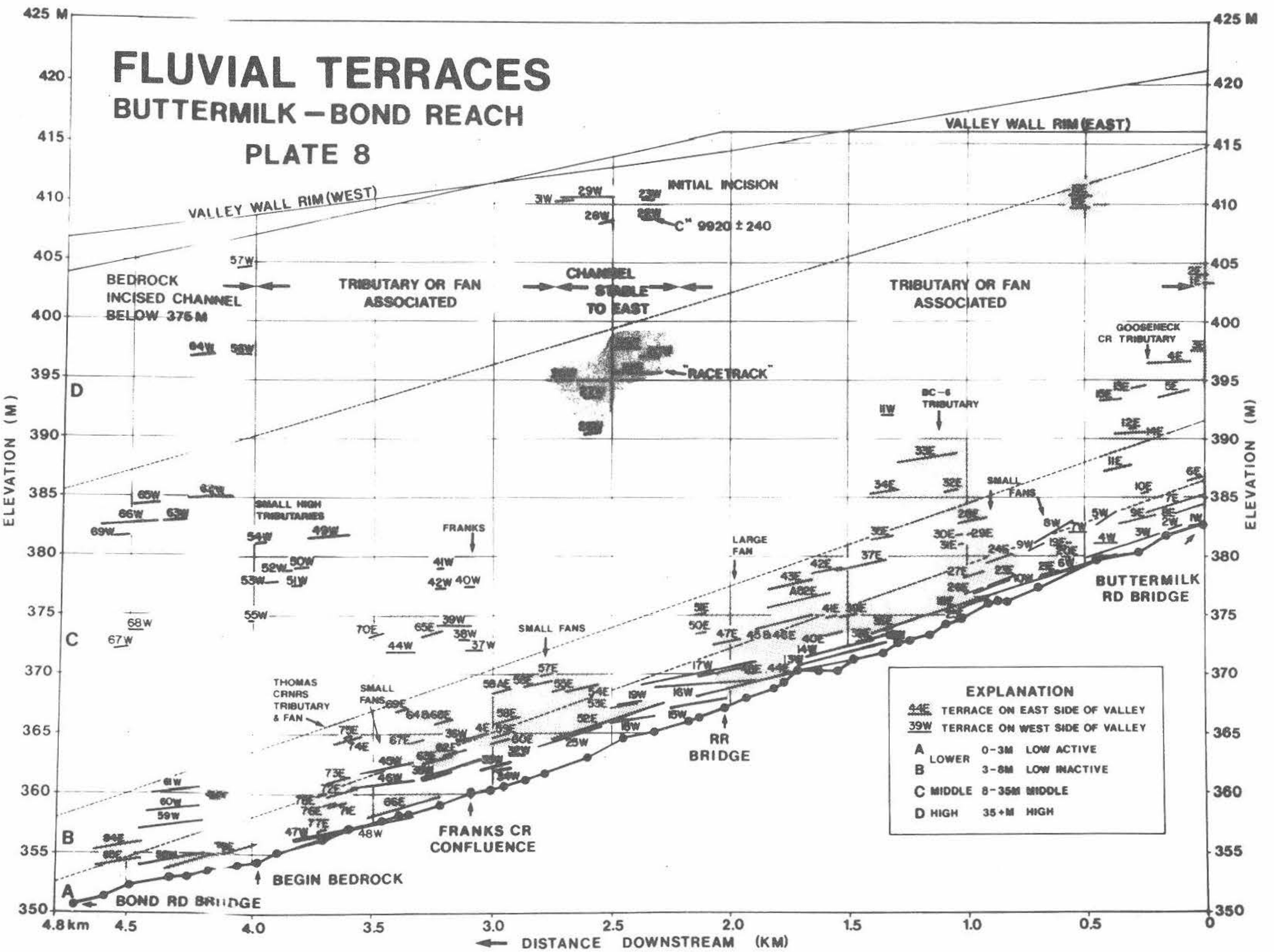

Figure 8. Fl[illegible]al terraces of Buttermilk Bond reach projected to low-stage thalweg. Note: C-14 date and initia[illegible] [illegible]ncision; location of BC-6 tributary.

deposited as slightly dipping, fan-shaped bar complexes at the mouths of the tributaries. The fans are skewed in downstream direction relative to Buttermilk Creek, due to redistribution by Buttermilk bedload processes. Continued incision of Buttermilk Creek and the associated tributary leads to the abandonment of the bar complexes, and they become terraces by definition. The excess of gravel supplied over transport capacity may temporarily, or permanently, retard the lateral sweep of the Buttermilk channel and destruction of the terrace array. Other terraces were deposited in a similar manner at the base of, and adjacent to, alluvial fans that developed within Buttermilk valley. Some fans are small, such as the BC-3 fan, whereas others are larger, with upper drainages well incised into the plateau above Buttermilk valley.

Some terraces at the lower end of the Buttermilk-Bond reach are bedrock defended; that is, the channel of Buttermilk is incised into Devonian bedrock on the west side of the valley preventing further channel sweep. We speculate that a third array of terraces, including the set that contained the dated wood fragmants and the set that includes the "Racetrack" (Fig. 8) have been preserved because the Buttermilk channel has remained stable on the east side of the valley for long periods of time. We do not know the cause for this channel behavior.

Reservoir Sedimentation

Sedimentation in two WVNSC reservoirs situated on tributaries of Buttermilk Creek (Fig. 2) was assessed to gain insight into the overall denudation rate and landscape evolution discussed in the next section.

The reservoirs are contained by earth dams constructed across separate tributary streams, with water accumulation beginning in 1963. The full stage for both is 412.4 m (1353 ft). A dredged channel connects the reservoirs allowing free flow between them and for the stages in each to equilibrate. Flood discharge is released through a pipe beneath the north reservoir, down the tributary, and into Buttermilk Creek just south of the Buttermilk Hill Road bridge. Extreme flooding results in overflow across a wide sluiceway east of the south reservoir and directly into Buttermilk Creek. The south reservoir is shown in plan view in Figure 9A.

The pre-reservoir valley cross-sections show a V-shaped form eroded in Lavery till, probably not unlike the present Frank's Creek. Sedimentation from 1963-1980 has been by: 1) progradation of a delta at the south end of reservoir; 2) density underflow of fine-grained material down the delta front and prodelta slope onto the reservoir floor, and 3) slumping and debris flow of the submerged valley walls down the side slopes.

Inspection of Figure 9A indicates that the delta plain has prograded about 140 m into the reservoir. The cross profiles near the delta front (8/21 and 9/22, Fig. 9B) show a flat to gently concave up reser-

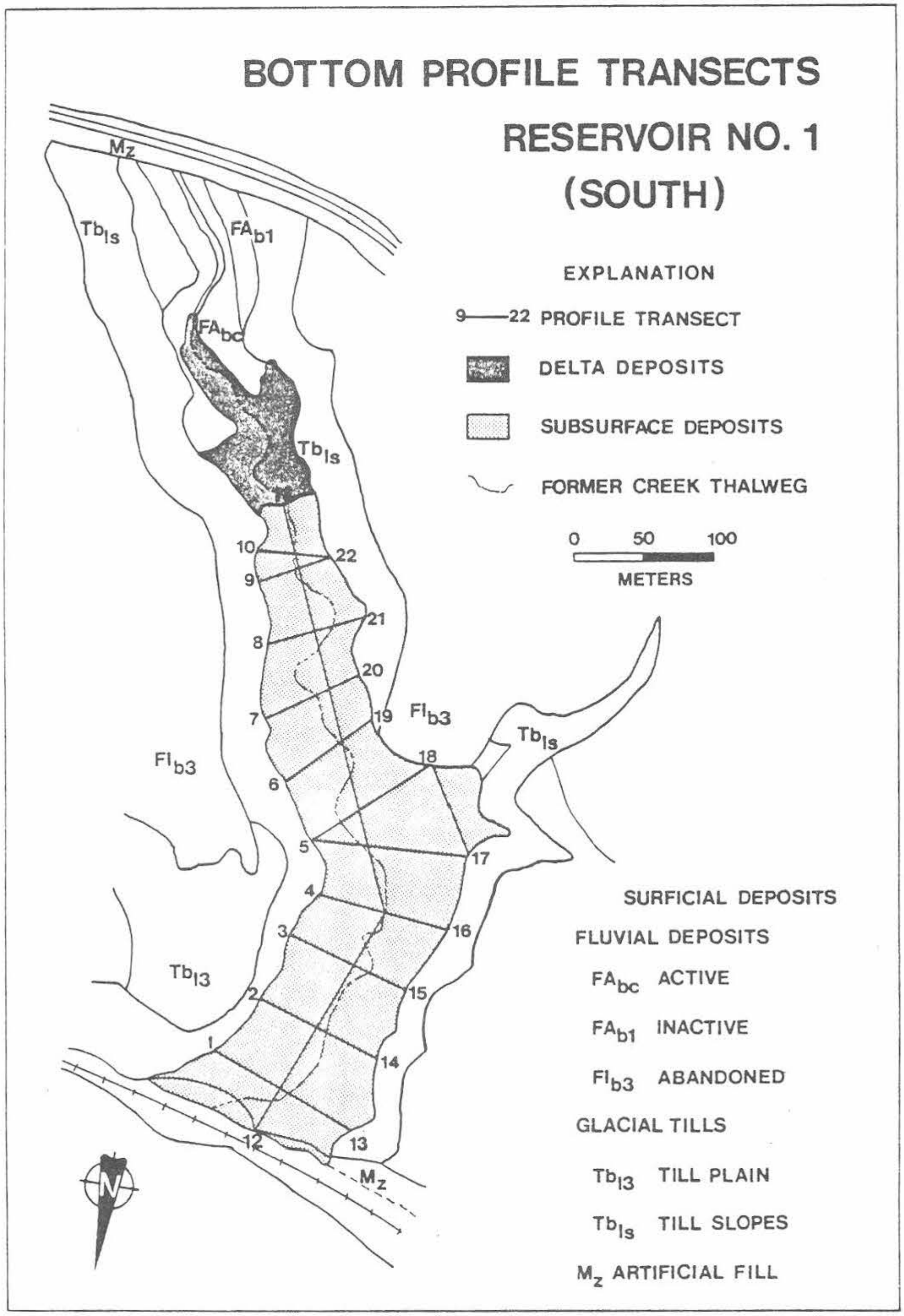

Figure 9. WVNSC storage reservoir No. 1 (south). A. Map with cross-profile locations.

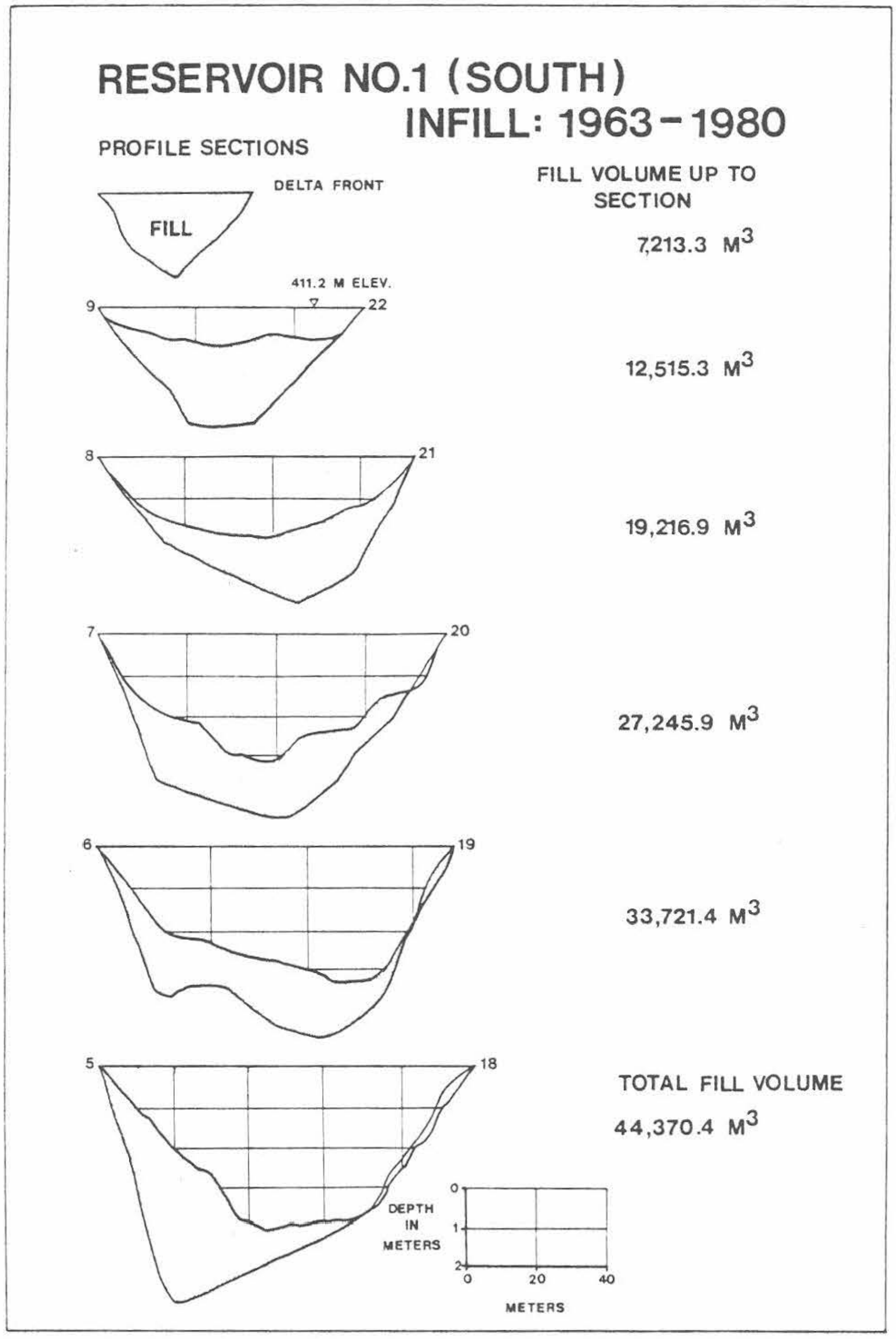

Figure 9. WVNSC storage reservoir No. 1 (south). B. Selected cross-profiles near delta front showing infill since 1963.

voir floor, whereas the cross profiles farther away show a more U-shaped section with terraces and uneven filling. The flat profiles probably reflect fill by density underflow, and the others, fill by a combination of slumping and underflow.

