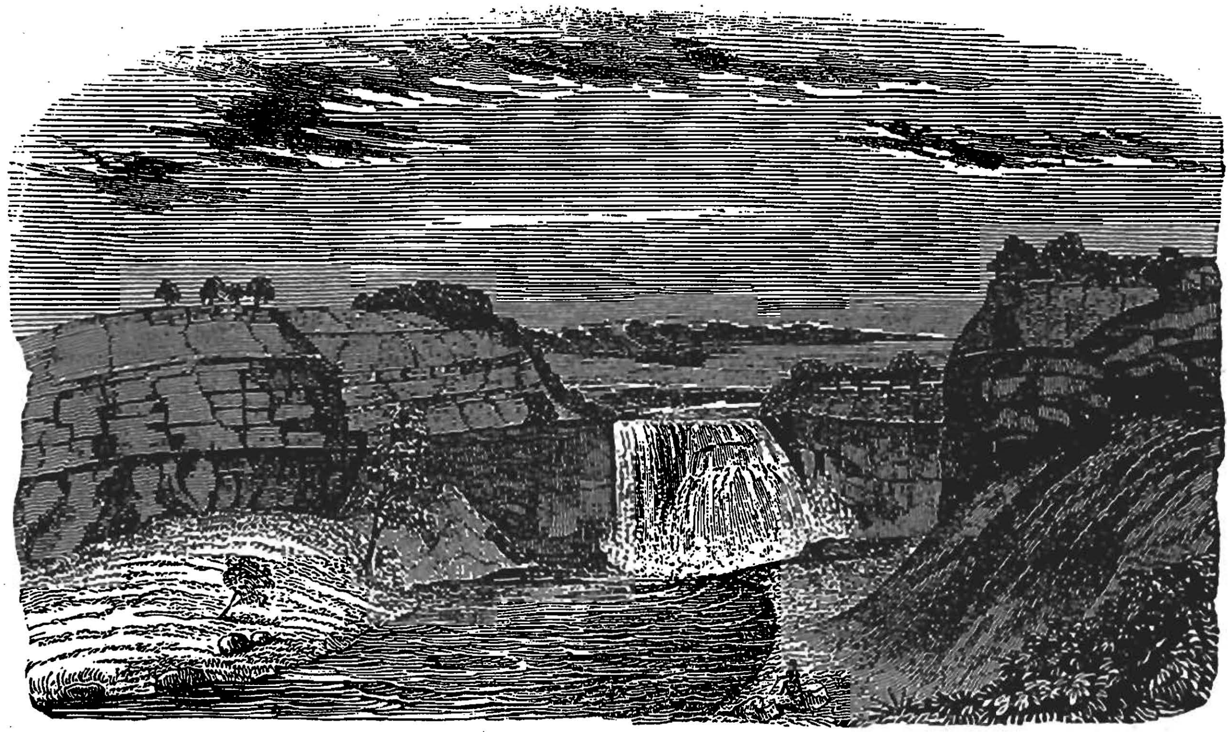
FIELD TRIP GUIDEBOOK

NEW YORK STATE GEOLOGICAL ASSOCIATION

66th Annual Meeting

October 7-9, 1994



Lower falls of the Genesee at Rochester. From a sketch by Mrs. Hall.
[From James Hall, 1843, p. 382, figure 185]

HOSTED BY:

DEPARTMENT OF EARTH AND ENVIRONMENTAL SCIENCES
THE UNIVERSITY OF ROCHESTER
ROCHESTER, NEW YORK 14627

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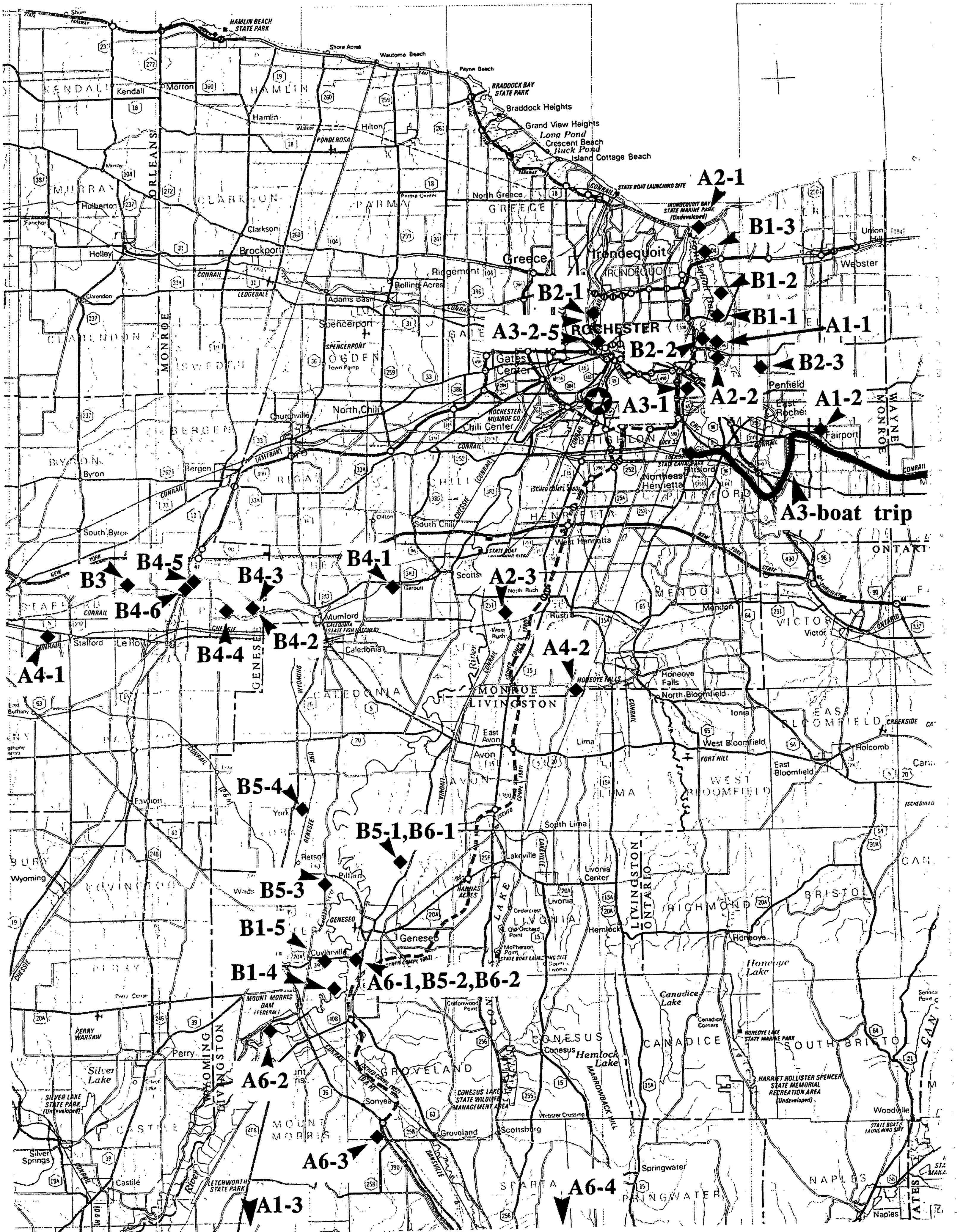
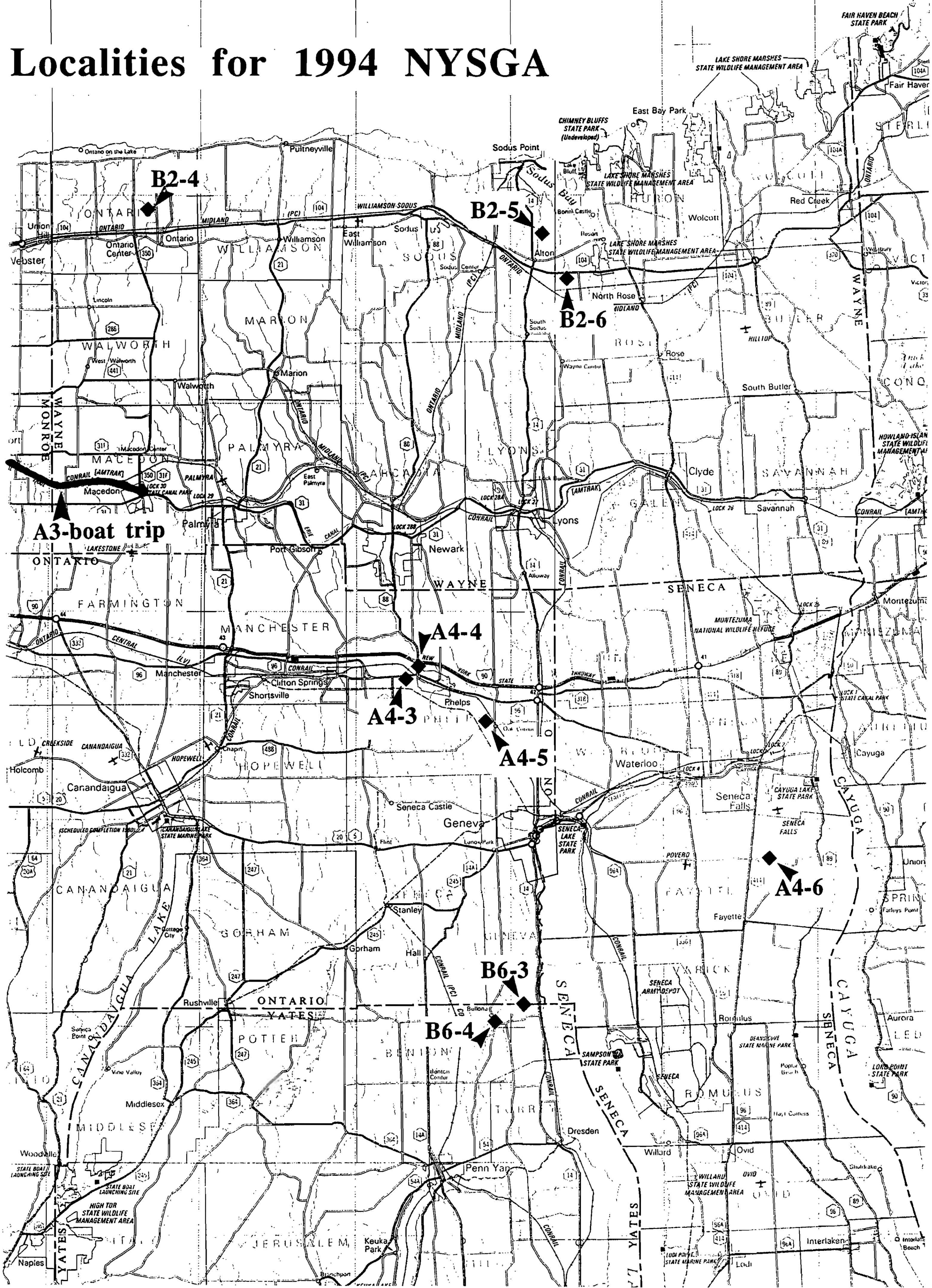
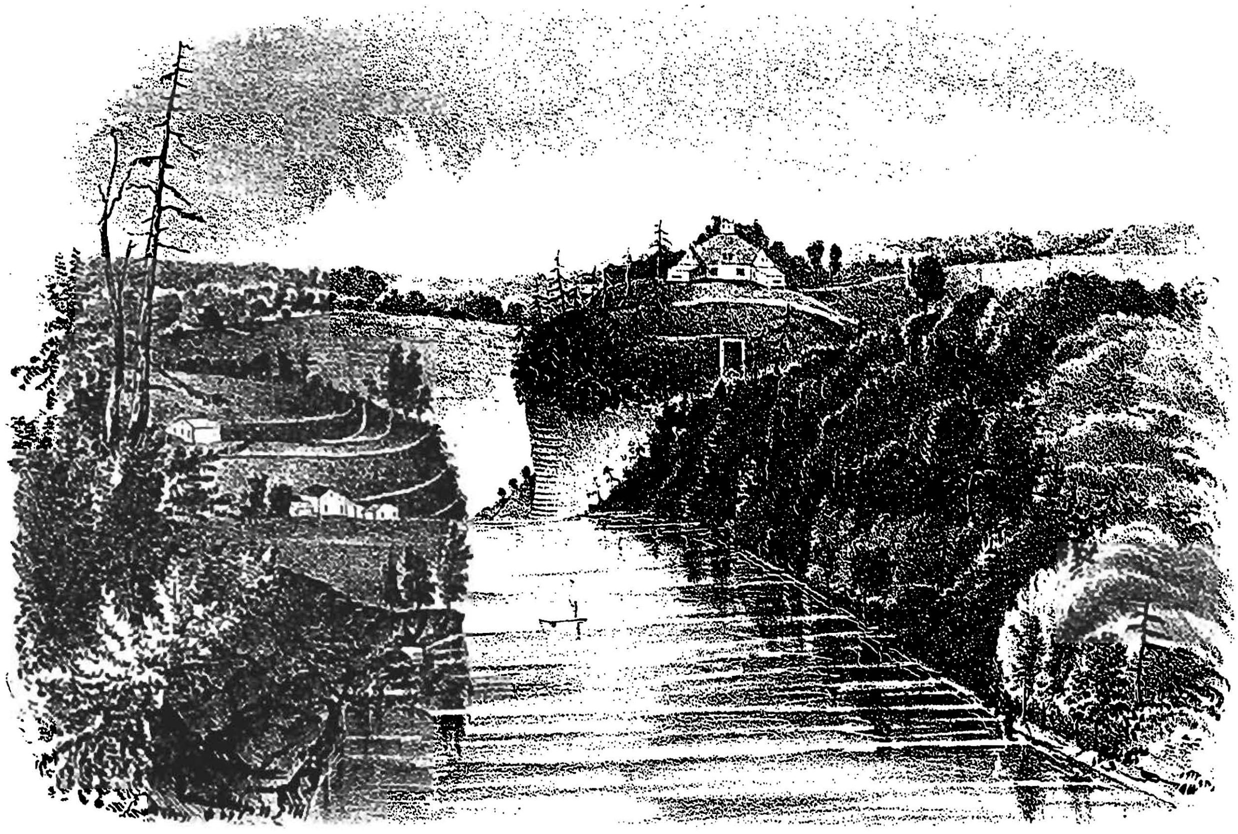


A2-1
B1-3
B1-2
B2-1
B1-1
A1-1
A3-2-5
ROCHESTER
B2-2
B2-3
A3-1
A2-2
A1-2
A3-boat trip
B4-5
B3
B4-3
B4-1
A2-3
B4-6
B4-2
B4-4
A4-1
A4-2
B5-4
B5-1,B6-1
B5-3
B1-5
B1-4
A6-1,B5-2,B6-2
A6-2
A6-3
A6-4
A1-3
HAMLIN BEACH STATE PARK
Kendall
Hamlin
Hilton
BRADDOCK BAY STATE PARK
Braddock Heights
Grand View Heights
Long Pond
Crescent Beach
Buck Pond
Island Cottage Beach
Greece
Irondequoit
Webster
Brockport
Holley
Spencerport
Gates Center
Churchville
North Chili
Chili Center
Bergen
Byron
South Byron
Penfield
Fairport
Pittsford
Northeast Henrietta
West Henrietta
Mendon
Victor
Stafford
Le Roy
Mumford
Caledonia
Scottsville
North Rush
Rush
West Rush
Honeoye Falls
North Bloomfield
MONROE
LIVINGSTON
East Avon
Avon
Lima
West Bloomfield
East Bloomfield
Holcomb
York
South Lima
Retsof
Lakeville
Livonia Center
Livonia
Hemlock
Geneseo
Cuylerville
Honeoye
Honeoye Lake
Canadice Lake
Perry
Silver Lake
Hemlock Lake
Conesus
Sonyea
Groveland
Scottsburg
Springwater
Castile
Silver Springs
Wyoming
Naples
Woodville

Localities for 1994 NYSGA





Deep Gorge of the Genesee below the Middle Falls at Portage.

[From Hall, 1843, Geology of the Fourth District, Plate XIX]

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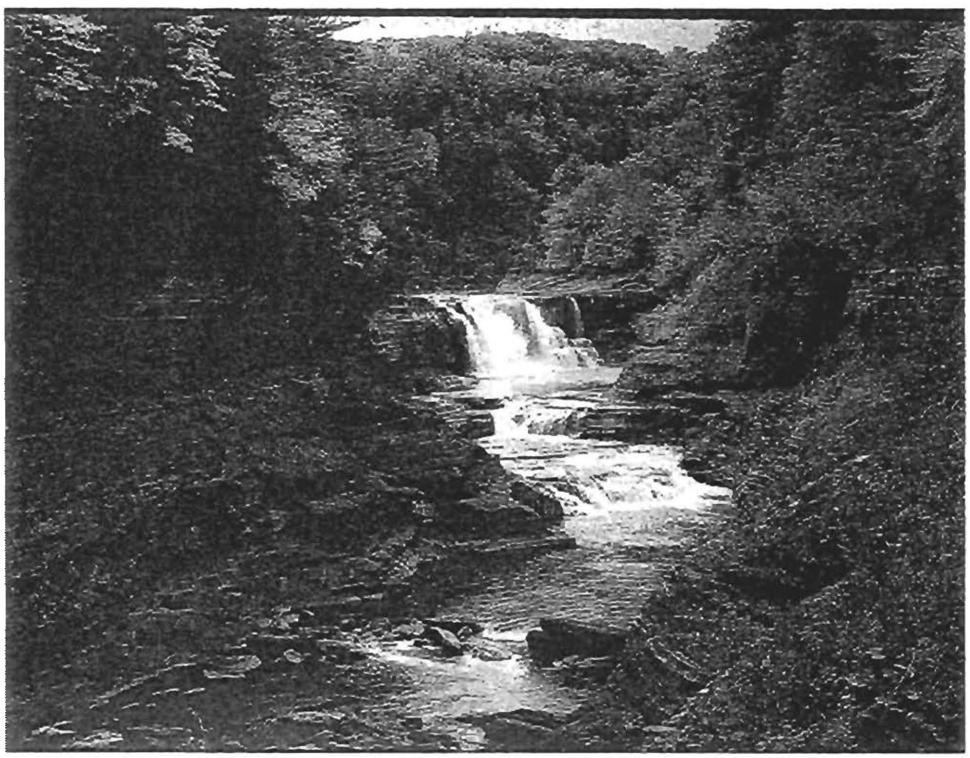
Ordovician-Silurian stratigraphic section at Lower Falls, Rochester, New York. Upper Ordovician Queenston Shale at base (contact above white reflectors), overlain by red Medina Sandstone; Kodak Sandstone (prominent white band) is overlain by greenish-grey Maplewood Shale.

Meander on the Genesee River near Jones Bridge Road.



Lower falls of Genesee River at Letchworth State Park.

TABLE OF CONTENTS

PREFACE

The Genesee Valley is a classic region for the study of sedimentary geology. The Genesee gorges at Rochester and Letchworth were explored geologically by many of the great early American geologists, particularly James Hall. Herman Leroy Fairchild, an early professor of Geology at the University of Rochester, and a founder of the *Geological Society of America*, made a career of studying the Genesee Valley glacial history.

The Genesee Valley area features an excellent cross-section of mid Paleozoic sedimentary rocks, ranging in age from the Late Ordovician to the Late Devonian. Upper Ordovician Queenston redbeds and overlying Lower to Middle Silurian mixed carbonates, shales, and sandstones, including the well known Clinton hematite, are exposed in the heart of Rochester itself. To the south are exposures of Late Silurian evaporites, shales and eurypterid-bearing dolostones, succeeded by fossiliferous beds of the Middle Devonian Onondaga Limestone. Still younger black and dark gray, exceptionally fossiliferous shales of the Middle Devonian Hamilton Group are exposed in tributaries of the Genesee River south of the Thruway. Outstanding exposures of Upper Devonian shales and siltstones of the "Portage" facies crop out in the spectacular cliffs of the Genesee River gorge at Letchworth Park, "the Grand Canyon of the East". Many of the outcrops in the Genesee area and the immediate vicinity are extraordinarily fossiliferous, and the strata themselves have been subject to substantial reinterpretation in the light of event and sequence stratigraphy, taphofacies, and models of foreland basin dynamics.

The Genesee region is an outstanding area for the study of surficial and glacial geology. The Genesee Valley was occupied by an extraordinary sequence of proglacial and moraine dammed lakes. Classic examples of eskers, kames, kettles and drumlins can be viewed in areas such as Mendon Ponds Park and also in the Pinnacle Hills moraine within the city of Rochester itself. Within the past decade, Pleistocene sand and marl deposits in the Genesee Valley region have yielded two exquisite mastodon skeletons, as well as other vertebrate and invertebrate fossils. The basic outlines of the complex Pleistocene geologic history of the Genesee Valley were established by Fairchild. However, renewed interest in surficial deposits, together with radiocarbon dating, has yielded a modified picture in the past several decades.

The Rochester area is also an ideal region for the study of urban geology and environmental hazards and problems. These include the highly publicized problem of roof collapse of the Akzo salt mine, landslides and slumps within Pleistocene sediments around the Irondequoit Bay region, problems of river erosion and flood control, including spectacular examples of migrating meander loops in the modern Genesee River, and the development of hazardous waste dumps in several places in the region.

This guidebook provides updated synthesis of several of these aspects of regional geology in the Genesee area. Bedrock geology of the Upper Ordovician to Upper Devonian is covered in a series of papers. Three other articles provide new syntheses and insights into aspects of surficial geology and environmental geology. Inevitably, there are gaps. There is no paper on the Salina Group, for example, which is unfortunate given recent interest in salt mining! Also, no tectonic or petrologic studies are included. However, on the whole, I believe that the guidebook presents a reasonable cross-section of Genesee Valley geology.

On a personal note, I cannot help but reflect that the first New York State Geological Association meeting that I attended as a student was held over twenty years ago in 1973, the last time NYSGA met in the Rochester area. (In fact, it was actually hosted by SUNY at Brockport.) It was on that occasion, while I was camped at Hamlin Beach State Park with some other students, that my friend, Gerry Kloc, came running back from the Saturday evening banquet full of excitement about a unique individual, named Gordon Baird, who knew about the geology of Erie County and other parts of New York State. I said "No way; nobody else really works on this stuff!" Nonetheless, the next day, accompanied by my fiancée (now wife), Betty Lou Hilton, I went to Old Dewey Hall at the University of Rochester to meet with this "Wunderkind". Sure enough, Gordon did know about Penn Dixie quarry, the North Evans conodont bed, and a great deal more that I thought that I had discovered. We had been working on the same stratigraphic sections for over five years without crossing paths. I was, of course, greatly impressed by Gordon's extensive collections of weird and wonderful geological specimens. More than that, I was struck by his incredible store of knowledge, wit and enthusiasm. Here, was a kindred spirit. Under other circumstances, Gordon and I could have become bitter rivals, like Cope and Marsh. That was not to be, however, for we both preferred cooperation to competition. From that September day onward, we became coconspirators in the "plundering of northeastern geology". By the time of the 46th Annual NYSGA, we had worked together enough to collaborate for the very first time in leading a field trip. Ironically, the trip dealt with the Windom Shale, a topic we again address in this guidebook, with considerable updating, some 20 years later. In the meantime, we have jointly studied over 1,000 streams and other outcrops in New York, have made many exciting discoveries, and have had countless conversations on the geology of the northeast during thousands of miles of travel on New York's Thruway and back roads. It has been an exciting ride, and a considerable part of it is logged in NYSGA guides.

I also note that the last time NYSGA officially met at the University of Rochester, in 1956, the guidebook, a thoroughly valuable contribution that was copied numerous times, was in more or less the form of handouts for student geological field trips. Since then, and even since 1974, the guidebooks have grown larger (yes, we have contributed to that), and of increasing quality. The NYSGA Guidebooks over the years, have become an invaluable trove of explicit information about New York's geology, coupled with road logs that enable students and professionals to assess for themselves the ideas presented in articles. These guidebooks present a great deal of information that is unavailable in any other source. I personally feel that the guidebooks have several very significant roles. The first, clearly, is that they are a vehicle for dissemination of often new, sometimes preliminary information. Certainly they continue to have a very critical teaching function. How many of us have at one time or another copied the road logs, figures or texts of NYSGA articles to supplement our class field trips? The articles in a guidebook, such as this one, present information, and more importantly, ideas, about geology, some correct, some perhaps incomplete or incorrect, but available for students in all categories of New York geology. In my view, the articles should present material in a variety of ways. Certain articles are geared for the more general audience, others are available for advanced students and for professionals specializing in particular areas. Another function which has not always been emphasized, but one which we have taken liberal advantage of, is that the guidebooks provide a vehicle for publication of fairly detailed information about regional geology. Where else can such information be published or found by the student interested in local details? Certainly, most of the professional journals, such as *Geological Society of America, Journal of Sedimentary Research* and others tend to publish articles that are broad, and of general interest, but typically, through editorial dictate, are "cleansed" of detailed outcrop-based information. In that regard, I feel that the NYSGA Guidebooks have a very important role to play as archives for such information.