Buttermilk Stage and Discharge

USGS Gaging Data. The USGS operated a gauging station on Buttermilk Creek near the Bond Road bridge during water years 1962-1968 (USGS, 1962-1968). Stage-discharge rating tables and curves, and indirect measurement calculations for a large flood event are included in these publications and in a special report by the USGS (1968b).

A stage-discharge rating curve for Buttermilk Creek (Fig. 10), adapted from those compiled by the USGS (1968b), includes the highest discharge events for each year the stage-height recorder was in place. These readings (max = 110 m^3 sec) of instantaneous discharge indicate that peak flow events are much higher than those that appear as the daily summation (USGS, 1963-68). This means that peak discharge events are of extremely short duration, several hours in length.

NYSGS Gauging Data. The stage recorder installed at Thomas Corners Road bridge by the NYSGS in August 1978 was removed by Hurricane Fredric flooding in September 1979 and was reinstalled in July, 1980. Selected stage-height records for the summer and fall of 1980 are shown in Figure 11. Velocity-area information collected during the summer-fall period, along with suspended sediment samples, are shown on the stage-discharge, suspended sediment concentration-discharge plot (Fig. 10).

Flood Events. The hydrographs of three flood events are illustrated in Figure 11; a relatively low-discharge event (Oct. 12), a moderate event (Aug. 11), and a high-discharge event (Oct. 25-26). The moderate and high events show the "spikey" nature of the flooding, particularly the rapid rise in stage to peak flow in a matter of hours. A review of the USGS stage-discharge data and rating curve (USGS, 1968b) reveals that the October 25-26 flood is within the range of the yearly maximum discharge event as determined by the USGS for 1962-1968. The Hurricane Fredric flooding that carried away the stage recorder was probably equal to, or greater than, the indirect measurement of 110 $m^3 sec^{-1}$ determined by the USGS (1969b) for a large flood in 1967. The flood level, as determined by debris in trees, is shown in Figure 12 for bar complex 4-6, transect 5. Also show is the base-flow water-surface elevation, and the flood flow of October 25-26, 1980.

The suspended sediment concentration at a given discharge increases rapidly with increase in discharge during a flood event, peaks early, and then falls off more rapidly than a proportional decrease in discharge (Fig. 11C). This relationship is common to small streams with rapid runoff and little infiltration (Gregory and Walling, 1973).

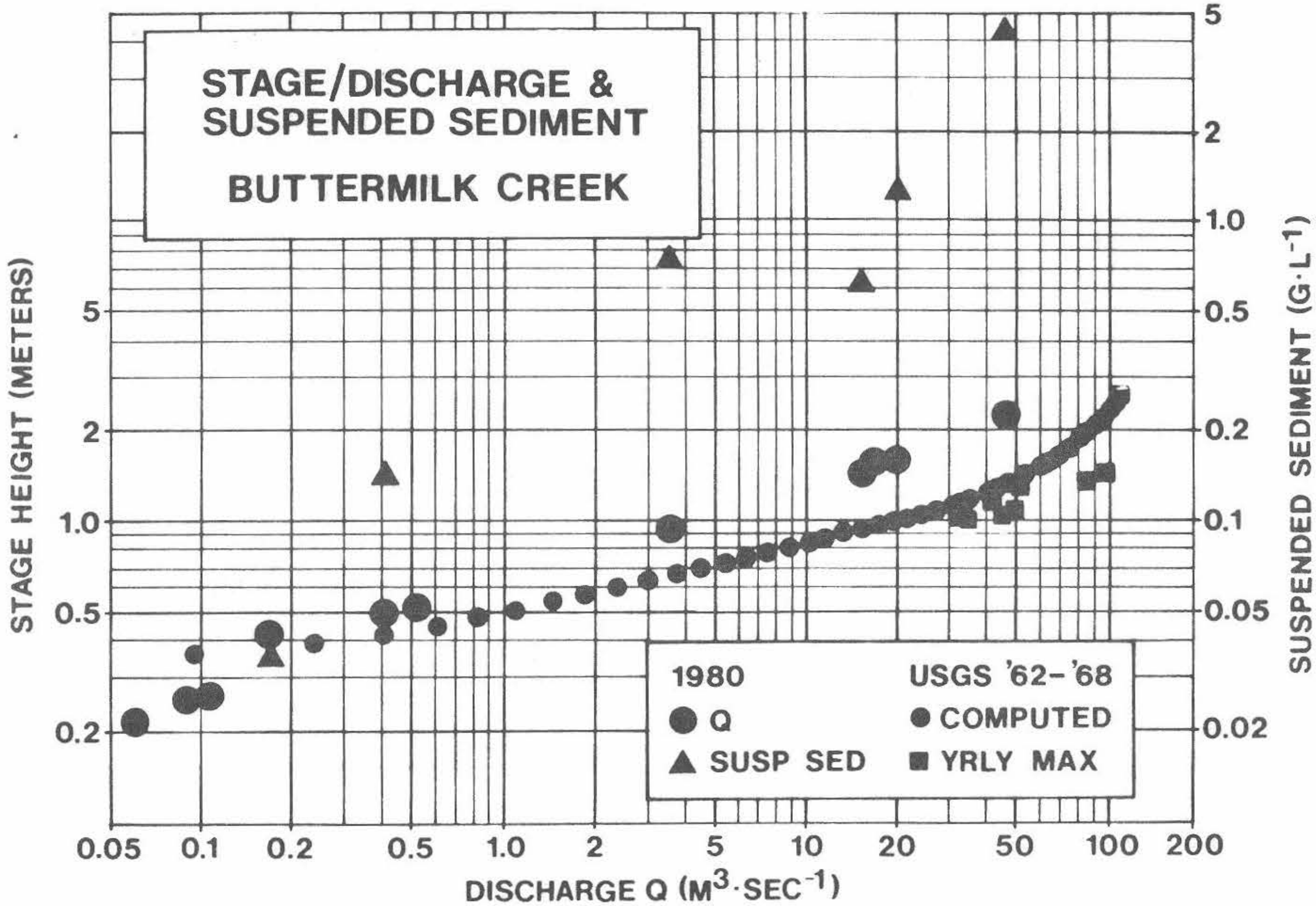

Figure 10. Stage-discharge, and suspended sediment data, Buttermilk Creek. USGS gauge at Bond Rd. bridge; 1980 data obtained from Thomas Corners Rd. bridge (Fig. 2).

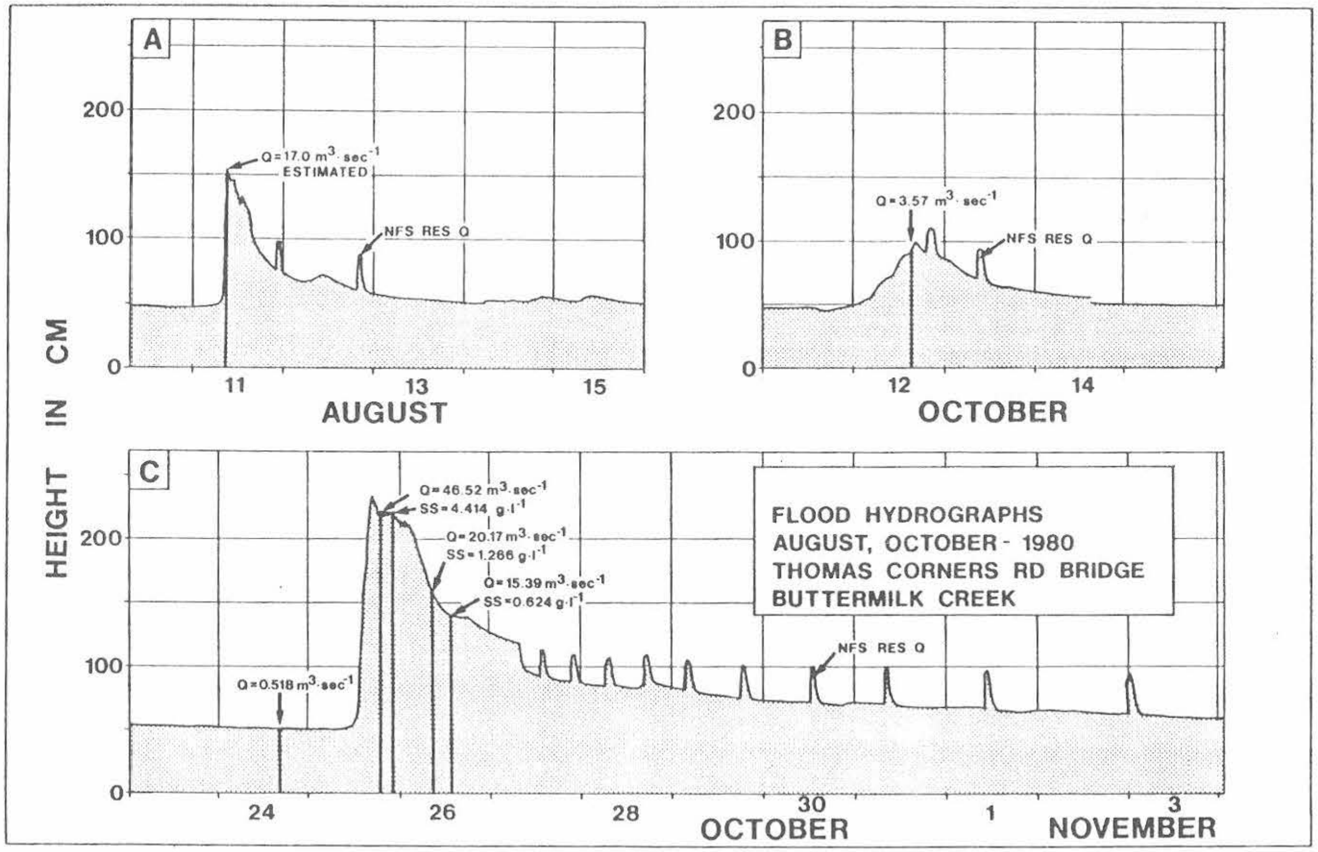

Figure 11. Flood hydrographs, Buttermilk Creek, 1980. Note suspended sediment values and reservoir slug discharge. Hydrograph C estimated to be mean annual flood.

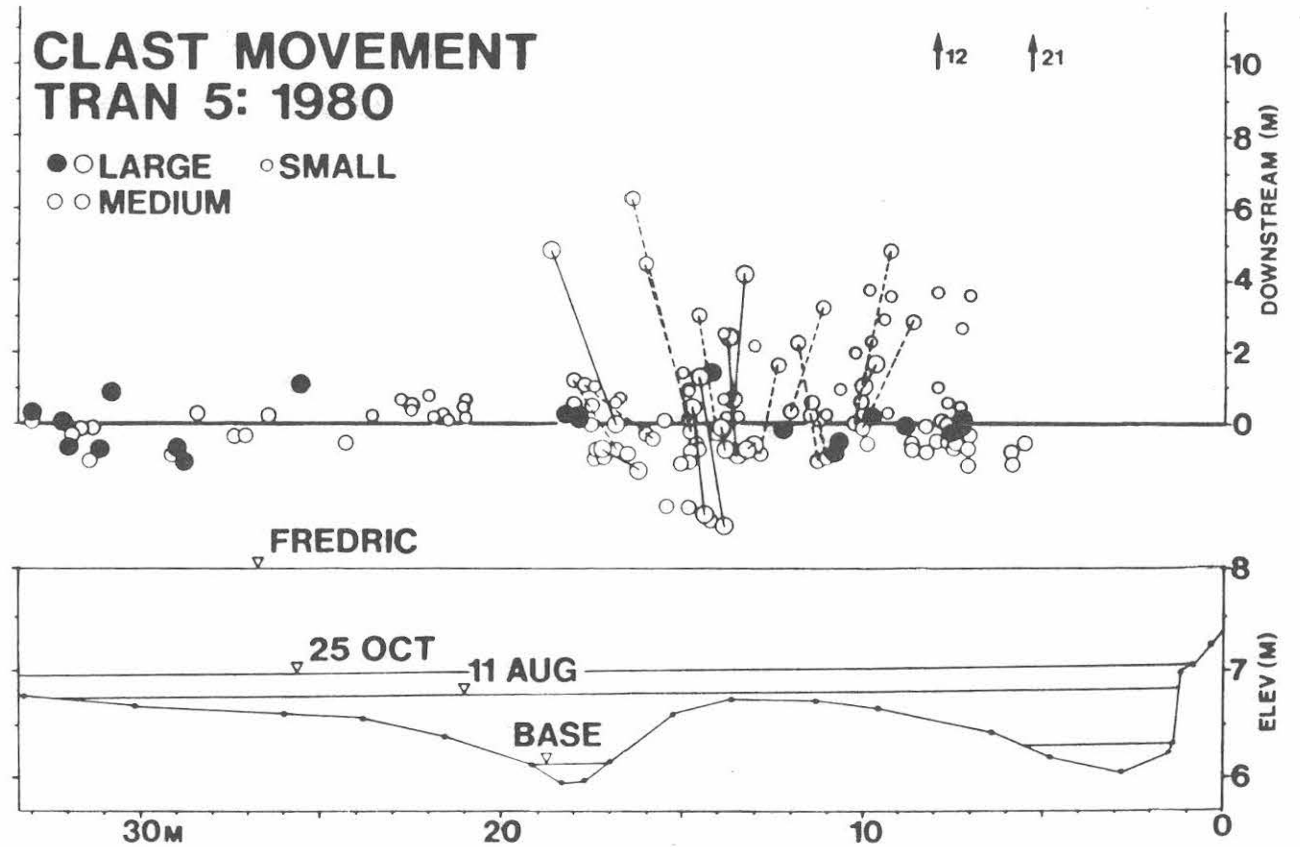

Figure 12. Clast movement, transect 5, bar complex 4-6, that occurred during Oct., 1980 flood event.

The sharp spikes on the hydrograph of about one hour duration represent controlled releases from the WVNSC reservoirs (Fig. 2,11). The gate at Dam No. 2 (north) opens automatically when the reservoir stage rises 30 cm (1 ft) above full 412.4 m (1353 ft). Discharge is through a 91 cm (36 in) pipe at a rate of 5.66 m^3sec^{-1} (200 cfs) until the water level in the reservoirs is lowered enough to allow the gate to close.