I would like to express my sincere appreciation to numerous individuals who helped in the organization of this meeting and in the writing, editing and compiling of the

present guidebook. First of all, I owe a debt of gratitude to the authors of the several articles and leaders of the trips. They all came through and worked diligently to produce meaningful and interesting articles. Final preparation of manuscripts and compilation of the guidebook were aided greatly by the secretarial expertise of Susan Todd and Heidi Kimble of the University of Rochester. Mary Nardi, David Lehmann and James Scatterday helped greatly in the editorial process. Numerous reviewers, acknowledged in the individual papers, aided in the improvement of the final product. Dr. William Kelly of the New York State Geological Survey and presently head of the NYSGA, provided strong support and encouragement in the preparations for this meeting. Don Parry and Keith Kurz of the University's Conference Office also helped greatly in organization aspects of the registration and meeting logistics.

Finally, several past and present students of the Department of Earth and Environmental Sciences helped in numerous ways to prepare for the meeting. I particularly wish to recognize the efforts of Wendy Taylor, Robyn Hannigan, Chuck Ver Straeten and Gerald Kloc. Richard Hamell of Monroe Community College helped me with design of the cover; we chose a pale purple in recognition of Rochester's lilac traditions. My colleagues in the Department of Earth and Environmental Sciences, especially Curt Teichert and Robert Sutton, have always been supportive and encouraging to me in my various projects, including this one. Gordon Baird has helped me over the past two decades in ways too numerous to mention. Finally, I should acknowledge the patience and support of my family, particularly Dr. Betty Lou Hilton Brett, for without that, none of this would be possible. She cannot say she was not forewarned! You may note that she was with me on that famous first meeting with Gordon Baird; and we were engaged within the same week as that fateful meeting of the NYSGA in the Rochester area in 1973.

The Empire Soils Corporation of Huntington Enterprises, Inc. of Rochester generously provided $200.00 in support of the meeting.

Carlton E. Brett, President
New York State Geological
Association - 1994

This guidebook is affectionately dedicated to Dr. Robert G. Sutton, former professor of geology at the University of Rochester, in recognition of his contributions to the understanding of western New York geology and as an outstanding educator.

THE HYDROGEOLOGY OF LANDFILL SITES IN WESTERN NEW YORK

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INTRODUCTION

Because existing state environmental regulations require extensive hydrogeologic reports for landfills, waste disposal sites provide a wealth of penetrative data that may be synthesized for characterization of regional surficial and bedrock geologic units. Access to penetrative data may be accomplished by filing a Freedom of Information request with the New York State Department of Environmental Conservation (NYSDEC).

Landfill siting and operation are governed by 6 NYCRR Part 360 regulations. This regulatory document contains a defacto table of contents for a comprehensive hydrogeologic investigation of landfill sites. Furthermore, the regulations require a literature search and analysis of broader, regional data in order to provide a context for definition of site-specific, hydrostratigraphic units. When properly completed, resultant hydrogeologic reports contain a plethora of penetrative and water quality data, as well as hydrogeologic interpretations, that academic researchers would find difficult to fund through grant-lending agencies. Consequently, in the course of local geologic or water resource investigations, hydrogeologic reports for landfill sites should not be ignored. When reconciled with data published by the U.S.G.S., U.S.D.A. Soil Conservation Service, NY State Geological Survey, and the academic community (e.g. in NYSGA guidebooks!), these hydrogeologic reports add considerable volumes of quantitative information to any geologic research database.

The purpose of this field trip is to provide an overview of the hydrogeologic aspects of the Part 360 regulations and to illustrate the diversity of hydrogeologic data collected for landfill sites. Furthermore, this article is intended to demonstrate how Part 360 data may be synthesized for characterization of geologic units in western New York. The field trip will also provide an opportunity for landfill operators to demonstrate how modern, secure facilities are planned, designed, operated and closed to minimize negative environmental impacts.

PART 360 LANDFILL SITING PROCESS

Landfill siting is a controversial and expensive proposition in New York State. The process had been particularly controversial for the private sector under previous versions of Part 360. One particular problem would commonly arise when a potential landfill site was selected, based upon economic and/or geographic characteristics, before performance of a formal Part 360 site selection study. Site selection studies, both under previous and current versions of Part 360, involve comparison of potential sites on the basis of hydrogeologic, engineering and socio-economic properties. Ideally, the site selection process provides a means to determine the "best" site out of many potential landfill locations. Prior to the October 9, 1993 revision to the Part 360 regulations, when a preferred site would be chosen before other candidate sites were identified, expensive siting studies would subsequently have to be "retrofitted" to produce the desired conclusion that the preferred site is the most appropriate location among other "strawman" candidate sites for the facility.

The New York State Department of Environmental Conservation apparently recognized the quandary in which project sponsors found themselves when the preferred, suitable site was identified prior to completion of the formal site selection study. Furthermore, the NYSDEC also apparently recognized that, because siting studies require analysis of non-hydrogeologic variables (e.g. transportation, population, utilities, etc.), less hydrogeologically suitable sites could potentially be promoted over sites with more desirable subsurface characteristics for reasons other than susceptibility to groundwater and surface water contamination. Apparently for these reasons, the NYSDEC recently revised the Part 360 siting process for project sponsors who identify sites that exceed minimum hydrogeologic and engineering criteria for landfill construction. Under certain specifically stated hydrogeologic conditions, an expensive site selection study may not be required to defend the obvious merits of a highly suitable site. The revised regulations ease the financial burden and the logistical difficulty in objective compliance with landfill siting requirements when specified hydrogeological conditions are met. By establishing standards which must be met in order to waive the requirement for a site selection study, the NYSDEC can still guarantee that landfills can be constructed and operated in a manner which minimizes the potential for negative impacts to humans, wildlife or the environment.

The following is a synopsis of the landfill siting regulations as stated in the October 9, 1993 revision to the Part 360 regulations:

Siting Prohibitions and Restrictions

1. Prime agricultural land, within an agricultural district formed pursuant to the Agricultural and Markets Law, is excluded from siting if the landfill site is proposed to be taken through the exercise of eminent domain.

2. Flood plains are excluded from siting unless provisions have been made to prevent encroachment of flood waters upon the facility and unless the facility will not pose a significant hazard to humans, wildlife, or land or water resources.

3. Critical habitat for endangered species is excluded from siting.

4. Regulated wetlands are excluded from siting.

5. No landfills may be constructed over principal or primary aquifers[1], or within the cone of a public depression water supply well.

6. Proximity of landfills to airports is a concern because of the hazards associated with bird/plane impacts. Therefore, no landfill containing putrescible waste may be constructed within 5,000 ft. from an airport runway used by piston-powered aircraft or within 10,000 ft. from an airport runway used by turbine-powered aircraft.

7. No landfills may be constructed over unstable areas. According to the Part 360 regulations, unstable areas are those susceptible to natural or human-induced events or forces capable of impacting any structural component of the landfill responsible for leachate containment. Lands suspectable to landslides or sink holes are examples of unstable areas.

8. No landfill may be located in an area that is unmonitorable or unremediable. For example, groundwater flow rates and directions must be predictable. Site conditions must permit the placement of groundwater monitoring wells both upgradient and downgradient of the facility. Furthermore, site conditions must not preclude the ability to remediate in the event of a contaminant release.

9. Landfills cannot be constructed within 200 ft. of a fault that has had displacement during Holocene time unless the owner or operator demonstrates that the facility will not suffer structural damage in the event of fault displacement.

10. Landfills must not be sited in seismic risk zones, unless the owner/operator demonstrates that the integral containment structures can resist the maximum horizontal ground acceleration. Seismic risk zones are areas where a 10 percent or higher probability exists that the maximum horizontal ground acceleration in lithified earth material, expressed as a percentage of earth's gravitational pull (g), will likely exceed 0.10 g in 250 years.

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[1] *Principal* and *primary* aquifers are NYSDEC designations. A *principal* aquifer is one which has potential for development but is not currently exploited. A *primary* aquifer is one which is presently utilized for municipal water supply.

Landfill Siting Requirements

A formal landfill siting study involving penetrative investigation of multiple candidate sites will not be required if a preferred site is identified which does not conflict with the preceding siting prohibitions and restrictions and also exhibits the following characteristics:

1) The site is not underlain by bedrock subject to rapid or unpredictable groundwater flow unless the project sponsor demonstrates that a failure of the facility's containment system would not result in contamination entering the bedrock system.

2) The site is not in close proximity to any mines, caves or other anamolous features that may alter groundwater flow.

3) The site must contain an unconsolidated overburden thickness of 20 ft. or greater beneath the constructed liner system.

4) More than 50 percent of the vertical section through the upper 20 ft. of overburden must consist of soils with a permeability of less than $5x10^{-6}$ cm/s with no appreciable, continuous deposits exhibiting a permeability greater than $5x10^{-4}$ cm/s. The top five ft. of soil beneath the constructed liner must be able to achieve a permeability of $5x10^{-6}$ cm/s or less (Figure 1).

New landfills may be located on parcels that do not exhibit the above characteristics if two conditions are met.

1. The proposed facility is identified in a NYSDEC-approved local solid waste management plan; and

2. A formal site selection study involving multiple candidate sites is performed.

Waiver of an expensive site selection study provides a prime motive to identify a preferred site that exhibits suitable hydrogeologic conditions for landfilling. Site selection studies are costly, comprehensive analyses that evaluate hydrogeologic, economic, technologic, and public safety factors. The site selection study must demonstrate that, in spite of not meeting the previously stated siting requirements, operation of a facility on the preferred site will have no adverse impacts on public health, safety, or welfare, the environment or natural resources and will be consistent with the provisions of the State Environmental Conservation Law.

Site selection studies are expensive, because the process requires that alternative sites are evaluated to a comparable degree as the preferred site. Penetrative hydrogeologic investigations are required to determine depths to water and bedrock as well as the hydraulic conductivities of the various surficial and bedrock geologic units. The goal of the penetrative investigations is to ensure that the following siting criteria are satisfied:

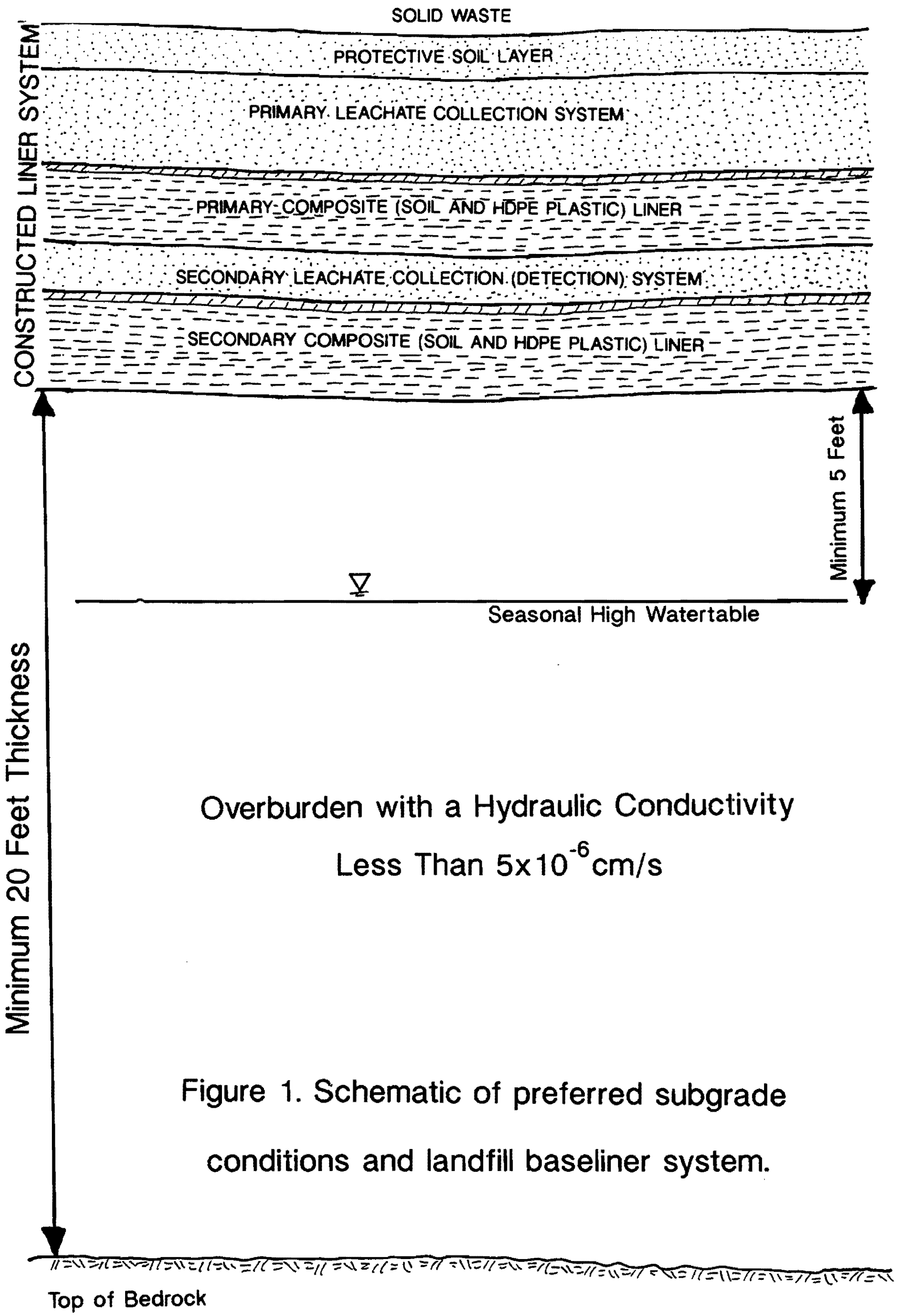

Figure 1. Schematic of preferred subgrade conditions and landfill baseliner system.

1) Candidate sites are those with the greatest possible thickness of unconsolidated deposits exhibiting hydraulic characteristics that permit them to serve as a barrier to migration of contaminants into rock.

2) Groundwater flow in bedrock must not be rapid or unpredictable unless it can be demonstrated that a designed containment system would not allow fugitive leachate to produce a contravention of groundwater quality standards.

3) Groundwater flow and quality must be such that containment failure would do the least environmental damage and could be easiest to correct.

4) Proximity and hydrogeologic relationship to water supply sources should be negligible.

5) Natural topography cannot be so steep that the engineered baseliner is unstable.

6) Relationship to mines, caves, or other anomalous hydrogeologic features that might alter groundwater flow should be negligible.

So, as can be discerned, if a project sponsor is going to promote a site as a possible host for a landfill, it is in his best interest to identify the site that truly meets the hydrogeologic standards necessary to avoid the requirement for a site selection study.

Waiver of the need for a site selection study does not, however, waive the requirements under the State Environmental Quality Review Act (SEQRA) for the project sponsor, the lead agency (usually the NYSDEC), and the concerned public to participate in an exhaustive analysis of environmental conditions, risks, impacts, and potential mitigation measures. The recent revisions to the Part 360 regulations save the project sponsor from unnecessarily consuming financial resources to defend the obvious merits of a highly suitable site so that appropriate emphasis can be placed on the SEQRA Environmental Impact Statement (EIS) and other engineering and hydrogeological documents required for the Part 360 permit application to construct and operate the facility.

Can Optimal Hydrogeologic Settings Be Defined?

The Part 360 process clearly promotes research to locate sites within hydrogeologic settings most likely to satisfy the stringent siting requirements. Recent efforts to place some geographic constraints on some of the hydrologic parameters necessary to streamline the siting process yielded a preliminary "terrain suitability map for landfill siting" (Goodman and others, 1992).

The map and its component "layers" (bedrock, surficial geology, aquifers and wetlands maps), that are based upon available published maps from the state and federal geologic surveys and agencies, can be used to determine regions (at the resolution of 1:250,000 scale data) that may exhibit characteristics unsuitable for landfill siting. Such characteristics that are discernible on the regional-scale maps include major limestone formations which could potentially exhibit karstic features, areas of exposed bedrock, faults, coarse-grained unconsolidated deposits, both documented and potential aquifers, and large-scale wetland areas (Figures 2A-D). Terrains possessing relatively unsuitable characteristics were shaded on the state-wide maps; conversely, relatively suitable areas were shown in white. When the layers are superimposed, a terrrain suitability map is the product (Figure 3). The map provides a degree of geographic tangibility to some of the hydrogeologic conditions which restrict or prohibit siting. This type of map may be useful for public awareness seminars, because the citizens of many low population regions of the state perceive that they are being "dumped on" by more populous areas. In actuality, there is a hydrogeologic rationale for prioritizing some areas of the state over others in the preliminary siting stage.

The results of the initial analysis suggest that, on the basis of hydrogeology alone, the Appalachian Plateau region contains the least sensitive and, therefore, most suitable terrains for landfill siting. The Erie-Ontario Plain also contains suitable hydrogeologic settings. Both of these physiographic provinces contain shale-rich bedrock units which are overlain by variably thick, fine-grained, glacial till and/or lacustrine deposits. It should be noted, however, that the terrain suitability map of Goodman and others (1992) was designed to address only hydrogeological siting criteria. Other demographic, economic, and transportation issues remain to be evaluated on a case-by-case basis.

The terrain suitability analysis may be useful for developing regional landfill siting strategies. Because of the generalized data used to construct it, however, the map does not serve as a substitute for site-specific hydrogeologic analysis (Cloyd and Concannon, 1993). In fact, key landfill construction requirements for bedrock and groundwater separation obviously cannot be evaluated using 1:250,000 scale data. Therefore, division of the major physiographic provinces into discrete hydrogeologic settings using all available maps and penetrative data is necessary to begin to produce "terrain suitability maps" at the appropriate scale for evaluation of local or site-specific conditions (Smith and others, 1993; Cole and others, 1993; Goodman and Stanwix, 1994). Only after evaluation of the compatibility of small-scale hydrogeologic settings with Part 360 siting and construction criteria can "optimal" subsurface conditions for landfill siting be mapped.

HYDROGEOLOGIC SETTINGS OF WESTERN NEW YORK

Western New York State is situated at the eastern limits of the Central Glaciated Groundwater Region of Health (1984). Aller and others (1987) have defined sixteen discrete hydrogeologic settings for the region (Table 1) as a foundation for

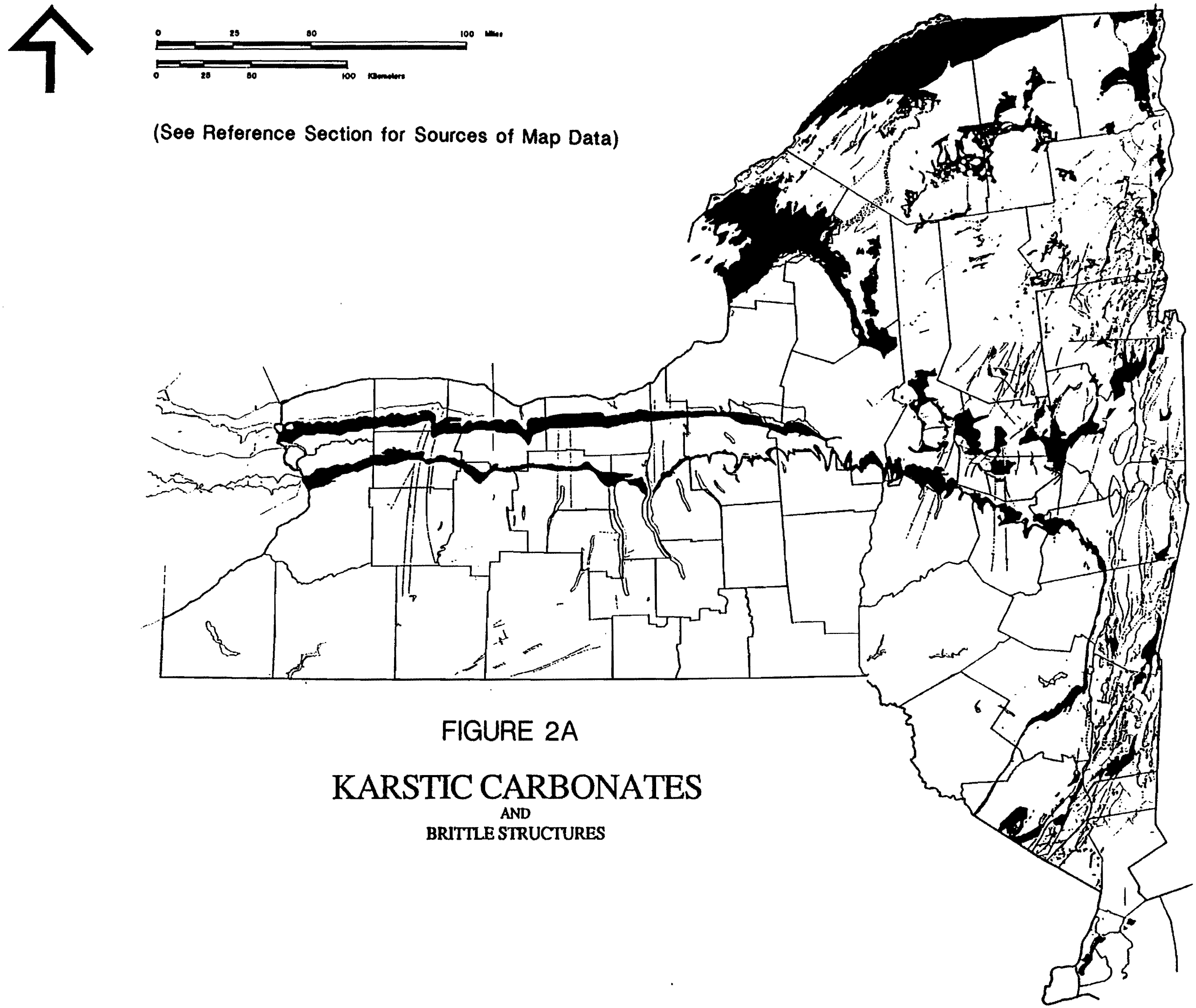

FIGURE 2A

KARSTIC CARBONATES
AND
BRITTLE STRUCTURES

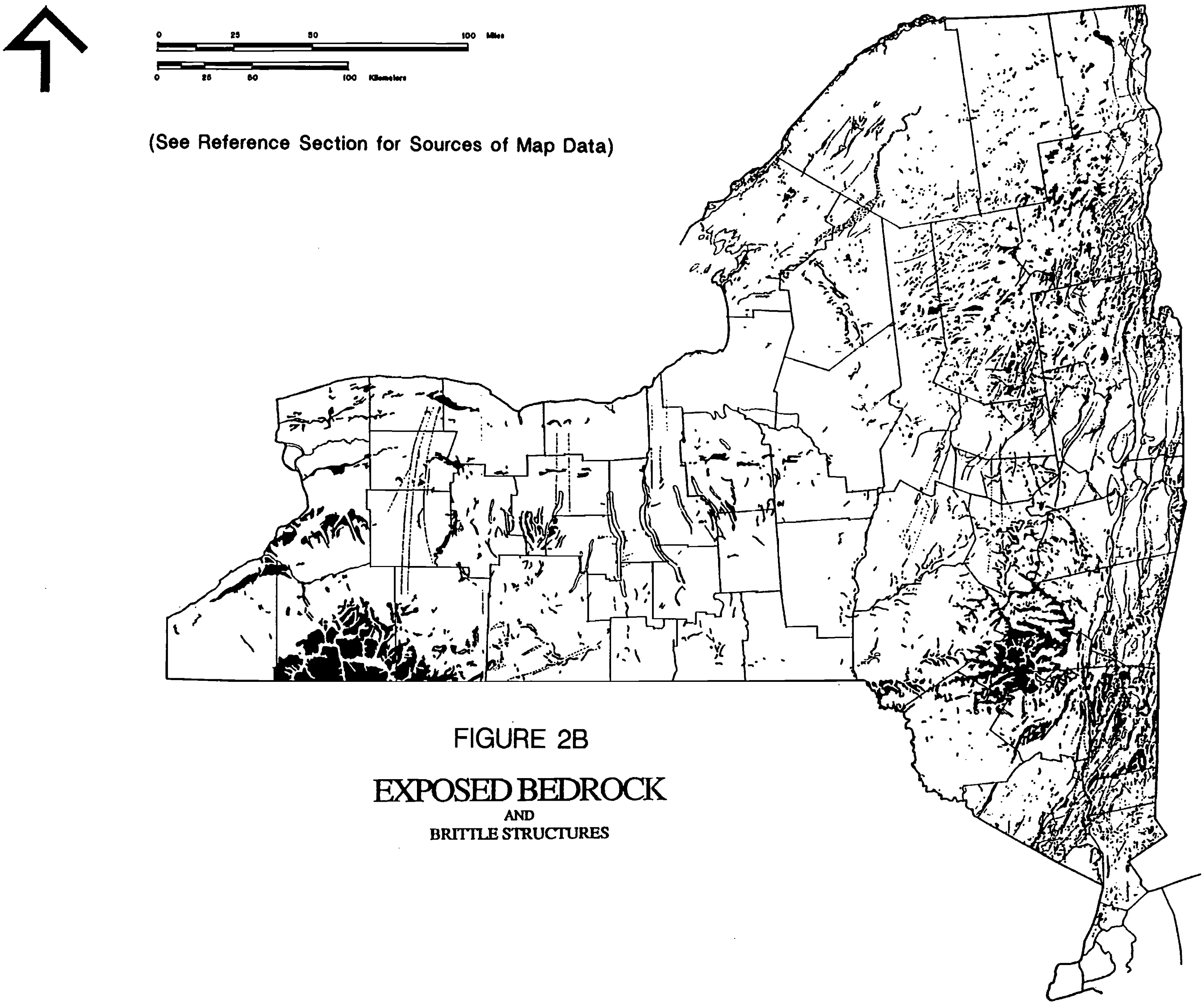

FIGURE 2B

EXPOSED BEDROCK
AND
BRITTLE STRUCTURES

(See Reference Section for Sources of Map Data)

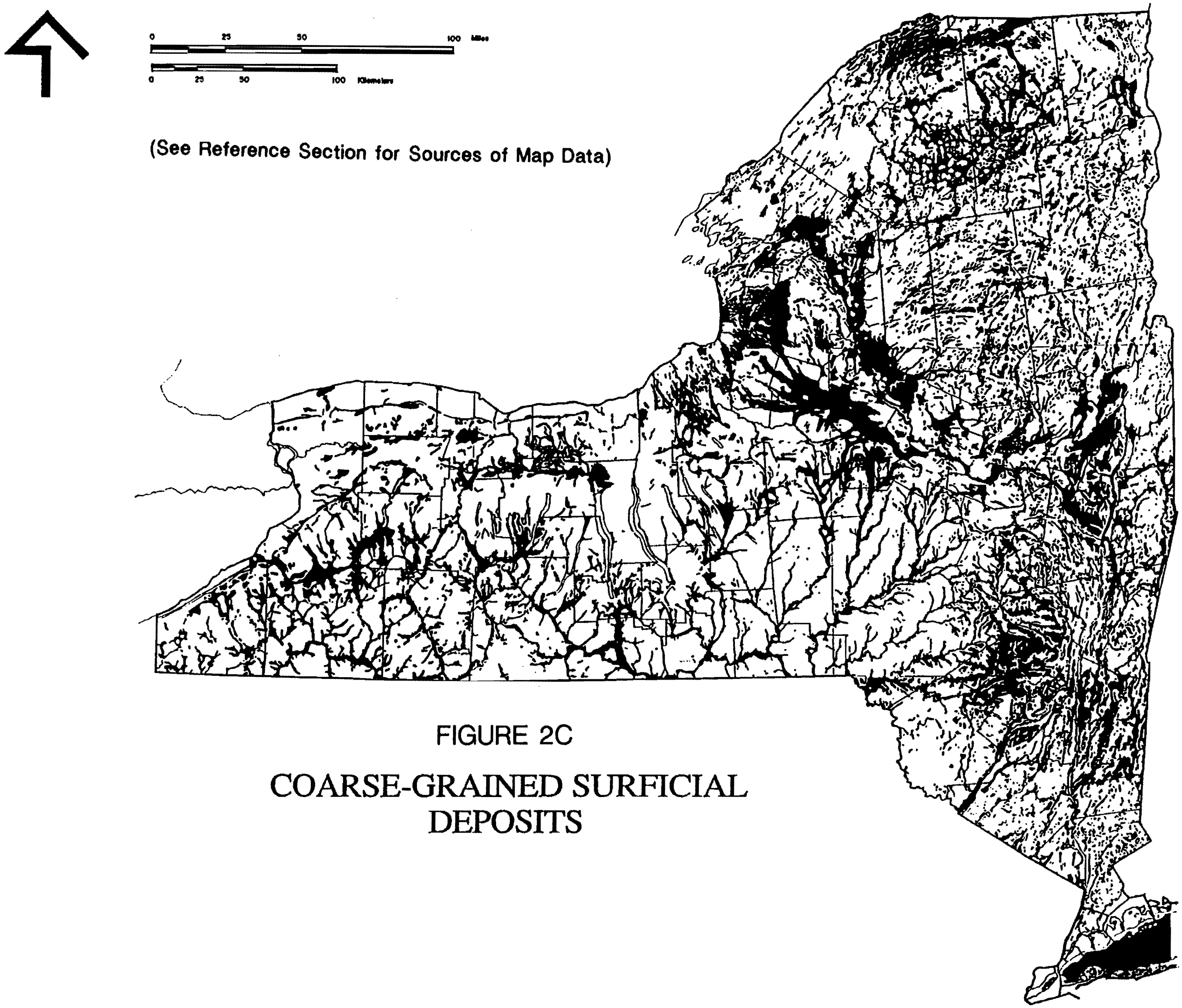

FIGURE 2C

COARSE-GRAINED SURFICIAL DEPOSITS

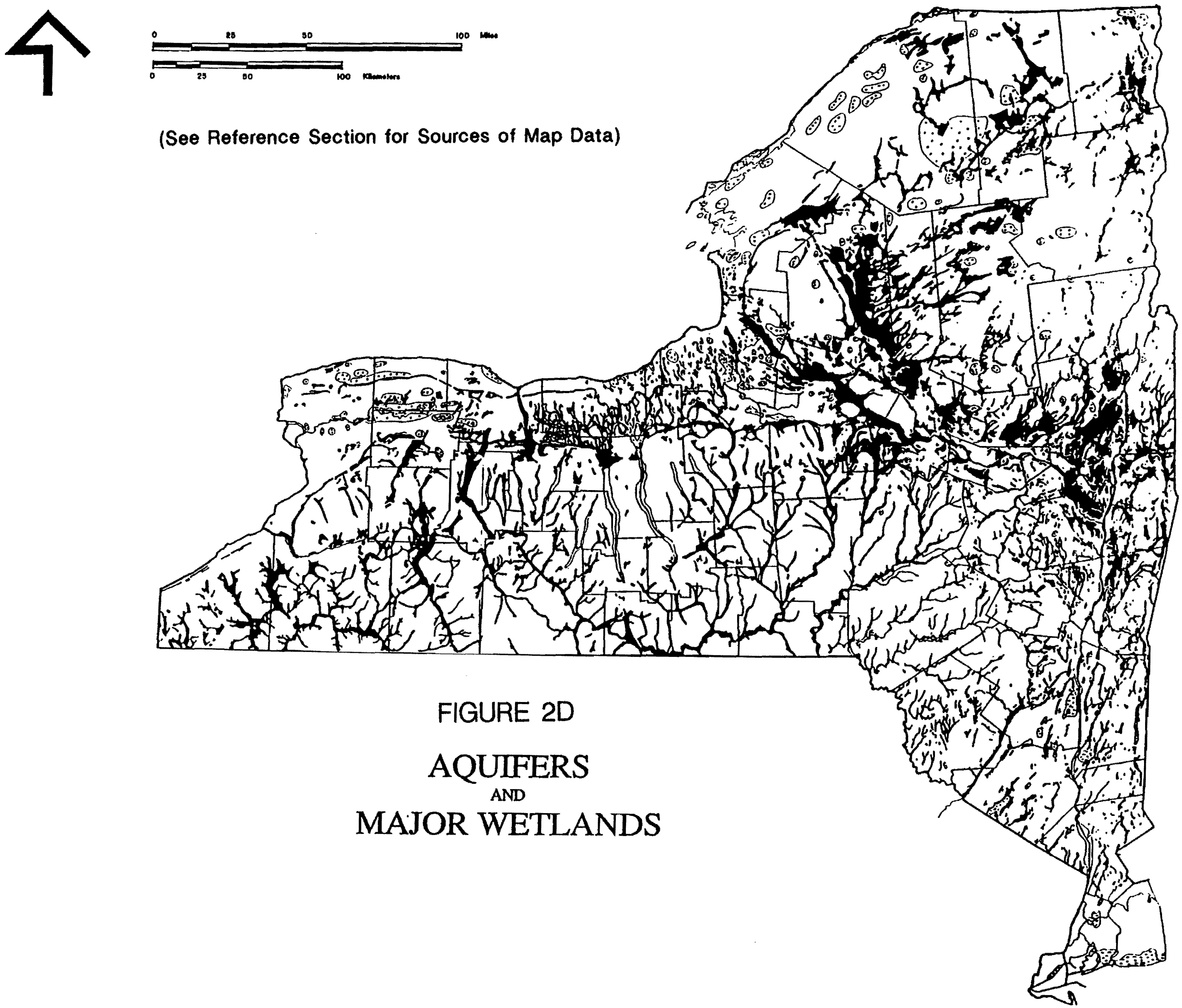

FIGURE 2D

AQUIFERS
AND
MAJOR WETLANDS

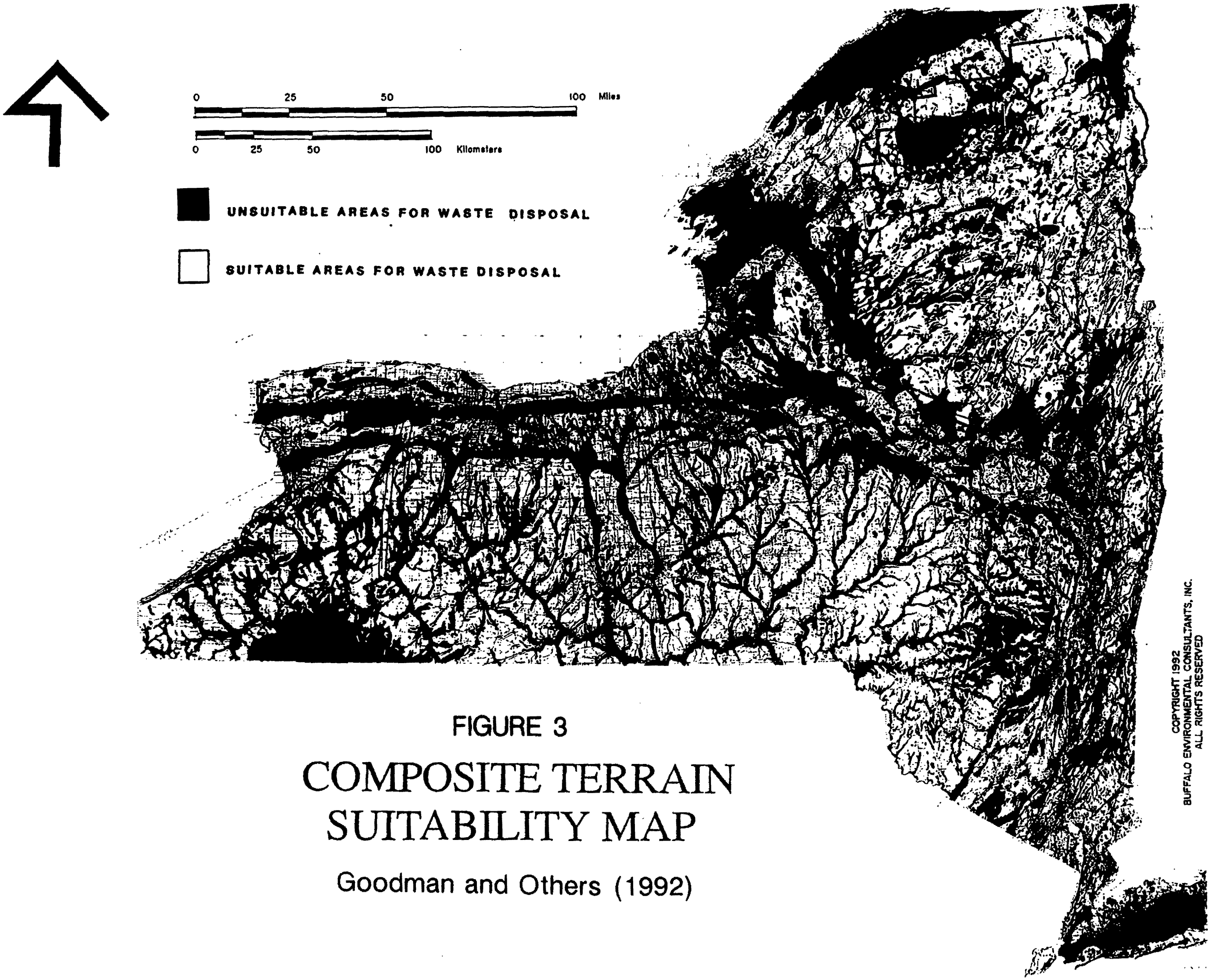

FIGURE 3

COMPOSITE TERRAIN SUITABILITY MAP

Goodman and Others (1992)

AFI ENVIRONMENTAL
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Table 1
Hydrogeologic Settings and Representative Drastic Indices[1] of the Central Glaciated Region
(Aller and others, 1987)

Setting	Drastic Index
Glacial Till over Bedded Sedimentary Rock	103
Glacial Till over Outwash	137
Glacial Till over Solution Limestone	139
Glacial Till over Sandstone	109
Glacial Till over Shale	88
Outwash	176
Outwash over Bedded Sedimentary Rock	156
Outwash over Solution Limestone	186
Moraine	135
Buried Valley	156
River Alluvium with Overbank Deposits	134
River Alluvium without Overbank Deposits	191
Glacial Lake Deposits	135
Thin Till over Bedded Sedimentary Rock	121
Beaches, Beach Ridges and Sand Dunes	202
Swamp, Marsh	160

1 Drastic Indices provide a relative gauge of sensitivity to point sources of pollution. The higher the index, the higher the vulnerability of the hydrogeologic setting.

their DRASTIC model[2]. Each hydrogeologic setting has been assigned a representative DRASTIC index which is a relative gauge of the vulnerability of groundwater resources to contamination from point sources at ground surface. The higher the assigned DRASTIC index value is, the more vulnerable the hydrogeologic setting is to groundwater contamination.

Existing, operating landfills, proposed facilities and a limited number of closed facilities in Western New York (NYSDEC Regions 8 and 9) are shown in Figure 4. The sites, their hydrogeologic settings and key bedrock and surficial deposits are identified in Table 2. As can be discerned, the majority of landfill sites are situated in those hydrogeologic settings with low DRASTIC indices. For example, the majority of sites located in the Appalachian Plateau are situated in "Glacial Till Over Bedded Sedimentary Rock" hydrogeologic settings. The exception is CID Landfill which is situated on low permeability till and lacustrine deposits within the Valley Heads Moraine complex. Most sites located on the Erie-Ontario Plain occupy two hydrogeologic settings: "Glacial Till over Bedded Sedimentary Rock" and "Glacial Lake Deposits". The exceptions are the Schultz C&D Landfill ("River Alluvium with Overbank Deposits") and the Niagara County Landfill that is located in a limestone quarry in Lockport, N.Y. In the following sections, the hydrologic properties of the various key deposits comprising the most suitable hydrogeologic settings for landfill siting are presented in order to demonstrate their compatibility with Part 360 criteria as well as the nature and depth of data required for landfill siting and routine monitoring during operation and after closure.

APPALACHIAN PLATEAU

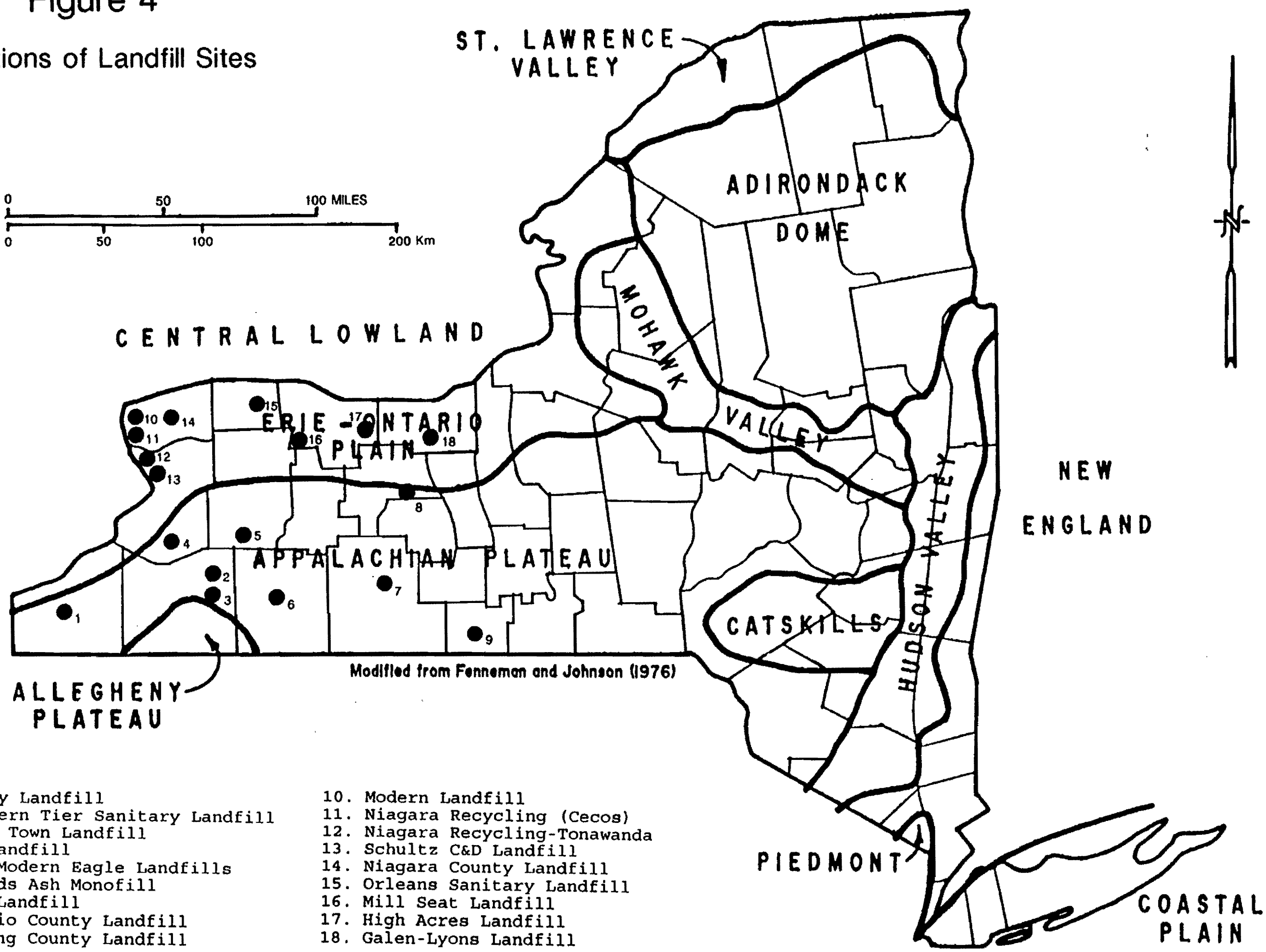
Based upon analysis of 9 landfill sites, a general model of the "Glacial Till Over Bedded Sedimentary Rock" hydrogeologic setting may be developed for the Appalachian Plateau of western New York (Fig. 5). Common elements among these sites include a bedrock hydrostratigraphy consisting of, in descending order, a glacitectonized rock aquitard, a fractured rock aquifer, and a competent rock aquitard. Other common elements include a low permeability surficial geologic profile consisting predominantly of lodgement and ablation till facies. A cross-cutting weathered zone, whose basal boundary is demarcated by a color change in tills and bedrock from brown above to gray below, probably reflects the slow, vertical migration of an oxidation front during the Holocene. The weathered zone cross-cuts glacigenic facies boundaries and also extends into bedrock beneath high elevation topographic divides that define local drainage basins. These common elements are described below.

Regional Bedrock Hydrostratigraphy

[2] DRASTIC is an acronym for seven hydrogeologic parameters that influence the vulnerability of groundwater to pollution from point sources at ground surface. These variables are 1) depth to water (D); 2) recharge rate (R); 3) composition of the local aquifer medium (A); 4) soil type (S); 5) topographic slope (T); 6) influence of the vadose zone media (I); and 7) hydraulic conductivity of the local aquifer (C).

Figure 4

Locations of Landfill Sites



1. Ellery Landfill
2. Southern Tier Sanitary Landfill
3. Olean Town Landfill
4. CID Landfill
5. BFI, Modern Eagle Landfills
6. Hylands Ash Monofill
7. Bath Landfill
8. Ontario County Landfill
9. Chemung County Landfill
10. Modern Landfill
11. Niagara Recycling (Cecos)
12. Niagara Recycling-Tonawanda
13. Schultz C&D Landfill
14. Niagara County Landfill
15. Orleans Sanitary Landfill
16. Mill Seat Landfill
17. High Acres Landfill
18. Galen-Lyons Landfill

zones.

Glacitectonized Bedrock Zone

Glacitectonized bedrock forms as stress imposed by moving ice deforms underlying bedrock units (Boulton and Paul, 1976; McGown and Derbyshire, 1977). Glacitectonized bedrock may commonly include small-scale, local folds and low-angle detachments. The competence of bedrock units (e.g. competent sandstone vs. incompetent shale) is the predominant determining factor controlling the degree of bedrock deformation beneath glacial ice. Less competent units tend to deform in a ductile, fold-forming fashion. More competent units tend to deform in a brittle fashion and, in some cases, form large, detached blocks (bedrock rafts) within poorly-sorted, variably comminuted matrix. Bedrock rafts render definition of local depth to bedrock (a key regulatory requirement) extremely difficult on landfill sites in the Appalachian Plateau of western New York.

Incorporation of significant volumes of till matrix between rotated bedrock rafts renders the glacitectionized bedrock zone readily identifiable. In areas where shale bedrock formations subcrop, the glacitectonized zone may be sampled using standard penetration tests with a split spoon sampler through hollow stem augers. Spoon samples will usually yield "disks" of weathered shale mixed with till.

In areas underlain by harder sandstone, the interval is more clearly observable in drill cores. Cores exhibit alternating zones of till matrix containing gravel-size, angular rock fragments and large, detached and rotated blocks of bedrock.

Generally, the glacitectonized bedrock zone is treated as the base of the lodgement till profile by most consultants because of the dislocation of the large bedrock rafts and the till matrix. The regulatory community, however, may prefer to use a more conservative definition of this interval as the top of the bedrock profile in order to insure that a minimum of ten feet of unequivocally defined, low permeability overburden is maintained beneath the constructed baseliner.

Given the gradational boundaries of the glacitectonized bedrock zone with the overlying lodgement till and underlying fractured bedrock aquifer, few sites contain wells that are screened discretely in this interval from which hydraulic conductivity values may be derived. At the Bath Landfill, however, six wells are screened in the zone and yield a range of K values between 2.8×10^{-6} cm/s and 2.2×10^{-5} cm/s and a geometric mean K value of 8.3×10^{-6} cm/s (Malcom Pirnie, 1994).

Seven wells are screened in the glacitectonized bedrock zone at the Southern Tier Sanitary Landfill site. These wells yield a range of K values between 1.9×10^{-5} and 5.0×10^{-4} cm/s and a geometric mean K value of 9.3×10^{-5} cm/s (AFI Environmental, 1992a).

Although thin till seams along bedding planes have been observed on some sites to depths of nearly 200 feet below ground surface, till injections into vertical fractures generally decrease progressively through the upper 30 feet of rock section and a gradual transition occurs from the till-choked, glacitectonized zone to the more permeable, fractured bedrock aquifer.

Fractured Bedrock Zone

The fractured bedrock zone is informally defined as the interval between the glacitectonized and competent bedrock zones. The interval is characterized by open, vertical fractures and numerous bedding parallel partings that may locally contain glacial detritus. Vertical fractures commonly are weathered indicating variations (probably seasonal) in degree of saturation. These fractures are also commonly lined by manganese oxides that, in conjunction with dissolution of calcareous fossils, render a blackened, decayed appearance to the joints.

On sites where the fractured and underlying competent bedrock zones are defined discretely, RQD data commonly reflect the differences in rock competency (Table 3). These data indicate that the RQD of a formation whose upper surface lies within the bedrock fracture zone may possess only 30% or less of the RQD value representative of the formation in the competent bedrock zone. The low RQD values of the aquifer zone reflect the numerous, bedding-parallel fractures that probably developed during unloading of glacially compressed bedrock. The combination of localized, differential slippage along bedding planes, perhaps induced in part by anomalously high subglacial hydrostatic pressure, and formation of unloading joints during stress relief, imposes a high secondary porosity on the top 10 to 30 feet of relatively till-free bedrock beneath most sites in the Appalachian Plateau. Consequently the hydraulic conductivity of the fractured bedrock zone is generally higher than corresponding values for the overlying glacitectonized zone and the underlying competent bedrock zone (Table 4).

Competent Bedrock Zone

In the competent bedrock zone, drill core samples are well-preserved, and strata retain their original gray tone colors, whereas the overlying glacitectonized and fractured aquifer zones may be oxidized to brown hues. Calcareous fossils remain intact and vertical fractures are tight and contain much less of the manganese oxide coating and fewer injected till seams that also are characteristic of the overlying horizons.

The hydraulic conductivity of the competent bedrock zone is slightly lower than that of the overlying bedrock hydrostratigraphic units (Table 4). Most sites possess a competent bedrock profile exhibiting a geometric mean K value in the low to mid 10^{-5} cm/s range although stratigraphic control on permeabilities at some sites may result in slightly higher mean values.

Table 3
RQD Values of Formations in the Fractured and Competent
Bedrock Zones
Appalachian Plateau

Site	Depth of Fractured Zone (ft.)	Ave RQD in Fractured Zone	Ave RQD in Competent Zone
Olean Landfill[1]			
Machias Shale	24	23%	46%
Cuba Sandstone	12	11%	46%
Wellsville Shale	12	0%	37%
Hylands Ash Monofill[2]			
Machias Shale	24	17%	53%
Cuba Sandstone	12	12%	65%
Wellsville	35	8%	65%

Notes:
(1) Earth Investigations LTD. (1990a)
(2) Earth Investigations LTD. (1990b)

Table 4. Hydraulic conductivity (cm/sec) of formations in the fractured and competent bedrock zones for the Appalachian Plateau.