Clast Movement. The August 11 flood was probably the threshold event for initiation of movement of medium-sized clasts on the lower bar surfaces. An event of this discharge (17 m^3sec^{-1}, estimated) (Fig. 11A) occurs several times a year based on the USGS (1968b) data. The October 25-26 event (46.52 m^3sec^{-1} measured; 60 m^3sec^{-1} estimated) moved some large clasts on the bar edges and shoulders an average of 3 m (Fig. 12). This event may be considered to be just above the threshold of movement for large clasts, although not all of them moved. A flood of this discharge falls in the range of events with a one year recurrence interval (USGS, 1968b).

BUTTERMILK VALLEY DENUDATION

Sediment transport rates have been calculated or estimated for the various processes that were described above. Space permits only a summary here; refer to Boothroyd et al. (1981) for a more detailed discussion. Table 1 is a summary of transport rates and volumes discussed below.

Simple Denudation Rate

The volume of sediment removed from Buttermilk valley as a function of time can be calculated using the age of terrace 22W (9920 ± 240 BP) (Fig. 8) as being close to the time of initial incision and downcutting of Buttermilk Creek. The total volume of sediment removed, neglecting tributaries, was: 65,923,331 m^3 (Table 1).

The simple denudation rate is:

$$\frac{65{,}923{,}331 \text{ m}^3}{10{,}000 \text{ yes}} = 6592 \ (6600 \text{ m}^3\text{yr}^{-1})$$

The denudation value represents the amount of bedload and suspended-load transport necessary, per year, by Buttermilk Creek, to remove valley fill and arrive at the present configuration. Variations in rate due to short or long-term climatic change have been ignored.

Gravel Movement

The Buttermilk valley sediment aggregate is composed of about 5 percent gravel, 85 percent fine sand, silt and clay, and 10 percent coarse and medium sand (Fig. 4) (Hoffman et al., 1980). Using the

TABLE 1. SEDIMENT TRANSPORT RATES

PROCESS	VOLUME (m^3 yr^{-1})	WEIGHT (kg yr^{-1})	TOTAL VOLUME (m^3)
Gravel bar migration	85		
Gravel volume deficit	116		
Suspended sediment			
Instant discharge		205 $kg\,sec^{-1}$	
Peak discharge	3000	4,800,000	
Total discharge	3770	6,033,000	
Landslide	150	250,050	10,500
Gravel, sand	23	37,510	
Fs, Si, clay	128	212,540	
Reservoir			
No. 1 south	736		
No. 2 north	379		
Buttermilk valley			
Simple denudation	6600		66,000,000
Basin sediment loss	6979		
Gravel denudation	330		3,300,000
Gravel terraces and bars			570,000 (1m) 1,140,000 (2m)
Fs,Si,Cl denudation	5610		56,100,000

denudation rate and sediment distribution, the volume of each available size can be calculated and a transport rate determined.

Volume of gravel available is:

$$66{,}000{,}000\ m^3 \cdot 0.05 = 3{,}300{,}000\ m^3$$

Gravel available per year for transport is:

$$6600\ m^3yr^{-1} \cdot 0.05 = 330\ m^3yr^{-1}$$

There is temporary storage of gravel in the bars and the low-active terrace systems. The gravel stored in a one meter thick section is: 570,000 m^3; and in a two meter section: 1,140,000 m^3.

A comparison of all the derived gravel transport rates reveals that: 1) The gravel bar migration rate plus volume deficit rate agrees quite well with the amount of gravel provided by simple gravel denudation.

The bar migration rate is low because it is based on movement of large clasts only. More information is needed on small-clast movement. 2) The amount stored in the bar and terrace system is about 20-35 percent of that made available by denudation per year. This material is recycled at an unknown rate, but the volume deficit for bar complex 4-6 may be a good indication of rate. If so, then there is a gravel deficit that must be up from more gravel-rich units upstream in Buttermilk or in the tributaries.

Suspended Sediment Transport

Utilizing the simple denudation rate and the selected grain-size distribution of till, the fine-grained material available per year can be calculated.

Volume of fine sand, silt and clay available is:

$$66{,}000{,}000\ m^3 \cdot 0.85 = 56{,}100{,}000\ m^3$$

Fine sand, silt and clay available for transport is:

$$6600\ m^3yr^{-1} \cdot 0.85 = 5610\ m^3yr^{-1}$$

The cumulative suspended-sediment discharge of the October 25-26, 1980 event (one year storm), a conservatively calculated value, was 67 percent of the simple yearly suspended-sediment denudation rate. Because fine-grained material is transported even during small floods, and most gravel is not, the total yearly transport of fine-grained material appears to balance that estimated to be eroded from the Buttermilk-Bond reach plus an added, unmeasured contribution from the tributaries and upper Buttermilk Creek. Additional information is needed on the tributary contribution, particularly the Frank's Creek drainage.

Sediment Loss in the Buttermilk Drainage Basin

A sediment loss value derived for the reservoir drainage basins can be applied to the total Buttermilk drainage basin with the understanding that the relationship of sediment loss to basin area may not be linear.

Reservoir No. 1 (South). The volume of fill, including delta plain to cross-profile 9/22, is: 12515 m^3 (Fig. 9B). Infill has occurred from 1963 to 1980 (17 yrs).

Volume of infill per year is:

$$\frac{12515\ m^3}{17\ yrs} = 736.2\ m^3yr^{-1}$$

The drainage basin of the south reservoir is 806.8 ha. A simple calculation of amount of sediment supplied per year per unit area indicates a sediment loss rate in the drainage basin (Gregory and Walling, 1973).

Drainage basin sediment loss per hectare per year is:

$$\text{South reservoir: } \frac{736.2\ m^3yr^{-1}}{806.8\ ha} = 0.91\ m^3ha^{-1}yr^{-1}$$

The average of both reservoir basins is: $0.89\ m^3ha^{-1}yr^{-1}$.

Buttermilk Creek. The sediment loss per unit area per year in the Buttermilk drainage basin is:

$$7841.5\ ha \cdot 0.89\ m^3ha^{-1}yr^{-1} = 6979\ m^3yr^{-1}$$

The sediment loss result compares well with the simple denudation rate (Table 1). The larger value is to be expected because it includes the tributary and upper Buttermilk Creek sediment contribution.

HOLOCENE LANDSCAPE EVOLUTION

Tributary Development

The larger tributaries of Buttermilk Creek are inherited from the late-glacial and early Holocene drainage systems. The segments of the tributaries aligned parallel to Buttermilk Creek originally flowed as separate streams down the 3 m km^{-1} paleoslope toward Cattaraugus Creek. These parallel segments are now entrenched and link with upper drainages that are incised within, or at the margin of, the Holocene alluvial fans of Lafleur (1979).

Some of the smaller tributaries head in the uplands adjacent to Buttermilk, but others began as small fans on the Buttermilk valley wall. Headward erosion of the upper drainage results in incision of the Lavery till plateau. Stream capture, such as may have occurred to the Frank's/Erdman system, can redirect stream patterns and result in rejuvenation due to lowered base-level.

Figure 3 illustrates a range of gradients of longitudinal profiles of streams in the Buttermilk basin from the steep BC-3 alluvial fan, to the lower-gradient Buttermilk Creek. The middle example, Frank's Creek, can be subdivided into morphologically distinct segments above and below the knickpoints of the Erdman Creek section. The valley above the knickpoints is not being actively incised at the present time. The valley walls appear to have mass-wasted, either by earthflow or soil creep, onto the valley bottom. The flat floor of the valley is not composed of gravel terraces, but consists of hummocky till with tension cracks. The incision will resume as the knickpoints progress up the valley.

Erdman Brook, below the knickpoints, and Frank's Creek are undergoing active incision resulting in an extreme V-shaped cross-profile. Terraces are rare along the Frank's Creek segment, but do exist along Erdman Brook. A small fan-shaped bar complex is present at the mouth of Quarry Creek, perhaps the forerunner of a terrace array. The reason for the steeper gradient along this section is unclear. As downcutting continues, both Frank's and Erdman valleys can be expected to widen by parallel retreat of slopes due to slumping of wall material and rapid removal by flood events.

Future Evolution

The base-level of Buttermilk Creek is controlled by the elevation of Cattaraugus Creek at the Buttermilk confluence. The Cattaraugus is entrenched in bedrock about one-half kilometer below the confluence, as is Buttermilk near the Bond Road bridge (Figs. 1,2,8). The bedrock retards downcutting of the active channel which in turn results in a decreased gradient and a decrease in sediment transport capacity. The effect of the bedrock temporary base-level is not yet reflected in the gradient of Buttermilk Creek and is interpreted not to be important over the 'middle' term (tens to hundreds of years).

We believe that tributary lowering and widening will occur somewhat independent of the lowering of Buttermilk Creek. The convex profile of Frank's Creek/Erdman Brook is interpreted to mean that it is unstable and will be subject to continued downcutting and widening even if the base-level at the confluence does not change. At some future date, as yet undetermined, a combination of tributary widening, alluvial-fanhead incision, and drainage capture will encroach on the waste-burial site and it will be destroyed.

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ROAD LOG

Itinerary

The trip will consist of an overview and orientation stop; a stop to park vehicles and assemble for the rest of the field trip; and one very long stop that will be a walking tour of selected features. We will cover 4-5 km on foot, including several stream crossings and a climb up and down steep slopes. Water depths may be ankle to knee deep depending on stream discharge. Plan to have wet feet or wear hip boots. We will be on NY State property adjacent to a nuclear reprocessing and waste-burial facility; prior permission to enter the property must be obtained.

Distance (in miles)		Route and Stops
Pt. to pt.	Cum.	
	0.0	Start at Buffalo Marriott Inn parking lot. Turn right onto US 263 south (Millersport Highway).
0.3	0.3	Turn right onto I-290 east. I-290 merges into I-90 (NY Thruway). Follow signs for: Thruway Westbound, Erie).
11.4	11.7	Exit 55. Exit right to US 219 south, Orchard Park, Springville.
23.8	35.5	Expressway ends at junction of US 219 south and NY 39 near Springville. Turn left and proceed east on NY 39.
0.2	35.7	Turn right onto US 219 south.
2.7	38.4	Cross bridge over Cattaraugus Creek gorge, go 200 meters, turn left on Rock Springs Rd.
2.9	41.3	West Valley Nuclear Service Center (WVNSC) entrance. Continue southeast on Rock Springs Rd.
0.3	41.6	Stop on Rock Springs Rd. for overview of WVNSC and associated waste-burial trenches.

STOP 1. Overview of the West Valley Nuclear Service Center. The buildings housing the nuclear fuel reprocessing facility are to the

left, waste-burial trenches are to the right. The land and buildings inside the high-security fence are owned presently by the U.S. Dept. of Energy and operated for them by the Westinghouse Corp. Westinghouse is carrying out the West Valley Demonstration Project, a program to solidify the high-level liquid waste stored on site, for later isolation in a still-to-be chosen facility. The NRC-licensed intermediate-level burial trench (left) is also operated by Westinghouse; the low-level trenches (right) are owned by the NYSERDA and are no longer accepting waste material.

The buildings are situated on an early Holocene alluvial fan; the waste-burial trenches are excavated in late Woodfordian Lavery till (Lafleur, 1979). The plateau containing the trenches marks the surface elevation of Buttermilk valley in early post-glacial time. The plateau beyond the trenches is covered by a thin veneer of fluvial gravel deposited in the ancestral late-glacial to early post-glacial Buttermilk Creek. Initial incision and downcutting began sometime before 9900 BP (Fig. 8), and continues to the present. The field trip discussion will focus on the manner and rate of erosion that will, in time, remove the plateau, and with it the waste-burial trenches.

0.2	41.8	Old schoolhouse at junction of old Buttermilk Creek Rd.
0.2	42.0	Turn left on Thornwood Drive. Proceed toward the hamlet of West Valley.
1.1	43.1	Look for south reservoir in valley on the left.
0.1	43.2	Turn left on Fox Valley Rd. Cross Baltimore and Ohio railroad tracks.
1.0	44.2	Turn left on NY 240 north. Go to hamlet of Riceville.
0.1	44.3	Turn left on Buttermilk Rd. Dead end.
0.8	45.1	New York State Plutonium Storage Facility. End of paved road. STOP. Obtain prior permission to proceed on dirt road. CAUTION. This road is rough and steep. Trucks and/or 4 WD vehicles needed in wet weather. Go downhill and cross B & O RR tracks.
0.4	45.5	Old Buttermilk Rd. and B & O RR work road. Watch for trains, this is a well-traveled way.

STOP 2. Assembly point. Park vehicles so they do not block the B & O work road or access up and down the hill. Old Buttermilk Hill Rd. continues to Buttermilk Creek (100 m), the bridge has been removed. The surveyed section (Buttermilk-Bond reach) begins at the former site of the bridge and extends downstream to the north. Reservoir slug discharge enters from the west just below the bridge site.

	0.0 km	Walk north along the B & O RR tracks. Pass section of double track.
0.35	0.35 km	Active channel of Buttermilk is undercutting railroad embankment. View of bar complexes and low terraces.
0.75	1.10 km	Small railroad bridge over tributary stream. Turn left (downstream) and follow stream to Buttermilk Creek.
0.10	1.20 km	Buttermilk Creek valley bottom.

STOP 3. Gravel-bar complexes, valley-wall landslides and alluvial fans, and Buttermilk fluvial terraces. The tributary stream enters Buttermilk Creek at bar complex 6, opposite the BC-6 landslide (Fig. 7). Extending 250 m upstream, to your left or south, is the BC 4-6 bar complex (Fig. 5). Just upstream of BC 4-6 is the BC-3 alluvial fan, situated on the west wall of the valley (Fig. 6). Fluvial terraces at several levels are preserved on the east side of the valley at the BC-6 tributary and on the west side adjacent to the BC-3 alluvial fan (Fig. 8). We shall discuss each of these features in turn, including gravel movement on the bars (Fig. 12). Please take care not to disturb any of the painted clasts.

Return up the tributary and along RR tracks to the vehicles. End of trip.