Site Formations	Fractured Zone Mean	Max	Min	N	Competent Zone Mean	Max	Min	N
Bath Landfill[1] Wiscoy, Caneadea	1.4×10^{-4}	1.0×10^{-3}	6.0×10^{-5}	9	4.3×10^{-5}	2.3×10^{-4}	1.4×10^{-5}	7
Hylands Ash MonoFill[2] Machias, Cuba, Wellsville	5.6×10^{-5}	3.7×10^{-4}	3.3×10^{-6}	9	1.7×10^{-5}	2.3×10^{-4}	1.2×10^{-6}	16
Olean Landfill[3] Cuba, Wellsville	2.6×10^{-5}	5.1×10^{-4}	9.6×10^{-6}	4	2.5×10^{-4}	1.2×10^{-3}	5.5×10^{-6}	20
Southern Tier Saniitary Landfill[4] Lower and Upper Canadaway Group	--	--	--	--	5.8×10^{-5}	9.1×10^{-3}	3.4×10^{-9}	77
Ellery Landfill[5] Ellicott Group	3.5×10^{-4}	2.0×10^{-4}	6.7×10^{-5}	10	2.4×10^{-4}	4.3×10^{-4}	3.5×10^{-5}	26
Modern-Eagle[6] Lower Canadaway Group	--	2.5×10^{-3}	8.4×10^{-5}	--	--	--	--	--

[1]Malcolm-Pirnie (1994b)
[2]Earth Investigations LTD (1990b)
[3]Earth Investigations LTD (1990a)
[4]AFI Environmental (1992a)
[5]Dunn Geoscience (1988)
[6]Malcolm-Pirnie (1994a)

Regional Surficial Hydrostratigraphy

Surficial geologic maps covering the study area are provided by Tessmer (1975), Muller (1977), and Cadwell (1988). The general distribution of glacial deposits in the Appalachian Plateau is such that hills supported by Upper Devonian sedimentary bedrock are covered by sheets of poorly-sorted, low permeability, glacial till and intervening valleys are filled by up to 500 feet of mixed glacial till, outwash and stream alluvium (Muller, 1977; LaFleur, 1979; Prudic, 1982). Stream valleys are typically occupied by perennial streams and can contain alluvium consisting of permeable sand and gravel (Crain, 1966; Miller, 1988). The northeast orientation of many of the stream valleys suggests that they reflect preglacial fluvial drainages that were subsequently deepened and widened by continental valley glaciation (Calkin, 1982; Prudic, 1986).

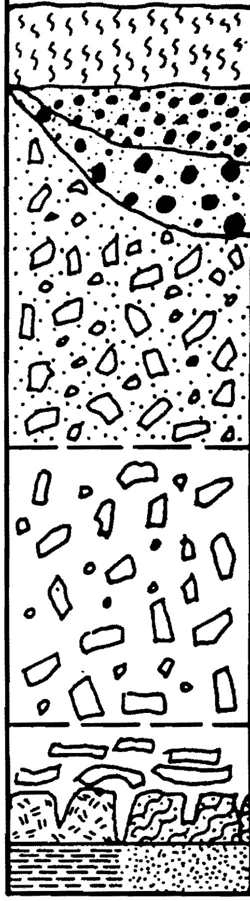
The stratigraphy and general properties of glacial deposits that cover landfill sites in the Appalachian Plateau of southwestern New York may be evaluated within the context of a continental glacial facies model. The importance and relevance of developing depositional models to understand the stratigraphic and geotechnical properties of glacial deposits in land-use evaluations (e.g. landfill siting analyses) is emphasized by Boulton and Paul (1976), Eyles and Sladen (1981), and Eyles (1983).

As indicated on surficial geologic maps, glacial tills are the most prevalent deposits in the high elevation terrains of the region. A genetic definition for glacial till is "an aggregate whose particles have been brought into contact by the direct agency of glacier ice and which, though it may have undergone glacially-induced flow, has not been significantly disaggregated" (Eyles, 1983, p. 11). Several types of till are encountered on many of the landfill sites. A typical stratigraphy of glacial tills is illustrated in Figure 7. Subglacial deposits include the deformation tills previously discussed as part of the bedrock hydrostratigraphy and overlying, densely-compacted lodgement tills. Englacial and supraglacial deposits include ablation and flow tills that contain discontinous lenses of water-sorted deposits, and a thin, mottled silt capping layer of uncertain origin.

A model for the distribution of subglacial deposits over a bedrock substrate is illustrated in Figures 8a and 8b. The subglacial tills found overlying glacitectonized bedrock at sites in Allegany and Cattaraugus Counties are generally interpreted in most hydrogeologic reports as lodgement tills. Further division into deformation till, comminution till, and lodgement till (sensu stricto) horizons may be possible, but these refinements are difficult to achieve because of the limited diameter and disturbance of deep till samples available from split spoons.

Lodgement Till

As observed in split spoon samples, the tills overlying glacitectonized bedrock are generally poorly sorted, gray to brown, silt- and clay-rich, channery deposits. As would be expected, the tills consist of detritus ranging is size from clay to boulders (Fig. 9). A typical till may contain roughly 25% gravel, 22% sand,



UNIT		AVERAGE GRAIN SIZE % Fines/% Sand/ % Gravel	FABRIC/STRUCTURES
SOIL		20/65/15	Massive silt loams
MOTTLED SILT		75/20/5	Massive silt with scattered sand and few pebbles
FLOW TILL	MUDDY FLOW TILL	55/10/35	Mud and silt-rich sand and gravel.
	DIAMICTIC FLOW TILL	15/30/55	Sandy and silty gravel, minor mud
BROWN LODGEMENT TILL		50/30/20	Massive to weak normal-grading. Many elongate bedrock rafts aligned parallel to bedding
GRAY LODGEMENT TILL		50/30/20	Same as brown lodgement till, but more cohesive and more compacted
DEFORMATION TILL			Elongate bedrock rafts that are folded and/or faulted
DECOMPOSED BEDROCK			Highly fractured bedrock with till injections; rafts in silt/clay matrix
FRACTURED BEDROCK			Massive to thin beds, cross-bedded sandstones, fissile shales

Figure 7.

Generalized Till Stratigraphy of the Landfill Sites in the Appalachian Plateau. Ablation Till is a common substitution for flow till. (after AFI Environmental, 1992b)

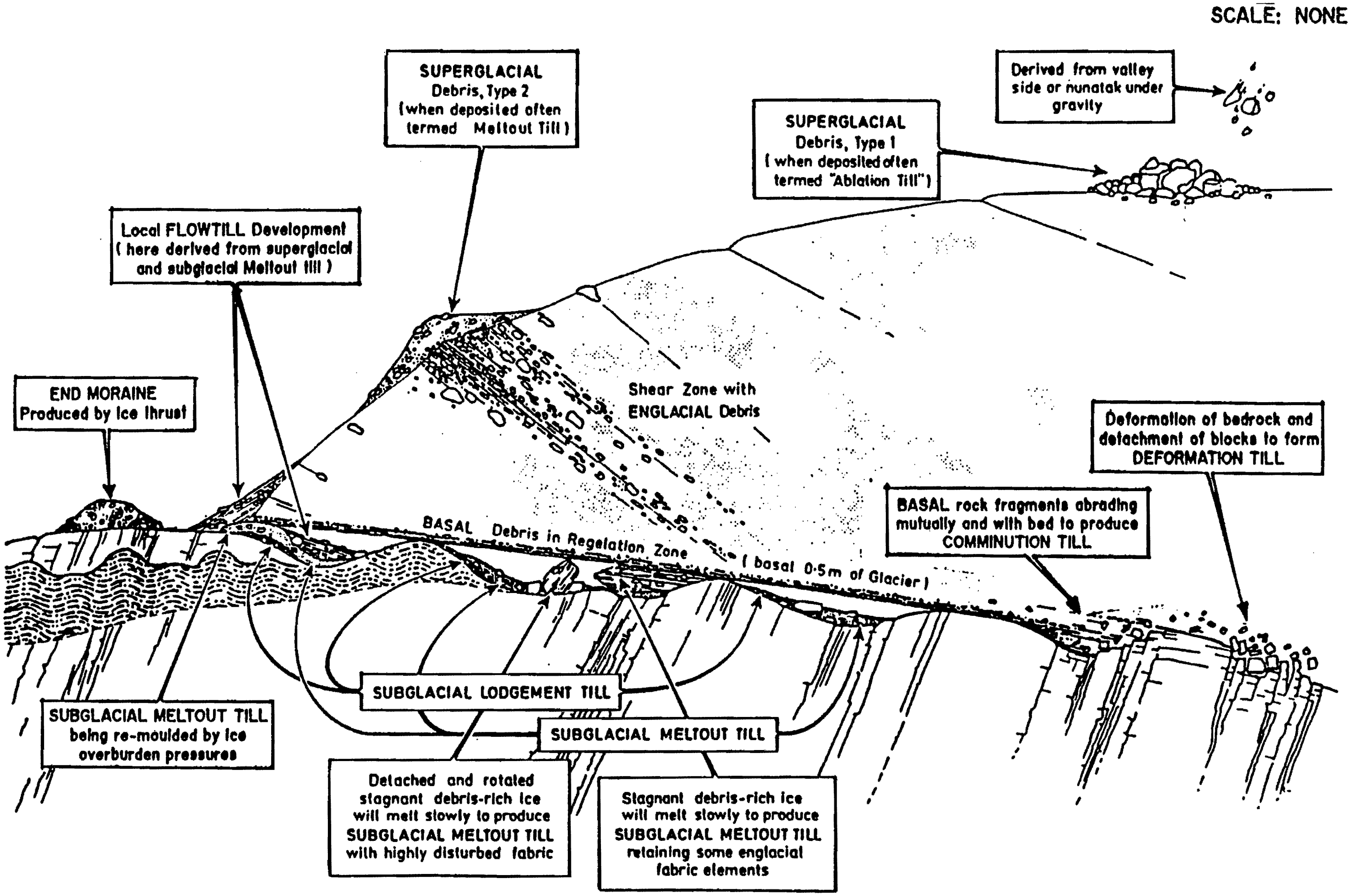
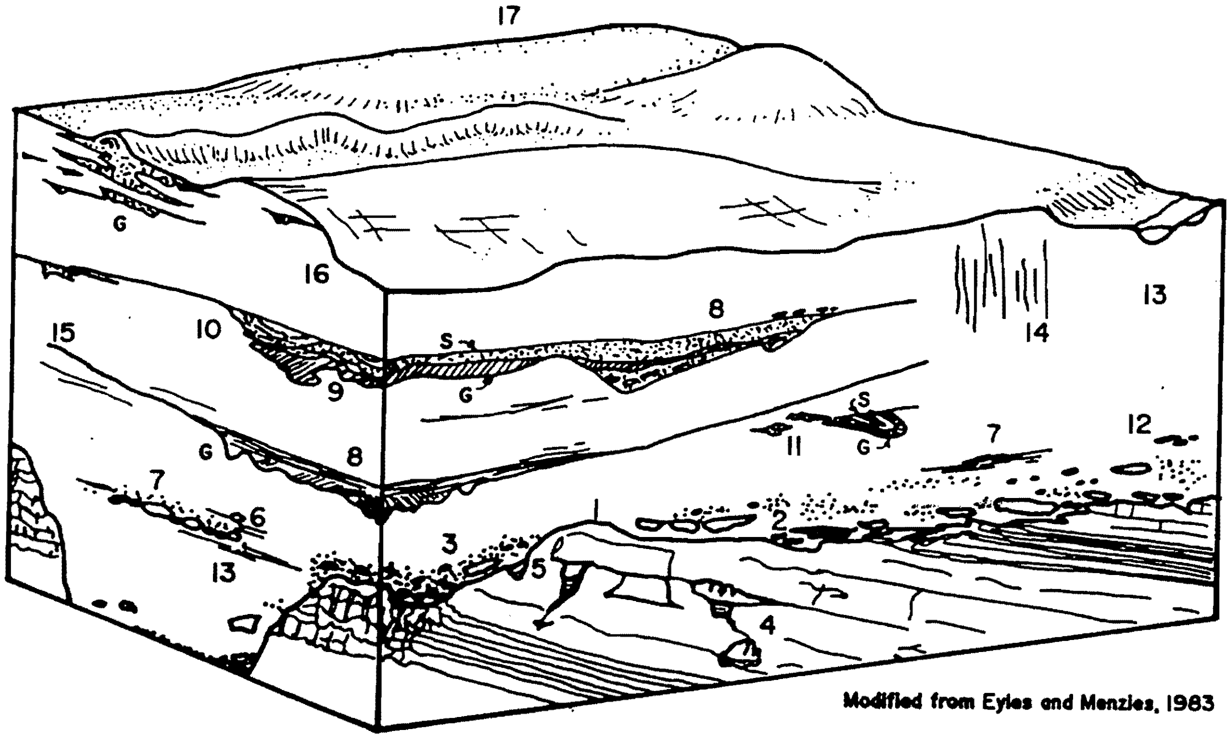
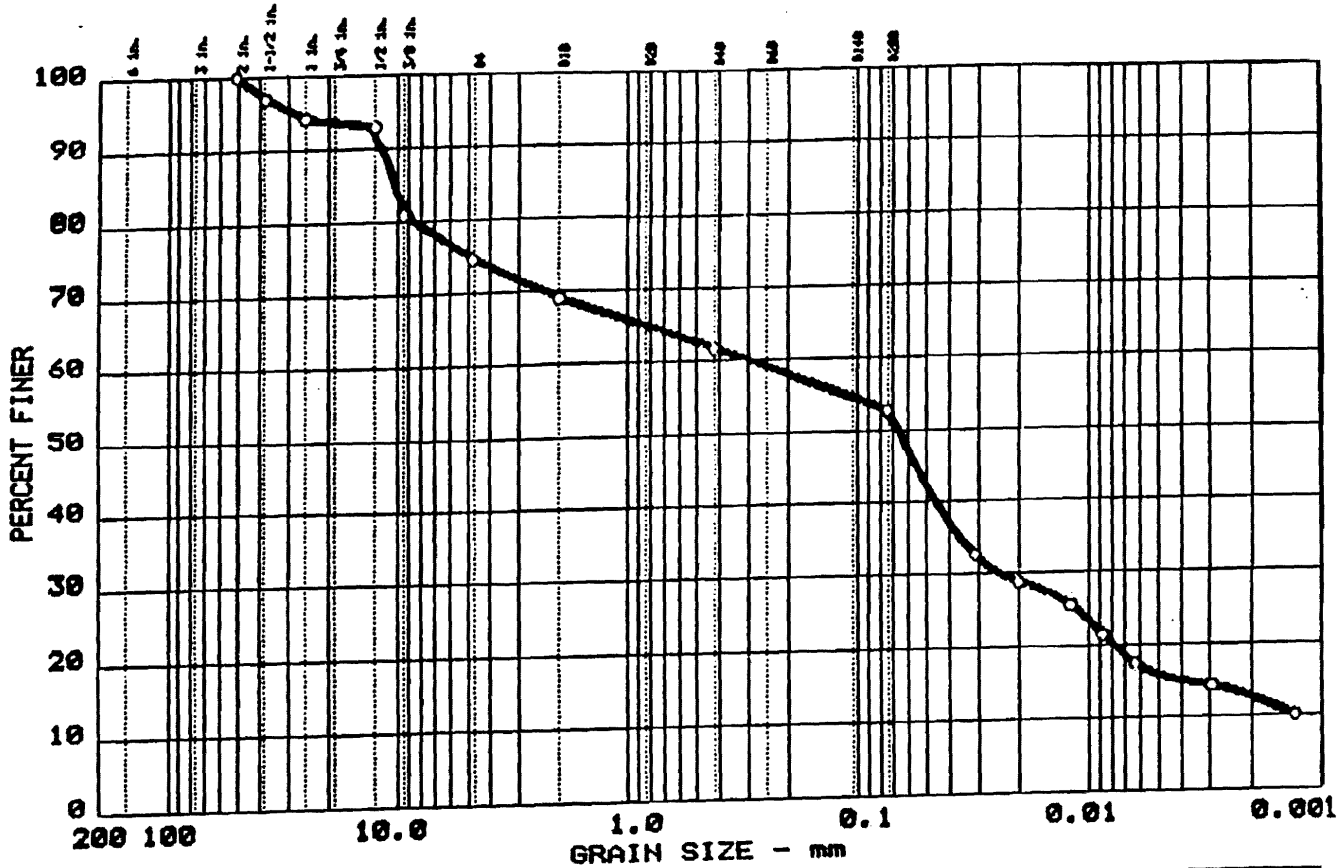


Figure 8A. Glacial processes and resultant till types (modified from McGown and Derbyshire, 1977).



1) Striated rockhead surface locally overdeepened below sea-level by subglacial erosion.
2) Rock "rafts", glacitectonized rockhead and deformation till.
3) Bouldery unit of scree-like debris filling lee-side cavities in rockhead.
4) "Cold water" karst from enhanced solution of limestones by subglacial meltwaters.
5) Intrusion of till into joints in rockhead.
6) Preferentially oriented clasts (long axes parallel to flow direction).
7) Distinct flat-iron shaping of fine-grained lithologies; coarse-grained lithologies produce faceted clasts of higher sphericity, frequently found as boulder pavements.
8) "Cut and fill" fluvial sediments deposited as sand (S) and gravel (G) in interconnected subglacial channels or as laminated clays in subglacial ponds. Often contain coherent debris masses dropped from ice roof.
9) Till masses diapirically intruded up into the base of fluvial channels.
10) Lenses of resedimented till extending into channel fills resulting from sidewall erosion and collapse.
11) Upper surfaces of cut and fill channels partially eroded by ice flow and resulting in deformed and folded inclusions in overlying till. Smaller channels folded.
12) Shear lamination caused by shearing out of soft, incompetent bedrock lithologies ("smudges").
13) Slickensided bedding plane shears resulting from subglacial shear.
14) Near vertical en echelon joints systematically oriented with respect to glacier flow direction or joint pattern in underlying bedrock.
15) Base of till units may be fluted; orientation of clasts with long axes parallel to flow direction.
16) Post-depositional sheared upper surface, frequently redeposited by solifluction.
17) Drumlinized, streamlined low relief surface; where rockhead is close to surface, rock-cored drumlins and "crag and tall" forms can be mapped. Subglacially engorged eskers are frequently related to the "cut and fill" sequences of (8) at depth.

Figure 8B. Typical till stratigraphy over a glacitectonized bedrock substrate.



% +3"	% GRAVEL	% SAND	% SILT	% CLAY
0.0	25.5	21.6	37.2	15.7

LL	PI	D_{85}	D_{60}	D_{50}	D_{30}	D_{15}	D_{10}	C_c	C_u
23	8	10.52	0.30	0.07	0.025	0.0040	0.003		

Figure 9. Representative grain-size distribution for lodgement tills of the Appalachian Plateau.

43% silt and 10% clay (Table 5). The large clasts are often flat channers of local sandstone. In test pits, the channers in lodgement till are typically aligned with long axes parallel to depositional surfaces. Most of the large clasts are derived from local Devonian formations, although a small percentage of transported Silurian Medina Sandstone and Precambrian metaplutonic clasts are present (Muller, 1977; Prudic, 1986). Bulk x-ray analysis of till deposits at the West Valley nuclear repository indicate that quartz, illite and chlorite are the major mineralic constituents of the fine-grained till matrix (Prudic, 1986).

The lodgement tills of the landfill sites in the Appalachian Plateau are densely compacted. Average N-values based upon standard penetration tests are approximately 68 blows per foot for the gray, unaltered till and approximately 52 blows per foot for the brown, weathered till (Table 6). Seismic velocities for the lodgement tills range between 5000 and 7000 feet per second (fps) for the unaltered till and between 3500 and 7000 fps for the weathered till (Harding Lawson Associates, 1992; Kick, 1992; Gartner Lee, 1993a, b).

Given these physical properties, the lodgement tills form low permeability aquitards (Table 7). Reported values for the gray, unaltered till range between 3.2×10^{-9} cm/s and 8.5×10^{-5} cm/s and average approximately 6.6×10^{-6} cm/s. Reported values for the brown, weathered till range between 3.4×10^{-9} cm/s and 2.9×10^{-4} cm/s and average 2.3×10^{-5} cm/s.

Ablation Till

The lodgement tills within the study area may be overlain by less densely compacted till that on some sites contains discontinous lenses of glaciofluvial deposits up to 6 feet thick. These variably textured deposits are likely to be ablation tills. Ablation till, a type of melt-out till, is deposited by the slow release of glacial detritus from ice that is neither sliding nor deforming internally (Dreimanus, 1988). Common properties of melt-out tills include the following: 1) banding of debris, bedrock blocks and rafts; 2) alignment of elongate clasts parallel to glacier flow; and 3) the retention on englacial fabrics (McGown and Derbyshire, 1977; Boulton and Paul, 1976; Dreimanis, 1988). These properties are consistent with the characteristics of the upper portions of till profiles observed in test pits and test borings on landfill sites in the Appalachian Plateau.

The hydrogeologic reports for most sites do not identify ablation tills explicitly, although water-sorted lenses are mentioned frequently. The ablation till/lodgement till boundary is commonly obscured by the cross-cutting oxidation front and may be best defined by contrasting N-values and seismic velocities, and the depth of water-sorted lenses in local vertical sections. Ablation tills generally have lower N-values than lodgement tills. An estimated average N-value for ablation tills is 13 blows per foot whereas the average N-values for regional lodgement tills are 52 and 68 for the weathered and unaltered profiles, respectively. Typical seismic velocities for ablation tills range between 3300 and 4800 feet per second (Kick, 1992). The range of values is narrower and at the low end of the spectrum (3500-7000 fps) for regional lodgement tills. Lastly, test boring logs for

Table 5. Grain size trends of lodgement tills of the Appalachian Plateau.

Site/Deposits	% Gravel	% Sand	% Silt	% Clay	% Fines*	n
Hylands Ash Monofill(1)	22.1	20.3	38.7	18.9	57.6	7
BFI-Eagle south(2)	27.3	28.0	NA	NA	44.7	14
BFI-Eagle north(2)	19.2	25.1	NA	NA	55.7	6
Ellery Landfill(3)	22.0	22.4	46.5	9.1	55.6	14
Southern Tier Sanitary Landfill(4)	35.0	15.0	NA	NA	50.0	21
Average	25.0	22.0	43.0	10.0	53.0	62

[1]Earth Investigations LTD (1990b)
[2]TAMS Consultants (1994)
[3]Malcolm-Pirnie (1991)
[4]AFI Environmental (1992a)
*Silt & Clay

Table 6. N-values for Lodgement Tills of the Appalachian Plateau

Site	Average	Range	Number
Southern Tier Sanitary Landfill[1]			
Gray	112	15-198	71
Brown	68	6-162	106
Hylands Ash Monofill(2)			
Gray	51	16-182	41
Brown	45	13-120	75
Olean Landfill(3)			
Brown	43	20-100+	33
BFI-Eagle(4)			
Gray	51	43-60	6
Brown	67	8-189	42
Ellery Landfill(5)			
Gray	59	25-166	45
Brown	35	5-140	52

1 AFI Environmental (1992a)
2 Earth Investigations LTD (1990b)
3 Earth Investigations LTD (1990a)
4 TAMS Consultants (1994)
5 Dunn Geoscience (1988)

Table 7
Lodgement till hydraulic conductivities (cm/sec) of the Appalachian Plateau

Site	Geometric Mean	Min	Max	N
Hylands Ash Monofill[1]				
Gray	3.6×10^{-9}	3.2×10^{-9}	2.0×10^{-8}	4
Brown	3.6×10^{-7}	3.4×10^{-9}	6.9×10^{-9}	4
Ellery Landfill[2]				
Gray	2.5×10^{-5}	3.8×10^{-6}	8.5×10^{-5}	8
Brown	9.9×10^{-5}	2.9×10^{-5}	2.9×10^{-4}	
Eagle-Modern[3]				
Gray	3.2×10^{-8}*	--	--	1
Brown	5.1×10^{-7}*	--	--	1
Olean Landfill[4]				
Brown	8.4×10^{-6}	4.5×10^{-7}	1.6×10^{-5}	2
Southern Tier Sanitary Landfill[5]				
Gray	1.6×10^{-6}	1.8×10^{-7}	2.2×10^{-5}	3
Brown	5.9×10^{-6}	1.8×10^{-7}	1.6×10^{-4}	8

1 AFI Environmental (1992a)
2 Earth Investigations LTD (1990b)
3 Earth Investigations LTD (1990a)
4 TAMS Consultants (1994)
5 Dunn Geoscience (1988)

* Recompacted permeability values.

many of the landfill sites indicate water-sorted lenses in tills to depths typically on the order of 30 feet below ground surface in the lower elevation portions of the valleys. When these lenses contain pockets of water, slug testing of wells yields an average hydraulic conductivity value of approximately $5x10^{-5}$ cm/s which is slightly higher than geometric mean values calculated for both weathered and unaltered lodgement tills.

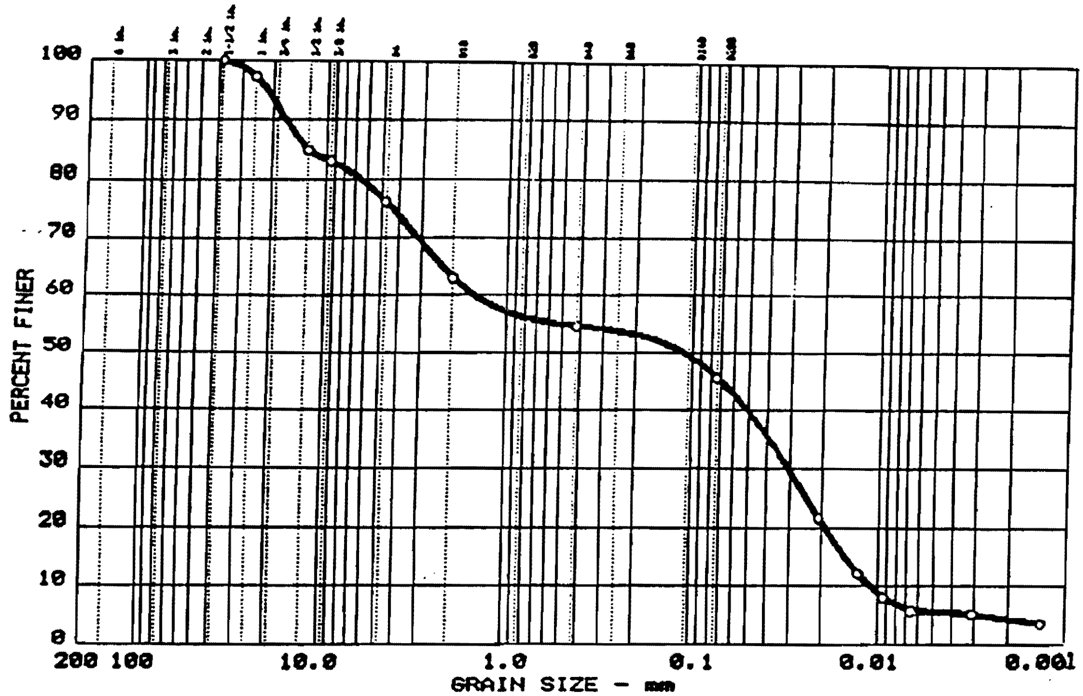
Flow Till

In the low elevation area of the Southern Tier Sanitary Landfill site in Farmersville, Cattaraugus County, a localized pod of normally graded, poorly sorted, silty sand and gravel was encountered. Clasts in this unit are noted by AFI Environmental (1992a, b) to be angular to subrounded. This unit has been interpreted as a flow till on the basis of internal characteristics and the lateral and vertical association with lodgement tills (Cole and Others, 1993). Flow tills are formed when glacial debris is remobilized downslope by gravity and may be deposited supraglacially, subglacially, subaerially or subaqueously (McGown and Derbyshire, 1977; Boulton, 1968; Lawson, 1982; Dreimanis, 1988). Flow till deposits are typically discontinous lenses of variably sorted detritus. The debris that comprises the flow till may have been reworked from other tills or may have been derived from the release of detritus from glacial ice. Overall, the poor sorting, massive nature, and the weak grading of flow till deposits at Farmersville suggest deposition from viscous sediment gravity flows (mudflows or debris flows).

The flow till unit consists of an upper silty gravel facies and a lower sand and boulder unit (Figs. 10a, b). The hydraulic conductivity values reported for the upper silty gravel range between $1.0x10^{-3}$ and $1.0x10^{-5}$ cm/s with a geometric mean of $2.0x10^{-4}$ cm/s. The lower sand and boulder facies is more permeable. Slug tests on monitoring well yield values averaging approximately $3.0x10^{-3}$ cm/s. A full-scale pumping test, however, yielded a hydraulic conductivity value of approximately $2.0x10^{-2}$ cm/s for the localized, permeable, lower horizon (AFI Environmental, 1992b).

Mottled Silt

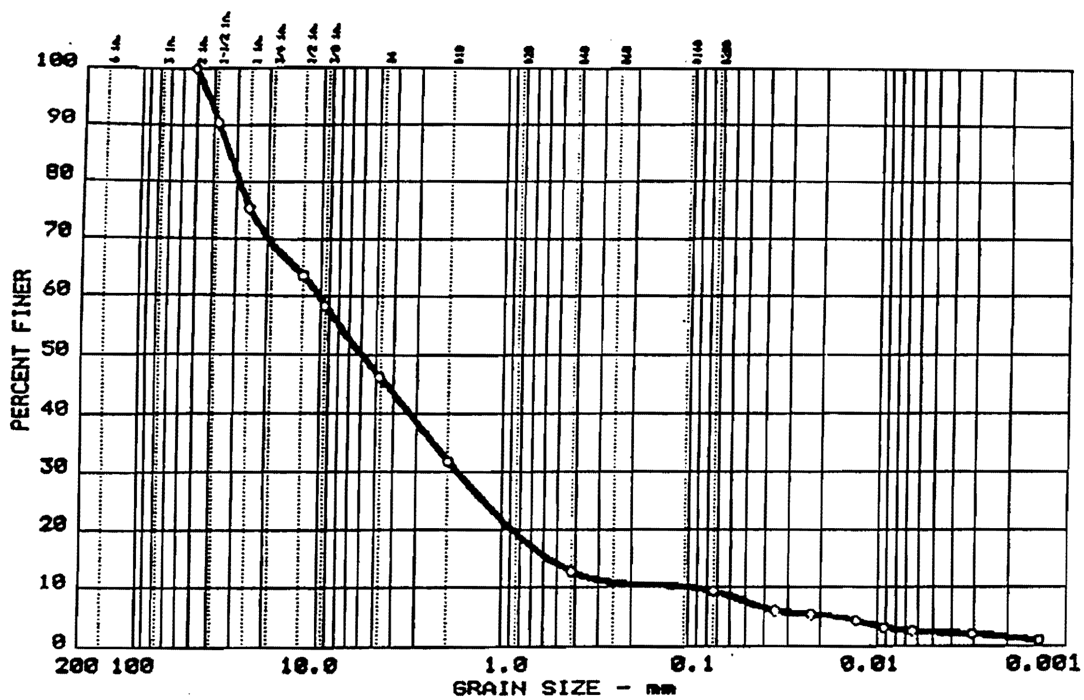
A surficial mottled silt bed has been documented at several sites in the Appalachian Plateau. The facies is typically gray and brown mottled silt with small percentages of sand and gravel (Fig. 11). The beds range in thickness between 1 and 6 feet. The unit apparently disconformably overlaps underlying lodgement and ablation till facies suggesting that the mottled silt accumulated during and/or following the waning phases of glacial recession. Furthermore, typical N-values based upon standard penetration tests average approximately 15 blows per foot, a value much lower than representative of the underlying, over-consolidated, lodgement till facies but similar to the mean value calculated for ablation tills. The mottled silt has a seismic velocity ranging between 1700 and 3500 feet per second. This range of values is also lower than the range for regional lodgement and ablation till facies. The genesis of the mottled silt remains uncertain. Working hypotheses are that the facies may be any of the following: 1) a bioturbated,



% +3"	% GRAVEL	% SAND	% SILT	% CLAY
0.0	23.9	30.4	40.0	5.7

LL	PI	D_{85}	D_{60}	D_{50}	D_{30}	D_{15}	D_{10}	C_c	C_u
nv	none	12.88	1.51	0.11	0.031	0.0151	0.0109	0.06	138.

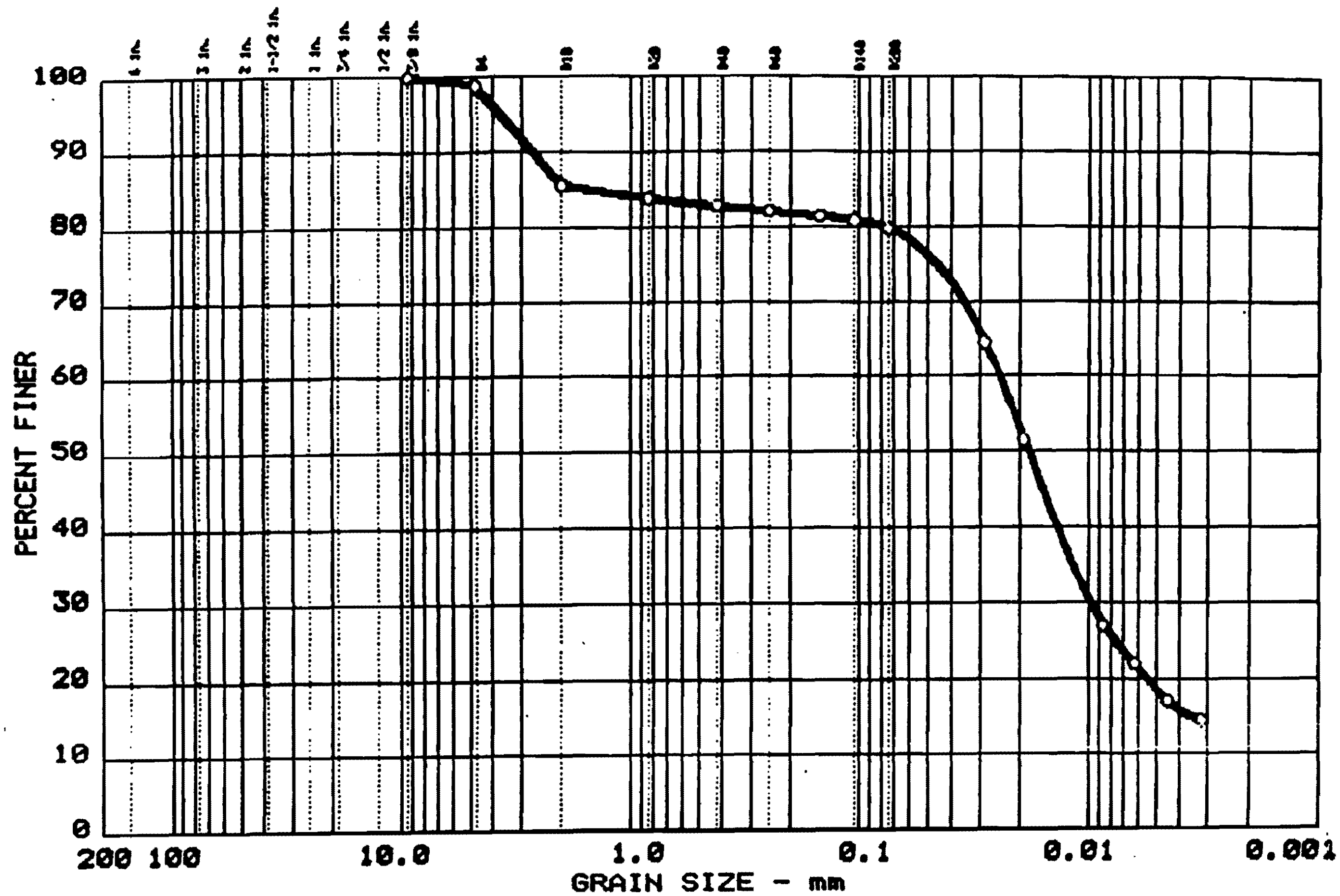
Figure 10A. Grain-size distribution for upper flow till horizon.



% +3"	% GRAVEL	% SAND	% SILT	% CLAY
0.0	53.8	36.8	6.9	2.5

LL	PI	D_{85}	D_{60}	D_{50}	D_{30}	D_{15}	D_{10}	C_c	C_u
26	10	33.88	10.34	5.88	1.776	0.5617	0.0922	3.31	112.2

Figure 10B. Grain-size distribution for lower flow till horizon.



% +3"	% GRAVEL	% SAND	% SILT	% CLAY
0.0	1.2	19.0	61.3	18.5

Figure 11. Representative grain-size distribution for the mottled silt horizon.

humid climate loess (Donald Owens, personal communication); 2) a till-derived colluvium (Parker Calkin, personal communication); 3) a late stage, melt-out "slurry till" (Donald Cadwell, personal communication); or 4) an incipient, hydric soil profile.

Regardless of origin, the mottled silt possesses a low permeability. Although the unit is typically too thin to discretely screen wells or piezometers across, vertical hydraulic conductivity values have been obtained for the deposit at the Southern Tier Sanitary Landfill site based upon triaxial permeability tests on Shelby Tube samples. These tests yielded a mean vertical hydraulic conductivity of approximately 1.0 x 10^{-6} cm/s (Blasland, Bouck & Lee, 1994). The horizontal hydraulic conductivity may be one to two orders of magnitude higher if bioturbation has not completely homogenized the deposit.

Groundwater Flow Trends

The general groundwater flow conditions on landfill sites in the Appalachian Plateau are characteristic of high elevation, bedrock-supported terrains with limited recharge areas (Fig 12). Recharge occurs on high elevation ridges where weathered glacial tills are typically less than 10 feet thick. Flow vectors are oriented vertically downward in the recharge areas until groundwater intercepts one of several potentially transmissive zones. Preferred flow paths become more numerous on valley side-slopes as the glacial till profile thickens and the influence of till aquicludes and aquitards becomes increasingly pronounced.

The uppermost preferred flow path occurs at the base of the mottled silt layer. This hydrostratigraphic unit typically contains a throughflow or interflow system which diverts a significant volume of vadose water (sensu Fetter, 1994, p. 5) from deeper, transmissive units below the water table. Throughflow systems typically discharge as diffuse spring lines at the base of the valley side slopes. The vadose water then flows overland toward surface water bodies. The interflow systems discharge directly to surface water bodies without the overland flow component.

A shallow, unconfined flow system typically develops in the weathered till profile. Due to the low permeability of surficial units, the seasonal high groundwater table typically occurs within 5 feet of the ground surface beneath the lower portions of the local drainage basins. Consequently, most landfills are designed with a groundwater suppression system to maintain separation between the water table and the baseliner until hydrostatic pressures equalize beneath the facility.

On some sites, the shallow-occurring "groundwater" may be restricted to dessication crack networks and water-sorted lenses in the valley-center ablation till profiles. Observations from test pits indicate that the bulk of the free water available within the top 10 to 20 feet below ground surface on some sites emanates from the throughflow/interflow horizons. Except for pockets of water in small lenses of granular material, the till matrix commonly appears to be unsaturated. None-the-less, shallow monitoring wells often produce small volumes

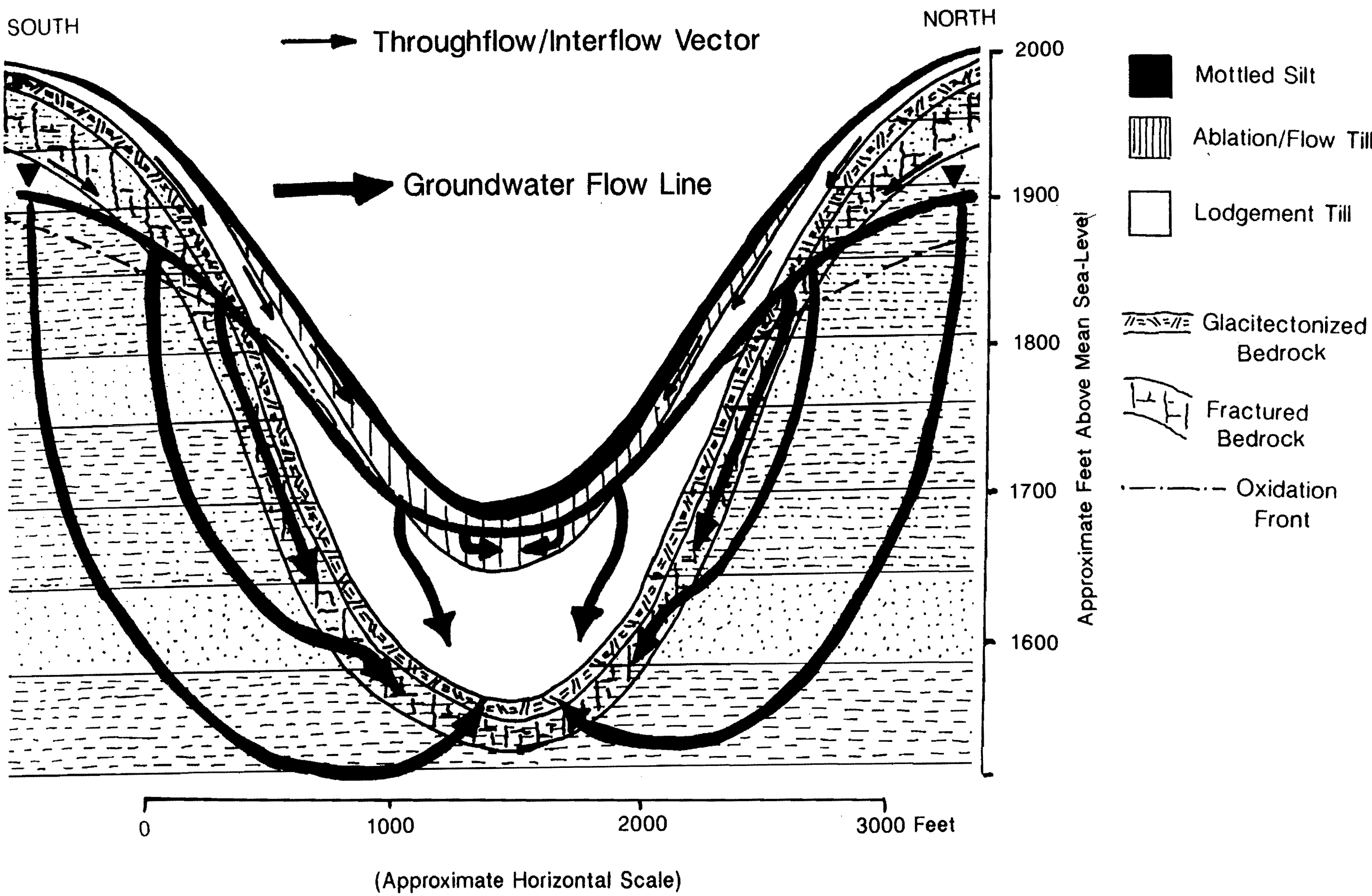


Figure 12. Generalized Groundwater Flow Patterns in a Till-Filled Valley of the Appalachian Plateau

of water. The source of this water must be either vertical leakage through dessication cracks or leakage from the interflow/throughflow zone along the well casings. Thus, definition of the shallow water table becomes equivocal on some sites, because it is difficult to reconcile the test pit observations with interpretation of a continuous phreatic surface in the weathered till profile.

A groundwater invert sometimes defines the base of a shallow "perched water table" zone in the weathered profile. Because the deeper, unaltered lodgement till is lower in permeability than the shallower, weathered ablation and lodgement till profile, the gray color marks the top of the first aquiclude. These gray tills may be at least seasonally unsaturated on some sites and may be perennially unsaturated at the Hylands site (Earth Investigations Ltd, 1990b). Regardless of whether saturated conditions extend from the weathered zone downward through the unaltered profile or not, the gray till/glacitectonized bedrock aquiclude separates the upper groundwater flow zone from the bedrock flow zone. Commonly, the aquiclude imposes artesian pressure on the flow zone in the underlying fractured bedrock aquifer just above and within valley center areas. Water table conditions, defined by a lack of pressure head in monitoring wells, commonly can be documented both above and below the aquiclude in the intermediate elevation areas of the local drainage basins. In the uppermost reaches, the aquiclude does not exist, because the oxidation front generally extends through the thin till profile and into bedrock. The water table resides in bedrock at the crests of the hills.

Background Water Quality

Representative background water quality has been compiled from several sites and is presented in Table 8. Values for field parameters (specific conductance and pH) as well as major metals, major anions and Total Organic Carbon (TOC) are provided. Most of these waters may be classified as calcium bicarbonate geochemical facies and reflect the solubility of calcareous body fossils in local bedrock units. Fossiliferous strata have been incorporated into the local tills and are directly observable in drill cores of both dislodged bedrock rafts in the glacitectonized zone and in in-situ bedrock.

The concentrations of total dissolved solids (TDS) are remarkably low for sites in the Appalachian Plateau. The low TDS values reflect the position of the plateau in the headland areas of several regional-scale groundwater basins. As will be illustrated in the following sections of this paper, these low TDS groundwaters differ significantly with water monitored near base-level of the regional flow system beneath the Erie-Ontario Plain. Near the sublacustrine groundwater outflow zones, water quality is significantly affected by naturally occurring, high concentrations of dissolved solids.

ERIE-ONTARIO PLAIN

Review of Table 2 reveals that two hydrogeologic settings within the Erie Ontario Plain have been utilized for host solid waste management facilities. These

Table 8
Representative Water Quality
Appalachian Plateau

Parameter*,(6)	Hylands Ash Monofill(1)				Southern Tier Sanitary Landfill(2)					
	Oxidized Ablation Till	Unaltered Lodgement Till	Bedrock Overburden Contact	Deep Bedrock	Flow Till	Oxidized Lodgement Till	Unaltered Lodgement Till	Shale Bedrock	Sandstone Bedrock	Carpenter Brook
pH (pH units)	6.79	7.20	7.71	7.58	7.88	7.71	8.16	7.75	7.83	7.60
Specific Conductance (umhos/cm)	540	530	468	370	348	279	485	242	250	231
Total Dissolved Solids	294	312	277	228	268	177	313	155	191	161
Hardness	252	258	216	207	173	153	157	146	183	144
Alkalinity	213	NA	230	195	103	150	157	145	151	126
Chloride	5.20	0.57	1.40	92.2	59.0**	2.18	6.64	1.50	2.42	23.3
Sulfate	43.1	42.4	33.0	28.0	54.0	26.2	110	14.9	28.7	12.9
Nitrate	0.27	0.18	ND	0.35	1.08	0.66	0.79	0.09	0.30	0.08
Ammonia	0.15	0.06	0.13	0.08	<0.05	0.07	0.08	0.07	0.06	0.05
Manganese	0.32	1.15	0.23	ND	0.25	0.33	0.38	0.14	0.34	0.04
Iron	3.00	3.00	0.54	0.40	3.04	1.91	27.9	1.27	27.6	0.23
Magnesium	16.4	20.4	17.0	0.20	6.91	8.40	5.65	8.70	6.20	3.98
Calcium	51.3	48.0	55.0	59.0	36.0	64.2	51.2	35.3	43.2	32.1
Sodium	11.0	11.6	16.3	5.10	33.0	13.4	52.3	6.50	10.6	10.3
Potassium	2.20	3.40	3.40	5.00	2.04	3.58	2.60	1.82	2.17	1.08
TOC	2.55	1.50	1.45	1.40	1.64	1.67	1.62	1.87	1.58	3.43

* all values in mg/l unless otherwise specified
** possible road salt influence
*** acid shale exposed at ground surface
(6) Values reported are mostly arithmetic means for each geologic interval.

(1) Source: Earth Investigations Ltd. (1990b)
(2) Source: AFI Environmental (1992a)
(3) Source: Earth Investigations Ltd. (1990a)
(4) Source: Malcolm-Pirnie (1994)
(5) Source: Malcolm-Pirnie (1991)

Table 8
Representative Water Quality
Appalachian Plateau

Parameter*,[6]	Olean Landfill[3]		Modern-Eagle[4]		Ellery Landfill[5]	
	Shallow Bedrock	Surface Water	Bedrock Overburden Contact	Shallow Bedrock	Bedrock Overburden Contact	Oxidized Ablation Till
pH (pH units)	4.68***	6.96	7.56	7.40	7.30	6.60
Specific Conductance (umhos/cm)	125	191	470	488	283	1280
Total Dissolved Solids	94.8	126	260	255	177	687
Hardness	10.0	39.0	173	166	101	700
Alkalinity	12.0	71.9	222	204	384	718
Chloride	2.53	5.60	2.00	3.50	8.86	8.01
Sulfate	9.61	12.7	21.5	29.0	17.4	25.7
Nitrate	0.01	0.31	0.68	5.00	ND	ND
Ammonia	0.12	0.92	NA	NA	0.39	0.10
Manganese	0.32	0.38	0.28	0.44	0.13	0.89
Iron	6.53	1.00	1.36	0.80	0.51	1.34
Magnesium	5.69	5.64	12.4	11.2	NA	NA
Calcium	1.58	13.7	49.0	48.0	34.3	140
Sodium	1.09	2.85	21.0	17.0	21.5	14.0
Potassium	3.47	2.10	NA	NA	NA	NA
TOC	4.24	7.72	15.5	17.0	4.33	7.40

* all values in mg/l unless otherwise specified
** possible road salt influence
*** acid shale exposed at ground surface
(6) Values reported are mostly arithmetic means for each geologic interval.

(1) Source: Earth Investigations Ltd. (1990b)
(2) Source: AFI Environmental (1992a)
(3) Source: Earth Investigations Ltd. (1990a)
(4) Source: Malcolm-Pirnie (1994)
(5) Source: Malcolm-Pirnie (1991)

two terrains are the "Glacial Till Over Bedded Sedimentary Rock" and "Glacial Lake Deposits" hydrogeologic settings of Aller and others (1987). The terrains dominated by glacial tills fall into three categories based upon depositional settings: 1) ground moraine (Fig. 13); 2) drumlin (Fig. 14); and 3) water-lain till/subaqueous flow till (Fig. 15).