DEVONIAN BLACK SHALES OF WESTERN NEW YORK

DOUGLAS G. PATCHEN
KATHARINE LEE AVARY
West Virginia Geological Survey
Morgantown, WV 26507

INTRODUCTION

The road log that follows is essentially a condensed version of the final day of a 5-day trip through the Middle and Upper Devonian clastics of central and western New York (Patchen and Dugolinsky, 1979). That particular trip was organized to give various contractors in DOE's Eastern Gas Shales Project (EGSP) the opportunity to examine, in the field, stratigraphic relations and facies changes hidden deep in the subsurface in the central Appalachian basin. The stops on the fifth day were designed to simulate a 2,200 to 2,500 foot stratigraphic section of "Devonian shale" in a typical "Brown Shale" gas well in West Virginia, Kentucky, or Ohio. The stops on this trip will enable us to visualize what the lowest 1,000-1,100 feet of section in such a well would look like. The main difference will be that the Dunkirk Shale, which is only 40-50 feet thick in this area, thickens westward and merges with the basal black shale of the Huron Shale. In southwestern West Virginia this black shale ranges from 300-400 feet in thickness and serves as both source bed and reservoir in more than 30 gas fields. More recently, prolific oil wells with initial potentials (IP's) ranging from 50 to 1,000 barrels per day have been drilled to this and other shales in the Devonian section in several counties along the Ohio River (Fig. 1). Thus, even though the EGSP is now defunct, interest in the Devonian Shales, both in the surface and subsurface, remains high.

When we agreed, at a rather late date, to organize this trip, we secured from Art Van Tyne a promise to submit a companion article for the guidebook discussing surface to subsurface correlations and gas production from Devonian shales in New York. Furthermore, Wallace de Witt informed us that he would present a paper on Devonian shale stratigraphy at the Eastern Section meeting of the American Association of Petroleum Geologists that immediately precedes the NYSGA field trips. Therefore, we refer the readers to their articles and will present only the barest of stratigraphic summaries herein.

STRATIGRAPHY

Devonian rocks in New York have been studied for more than 100 years and have emerged as the standard reference section for the eastern United States. The resulting physical stratigraphic framework, biostratigraphic zonation, and facies relations of the Catskill clastic wedge are all illustrated in the most recent Devonian correlation chart (Rickard, 1975). On this chart black shales are shown to parallel biostratigraphic zones and, therefore, can be used not only as time lines but to define the bases of thick time-stratigraphic units as well (Fig. 2).

In western New York, each of the main stratigraphic units (Hamilton, Genesee, Sonyea, West Falls, Java, and Canadaway) to be observed on the trip is defined at the base by a black shale (Marcellus, Geneseo, Middlesex, Rhinestreet, Pipe Creek, and Dunkirk, respectively). However, due to time restrictions, the 2 oldest black shales (Marcellus and Geneseo) will not be observed.

The oldest time-stratigraphic unit that we will see on this trip, the Hamilton Group, is divided farther to the east by 3 persistent limestones (Stafford, Centerfield, and Menteth/Portland Point) into 4 formations: the Marcellus, Skaneateles, Ludlowville, and Moscow. In the Buffalo area, the Hamilton units have thinned drastically to less than 225 feet and are cut by at least 2 unconformities. Black shales, the Oatka Creek Member of the Marcellus and Levanna Member of the Skaneateles, are confined to the lower half and are difficult to observe in this relatively flat area. Exposures of the younger Ludlowville and Moscow are more accessible, and 3 members (Wanakah Shale and Tichenor Limestone Members of the Ludlowville; Windom Shale Member of the Moscow) will be observed in the gorge of Eighteenmile Creek (Fig. 3, Stop 1).

The base of the Genesee Formation is defined by the black Geneseo Shale Member east of Buffalo. However, in the Buffalo area the entire formation ranges from 10-20 feet in thickness (Tesmer, 1966) and consists of the black Geneseo Shale (0-2 feet), dark gray Penn Yan Shale (0-1 feet), Genundewa Limestone (2 inches to 2 feet), and gray West River Shale (8-14 feet). A local limestone facies of the lower Penn Yan Shale, the North Evans Limestone, can be observed near the mouth of Pike Creek where it is less than 1 foot thick and unconformably overlies the Windom Shale, the Geneseo Shale being absent. The Geneseo also is absent in the gorge of Eighteenmile Creek where only the Penn Yan, Genundewa, and West River will be observed.

The base of the overlying Sonyea Formation is marked by the black Middlesex Shale Member, which is only 6-8 feet thick in the Buffalo area (Tesmer, 1966). The overlying Cashaqua Shale Member is greenish-gray with calcareous concretions, and is about 40 feet thick along Eighteenmile Creek.

The Cashaqua Shale is overlain in the Buffalo area by the black Rhinestreet Shale, the basal member of the West Falls Formation. The Rhinestreet is about 195 feet thick in the area (Pepper, et al, 1956), and is typically very dark, platy, and organic-rich. Very large concretions, 2-3 feet in diameter, are characteristic of the Rhinestreet in this area. The younger Angola Shale Member of the West Falls is about 250 feet thick in the area and is greenish gray with smaller concretions.

The thin Pipe Creek Shale Member overlies the Angola Shale and marks the base of the Java Formation. Whereas the Pipe Creek is 20 feet thick farther east near Java village, it is less than 2 feet thick near Lake Erie. The overlying Hanover Shale Member consists of greenish-gray, thinly bedded shales with concretions, and becomes more silty in the upper one third.

The black Dunkirk Shale overlies the Hanover Shale and is the basal member of the Perrysburg Formation. As defined by Pepper and de Witt (1951) the Perrysburg consists of the Dunkirk, South Wales, Hume, and Gowanda Members, although Rickard's Devonian correlation chart (1975) includes only the Dunkirk and Gowanda.

The overlying Laona Sandstone marks a significant change in the Upper Devonian rocks along Lake Erie. Below this, the Genesee, Sonyea, West Falls, Java, and lower Canadaway rocks consist mainly of black or greenish gray shales with minor limestones and a few siltstones. Above the Laona, the rocks of the Canadaway, Conneaut and Conewango Groups are predominantly gray shales and siltstones with increasing numbers of sandstone beds toward the top. Thin black shales are common only in the upper Canadaway, and there they are interbedded with gray shales, none of which is thick enough to warrant member status. These youngest black shales probably thicken westward, like the Dunkirk, and eventually merge with the black Huron Member of the Ohio Shale.

REGIONAL CORRELATIONS

Subsurface correlations by Schwietering (1970, 1979) between Devonian outcrops in central Ohio and western New York have established the physical correlation of the Dunkirk and basal Huron, and the facies relations of the Canadaway, Conneaut and Conewango with the Ohio Shale. These correlations can be extended southward through Ohio into eastern Kentucky and southern West Virginia. In those areas the youngest, coarsest, gray rocks (sandstones, siltstones, and shales) are usually termed "Upper Devonian undifferentiated" (Neal, 1979; Dowse, 1980), but the black shales, because they can be traced physically on gamma-ray logs, are correlated and named with either the Ohio or New York outcrop terminology (Fig. 4).

The older black shales in western New York, the Geneseo and Middlesex, thin to the west and southwest, and thicken eastward. However, these units can be recognized on gamma-ray logs in western West Virginia above the Middle Devonian unconformity. The thin Pipe Creek Shale also has been found to be a useful marker bed in the subsurface (Neal, 1979; Dowse, 1980). The oldest black shale, the Marcellus, is not present, due to unconformity, in western West Virginia, eastern Kentucky, and Ohio. The thick Rhinestreet Shale section in part of that area has traditionally been called "Marcellus" by drillers because of its close proximity to the top of the Onondaga Limestone where the Hamilton to Sonyea equivalents are missing. Farther west, in Ohio, the Rhinestreet also is cut out by the unconformity, leaving only the Ohio Shale and Olentangy Shale present above the Onondaga equivalents (Schwietering, 1970; 1979).

PETROLEUM PRODUCTION

Natural gas has been produced from the Huron (Dunkirk), Rhinestreet, Geneseo, and Marcellus in West Virginia, the Huron in Ohio, and the Huron and probably Rhinestreet in eastern Kentucky. Commercial production, particularly of oil in recent drilling, is due to the extensive development of a fracture system within the black Devonian shales. As you can observe on this trip, the fracture systems change vertically. Dark, organic-rich, lower density black shales are more highly fractured than overlying and underlying greenish-gray "normal" density shales.

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ROAD LOG FOR DEVONIAN BLACK SHALES OF WESTERN NEW YORK

STARTING POINT: The bus will leave the parking lot at the Marriott Inn promptly at 8:00 a.m. The road log begins at the traffic light on Rt. 263 (Millersport Highway) at the Marriott entrance.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	
0.0	0.0	Turn right at the traffic light onto Rt. 263 (Millersport Highway).
0.4	0.4	Turn right after overpass onto the parallel feeder for I-290.
1.4	1.0	Stay left at Sheridan Drive (Rt. 324) exit.
1.7	0.3	Merge left onto I-290 (Youngmann Memorial Expressway)
3.3	1.6	Exposures of Onondaga Limestone along both sides of the highway. The black Marcellus Shale that overlies the Onondaga cannot be readily observed in the Buffalo area due to the relatively flat topography near Lake Erie.
3.8	0.5	Stay right for I-90 west toward Erie; continue on I-90 west.
9.5	5.7	Stay left on I-90 as I-190 goes to the right.
14.1	4.6	Stop; get ticket for the NY Thruway.
18.4	4.3	Town of Hamburg water tower. Good exposure of black Rhinestreet Shale with large concretions.
19.5	1.1	Exit 57 (Hamburg); leave I-90.
20.1	0.6	Stop; pay toll and then turn right toward Rt. 75 north.
20.4	0.3	Merge onto Rt. 75 northbound.
20.9	0.5	JCT with U.S. 20; continue straight through intersection.
22.2	1.3	Stay left on Camp at traffic light as Rt. 75 turns right.
22.4	0.2	JCT with Rt 5. Turn left (south).

23.1	0.7	Entrance for Hamburg Park. Exposures of the Ledyard and Wanakah Shales of the Ludlowville Formation can be observed near lake level in the park but not from the highway. Stop 1 of Tesmer (1966); Stop 47 of Patchen and Dugolinsky (1979).
25.2	2.1	Wanakah
26.4	1.2	Turn right on Lake Shore Road at the traffic light (Rt. 5 goes straight).
29.6	3.2	Bridge over Eighteenmile Creek.
29.7	0.1	Turn left on South Creek Road; park on right side of road. Walk carefully back along the road and over the bridge.

STOP 1: EIGHTEENMILE CREEK GORGE

Upstream from the bridge over the main highway are exposures of the Wanakah Shale, Tichenor Limestone, Windom and Penn Yan Shales, Genundewa Limestone, and a few feet of West River Shale. The black shale near the top of the outcrop, which is weathered a reddish color and displays well-developed fractures, is the base of the Middlesex Shale. Notice the weathered pyrite associated with the base of the Tichenor Limestone, along with the abundant fossils, both characteristic of the Tichenor.

It is possible to walk from the exposures on the beach at the mouth of this stream where the Wanakah and Tichenor are present, upstream past this point and see the Middlesex, Cashaqua and Rhinestreet. Thus, continuous exposures from the Ludlowville Formation of the Upper Hamilton Group, through the Moscow Formation, the Genesee and Sonyea Formations, and into the West Falls Formation are present and all visible, representing a large proportion of Middle and Upper Devonian time.

The following 2 stops will be farther upstream from this locality.

Continue east on South Creek Road.

30.0	0.3	JCT with Rt. 5. Cross and park on right. Descend gorge under the bridge.

STOP 2: RT. 5 BRIDGE OVER EIGHTEENMILE CREEK

The Wanakah Shale is exposed in the stream bed and may be under water, depending on water level. This is followed upward by one or two feet of Tichenor Limestone, about ten feet of Windom and Penn Yan Shale, some thin Genundewa Limestone, about 3 feet of West River Shale, a few feet of dark, platy Middlesex Shale, and the lower few feet of Cashaqua Shale which cap the section. The entire section is perhaps 35 to 40 feet thick and all is accessible.

Continue east on South Creek Road.

30.9	0.9	Railroad Bridge. Short walk along path to left before RR bridge offers a view of upper Cashaqua and lower Rhinestreet (Stop 3 of Tesmer, 1966). Continue on South Creek Road.
31.8	0.9	JCT with U.S. Rt. 20. Turn left, drive slowly across bridge in right lane; turn around in parking area to right and drive back. Observe excellent, thick exposures of black Rhinestreet Shale in stream banks.
32.1	0.3	Turn right onto South Creek Road.
32.4	0.3	Turn right on Versailles at intersection. Drive down into gorge.
32.6	0.2	Park at Town of Evans Nature Park.

STOP 3: TOWN OF EVANS NATURE PARK

Exposures of the Rhinestreet may be seen along the road and adjacent to the parking lot by the pedestrian bridge. The base of the Rhinestreet is exposed farther downstream but not here. The shale here is the typical dark, platy shale characteristic of the Rhinestreet, with reddish weathering in places and concretions being very common, although not as evident at this locality as they are in Rhinestreet exposures along I-90 southeast of Hamburg.

		Return to South Creek Rd. on Versailles.
32.8	0.2	Turn right onto South Creek Rd. Return along Eighteenmile Creek, across Rt. 5 to Lake Shore Road.
34.7	1.9	Lake Shore Rd. Turn left.
34.8	0.1	Entrance to home of Piarist Fathers, Stop 48 of Patchen and Dugolinsky (1979); Stop 2 of Tesmer (1966). Exposures of the Wanakah Shale, Tichenor Limestone and Windom Shale are present along the beach. If anyone is interested in examining this site, permission should be obtained from the residents. Also, access to the beach is by way of an old metal stairway. Extreme caution should be used.
36.5	1.7	Gibralter on the Lake (PRIVATE ROAD). Stop 52 of Patchen and Dugolinsky (1979). The Windom Shale to Middlesex Shale interval is exposed along the beach.
36.6	0.1	Bear right at the "Y" intersection.
39.5	2.9	Lake Shore Rd. turns right at Dennis Rd. Stay on Lake Shore Rd.