Although the Lake Ontario Plain incorporates outcrop belts of diverse strata ranging in age from Ordovician to Lower Devonian, the upper 100 feet or so of bedrock may be divided into hydrostratigraphic units comparable to the intervals defined for the Appalachian Plateau. These hydrostratigraphic units include: 1) an uppermost, decomposed rock aquitard; 2) a medial, fractured rock aquifer; and 3) a deeper, competent rock aquitard.

Weathering (oxidation) of both surficial deposits and bedrock is also common to depths of 10 to 30 feet. As is often the case with the sites in the Appalachian Plateau, the weathered profile typically obscures lateral and vertical glacigenic facies boundaries in the shallow subsurface zone.

General characteristics of landfill sites on the Erie-Ontario Plain are more fully discussed in the following sections.

Regional Bedrock Hydrostratigraphy

Hydrogeologic studies for a variety of environmental projects including landfill siting have contributed significantly to the understanding of regional bedrock stratigraphic trends on the Lake Ontario Plain. Major hydrogeologic investigations such as the recent study of the bedrock hydrology "in press; also in this volume" by the USGS Water Resources Division have produced valuable drill core sections that have facilitated regional correlations of Ordovician and Silurian strata (see Brett and others, 1990a,b; impress). Drill cores from landfills near Middleport (Niagara County), Albion (Orleans County) and Webster (Monroe County) as well as from sewer projects in Rochester and the defunct super collider project in Walworth (Wayne County) have been instrumental in extending correlations between Niagara County and Wayne County, a distance in excess of 100 miles (150 km). Based, in part, upon these drill cores, cross-sections of Upper Ordovician through Upper Silurian strata may be developed (Fig. 16a-c).

The hydrostratigraphy of the bedrock units is profoundly influenced by past and present surficial processes. Chemical weathering and post-glacial isostatic rebound have resulted in the tripartite division of the shallow bedrock profile into three hydrostratigraphic units: 1) an upper aquitard consisting of disaggregated (decomposed) rock; and medial, fractured rock aquifer; and 3) a lower, competent rock aquitard.

Decomposed Rock Zone

The decomposed rock aquitard is typically overlain directly by a lodgement till facies. The hydrostratigraphic unit is best defined atop shale formations.

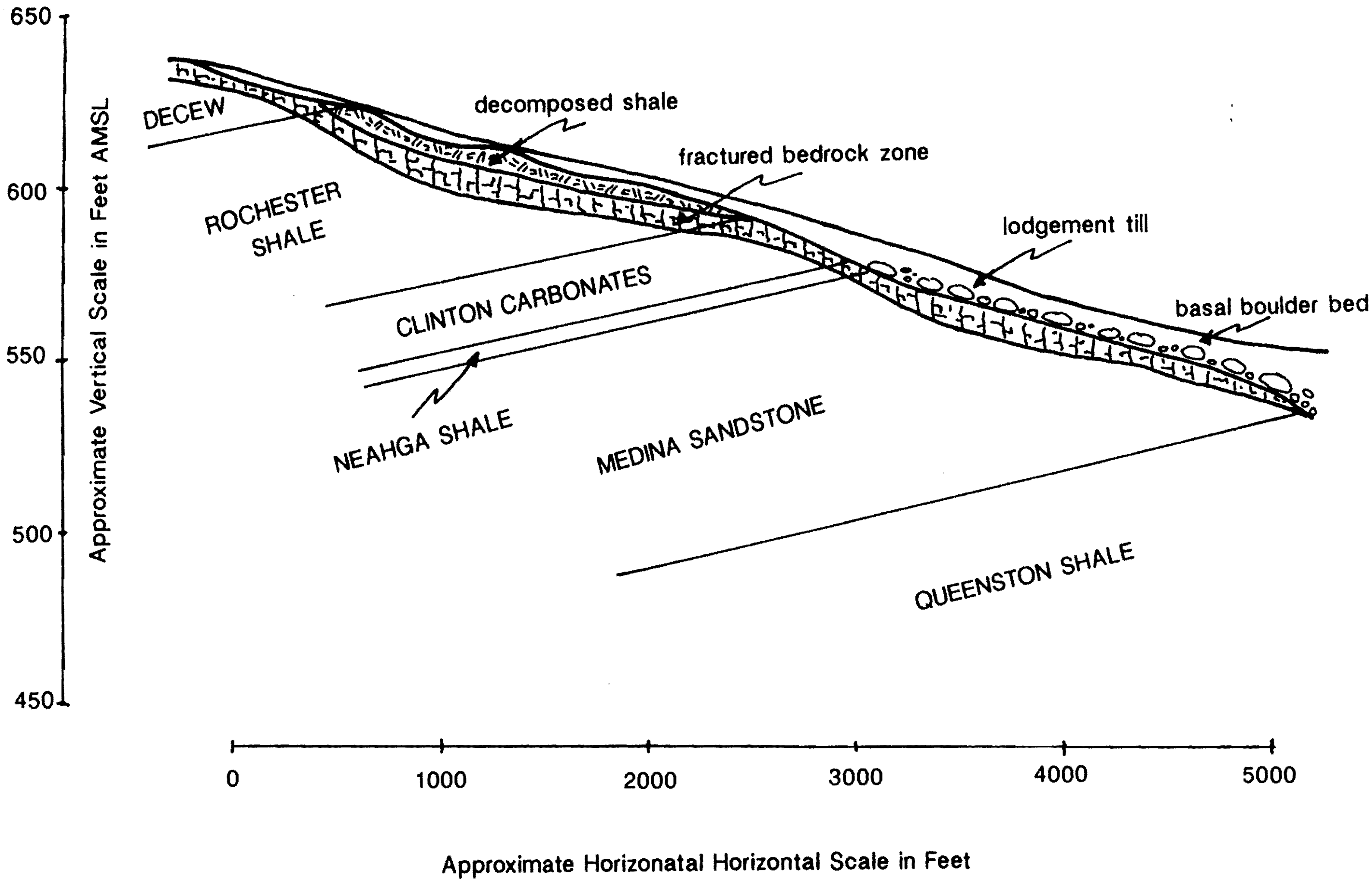


Figure 13. Geologic cross-section of ground moraine and underlying bedrock strata on the Lake Ontario Plain.

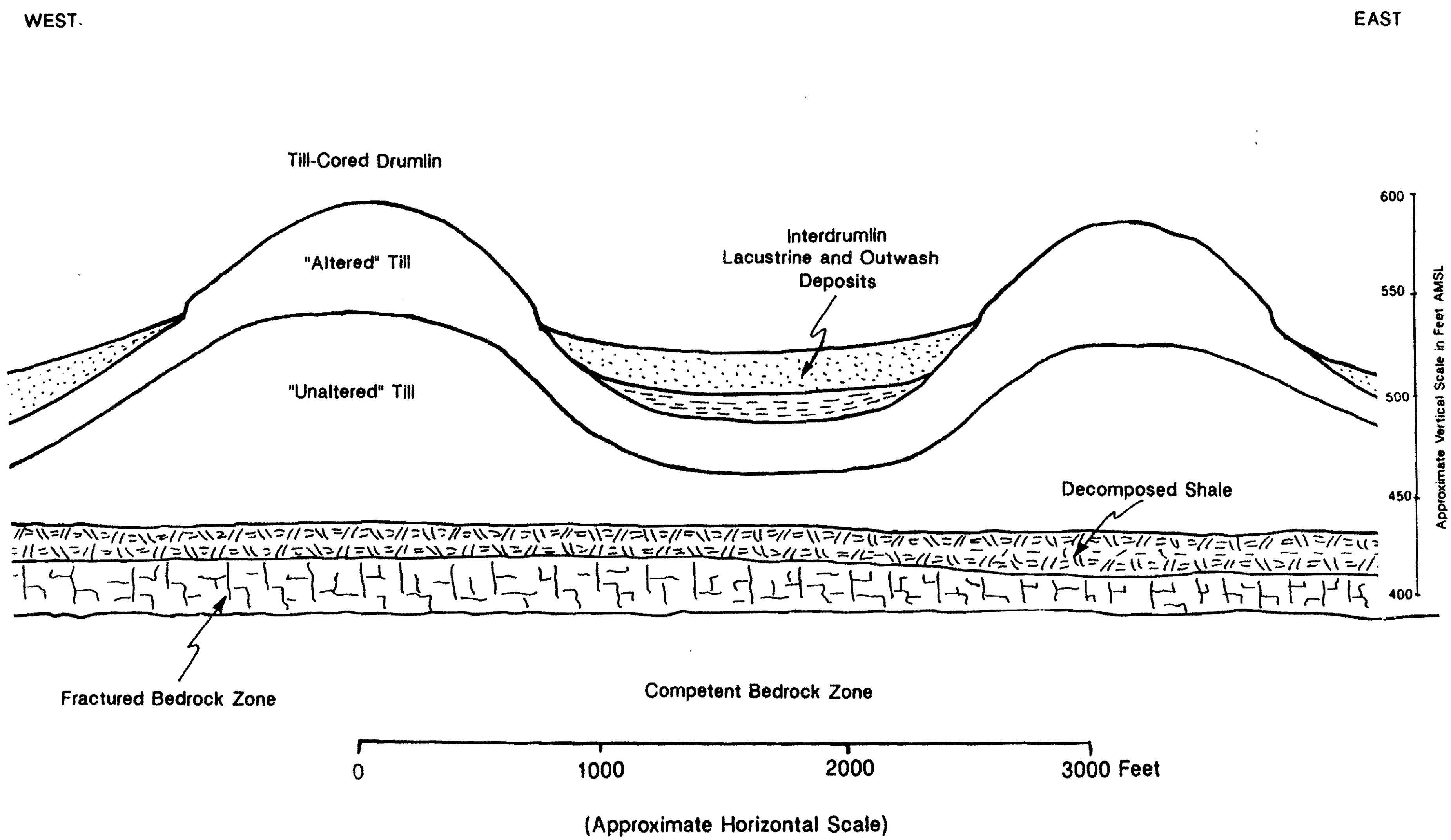
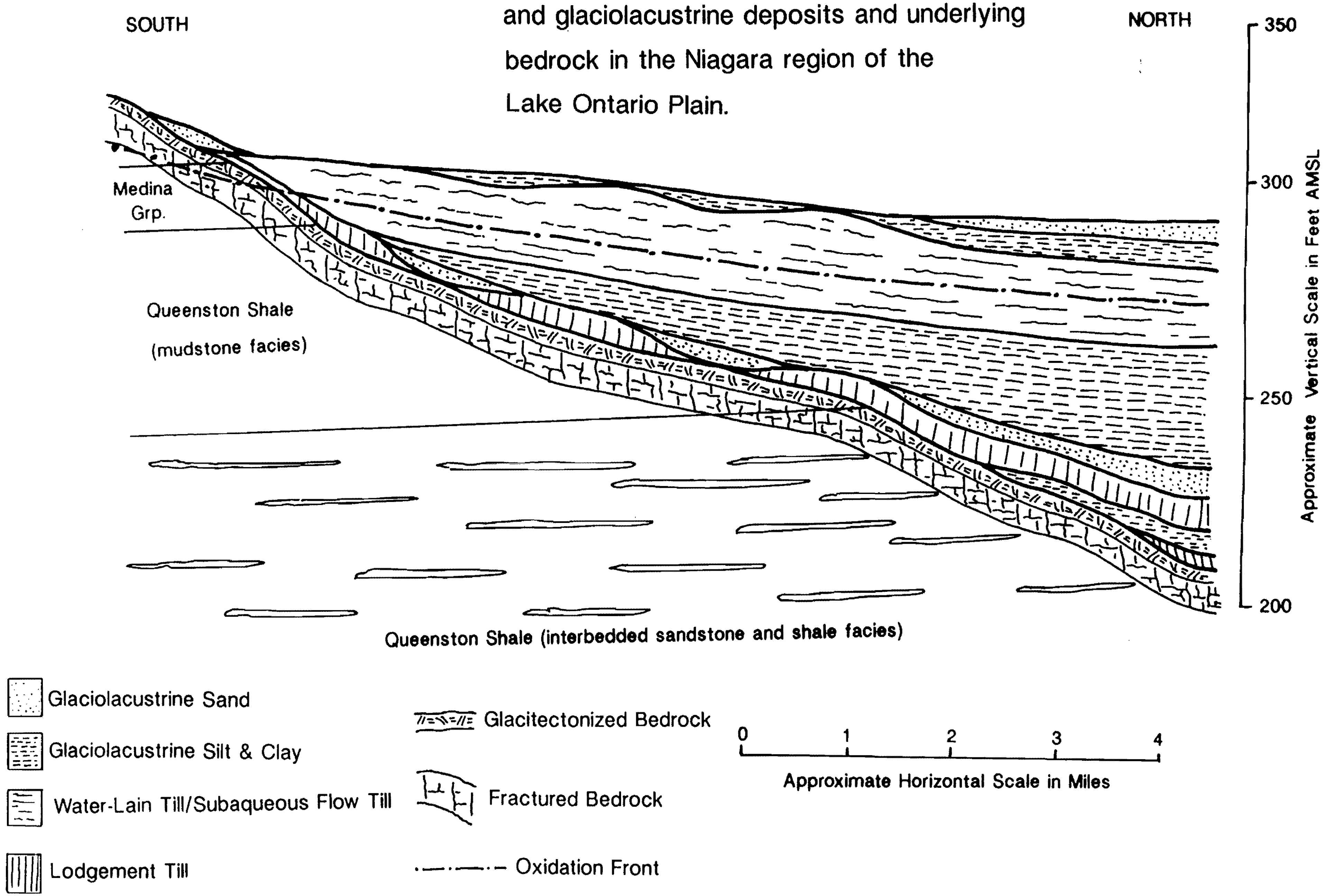


Figure 14. Geologic cross-section of drumlin and bedrock hydrostratigraphic units on the Lake Ontario Plain.

Figure 15. Geologic cross-section of interstratified till and glaciolacustrine deposits and underlying bedrock in the Niagara region of the Lake Ontario Plain.



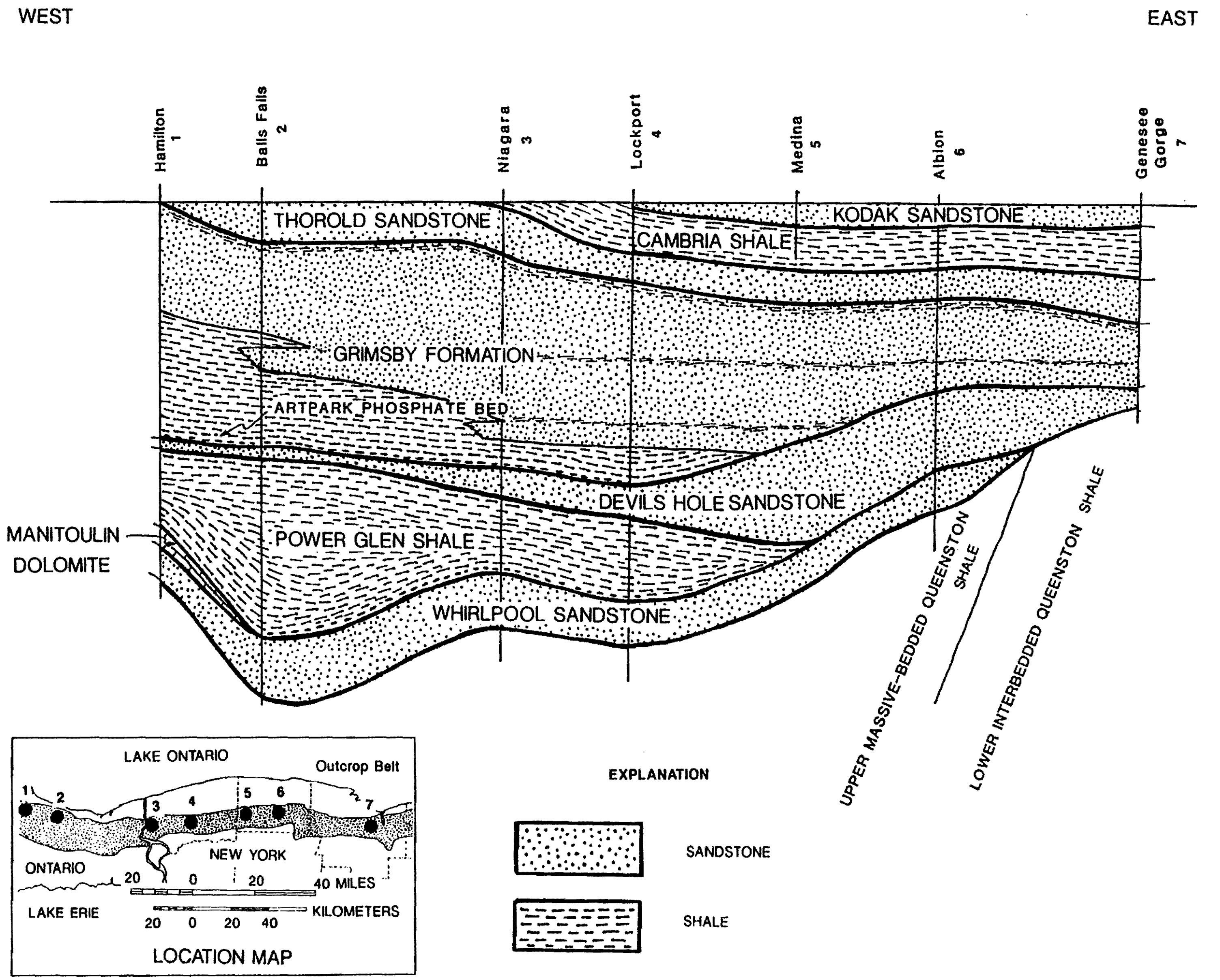


Figure 16A. Geologic cross-section of Medina Group strata on the Lake Ontario Plain.

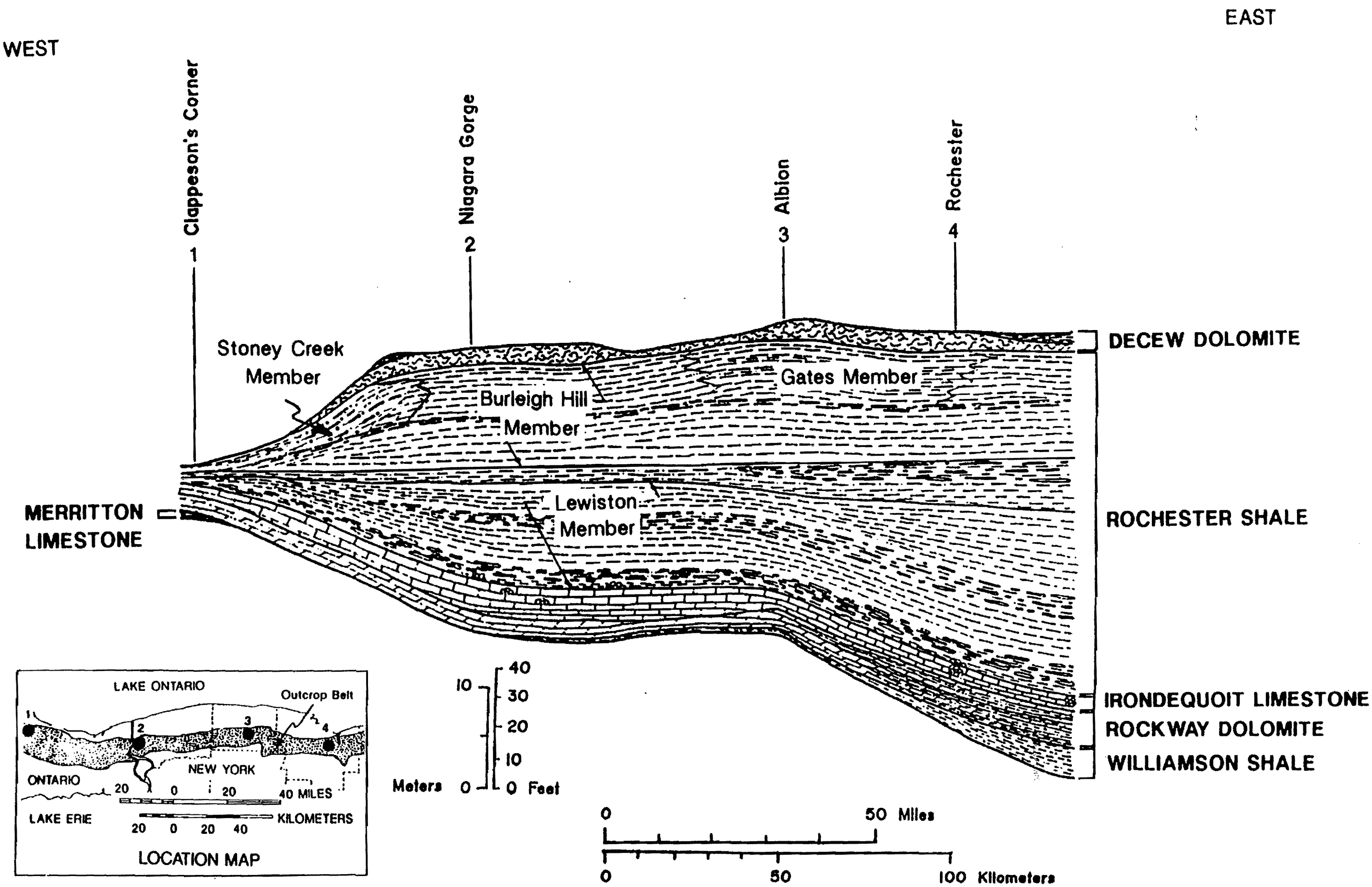


Figure 16B. Geologic cross-section of upper Clinton Group strata on the Lake Ontario Plain.

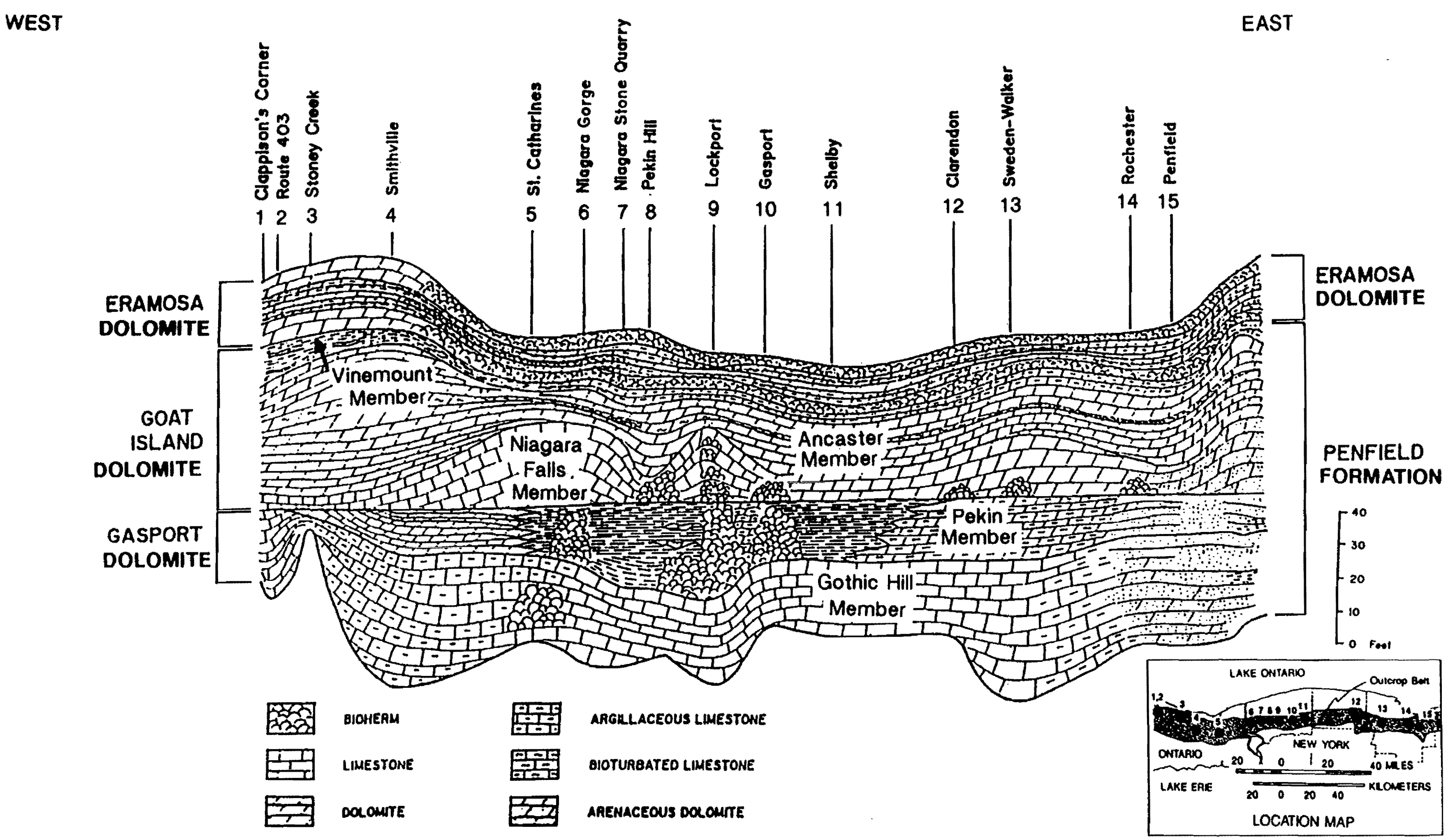


Figure 16C. Geologic cross-section of lower Lockport Group strata on the Lake Ontario Plain.