41.2	1.7	Prominant sand ridges to the right.
45.5	4.3	Turn left at entrance to Evangola State Park. This was Stop 53 of Patchen and Dugolinsky (1979); Stop 5 of Tesmer (1966).
46.7	1.2	JCT with Rt. 5. Turn right.
47.3	0.6	Farnham
49.4	2.1	JCT with U.S. Rt. 20. Go right on U.S. Rt. 20 and Rt. 5.
49.5	0.1	Cattaraugus Creek. Chautauqua County line.
52.2	2.7	Rts. 5 and 20 split. Go straight on U.S. Rt. 20.
52.6	0.4	Silver Creek
53.2	0.6	Turn left at traffic light in center of village (Rt. 5 goes straight).
53.5	0.3	JCT with Rt. 428; continue on U.S. Rt. 20.
53.7	0.2	Bear right on Main St.
53.8	0.1	Turn right on Ward St.
53.9	0.1	Turn right on Parkway.
54.0	0.1	Park along Parkway near fire hydrant. Walk through small lot to edge of Walnut Creek. Descend into creek bed. Observe Pipe Creek Shale above Angola Shale. Wade under bridge to 60 ft. exposure of Hanover. Dunkirk Shale is exposed 2 miles upstream.

STOP 4: WALNUT CREEK IN TOWN OF SILVER CREEK

The contacts of the Pipe Creek Shale with the underlying Angola Shale and the overlying Hanover Shale are exposed in the stream bed of Walnut Creek in the Town of Silver Creek. The top 5 feet or so of the Angola is all that is clearly exposed, but the entire Pipe Creek and the lower 60 feet of Hanover are accessible, although wading boots are suggested. Lithologies in the Hanover range from green-gray, thinly bedded shales to silty beds which are more prominent due to higher resistance to weathering.

		Continue west on Parkway.
54.5	0.5	JCT Rt. 5 (Central Avenue). Turn left (south). Continue south along Lake Erie into Dunkirk.
63.8	9.3	JCT Rt. 60 (to left) with Rt. 5. Continue straight on Rt. 5.

64.9	1.1	Turn right at 3rd traffic light where sign says "Point Gratiot ahead".
65.3	0.4	Turn right (just after paved entrance to power station) into unpaved parking area and park.

STOP 5: POINT GRATIOT, DUNKIRK

The base of the Dunkirk is well exposed on the beach just south of the power plant. The distinct transition zone between the Hanover and the Dunkirk is at, and just beneath (but visible in clear water), the water at normal lake level. The greenish-gray beds of the Hanover and the black beds of the Dunkirk are separated by two black and two greenish-gray transition beds. A small flexure may be seen, particularly if the water is calm, by tracing the lighter colored units along the beach and noticing a slight northward dip in this location.

Feeding trails and carbonized plant fragments may be seen on the tops of the beds within the transition zone. Pyrite nodules are common. A freshly broken surface of Dunkirk Shale provides a strong petroleum odor.

65.8	0.5	Return to Rt. 5. Turn left.
66.8	1.0	Turn right on Rt. 60. Stay on Rt. 60.
68.4	1.6	Cross over the thruway. Turn left and enter the thruway north toward Buffalo (Interchange 59). Along the route notice new gas wells in vineyards to east. These are in the Medina sands (Lower Silurian).
77.7	8.3	Cross bridge adjacent to an old bridge on a parallel route to the right (east). This was Stop 55 of Patchen and Dugolinsky (1979). About 55 feet of Dunkirk Shale are exposed in the gully under the bridge. The transitional zone between the Hanover and Dunkirk can be observed in the stream bed.
82.1	4.4	Exit 58 (Silver Creek). Observe excellent Hanover-Dunkirk contact on the right.
90.2	8.1	Cross gorge with fairly poor exposure.
95.5	5.3	Cross gorge with a good section exposed.
101.3	5.8	Exit 57 (Hamburg).
102.8	1.5	Excellent Rhinestreet exposure along both sides of the thruway near the town of Hamburg water tower. The large concretions are characteristic of the Rhinestreet in this area. (Same as mile point 18.4.)

104.0	1.2	Rhinestreet-Cashaqua contact in the ditch to the east.
105.3	1.3	Blasdell Interchange (exit 56).
111.2	5.9	JCT with I-190 at exit 53. Continue on I-90 (free portion).
116.9	5.7	Exit I-90 at Interchange 50 for I-290 north (Youngmann Memorial Expressway). Return to the Marriott on I-290 and Rt. 263 (Millersport Highway).
120.7	3.8	Marriott Inn. End of trip.

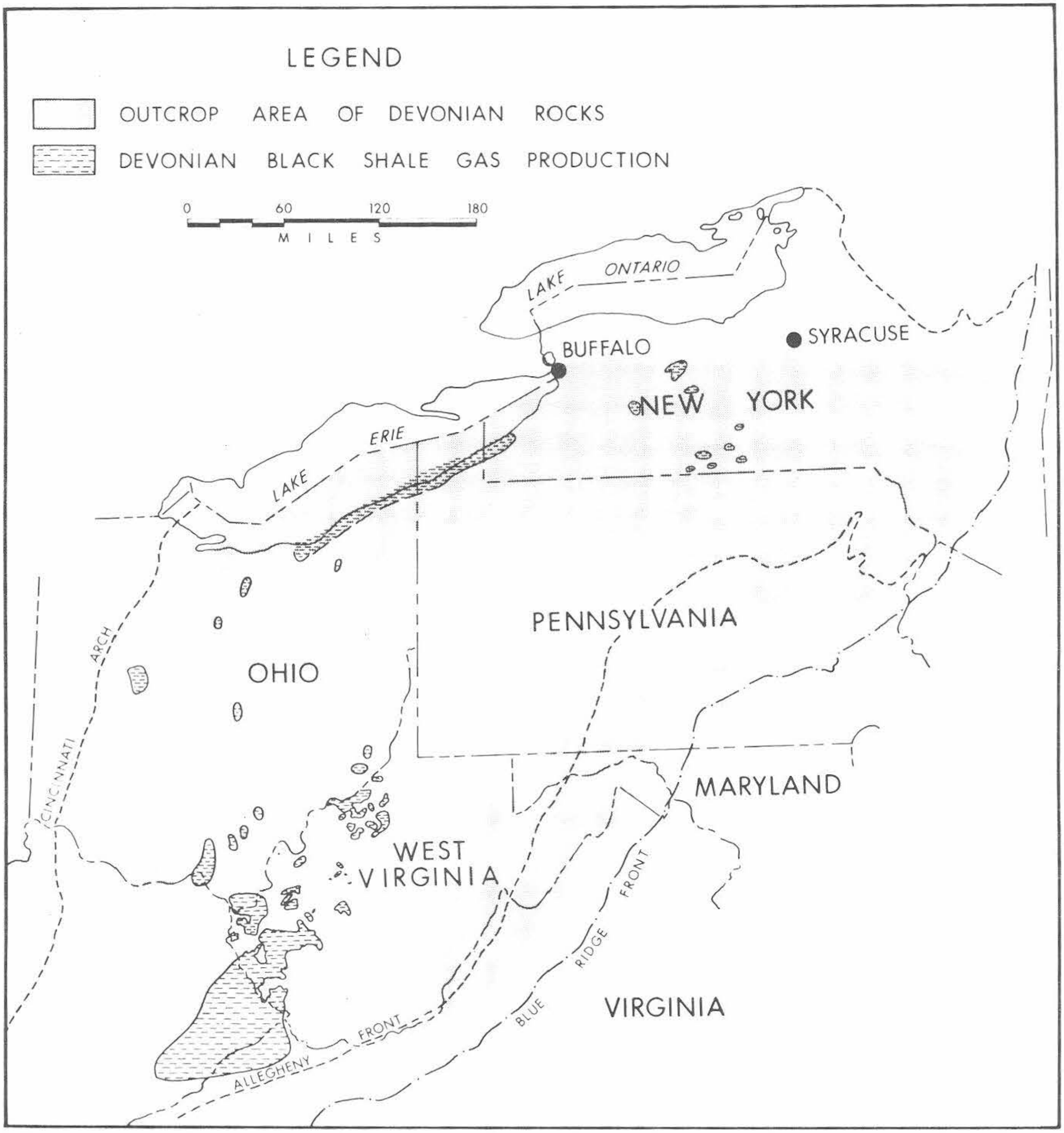

Figure 1. Regional outcrop pattern of Middle and Upper Devonian clastics and generalized locations of Devonian shale gas (and oil) fields.

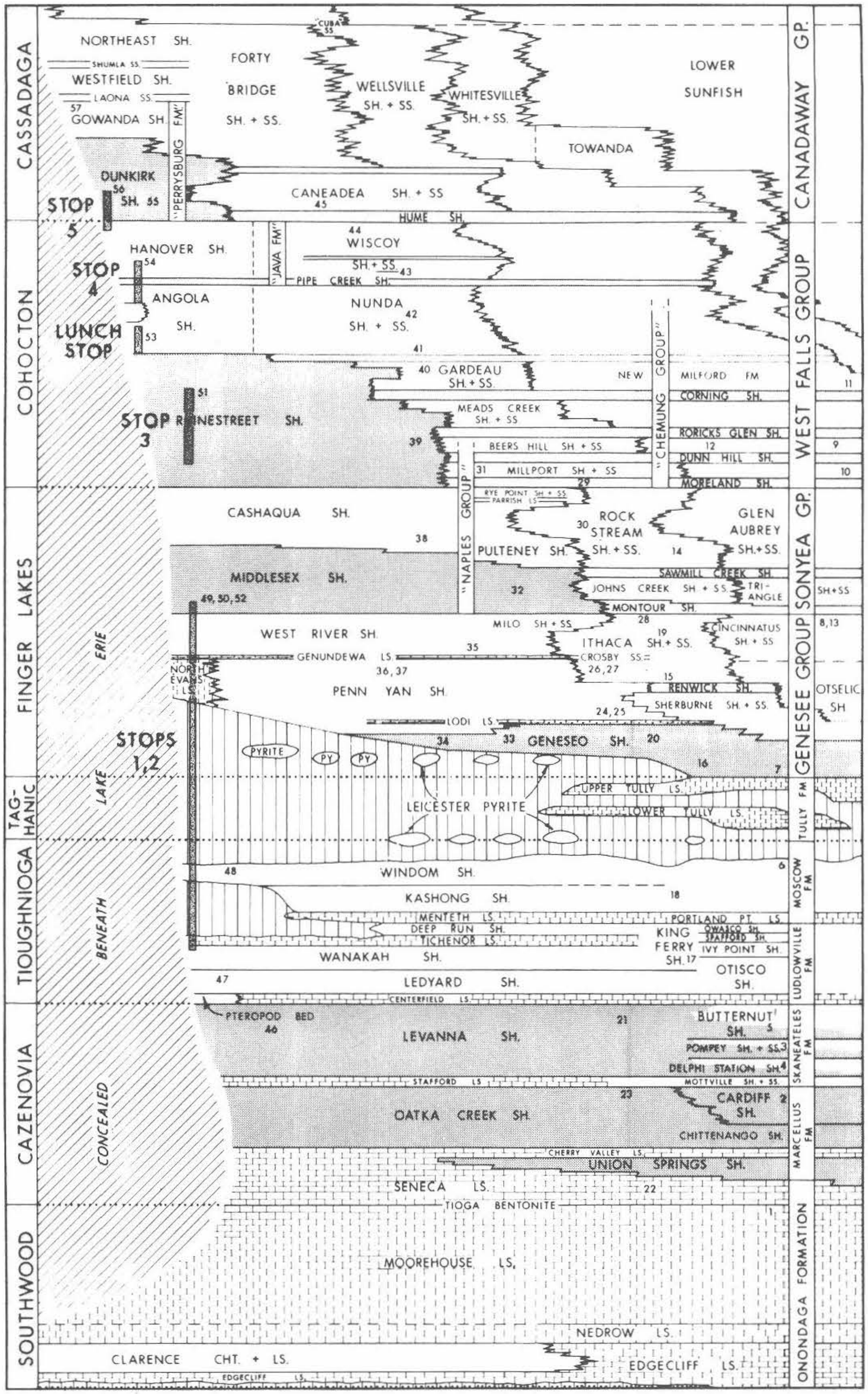

Figure 2. Middle and Late Devonian correlation chart for central and western New York (modified from Rickard, 1975). Stratigraphic ranges of stops are indicated; vertical scale is time, not thickness.

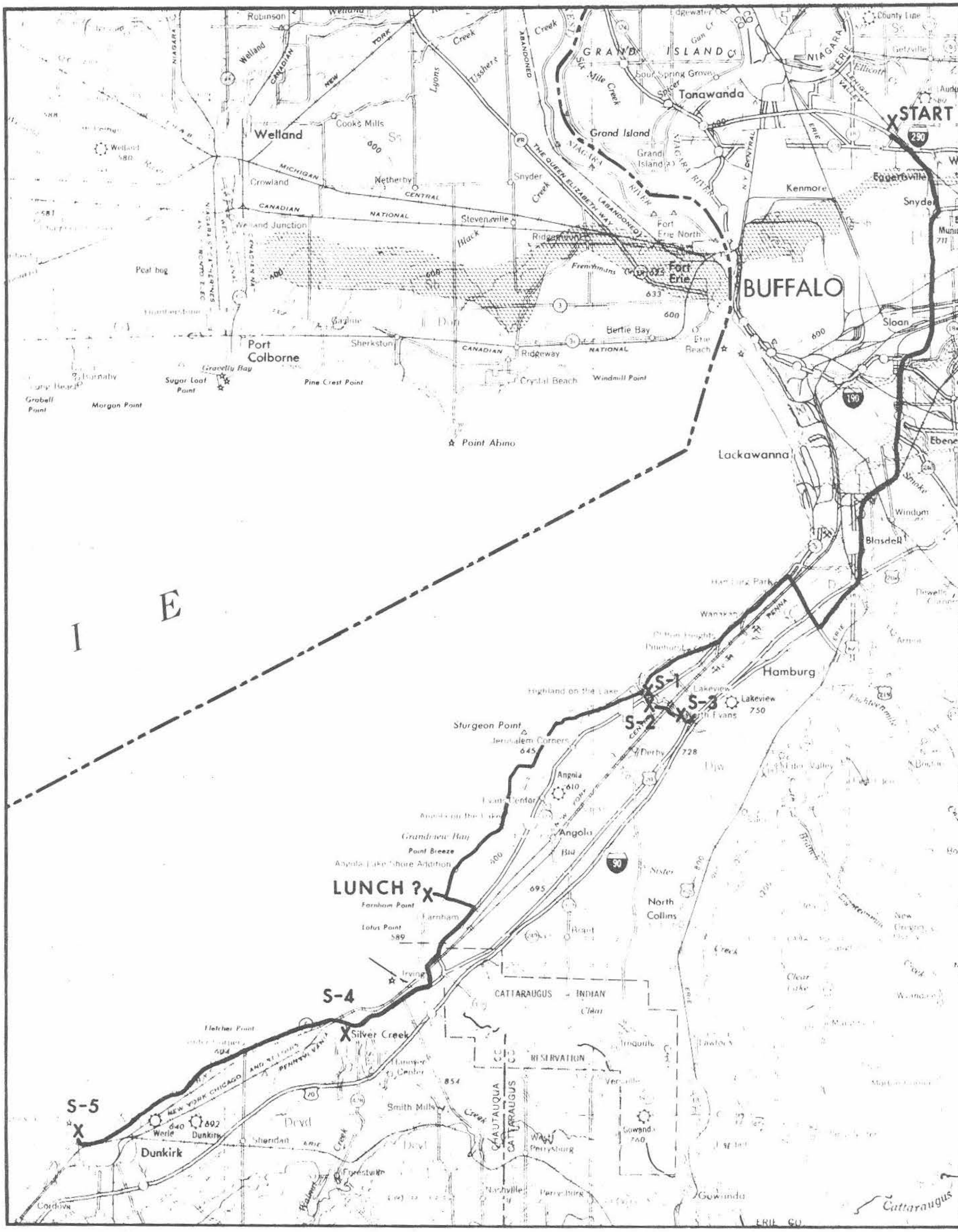

Figure 3. Field trip route and stop locations.

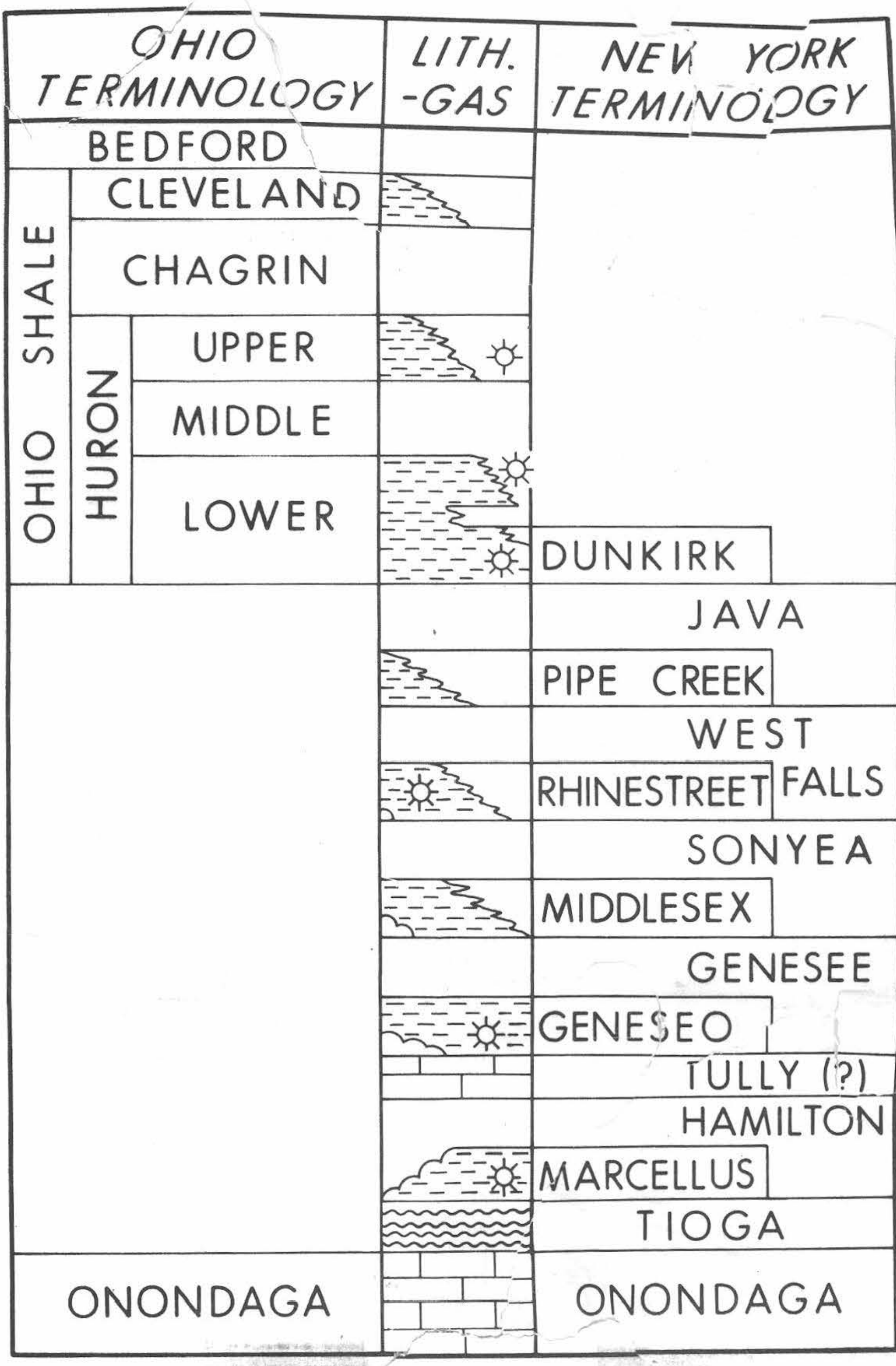

Figure 4. Terminology used in subsurface mapping of black Devonian shales in the central Appalachians. New York names are used for older shales; Ohio names for younger shales.

FIGURE CAPTIONS

Figure 1. Regional outcrop pattern of Middle and Upper Devonian clastics and generalized locations of Devonian shale gas (and oil) fields.

Figure 2. Middle and Late Devonian correlation chart for central and western New York (modified from Rickard, 1975). Stratigraphic ranges of stops are indicated; vertical scale is time, not thickness.

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SUBSURFACE EXPRESSION AND GAS PRODUCTION OF DEVONIAN BLACK SHALES IN WESTERN NEW YORK

Arthur M. Van Tyne - Consulting Geologist, Wellsville, New York 14895

INTRODUCTION

Recent work attempting to refine our understanding of the Middle and Upper Devonian black shales of the Appalachian Basin has been undertaken by a number of agencies and individuals in several eastern states as part of the U. S. Department of Energy's Eastern Gas Shales Project. The EGSP was initiated in 1976 as part of the U. S. Department of Energy's Unconventional Gas Recovery Program. This multi-year project was designed to study the Devonian black shales of the eastern United States, to develop new and improved methods of exploiting their indigenous gas reserves, and to promote commercial development of this gas resource. The project is part of a long-range plan to increase the ultimate recovery of gas from these shales in an attempt to supplement local supplies and to lessen United States dependency on foreign energy sources.

The New York portion of this study was accomplished by a team under the direction of the author at the Alfred Oil and Gas Office of the Geological Survey, New York State Museum. Dr. L. Rickard of the Albany office of the survey was associated with this group during the early phases of the project. Approximately 18,000 square miles of the State are underlain by one or more Devonian black shale formations.

STRATIGRAPHY

The Devonian black shales of New York are included in the Middle Devonian Hamilton Group and the Genesee, Sonyea, West Falls and Canadaway Groups of the Upper Devonian (See fig. 1). The following discussion of these groups and their black shales summarizes the pertinent subsurface stratigraphy that we found from studies of gamma ray well logs and drill cuttings samples. It appears that the black shale tongues are a reasonable approximation of time planes. The tongues probably represent a single episode of deposition and make it possible to reconstruct the depositional history of this complex section. Sutton (1963) has previously suggested that the Upper Devonian black shale tongues, "appear to be the closest approach to time planes".

MIDDLE AND UPPER DEVONIAN STRATIGRAPHY OF WESTERN AND WEST-CENTRAL NEW YORK

SHOWING MAJOR BLACK SHALE UNITS

(NOT TO SCALE)

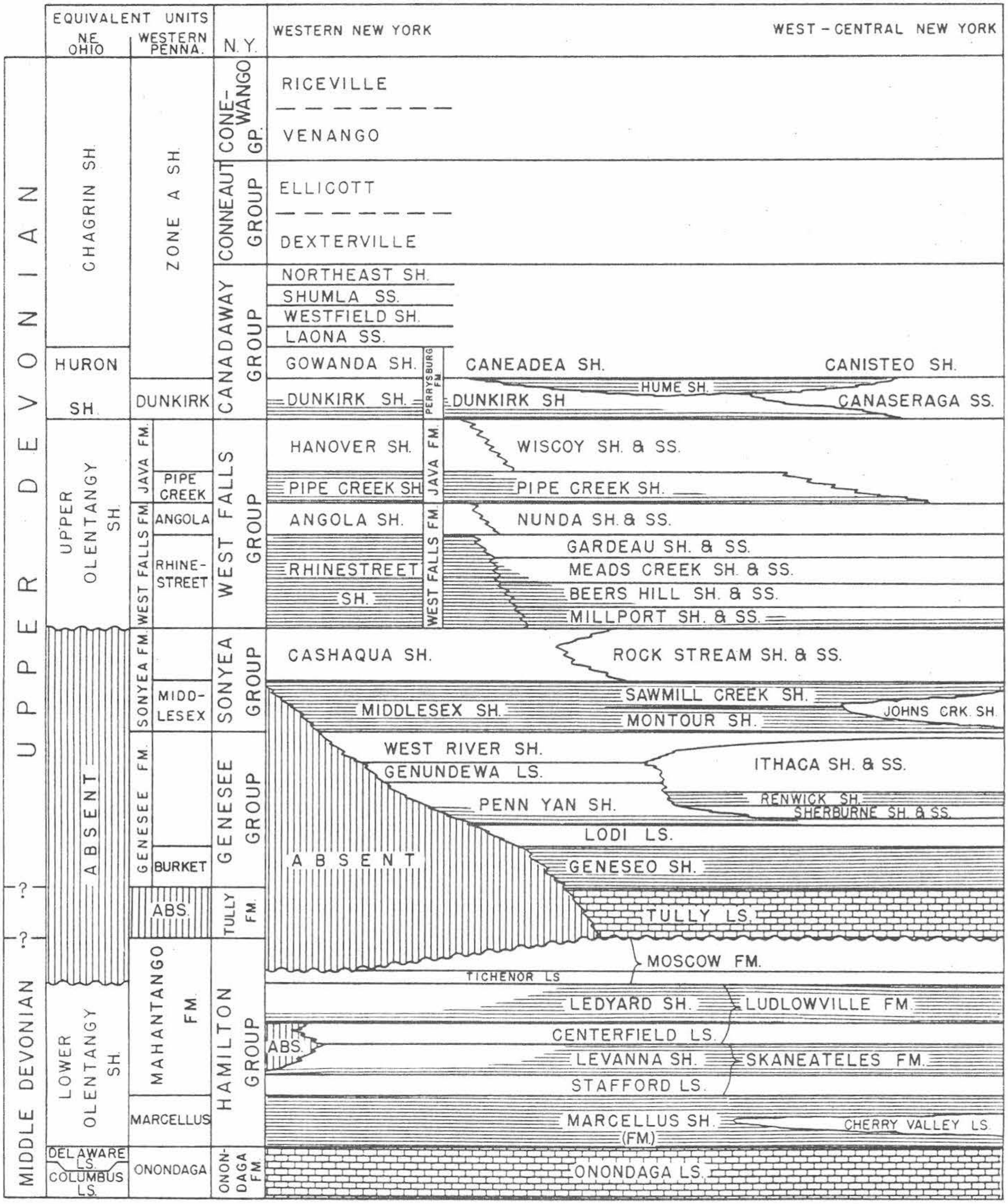

FIGURE 1

Hamilton Group

The base of the Hamilton (base Marcellus), the lowest of the units studied, lies conformably upon the Onondaga limestone (Rickard, 1975). In ascending order, the group is composed of the Marcellus, Skaneateles, Ludlowville and Moscow Formations. At the outcrop, the three formations above the Marcellus have been delineated by persistent basal limestone beds and these are also traceable in the subsurface. The Stafford limestone lies at the base of the Skaneateles and the Centerfield limestone forms the base of the Ludlowville. Recent work by Baird (1979) has shown that the Tichenor-Portland Point is the basal limestone bed of the Moscow. Previous work by Cooper (1930) had placed the Tichenor within the upper Ludlowville as shown by Rickard (1975). The top of the Hamilton (top of Moscow) is at a regional unconformity overlain by the Tully limestone from eastern to west-central New York (where the Tully wedges out) and by progressively younger beds westward to Lake Erie (fig. 1). The total thickness of the Hamilton increases from 235 feet at Lake Erie to more than 2,800 feet in Ulster and Greene Counties in eastern New York, a distance of about 300 miles.

The Marcellus Formation consists mainly of massive black shales. The basal Union Springs black shale is the most highly radioactive shale in the entire New York Devonian section. It is recognized by a distinctive strong righward deflection on a gamma ray well log. The Union Springs is overlain by the thin, dark Cherry Valley limestone which is quite distinctive in eastern and central New York but which passes westward into a calcareous shale zone. This occurs in the area of Allegany and Livingston Counties in western New York. At the outcrop it is not present west of the Geneva area in Ontario County.

Above the Cherry Valley in eastern New York, the massive Marcellus black shales are represented by the Chittenango shale. This shale passes westward into the Oatka Creek black shale which is overlain by the thin Stafford limestone, the basal member of the succeeding Skaneateles Formation. Our correlations indicate that the Oatka Creek thickens in Chautauqua County, the western part of Erie County, and northwestern Cattaraugus County. Eastward from Cattaraugus County the Marcellus gradually thickens.

The three thin limestone beds in the lower Hamilton (Cherry Valley, Stafford and Centerfield) are convenient for subdividing the group but have caused some confusion in gamma ray log correlation work. In eastern and central New York, the Cherry Valley splits the Union Springs from the overlying Chittenango-Oatka Creek. The sooty, black Union Springs shale thins to disappearance in western New York as does the overlying Cherry Valley limestone. As these units disappear, the next higher limestone in the sequence, the Stafford, becomes a prominent marker between the Oatka Creek and the overlying black Levanna shale. Because of its stratigraphic position and the similarity of the gamma ray log signatures, the Stafford has often been called Cherry Valley. From central Erie and Cattaraugus Counties westward, the Centerfield becomes less prominent and converges toward the Stafford as the Levanna shale thins. Some workers feel that they merge to the west with the Stafford thinning out and the Centerfield continuing. The data available for this study appeared to indicate that the Centerfield wedges out as the two limestones converge while the Stafford continues to the west. Further detailed work with better log control from later drilling in the western Erie and Cattaraugus County areas, where this critical change takes place, will probably resolve this disagreement.