Documented, low permeability clay and silt horizons consisting of decomposed shale occur atop the Queenston Shale at the Modern Landfill (Lewiston, Niagara County), the Rochester Shale in borrow pits immediately south of the Orleans Sanitary Landfill (Albion, Orleans County), and the Vernon Shale at the Mill Seat Landfill (Riga, Monroe County), the High Acres Landfill (Perinton, Monroe County), and the Galen-Lyons Sanitary Landfill (Galen, Wayne County). Although the regulatory definition of bedrock mandates that the decomposed shale be treated as bedrock, the unit may be sampled using standard penetrations tests through hollow stem augers, i.e. conventional soil sampling techniques.

For most sites the decomposed shale has been combined with the basal lodgement till profile to form a single layer in conceptual and numerical hydrologic models. Consequently, discrete hydraulic conductivity values are not widely available. However, a fairly comprehensive data base consisting of slug test results from 15 piezometers screened in the upper 9 feet of Queenston Shale at the Modern Landfill is reported by Wehran Envirotech (1991). These data are included in Table 9. Hydraulic conductivity values for decomposed Queenston Shale at the Modern Landfill range between 8.3×10^{-6} and 1.9×10^{-3} cm/s. The geometric mean of these values for decomposed Queenston Shale is 2.6×10^{-4} cm/s.

Limited data are also available from both the High Acres Landfill and the closed Galen-Lyons Sanitary Landfill for the decomposed Vernon Shale horizon (see Table 9). Two wells screened in the interval at High Acres yield a geometric mean hydraulic conductivity of 2.7×10^{-5} cm/s (Eckenfelder, 1992). Four wells screened in the interval at the Galen-Lyons Landfill yielded hyraulic conductivity values between 3.9×10^{-5} and 9.9×10^{-3} cm/s (Larsen, 1990). The geometric mean K value from the Galen-Lyons site for decomposed Vernon Shale is 2.6×10^{-4} cm/s.

Although hydraulic conductivity data have not yet been compiled on the decomposed Rochester Shale aquitard, its use at the closed Orleans Sanitary Landfill for the low permeability layer in the cover system suggests that the material may be recompacted to achieve a 1.0×10^{-7} cm/s or lower permeability. In-situ hydraulic conductivity values are likely to be within an order of magnitude of the recompacted permeabilities.

<u>Fractured Bedrock Zone</u>

On sites where the fractured and underlying competent bedrock zones are defined discretely, RQD data commonly reflect the differences in rock competency (Table 10). These data indicate that the RQD of a formation whose upper surface lies within the bedrock fracture zone is commonly considerably lower than the RQD value representative of the formation in the competent bedrock zone. The low RQD values of the aquifer zone reflect the numerous, bedding-parallel fractures that probably developed during unloading of glacially compressed bedrock. Consequently the hydraulic conductivity of the fractured bedrock zone is generally higher than corresponding values for the overlying glacitectonized zone and the underlying competent bedrock zone (see Table 9).

Table 9. Hydraulic conductivity (cm/sec) of formations in the fractured and competent bedrock zones for the Lake Ontario Plain.

Site Formations	*Weathered Zone &* Fractured Zone Mean	Max	Min	N	Competent Zone Mean	Max	Min	N
Modern Landfill[1] Queenston Shale	*2.6×10^{-4}* 1.1×10^{-2}	*1.9×10^{-3}* 1.5×10^{-2}	*8.3×10^{-6}* 7.3×10^{-5}	*15* 2	8.1×10^{-6}	4.8×10^{-5}	9.2×10^{-7}	5
Niagara Landfill-Tonowanda[2] Camillus Shale	--	3.0×10^{-4}	4.5×10^{-5}	--	--	--	--	--
Niagara Landfill-Cecos[3] Lockport Group	9.0×10^{-4}	--	--	--	--	2.0×10^{-5}	4.0×10^{-6}	--
Mill Seat Landfill[4] Vernon Shale	6.1×10^{-4}	2.4×10^{-3}	8.4×10^{-5}	12	4.1×10^{-5}	1.0×10^{-3}	4.9×10^{-8}	24
Niagara Landfill-Lockport[5] Rochester Shale	--	--	--	--	--	10^{-6}	10^{-8}	--
High Acres[6] Vernon Shale	*2.7×10^{-5}* 1.5×10^{-2}	-- 1.5×10^{-1}	-- 3.2×10^{-7}	-- 16	2.5×10^{-5}	5.2×10^{-4}	1.1×10^{-6}	20
Seneca Meadows[7] Camilus Shale and Bertie Group	1.7×10^{-3}	4.0×10^{-3}	1.8×10^{-5}	--	--	--	--	--
Galen Lyons[8] Vernon Shale	*7.5×10^{-5}* --	*1.0×10^{-4}* 10^{-2}(est)	*4.0×10^{-5}* 10^{-4}(est)	-- --	--	--	--	--
Orleans Sanitary Landfill[9]								
Devils Hole/Whirlpool Sandstone	1.1×10^{-5}	3.3×10^{-5}	1.9×10^{-6}	3	8.3×10^{-6}	5.8×10^{-5}	5.8×10^{-7}	10
Grimsby Sandstone	5.7×10^{-6}	1.4×10^{-5}	1.7×10^{-6}	4	1.8×10^{-5}	8.6×10^{-5}	5.8×10^{-6}	4
Thorold Sandstone	7.6×10^{-5}	1.0×10^{-4}	6.7×10^{-5}	3	1.8×10^{-7}	--	--	1
Cambria Shale	1.1×10^{-4}	2.0×10^{-4}	5.3×10^{-5}	3	1.5×10^{-5}	--	--	1

[1]Wehran-New York, Inc. (1991)
[2]RECRA Environmental (1988)
[3]RECRA Environmental (1985)
[4]H & A of New York (1987)
[5]GZA (1984)
[6]Eckenfelder (1992)
[7]Dunn Geosciences (1990)
[8]Larson (1990)
[9]AFI Environmental and Beak Consultants, Ltd. (1988)

Table 10. RQD values of formations in the fractured and competent bedrock zones for the Lake Ontario Plain.

Site	Average RQD Fractured Zone	Average RQD Competent Zone
Modern Landfill[1]		
Queenston Shale	27	72
Orleans Sanitary Landfill[2]		
Whirlpool/Devils Hole Sandstone	42	78
Grimsby Sandstone	64	65
Thorold Sandstone	43	77
Cambria Shale	9	60
Kodak Sandstone	34	No data available
Maplewood & Reynales fms.	10	No data available

[1]Wehran-New York, Inc. (1991)
[2]AFI Environmental and Beak Consultants, Ltd. (1988)

Table 11
General Surficial Stratigraphies
of Sites on the Erie-Ontario Plain

Modern Landfill upper glaciolacustrine cap upper till lower glaciolacustrine sequence basal till	Niagara Landfill-Tonawanda upper till lower glaciolacustrine sequence basal till
Mill Seat Landfill upper glaciolacustrine sequence upper till lower glaciolacustrine sequence basal till	High Acres outwash glacial till/outwash brown "weathered" till gray basal till
Galen-Lyons Landfill outwash "weathered" till basal till	Seneca Meadows upper glaciolacustrine sequence upper till lower glaciolacustrine sequence basal till

Competent Bedrock Zone

In the competent bedrock zone, drill core samples are well-preserved. The hydraulic conductivity of the competent bedrock zone is slightly lower than that of the overlying bedrock hydrostratigraphic units (Table 11). Most sites possess a competent bedrock profile exhibiting a geometric mean K value in the low to mid 10^{-5} cm/s range although stratigraphic control on permeabilities at some sites may result in slightly higher mean values.

Regional Surficial Hydrostratigraphy

Surficial geologic maps covering parts of the study area are provided by Kindle and Taylor (1914), Muller (1977), Young (1980), Yager and others (1984), Cadwell (1988) and Goodman and Stanwix (1994). In addition, a regional Pleistocene stratigraphy for the Erie-Ontario Plain has been synthesized by Calkin and Muller (1992).

Review of hydrogeologic reports for several of the 10 landfill sites situated on the Erie-Ontario Plain reveal general similarities in their stratigraphic successions (Table 11). Furthermore, at least some of these site-specific stratigraphies may be understood within the context of the regional synthesis recently published by Calkin and Muller (1992).

The general stratigraphy of the Lake Ontario Plain consists of two or more couplets of glacial till and recessional lacustrine/outwash facies (Fig. 17) that mostly record Port Huron stade and younger Quaternary history (Calkin and Muller, 1992). This general stratigraphy appears to be recorded in the surficial stratigraphy of several landfill sites on the Lake Ontario Plain (Table 11).

Lower Glaciolacustrine Beds

The oldest documentated surficial deposits on the Lake Ontario Plain appear to be glaciolacustrine silts and clays that are exposed along the shoreline near Somerset, Niagara County. As the bedrock topography increases to the south, this lowest unit probably pinches out against the bedrock subcrop and is beveled at its upper surface by an overlying lodgement till, designated the Furnaceville Till by Calkin and Muller (1992). At landfill sites south of the Onondaga Escarpment, lodgement tills that share the same general geotechnical and hydraulic properties with the Furnaceville Till occur immediately above bedrock except where deeply incised fluvial valleys created topographic lows in the bedrock surface. One such valley has been documented on the Ontario County Landfill site in the Town of Flint (Wehran Engineering, 1986). In the bedrock valley, a glaciolacustrine silt and clay unit attaining a maximum documented thickness of 58 feet overlies shales assigned to the Ludlowville and/or Moscow Formation of the Hamilton Group. Similarities in the succession of overlying till and glaciolacustrine units on sites both north and south of the Onondaga Escarpment suggests that the basal silt and clay deposit at the Ontario County Landfill occupies the same stratigraphic position as the basal silt and clay deposit documented along the Lake Ontario shore by Calkin and Muller

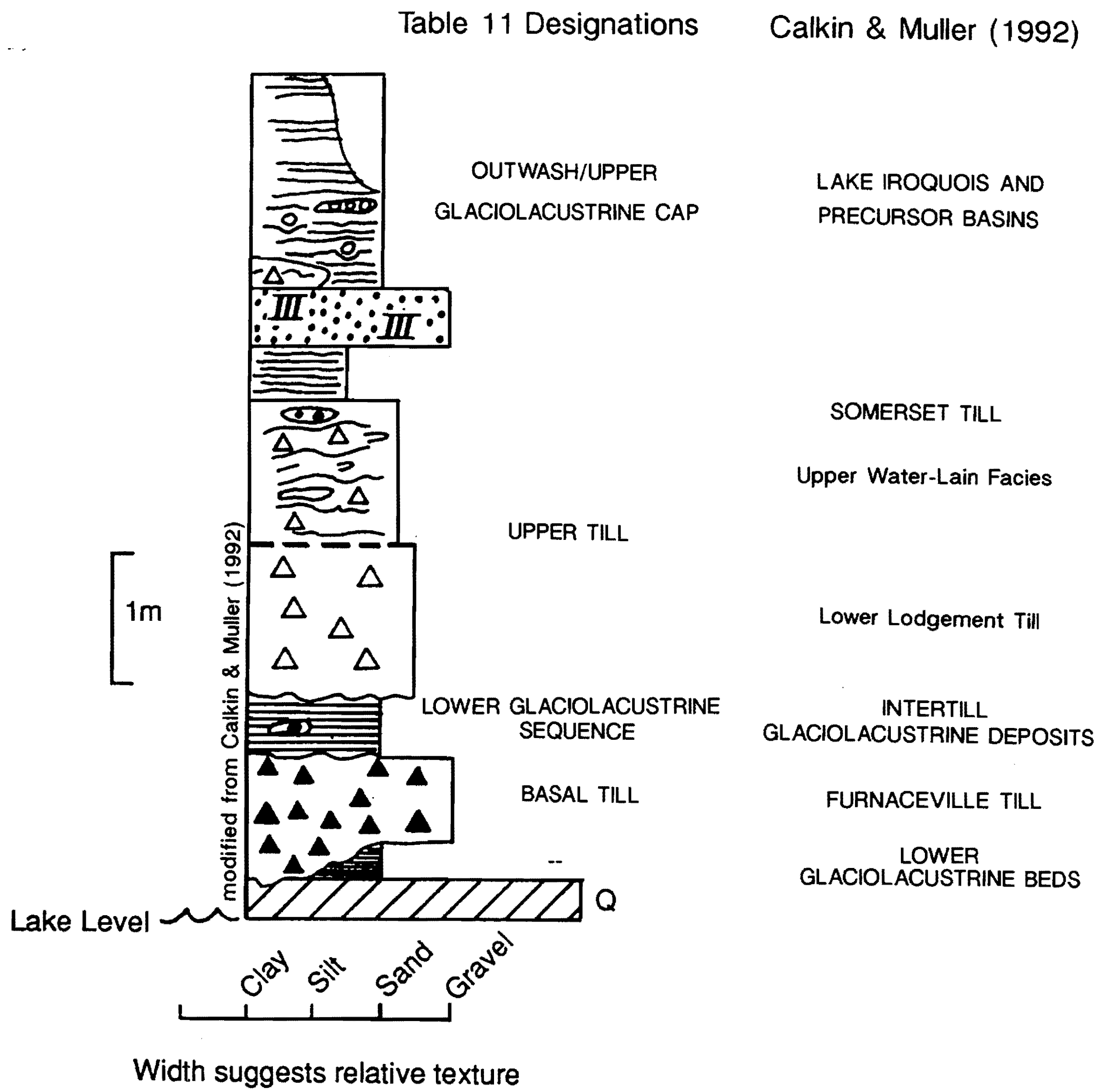


Figure 17. Composite stratigraphic section of regional surficial geologic units encountered on landfill sites on the Lake Ontario Plain.

(1992).

Furnaceville Till

The Furnaceville Till is described by Calkin and Muller (1992) as a stony, massive, red diamicton. Clasts are particularly concentrated near the base of the unit. Large clasts may comprise 50 to 80 percent of the unit by mass. Analysis of 37 till matrix samples by Brennan and Calkin (1984) yielded an average sand:silt:clay ratio of 45:32:23.

Calkin and Muller (1992) have been able to establish correlation of the Furnaceville Till between Niagara and Wayne Counties, a distance in excess of 150 km. Review of till stratigraphies on landfill sites within the same region supports correlation of the Furnaceville Till across ground moraines and drumlin cores.

As would be expected for a lodgement till, the unit is over-consolidated and densely compacted. N-values derived from standard penetration tests (ASTM-D-1586) are summarized in Table 12. Average N-values collected from 6 sites on the Erie-Ontario Plain range between 42 and 124. Caution must be used in evaluating these N-values, however. High blow counts may be attributable to splitspoon contact with cobbles and boulders and are not always an indication of the compaction of the till matrix.

The Furnaceville Till forms an aquitard-grade hydrostratigraphic unit. Average hydraulic conductivity values for this lodgement till at 6 landfill sites are presented in Table 13. The geometric mean K-values for these sites range between 1.0×10^{-7} and 4.1×10^{-5} cm/s.

Intertill Glaciolacustrine Deposits

Overlying the Furnaceville Till along the shoreline of Lake Ontario is a glaciolacustrine sand to clay sequence ranging up to approximately 8 feet thick. The glaciolacustrine sequence is erosionally overlain by a second diamicton designated the Somerset Till by Calkin and Muller (1992). Identical successions have been documented at the Modern Landfill, Niagara Landfill-Tonawanda, Mill Seat Landfill and Seneca Meadows Landfill.

At most sites, the intertill glaciolacustrine sequence may be divisible into lowstand sand and silt and highstand silt and clay intervals. The lowstand sands apparently record initial transgression and reworking of the Furnaceville Till substrate during glacial recession. At the Modern Landfill, the lacustrine lowstand deposit is primarily a sandy silt exhibiting a gravel:sand:silt ratio of 18:27:55 (Table 14). In constrast, the glaciolacustrine highstand deposit is a fairly lean, varved, silty clay.

On two sites in the Niagara region, the intertill glaciolacustrine clay is apparently normally consolidated and soft. The average N-values for the lower clay at the Niagara Landfill-Tonawanda and Modern Landfill sites are 8 and 6 blows per

Table 12
Surficial deposit N-values of the Lake Ontario Plain

Site	Average	Range	n
Seneca Meadows[1]			
Upper lsc*	14	wor*-46	60
Upper till	33	3-195	95
Lower lsc	27	wor-94	26
Basal sand/basal till	49	4-94	11
Niagara Landfill-Tonawanda[2]			
Upper till	35	11-87	31
Lower lsc	8	wor-20	53
Basal sand/basal till	67	7-120	13
Modern Landfill[3]			
Upper till	19	7-36	72
Lower lsc	6	2-21	73
Basal sand/basal till	78	10-185	77
MillSeat Landfill[4]			
Drumlin tills	85	6-205	146
High Acres Landfill[5]			
Upper till/outwash	60	--	60
Basal till	124		25

1 Dunn Geoscience (1990)
2 Recra Environmental (1988)
3 Wehran-New York, Inc. (1991)
4 H&A of New York (1987)
5 Eckenfelder (1992)

*lsc = lacustrine silt and clay
xwor = weight of rod

Table 13. Hydraulic conductivity of surficial deposits (cm/sec) of the Lake Ontario Plain

Site/Deposits	Geometric Mean	Max	Min	n
Seneca Meadows[1]				
upper lacustrine	5.6×10^{-5}	7.1×10^{-5}	1.1×10^{-6}	4
upper till	8.1×10^{-7}	2.9×10^{-4}	1.8×10^{-8}	17
lower lacustrine	1.5×10^{-5}	1.9×10^{-4}	1.1×10^{-7}	4
Niagara Landfill-Tonawanda[2]				
upper till	1.6×10^{-8}			1
lower lacustrine	1.6×10^{-8}			1
basal till	7.0×10^{-5}	9.4×10^{-5}	4.5×10^{-5}	2
Modern Landfill[3]				
upper till	5.8×10^{-6}	4.2×10^{-4}	2.6×10^{-7}	13
lower lacustrine	5.9×10^{-8}	5.9×10^{-8}	5.9×10^{-8}	3
basal sand	4.9×10^{-5}	3.4×10^{-3}	1.5×10^{-7}	25
basal till	1.0×10^{-7}	2.4×10^{-7}	1.3×10^{-9}	7
Mill Seat Landfill[4]				
coarse grained levels	2.8×10^{-5}	1.9×10^{-4}	4.2×10^{-6}	5
weathered till	1.4×10^{-6}	2.5×10^{-5}	3.2×10^{-9}	8
unweathered till	3.3×10^{-6}	2.1×10^{-5}	8.1×10^{-8}	11
High Acres[5]				
outwash	1.2×10^{-3}	4.7×10^{-3}	5.1×10^{-5}	7
weathered till	1.7×10^{-5}	7.9×10^{-5}	8.7×10^{-7}	11
Galen Lyons[6]				
alluvium	1.9×10^{-4}			1
upper till	8.6×10^{-6}	2.8×10^{-4}	2.1×10^{-8}	7
lower till	3.8×10^{-6}	1.4×10^{-4}	7.3×10^{-8}	5

[1]Dunn Geoscience (1990)
[2]RECRA Environmental (1988)
[3]Wehran-New York, Inc. (1991)
[4]H&A of New York (1987)
[5]Eckenfelder (1992)
[6]Larson (1990)

Table 14. Grain size trends of surficial deposits of the Lake Ontario Plain.

Site/Deposits	% Gravel	% Sand	% Silt	% Clay	n
Modern[1]					
brown till	0.2	9.2	29.4	61.2	3
grey till	2.2	8.4	35.4	54.0	8
lacustrine lowstand	18.4	26.5	55.1	0.0	1
Seneca Meadows[2]					
upper till	7	45	25	23	1
glaciolacustrine silt & clay	0	0	35	65	1

[1]Wehran-New York, Inc. (1991)
[2]Dunn Geoscience (1990)

foot, respectively. These values suggest that the unit has not been subjected to significant compaction by the overlying Somerset Till. Consequently, either the Somerset Till does not contain a true lodgement till facies beyond northern Niagara County or anomalously high hydrostatic pressures were able to offset the weight of the advancing ice sheet.

For the most part, hydraulic conductivity values reported for the intertill glaciolacustrine deposts reflect the relative percentages of sand and clay in local sections (see Table 13). Values representative of the basal sand and silt are generally in the 10^{-5} cm/s range. Values representative of the overlying silt and clay highstand unit are in the 10^{-8} cm/s range.

Somerset Till

The predominant diamicton of the Lake Ontario Plain in western New York is a gray to purplish-gray unit that overlies the intertill glaciolacustrine sequence, the Furnaceville Till or bedrock (Calkin and Muller, 1992). This unit has been designated the Somerset Till.

Calkin and Muller (1992) recognize two facies in sections along the shore of Lake Ontario: 1) a basal, compact, massive-bedded, homogeneous, silty till; and 2) an overlying, subaqueously deposited succession of diamictic beds and stratified, water-sorted facies. The basal contact of the Somerset Till is apparently sharp in the type area based upon consistent identification in test borings. Conversely, the upper contact appears gradational with the capping glaciolacustrine sequence, and a distinction is not always made in test boring logs.

Although the texture of the Somerset Till may be widely variable, Calkin and Muller (1992) indicate that the unit is generally finer-textured than the older Furnaceville Till. These workers report that the gravel component averages only 4 percent by mass. Grain-size data from the Modern Landfill and Seneca Meadows Landfill are consistent with the reported low gravel content (see Table 14).

Generally, the Somerset Till is less compact than the older Furnaceville Till. One reason may be that at least the upper half of the Somerset Till was deposited subaqueously. Varved interbeds are common at the Modern Landfill site (Wehran, 1991). The lower degree of compaction is reflected in lower average N-values for the upper till than the lower diamicton (see Table 12). Average N-values for the Somerset Till at four landfill sites range between 19 and 60 blows per foot compared to a range between 42 and 124 blows per foot for the Furnaceville Till.

Due to the high percentages of silt and clay in the Somerset Till, the diamicton forms a low permeability, aquitard-grade, hydrostratigraphic unit. Average hydraulic conductivity values from 5 landfill sites range between 1.6×10^{-8} and 8.6×10^{-6} cm/s (see Table 13).

Upper Glaciolacustrine Deposits

A progradational clay to sand sequence overlies the Somerset Till or ice contact facies along the shore of Lake Ontario (Calkin and Muller, 1992). Lacustrine sequences cap the upper till at many landfill sites both north and south of the Niagara Escarpment (see Table 11). North of Route 104, these deposits are attributed to Lake Iroquois. South of the escarpment, the glaciolacustrine sequence may be attributed to one of several Iroquois-precursor basins.

Because of the gradational contact with the water-lain till facies of the underlying Somerset Till, the glaciolacustrine sequence is not always treated discretely on landfill sites. At Seneca Meadows, however, a distinction between the two units can be made (Dunn Geoscience, 1990). N-values for the glaciolacustrine unit range between 0 (weight of drill rods) and 46 with an average N-value of 14 (see Table 12).

At some landfill sites in eastern Monroe and Wayne Counties, the upper glaciolacustrine silt and clay sequence appears to be replaced by coarse-grained, lacustrine shoreface to outwash facies (see Table 11). These coarse-grained materials are widespread in the Irondogenesee Valley (Yager and others, 1984). On the High Acres and Galen-Lyons Landfill sites, coarse-grained deposits occupy interdrumlin areas. Coarse-grained deposits also rim a portion of the drumlin on which the Mill Seat Landfill is constructed. The coarse-grained materials exhibit site-specific, geometric mean hydraulic conductivities ranging between 2.8×10^{-5} and 1.2×10^{-3} cm/s (see Table 13).

Groundwater Flow Trends

Groundwater flow conditions on landfill sites of the Lake Ontario Plain reflect gentle topography and the alternating succession of aquitards and transmissive horizons. Recharge of the shallow groundwater zone occurs atop flat-lying ground moraines and glaciolacustrine plains (Figures 18a, b). On drumlin sites, runoff and shallow interflow proceed radially from topographic highs to groundwater recharge areas between the drumlins (Figure 18c). Flow vectors are oriented vertically downward to potentially transmissive zones in the overburden or fractured bedrock zone. On ground moraines and glaciolacustrine plains, the first transmissive zone may occur within a glaciolacustrine sand and silt lowstand horizon atop the Furnaceville Till (Wehran-New York, 1991). On drumlin sites, the first transmissive zone may occur in interdrumlin outwash at very shallow depths. These deposits are typically removed on landfill sites in order to insure that landfill subgrades consist of low permeability materials.

On most flat-lying sites, the water table occurs within 5 feet of the ground surface. In drumlins, a shallow water table may be difficult to define because of the anisotropy of the upper till horizon. Groundwater may exist as perched lenses in an otherwise unsaturated till matrix (H&A of NY, 1989). A perched groundwater table may also exist in the weathered till profile on the drumlin (Eckenfelder, 1992).