The Skaneateles Formation, which overlies the Marcellus Formation, consists of the basal Stafford and the dark gray to black Levanna shale in the area from central to western New York (fig. 1). On a gamma ray log the Levanna is the most highly radioactive shale above the Marcellus in western New York. Eastward from central New York, the Levanna becomes less black and contains interbedded sandy shale and siltstone members.

The Centerfield limestone forms the basal unit of the succeeding Ludlowville Formation. From central to western New York, this is overlain by the Ledyard black shale. In that area, the Ledyard does not always show a high response on the gamma ray log, but sample studies show that the shale is dark gray to black in color. Shales above the Ledyard are lighter gray, calcareous and contain numerous thin limestone beds. For a discussion of these shales and the proposed regional paraconformity at the top of the Ludlowville, reference is made to the paper by Baird (1979).

The Moscow Formation is at the top of the Hamilton Group. As proposed by Baird, the Tichenor forms the base of the Moscow and is overlain by calcareous gray shales. In western New York, the Moscow is progressively cut out by the overlying basal Tully unconformity.

According to Heckel (1973), the disconformity at the Hamilton-Tully contact in central and western New York is entirely of submarine origin. In eastern New York, the Hamilton and overlying Tully clastic equivalent is conformable. Tully limestone was deposited in central and central-western New York but pinches out along a westward and southwestward trending line from Canandaigua Lake through Livingston and Wyoming Counties then southward through eastern Cattaraugus County (METC/EGSP Series 120). Gradual deepening of the sea and an increase in reducing conditions brought about the demise of carbonate producing organisms and the cessation of Tully limestone deposition. The black muds which were to become the Geneseo shale were deposited on top of the Tully (Heckel, 1973).

The isopach map of the total Hamilton section (METC/EGSP Series 121) shows a gradually increasing rate of thickening of 3 to 4 feet per mile in far western New York. This becomes 6 to 8 feet per mile in the Allegany and western Steuben County areas. At about the Seneca Lake meridian, the rate of thickening increases to 12 to 15 feet per mile. This rate is maintained for about 100 miles to the east into central Delaware County where it becomes 20 to 25 feet per mile. Several east-west oriented troughs and lobes are shown, and these may represent areas of turbidite flows.

Genesee Group

With the close of Tully deposition, clastics from the east began spreading over the Tully and the regional erosion surface to the west. The first sediments formed the black shales and mudstones of the Geneseo.

Overlying the Geneseo black shale is a zone of argillaceous, nodular limestone which has been called the Lodi limestone. This zone, which may be less than one foot to a few feet in thickness, separates the Geneseo from the overlying Penn Yan black shale. The Lodi makes a very distinctive limestone signature on the gamma ray log but is misleading as to actual limestone thickness because only a small zone of impure limestone nodules and limey shale is present. However, it makes a convenient break on the log between the two black shale units.

The Penn Yan shale consists of dark to light gray shale and mudstone with interbedded black shales. Thin interbedded siltstones occur in this unit in central New York where the lower Penn Yan interfingers with the Sherburne member (fig. 1). The middle and upper parts of the Penn Yan interfinger with Sherburne, Renwick black shale and Ithaca members in west-central to central New York (deWitt and Colton, 1978).

The Penn Yan is topped by the thin, argillaceous Genundewa limestone. This was originally defined as a 12 to 15 foot thick sequence of dark gray to black shales and interbedded dark gray to black argillaceous limestone nodules capped by a thin limestone containing abundant Styliolina fossils. DeWitt and Colton (1978) restrict the Genundewa to the top, thin styliolinid limestone bed. In the subsurface, this unit can be traced eastward to central Steuben and Yates Counties.

The uppermost member of the Genesee is the West River shale. This is a medium to dark gray shale and mudstone which contains interbedded siltstones in central New York. All of the Genesee Group units pinch-out to the west by non-deposition. These pinch-outs migrate westward with each stratigraphically higher unit.

Genesee Group thickness increases eastward from zero in the subsurface at Lake Erie in Chautauqua County to more than 1,500 feet in Broome County, a distance of about 200 miles. West of Steuben County, the change in thickness is subtle, but, in central Steuben County, there is a large increase in the thickening rate. This is due to the appearance of the Ithaca clastics which thicken rapidly to the east. A southeasterly sediment source is indicated by the NE-SW strike of the isopachs.

Some interesting sedimentological features are present on the Genesee isopach map (METC/EGSP Series 120). In central Steuben County, a trough greater than 100 feet deep and up to 3 miles wide occurs in Genesee Group rocks. This is believed to be an ancient submarine channel. Gamma ray logs from wells within this channel show that the channel occurs within the stratigraphic and geographic limits of the Ithaca Formation and show the Ithaca Formation rocks to be more clastic than in surrounding wells.

In a direct line with the channel to the northwest in Wyoming and Erie Counties, a thicker lobe of Genesee sediments occurs. This appears to be a sediment fan built out in front of the channel area.

Two other similar features occur in the same general area; one in southwestern Steuben County and the other in northwestern Schuyler County. The southwest Steuben feature is probably a channel. Thinning in the Schuyler County area may be due to upward salt core movement during Genesee time in a prominent structure located in that area (Wayne-Dundee).

Sonyea Group

The Sonyea conformably overlies the Genesee Group and is, in turn, conformably overlain by rocks of the West Falls Group.

In western New York, the Sonyea consists of the basal Middlesex black shale overlain by the calcareous, gray Cashaqua shale. In the Finger Lakes region of west-central to central New York, the Middlesex is split into a lower black shale tongue, the Montour, and an upper black shale tongue, the Sawmill Creek, by the Johns Creek gray shale and siltstone wedge. In the same area, the Cashaqua is split by a wedge of gray shale, sandstone, siltstone and mudstone known as the Rock Stream. The Rock Stream and Johns Creek thicken to the east.

The Middlesex black shale commonly consists of laminae of dark blackish brown organic matter alternating with clays and silt-size quartz grains. Upon fracturing it exudes a strong hydrocarbon odor.

In the subsurface, the Sonyea increases from 4 feet at Lake Erie near the New York-Pennsylvania state line to about 900 feet in central Tioga County, New York, a distance of 175 miles. The isopach (METC/EGSP Series 119) is similar to that for the Genesee Group. The highest thickening gradient is located in the area from central Steuben eastward into Chemung County. This is the area where the Johns Creek and Rock Stream turbidite wedges enter into the section. In the underlying Genesee Group this is also the area where the Sherburne and Ithaca wedges come into the section from the east.

The two submarine channels seen on the Genesee Group isopach are also present on the Sonyea isopach trending to the northwest through central and southwestern Steuben County. No sediment fan, such as that seen on the Genesee isopach, is evident northwest of the central Steuben County channel.

West Falls Formation

This unit forms the lower part of the West Falls Group and has been defined by Pepper, deWitt and Colton (1956). In western New York, it consists of the Rhinestreet massive black shale at the base overlain by the light to medium gray Angola shale and mudstone.

The contact of the base of the Rhinestreet with the underlying Cashaqua is sharp and conformable. However, the top of the Rhinestreet is not well defined. In western New York, the upper Rhinestreet contains interbedded medium gray shale and thin-bedded argillaceous limestones. The top was described by Luther (1903) as, "A concretionary layer 6 inches to 8 inches thick, the upper surface of which is a scraggly mass of angular fragments of impure limestone...". Some scattered black shale beds occur above this zone. The location of this zone on nearby gamma ray logs is questionable, but, for correlation purposes, the top may be picked where gray shale becomes predominant over black shale in the same manner as Pepper, deWitt and Colton when they mapped surface rock exposures (1956).

Eastward into west-central and central New York, the Rhinestreet has been divided into several members. Some of these contain black shales and others are mostly gray shales and siltstones (fig. 1). To the east, the Rhinestreet gradually loses the massive black shale characteristic of western New York as the black shale beds are split by thin gray shale and siltstone beds. The black shales also drop to lower positions in the section as the total Rhinestreet thickens. In eastern Steuben County, the black shales are pretty much confined to the lower one-third of the formation. Siltstone beds, which are interpreted to be turbidites, become more common and thicken in an eastward direction as the sediment source is approached.

The upper member of the West Falls Formation, the Angola, is a light to medium gray shale and mudstone in far western New York. It contains numerous concretions and calcareous nodules in this area and interfingers at the top with the greenish-gray Nunda sandstone. In an eastward direction, the Angola is rapidly replaced in the section by the Nunda sandstone. The Nunda is a sequence of very fine grained greenish to bluish gray quartz sandstone overlying gray silty shale and siltstone at its type section in west-central New York. In the subsurface, the Nunda sandstone thickens in northern Allegany and southern Wyoming Counties. This buildup may be the result of the dispersal of clastics from the central Steuben County submarine channel which is evident on the West Falls isopach. Sand buildup is also seen in various areas in Steuben County.

The West Falls isopach (METC/EGSP Series 118) shows a steady rate of thickening from west to east across the state. An abrupt change in the rate of thickening in Steuben County is related to a major zone of turbidite deposition. The submarine channels noted on the Genesee and Sonyea Group isopach maps are also present on this isopach. Their positions have migrated slightly to the southwest.

Java Formation

The Java forms the upper part of the West Falls Group. In the Lake Erie area, it consists of the basal Pipe Creek black shale and the overlying Hanover shale. The Pipe Creek is a thin black shale which increases from about one foot thick at Lake Erie to over 20 feet in central Erie County. The Hanover is a gray to greenish-gray shale containing numerous calcareous nodules. To the east, the Hanover interfingers with the greenish-gray siltstones, sandstones and gray shales of the Wiscoy.

The base of the Pipe Creek is in sharp contact with the underlying Angola (west) and Nunda (east). However, in the subsurface, local interfingering shales in these units sometimes make it difficult to pick the base. In these cases the density log (if available) is helpful as the Pipe Creek shows the lowest density because of its high organic content. The upper contact of the Pipe Creek is difficult to determine because of interfingering with the Hanover and Wiscoy. The Pipe Creek is thickest in eastern Erie County, western Wyoming County and south into central Cattaraugus County in the subsurface.

Above the Pipe Creek, in the Hanover-Wiscoy section, occur two black shale beds. These usually are about one-third and two-thirds of the distance between the Pipe Creek and the base of the Dunkirk black shale which overlies the Java. The lower black shale has been termed "Shale A" and the upper "Middle Java" by Rickard (personal communication). This sequence of black shales is sometimes useful in correlating the Pipe Creek-Dunkirk section but should be used with caution because the two middle black shale beds tend to vary in thickness and occurrence. The top of the Hanover-Wiscoy section makes a sharp contact with the overlying Dunkirk black shale.

The Java isopach (METC/EGSP Series 117) shows a gradual thickening southeast of Lake Erie where the thickness is about 100 feet. The rate of thickening increases in eastern Cattaraugus County approaching a large thick area in central Allegany County where the Java is 225 feet thick. To the east, in western Steuben County, the northwest-southeast trending submarine channel seen in the underlying units is also evident in the Java. Thin zones occur on both sides of the channel, to the north and south, while a thick pod of sediments is present to the northwest in the mouth of the channel. South and southeast of the channel the Java thickens again. At the Java outcrop in southeastern Steuben County the section consists largely of siltstones and fine grained sandstones. Examination of gamma ray logs shows an increase in siltstone and/or fine grained sandstone in the Wiscoy in wells in the channel area.

Thick silt and sand deposits in the Allegany-Steuben area may indicate a stillstand in the westward prograding movement of the Catskill delta in this area. During this time turbidites filled the submarine channels and poured out over the basin floor where they were spread to the southwest by currents. In north-central Steuben County, J.G. Sorauf (personal communication, 1980) found shallow water deposits and indications of proximity to the shoreline.

Perrysburg Formation

This formation was defined by Pepper and deWitt (1951) as comprising the rock section from the base of the Dunkirk black shale to the base of the Laona sandstone in Chautauqua County, New York.

In Chautauqua County, the Dunkirk is a massive black shale that forms a sharp contact with the underlying Hanover shale. The massive Dunkirk black shale is more than 40 feet thick in Chautauqua County but thins rapidly to the east (Van Tyne and Peterson, 1978). Eastward, the basal Dunkirk contact is harder to pick on gamma ray logs because the Dunkirk begins to lose its massive black shale aspect and becomes more gray in color. Two black shale beds which occur between the Pipe Creek and Dunkirk are useful in making this gamma ray log correlation (see discussion of Java Formation). In west-central New York, only a thin black shale is present at the base of the Dunkirk while its upper portion is medium to dark gray in color and silty.

The top of the Dunkirk, at the South Wales contact as defined by Pepper and deWitt (1951), can only be picked on gamma ray logs in Chautauqua County. To the east into Cattaraugus County, the upper Dunkirk becomes silty and grayer in color and this correlation is uncertain. Rickard (1975) has dropped the South Wales designation. The contact of the top of the Dunkirk (formerly South Wales) with the overlying Gowanda gray shales may be picked on gamma ray logs in Chautauqua County but cannot be picked east of there because of the similarity of the upper Dunkirk and lower Gowanda gray shales.

Pepper and deWitt (1951) described and named the Hume black shale and designated its type section in north-central Allegany County. They suggested that the Hume grades eastward into the Canisteo shale and westward lies below the Gowanda. Rickard (1975) saw the Hume as an eastward basal black shale tongue of the Dunkirk.

Our recent studies indicate that the Hume is probably an upper Dunkirk black shale pulse which did not quite reach Chautauqua County. Pepper and deWitt (1951) give a thickness of 70 feet for the Hume at its type locality. Log studies show a Hume thickness of 30 feet in central Cattaraugus County decreasing to 15 feet in the northwestern part of the County and to a questionable thin sliver in eastern Chautauqua County.

In the Cattaraugus-Erie County area, the lower Dunkirk black shale is separated from the Hume black shale by intervening gray shales. To the east in the Allegany-Steuben County area, the black shales are separated by the Canaseraga sandstone and shale. This cycle of black shale tongues separated by westward prograding clastics is repeated several times in the Upper Devonian section of New York. Other examples are: lower Penn Yan and Renwick black shales separated by the Sherburne; Montour and Sawmill Creek black shales separated by the Johns Creek. Similar cycles with the larger black shale units separated by clastic incursions from the east are also apparent (fig. 1). These cycles represent oscillations of the shoreline during Upper Devonian time.

In the Lake Erie area, the Gowanda gray shale overlies the Dunkirk section. The eastern equivalents are the Caneadea gray shale and siltstone in Allegany County and the Canisteo gray to olive shales and siltstones in Steuben County.

Pepper and deWitt (1951) designated the base of the Laona sandstone, which overlies the Gowanda, as the top of the Perrysburg Formation in Chautauqua County near Lake Erie. This siltstone, or zone of interfingering siltstone beds as shown by gamma ray log correlations, can be traced through most of Chautauqua County but is not correlative with any certainty southward into Pennsylvania or eastward in New York. Eastward from Chautauqua County, based on outcrop work, they designated the base of the Rushford sandstone as the top of the Perrysburg although calling attention to the possibility that the Rushford may be older than the Laona.

In the present study, we have not been able to trace the Laona eastward from Chautauqua County in the subsurface by the use of gamma ray logs and/or well sample studies. Although control is sparse, it appears that the Rushford does lie below the Laona as suggested by the above authors. Therefore, the top of the Perrysburg Formation, as defined above, drops downward in the section from Chautauqua County eastward to central Allegany County. Further to the east there is no discernible top to the Perrysburg. After the Dunkirk-Hume black shale deposition, no comparable long-range marker beds were laid down during the remaining time of Upper Devonian sedimentation in New York.

GAS PRODUCTION

Prior to 1980 there were eight Devonian black shale gas fields, or areas of gas production, in New York. The oldest of these is the Lakeshore shale gas belt located along the shoreline of Lake Erie in western Chautauqua County (fig. 2). This belt extends westward into Pennsylvania and Ohio. Gas production in the Lakeshore area is from the Dunkirk shale and possibly also from the overlying Gowanda. Most of the wells were drilled in the late 1800's and early 1900's and were only 100 to 300 feet deep. It is estimated that up to 300 wells may have been drilled here (Van Tyne, 1974), but records were not kept and the exact locations of most of these wells is unknown.

The Naples shale gas field (fig. 2) was discovered in 1880. Nineteen wells were eventually drilled in and around the Village of Naples and twelve of these produced gas from the Marcellus shale, the basal unit of the Middle Devonian Hamilton Group. The wells average 1,200 to 1,400 feet in depth. Eleven wells are said to be still delivering a small amount of gas into a local utility system.

The Rushville shale gas field (fig. 2) was discovered in the latter 1800's. Eventually, 23 wells were drilled in the Rushville area, but records are non-existent or poor, and it is not known how many of these wells were productive. Apparently, the depths of the wells averaged 700 to 900 feet, and gas was found in the Hamilton section at unspecified depths.

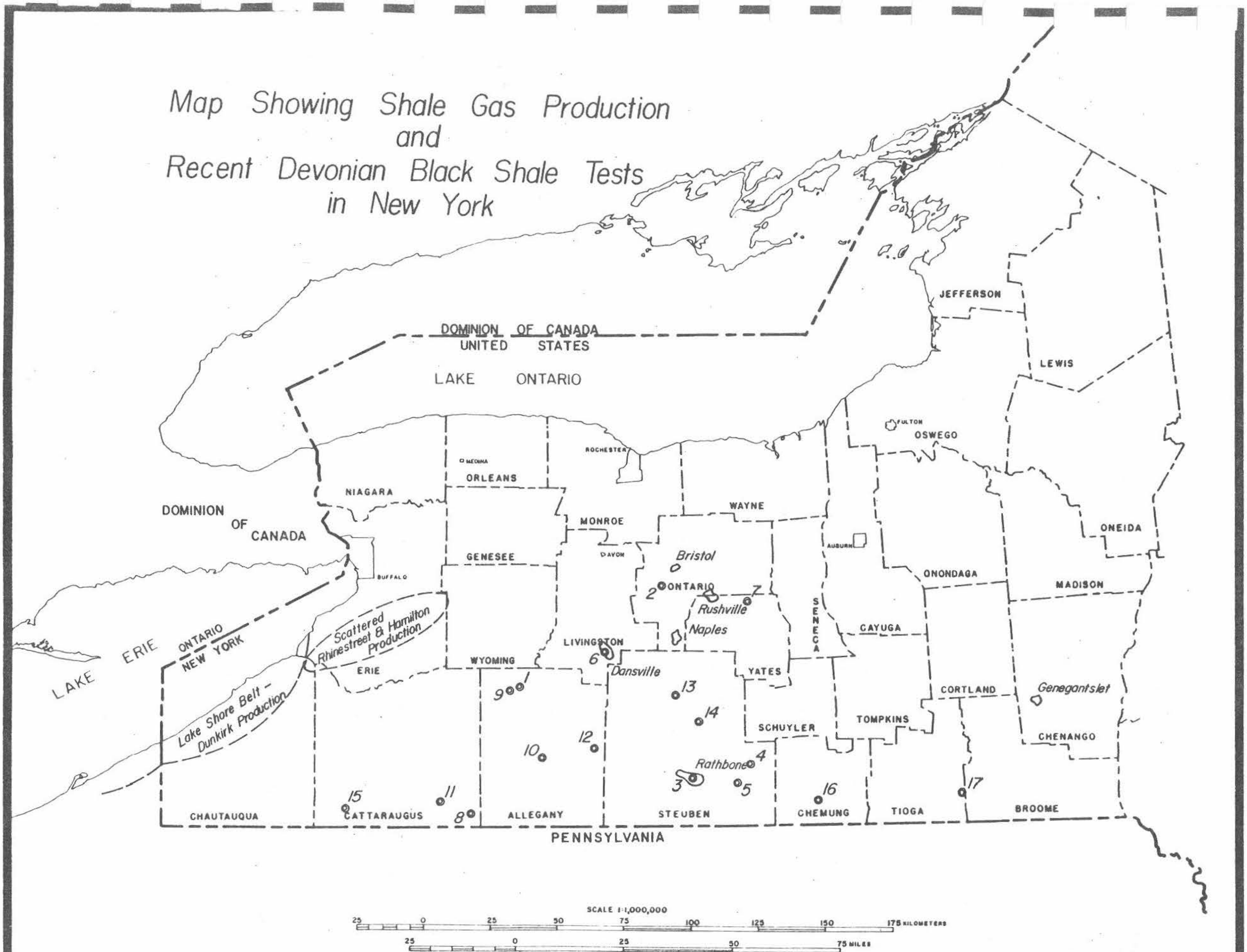
Map Showing Shale Gas Production
and
Recent Devonian Black Shale Tests
in New York
DOMINION OF CANADA
UNITED STATES
LAKE ONTARIO
DOMINION OF CANADA
LAKE ERIE
ONTARIO
NEW YORK
Lake Shore Belt - Dunkirk Production
Scattered Rhinestreet & Hamilton Production
JEFFERSON
LEWIS
OSWEGO
FULTON
ONEIDA
NIAGARA
ORLEANS
MEDINA
ROCHESTER
MONROE
WAYNE
GENESEE
AVON
BUFFALO
AUBURN
ONONDAGA
MADISON
Bristol
ONTARIO
Rushville
Naples
SENECA
CAYUGA
LIVINGSTON
Dansville
WYOMING
ERIE
YATES
CORTLAND
Genegantslet
SCHUYLER
TOMPKINS
CHENANGO
Rathbone
CHAUTAUQUA
CATTARAUGUS
ALLEGANY
STEUBEN
CHEMUNG
TIOGA
BROOME
PENNSYLVANIA
SCALE 1:1,000,000
25 0 25 50 75 100 125 150 175 KILOMETERS
25 0 25 50 75 MILES
MVT

FIGURE 2

The Dansville field (fig. 2) was discovered in 1881. Twenty seven wells were drilled in this field and most were small gas producers from the Hamilton section. The wells averaged 1,000 to 1,200 feet in depth. Recent drilling activity in this field is discussed in the next section of this paper.

The Bristol shale gas field (fig. 2) was discovered in 1914. Wells there average 700 to 900 feet deep and have produced gas from the Marcellus black shale formation of the Hamilton Group. Ten gas wells and six dry holes have been drilled in this field.

In southern Erie County, there is a northeast-southwest trending belt of gas production from the Rhinestreet and Hamilton black shales (fig. 2). Wells in this area also produce gas from the Akron section beneath the Onondaga and from the deeper Medina sandstone. Productive shale wells are scattered erratically throughout this area, but many wells which were completed to produce gas only from the Akron or Medina encountered good gas shows in the black shale sections. Most of the shale shows and production have been encountered above depths of 1,200 to 1,500 feet.

The Rathbone field (fig. 2) was discovered in 1931. Thirty two tests have been drilled in the field and nearby area. These wells average 1,200 to 1,600 feet in depth and encountered gas, and a little oil, in the Nunda siltstone and Rhinestreet shale section. Gas flows in this field have been rather spectacular but short-lived.

In mid 1964, a well drilled in the southern part of the Town of Smithfield in Chenango County encountered a good flow of gas in the upper Hamilton. This was the discovery well of the Genegantslet field (fig. 2), and subsequently, nine more wells were drilled there. Three of these were gas wells and six were dry holes. Except for a small amount of local farm use, the field has been shut-in since 1964.

RECENT DRILLING

By 1979, a good deal of information about the Devonian black shales of New York was becoming available from the ongoing studies, and interest was being stimulated to drill for possible shale gas production. In the summer of 1979, a Devonian black shale test was drilled on the campus of Houghton College in northern Allegany County (fig. 2, No. 1). The well was fracture treated in the Marcellus black shale (Middle Devonian) section just above 2,300 feet and responded with a strong flow and good pressure. The well was tested and completed in 1980 and hooked up to college buildings for use in heating systems. It continues to be a good gas producer. In mid 1979, Honeoye Storage Corp. drilled a shallower pool test in the Honeoye Medina storage field in central-western Ontario County (fig. 2, No. 2) to explore for possible gas production in the Hamilton shale. The well was located about four miles south of the previously discussed Bristol shale gas field. No shows were encountered in the Hamilton and the well was abandoned.

The Houghton College well was jointly financed by the U. S. Department of Energy and the New York State Energy Research and Development Authority. Following the success of this well, NYSERDA signed a contract with a private company to drill several black shale test wells in areas of south-central New York where gas production was likely to be obtained from Devonian black shales. Four wells were subsequently drilled in Steuben and southeastern Livingston County. NYSERDA's purpose in this, and later drilling, was to find out more about the Devonian black shale gas potential in New York and to stimulate drilling to develop this resource.

In Steuben County, the Valley Vista View well (fig. 3, No. 3) was drilled in the later summer and early fall of 1980 in the Rathbone field. The Marcellus and Rhinestreet sections were both fracture treated, but the treated zone in the Rhinestreet showed little or no response, so it was plugged off. The well was completed as a fair Marcellus gas producer in early 1981. The Scudder well (fig. 2, No. 4) was drilled through the Rhinestreet in July of 1980. A section which had produced some gas in nearby wells was cored, but no gas was found, so the well was not treated but was abandoned. The Dann well (fig. 2, No. 5) was drilled through the Rhinestreet in August 1980. A good gas show was encountered so the well was fracture treated. However, the well responded poorly to the treatment and was completed as a very small Rhinestreet gas producer. The Meter well (fig. 2, No. 6) was drilled in the Dansville field in southeastern Livingston County in late 1980. The well was completed in early 1981 and produces a small amount of gas from a fracture treated section in the lower half of the Hamilton.

In the fall of 1980, while drilling a Medina test located about four miles northwest of Penn Yan in Yates County (fig. 2, No. 7), Pennzoil encountered a large flow of gas from the Hamilton. The well was eventually completed as a small gas producer from the upper Hamilton shale. In early 1981, a second well was drilled nearby, but a commercial gas flow was not encountered, and the well was plugged and abandoned.

During 1981, NYSERDA contracted for five additional Devonian black shale tests to be drilled on various school properties in Allegany and Cattaraugus Counties. All five wells were drilled to the Onondaga limestone in June and July of 1981 and were fracture treated and completed by the fall of 1981 as small gas producers from the Marcellus shale. These wells are: Portville Central School (fig. 2, No. 8), Houghton College No. 2 (fig. 2, No. 9), Allegany County Board of Cooperative Educational Services (fig. 2, No. 10), St. Bonaventure University (fig. 2, No. 11) and Alfred University (fig. 2, No. 12).

Also in 1981, four Devonian black shale wells were drilled in central Steuben County. The C. L. Haines Manufacturing Company completed a good Marcellus shale well in early 1981 (fig. 2, No. 13) at Avoca. A second such well was drilled and completed during the summer of 1982. The Village of Bath, with financial assistance from the U. S. Department of Energy funds administered by the American Public Gas Association, had three Devonian black shale tests drilled at Bath during the summer of 1981 (fig. 2, No. 14).

Two of the wells were completed during the early fall as good gas producers from the Marcellus shale. The third well was completed as a poorer gas producer from this section.

In late 1981, a Devonian black shale gas well was drilled and completed by the Seneca Nation of Indians on their lands south of the Village of Steamburg in Cattaraugus County (fig. 2, No. 15). The Rhinestreet and Marcellus sections in this well were fracture treated, and the well is said to be a good gas producer from these zones.

In 1982, NYSERDA has contracted for four more Devonian black shale tests to be drilled in New York. Two of these wells have been drilled and fracture treated in the Marcellus black shale section, but final results are not known as yet. These two wells are located at Elmira (fig. 2, No. 16) and near Endicott (fig. 2, No. 17). The remaining two wells will be drilled in the Naples and Rushville shale gas fields.

The Devonian black shale studies and gas tests drilled in New York indicate that there is a large area of the state which has some productive gas potential from these shales. Some of the black shale zones have not been tested as yet for possible gas production. Such production tends to be small, by comparison with other producing zones, but it may be important locally as an auxiliary source of gas. The economics of a black shale drilling venture must be carefully considered, and the locations picked with the maximum possible available geologic knowledge of these formations.

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