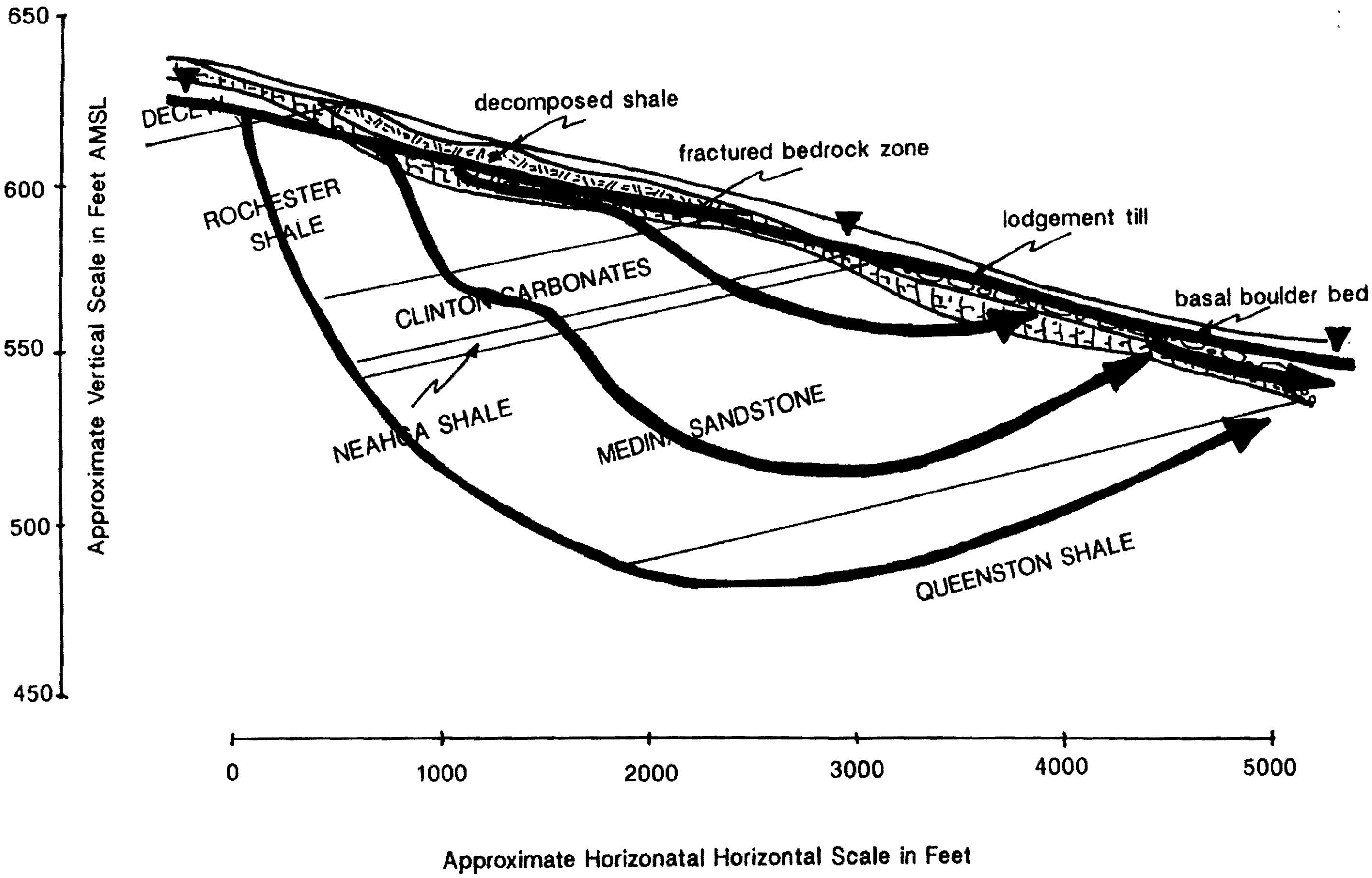


Figure 18A. Generalized groundwater flow patterns beneath the ground moraine hydrogeologic setting, Lake Ontario Plain.

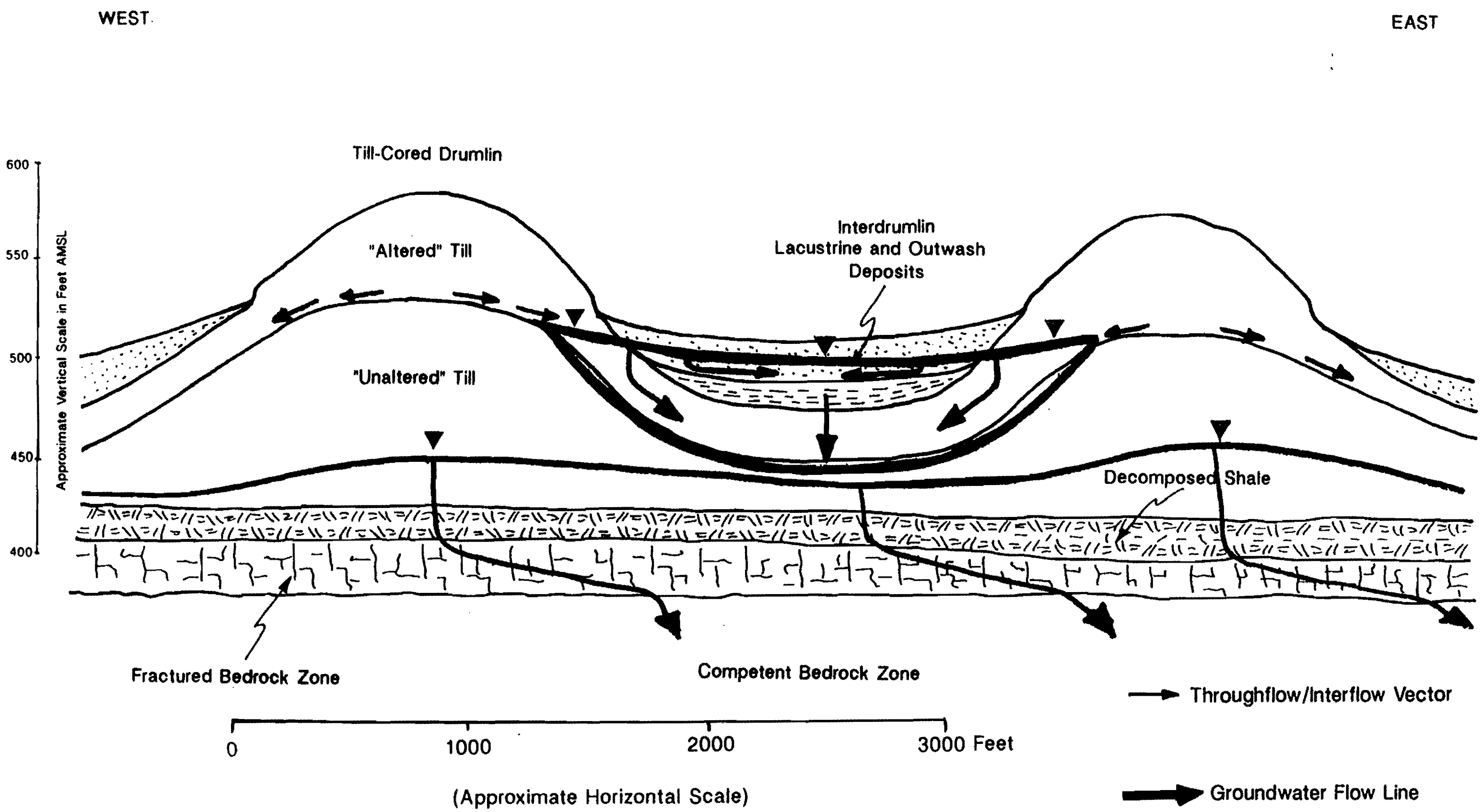
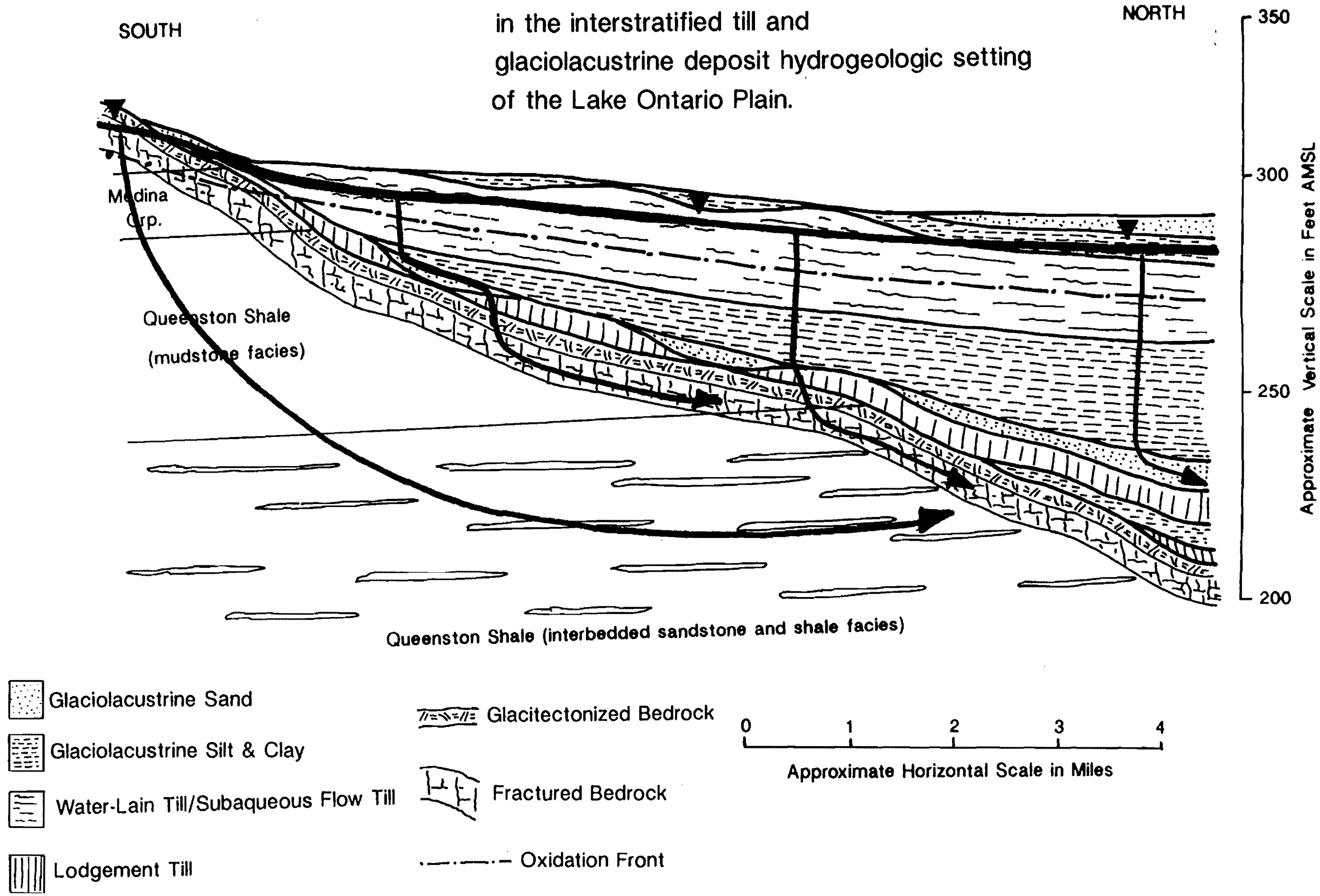


Figure 18B. Generalized groundwater flow patterns beneath the drumlin hydrogeologic setting, Lake Ontario Plain.

Figure 18C. Generalized groundwater flow patterns in the interstratified till and glaciolacustrine deposit hydrogeologic setting of the Lake Ontario Plain.



Regional Background Water Quality

A summary of representative background water quality for landfill sites on the Erie-Ontario Plain is provided in Table 15. These data reflect the position of the lake plain near regional base-level. The plain lies in close proximity to major sublacustrine groundwater discharge zones for regional-scale flow systems that may originate in the Appalachian Plateau. Thus, a remarkable contrast in concentrations of dissolved solids exists between water quality data in Tables 8 and 15.

Water quality is significantly impaired by high concentrations of naturally-ocurring calcium, sodium, sulfate and chloride. The majority of the dissolved ionic species are likely derived from the evaporites of the Silurian Salina Group. Discharging brine springs at the contact between the Silurian Medina Group and Ordovician Queenston Shale have been described as early as Amos Eaton (1824). These Medina brines may not be associated with the Salina evaporites, but may instead be connate formation waters or very old meteoric waters.

CONCLUSION

This study demonstrates that hydrogeologic reports for landfill sites contain valuable data for characterization of regional hydrostratigraphic units. Access to documents was provided by the NYSDEC after filing of a Freedom of Information Law request. Technical reports from 19 landfills were reviewed. The quantitative data available in these reports can be reconciled with and may compliment existing published regional surficial and bedrock stratigraphic syntheses.

Common elements of hydrogeologic settings that are appropriate for landfill siting may be used to construct general models for till-dominated terrains in the Appalachian Plateau and the Erie-Ontario Plain. These models may be used predictively and may be modified following further testing through future comparative studies.

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Table 15
Representative Background Water Quality
Erie-Ontario Plain

Parameter*,(6)	CECOS	Niagara County Landfill	Niagara Landfill Inc. Tonawanda		High Acres Landfill		Orleans Sanitary Landfill					
							Western Area			Eastern Area		
	Bedrock/ Overburden Contact	Shallow Bedrock	Shallow Bedrock	Lodgement Till	Lodgement Till	Shallow Bedrock	Lodgement Till	Shallow Bedrock	Deep Bedrock	Lodgement Till	Shallow Bedrock	Deep Bedrock
pH (pH units)	6.68	6.59	8.62	7.62	7.43	7.72	7.40	7.48	7.70	7.90	7.68	7.62
Specific Conductance (umhos/cm)	1750	1592	2973	3340	1000	2794	520	2450	14600	540	680	805
Total Dissolved Solids	N/A	1119	3438	7625	625	2500	370	1575	9300	260	390	528
Hardness	N/A	1310	N/A	N/A	496	1260	230	895	2300	350	375	383
Alkalinity	N/A	301	N/A	N/A	320	205	153	199	250	290	298	228
Chloride	340	171	372	1385	127	101	25.0	615	5600	11.6	27.3	74.4
Sulfate	2.10	541	1998	586	75.0	1402	126	77.0	135	7.30	54.0	70.6
Nitrate	N/A	0.31	N/A	N/A	1.31	0.70	0.01	1.21	0.10	0.02	0.08	0.58
TOC	62.0	19.9	35.0	284	0.80	1.6	9.5	1.25	18.0	21.0	0.67	0.67
Ammonia	N/A	N/A	N/A	N/A	N/A	N/A	0.08	0.02	1.93	0.23	0.02	0.37
Iron	N/A	10.8	0.09	5.63	7.10	2.50	0.04	0.03	N/A	10.5	0.04	0.12
Manganese	N/A	0.86	0.14	1.34	N/A	N/A	0.005	0.08	N/A	11.5	0.17	0.13
Magnesium	N/A	N/A	N/A	N/A	50.0	57.0	13.1	50.2	123	17.6	27.5	30.8
Calcium	113	N/A	N/A	N/A	113	409	59.0	244	700	110	105	94.4
Sodium	300	N/A	N/A	N/A	550	258	77.0	155	2200	15.2	13.9	17.0
Potassium	N/A	N/A	N/A	N/A	6.90	18.1	3.70	13.3	56.0	1.82	2.57	13.9

* all values in mg/l unless otherwise specified

(6) Values reported are mostly arithmetic means for each geologic interval.

Table 15
Representative Background Water Quality
Erie-Ontario Plain

Parameter*	Galen-Lyons Landfill			Mill Seat Landfill				Modern Landfill	
	Weathered Till	Unaltered Till	Shallow Bedrock	Weathered Till	Unweathered Till	Shallow Bedrock	Deep Bedrock	Lacustrine Silt & Sand	Shallow Bedrock
pH (pH units)	6.21	7.39	8.66	7.43	8.22	6.98	6.91	7.53	7.80
Specific Conductance (umhos/cm)	597	473	460	859	661	1491	1963	2356	3458
Total Dissolved Solids	451	362	142	543	389	1283	1776	3822	5818
Hardness	434	320	187	466	315	992	1325	2043	2075
Alkalinity	364	237	110	375	271	305	319	433	292
Chloride	37.9	11.4	1.98	14.9	13.1	37.5	59.7	190	460
Sulfate	29.3	80.0	327	122	77.0	638	928	2072	3190
Nitrate	1.50	0.11	<0.05	1.35	0.80	0.12	0.03	0.19	0.06
TOC	6.41	1.53	1.83	120	1.13	2.16	4.47	2.39	2.98
Ammonia	1.93	0.05	0.09	0.03	0.11	0.06	0.14	0.28	0.75
Iron	7.05	0.61	2.69	0.09	0.08	0.16	0.57	7.60	1.87
Manganese	0.49	0.03	0.07	0.08	0.02	0.06	0.04	0.27	0.24
Magnesium	N/A	N/A	N/A	60.0	43.3	41.6	43.7	271	235
Calcium	115	66.9	36.1	83.3	46.0	326	456	193	337
Sodium	11.8	10.2	14.5	16.3	37.1	11.2	13.7	222	628
Potassium	4.35	1.39	14.5	4.00	7.87	2.44	5.40	9.95	34.9

* all values in mg/l unless otherwise specified

(6) Values reported are mostly arithmetic means for each geologic interval.

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ROAD LOG
HYDROGEOLOGY OF LANDFILL SITES
IN WESTERN NEW YORK

TOTAL MILES	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Depart from U of R parking lot; turn left (south) onto Wilson Blvd.
0.1	0.1	Intersection with Elmwood Avenue; turn left (east) onto Elmwood Ave.
0.3	0.2	Intersection with Lattimore Road; turn right (south) onto Lattimore.
0.4	0.1	Intersection with Crittenden Blvd; turn left (east) onto Crittenden.
1.2	0.8	Intersection with Routes 15 and 15A; Bear right onto Route 15A.
2.2	1.1	Intersection with Route 390; bear left onto 390 south.
2.4	0.2	Junction with Route 590 north; bear left onto Route 590 north.
18.4	16	Browncroft Blvd. (Rt. 286) off-ramp; turn right (east) onto Browncroft.
18.5	0.1	Turn left into access road to Brighton Town Landfill.

STOP 1. BRIGHTON TOWN LANDFILL

Facility: Brighton Solid Waste Management Facility

Location: Town of Brighton, Monroe County (Rochester East 7.5" Quadrangle)

Hydrogeologic Setting: Glaciolacustrine deposits and glacial till over bedded sedimentary rock.

Site Description: The site is located on the northeast-facing side of the Irondequoit Creek valley. The site includes 75 acres; only 24 acres are considered suitable for waste disposal purposes. Topographic relief across the active portion of the site is approximately 150 feet; the average topographic slope is approximately 5 percent. The site is bounded on the west by Thomas Creek, a small tributary of Irondequoit Creek, and on the north and east by Irondequoit Creek. Irondequoit Creek drains into a major bay (Irondequoit Bay) of Lake Ontario.

The following items, listed in the Part 360 Permit Application for this facility, reflect current site conditions:

- There is no evidence of leachate derived from this facility.
- Natural vegetation is present on previous disposal areas.
- Vegetation is established on all slopes of disposal areas.
- There is no evidence of significant erosion occurring on site.
- The disposal area is adequately covered and graded to maintain safe and efficient storm water runoff.

Surficial Geology: Overburden consists of probable lodgement till overlain by interbedded flow tills and stratified silt- to sand-rich glaciolacustrine facies. The overburden ranges in thickness from 22 to approximately 100 feet. The overburden is thinnest at the topographically highest (southwestern) portion of the site and thickens towards Irondequoit Creek, to the northeast. The strike of the surficial deposits generally follows the topographic contours of the valley.

The interbedded, stratified glaciolacustrine facies and water-lain till deposits include evenly laminated silt (offshore lacustrine deposits), evenly- and cross-laminated coarse silt and very fine sand (nearshore lacustrine deposits), and gravelly, clay-rich silts (tills). Based on detailed boring logs, split-spoon samples (Huntingdon-Empire Soils Investigations, Inc., 1994), and outcrop sections, the glacial and proglacial strata include, in descending order:

Unit 1:	Evenly laminated silt (10 to 25 feet thick)
Unit 2:	Very fine, silty sand (5 to 15 feet thick)
Unit 3:	Gravelly, clay-rich silt (5 feet thick)
Unit 4:	Evenly laminated silt (> 5 feet)
Unit 5:	Undifferentiated deposits (No boreholes with detailed logs or available split spoon samples penetrated Unit #4. However, tills are present over bedrock throughout the lower Irondequoit Creek valley.)

This succession mostly records accumulation in a proglacial lake basin. The glaciolacustrine succession disconformably overlies lodgement till that accumulated during the earlier glacial advance. A minor glacial advance is recorded by Unit 3, a waterlain diamicton or subaqueous flow till. The stratigraphic record of this advance is asymmetrical: offshore silts are overlain by flow till, and the flow till, in turn, is overlain by nearshore sands. Unit 3 may correlate with the Somerset Till described by Calkin and Muller (1992) from sections along the Lake Ontario shoreline.

In general, lacustrine silts and sands are extremely well sorted with D_{10} values on the order of 0.05 mm or less for the very fine sands. N-values for these deposits generally range from 15 to refusal, with the lower N values correlating to shallowest subsurface sandy deposits.

Bedrock Geology: Based on reported depth to bedrock, strata assigned to the Lower Silurian Medina and Clinton groups probably subcrop beneath glacial sediments. These strata consist of sandstone, shale, and limestone. In ascending order, the following formations are likely subcrop beneath the site: Grimsby Sandstone, Cambria Shale, Kodak Sandstone, Maplewood Shale, Reynales Limestone, and Sodus Shale.

Groundwater Flow: Groundwater flow is topographically controlled. Groundwater equipotentials are generally parallel to topographic contours; groundwater flows to the northeast.

18.7	0.2	Return west on Browncroft Blvd to Rt. 590. Turn left onto Rt. 590 south.
20.2	1.5	Junction with Route 490; take Route 490 east.
23.4	3.1	Junction with Route 31F; turn right (east) onto Route 31F.
26.2	2.8	Route 31F to the Village of Fairport;
27.4	1.2	Route 31F crosses canal on east side of village.
29.9	2.5	Intersection of Perinton Parkway and Route 31F; turn right (south) onto Perinton Parkway.
30.8	0.9	Perinton Parkway to Resident's Entrance to High Acres Landfill.

STOP 2. HIGH ACRES LANDFILL

Site Location: Town of Perinton, Monroe County, New York
Fairport and Macedon 7.5 Minute Quadrangles

Site Description: The east side of the site contains a north-south trending drumlin. The existing operational facility is constructed on the drumlin. The west side of the site is relatively flat-lying, interdrumlin terrain. The western area is presently under construction as a lateral expansion area to the landfill.

Surficial Geology: The drumlin on the west side consists of two separate till horizons: a lower, gray, bouldery lodgment till and an upper, brown weathering "altered till". The upper till contains fewer cobble and boulder size clasts than the basal, unaltered till (Dan Coon, personal communication). Given the difference in grain-size composition and density, the upper till may actually represent a separate genetic unit. One hypothesis is that the basal till correlates with the Furnaceville Till, whereas the upper till is correlative to the Somerset Till.

In the interdrumlin areas of the west side of the site, the surficial stratigraphy is more variable. The basal till that directly overlies bedrock is also divisible into a lower, unaltered zone and an upper weathered zone. The presence of this horizon may suggest a period of subaerial exposure prior to deposition of the overlying recessional sequence. Such an interpretation, however, is difficult to reconcile with the regional glacial history. An alternative hypothesis is that the "weathered till" may represent the diamictic facies of the younger Somerset Till equivalent and that the mixed water-lain and till deposits ("glacial till/outwash" deposits of Eckenfelder, 1992) may correspond to the upper water-lain till facies of the Somerset Till as defined by Calkin and Muller (1992).

Bedrock Stratigraphy: The bedrock strata that subcrop beneath glacial tills on the site are assigned to the upper part of the Silurian Vernon Shale. Eckenfelder (1992) reports that the drill core sections consist primarily of red and green shale with varying amounts of gypsum. Gamma ray logs were used to correlate borehole sections. Through correlation of gamma ray logs, Eckenfelder (1992) was able to document the regional southward dip of bedrock strata.

Groundwater Flow Conditions: Shallow groundwater flow tills and glacial outwash is topographically controlled. On the east side of the site, groundwater in the brown till horizon flows radially off the drumlin. In flat-lying areas, groundwater flow generally proceeds to the south. The low permeability of the glacial tills results in vertically oriented groundwater flow vectors. Flow lines are deflected to a more horizontal orientation in lenses of glacial outwash.

The bedrock fracture zone provides a preferred groundwater flow path. Upward gradients are common between the fractured bedrock zone and the underlying competent bedrock zone.

30.8	0.0	Exit High Acres left (west) onto Perinton Parkway.
31.7	0.9	Intersection Perinton Parkway and Route 31F; turn left (west) onto Route 31F.
37.2	6.4	Route 31F west back through Fairport until junction with Route 490; turn right onto Route 490 west.
40.3	3.1	Route 490 west to Route 590; bear right onto Route 590 south.
56.3	16.0	Route 590 south to junction with Route 390; bear righ onto Route 390 south.

85.8	32.5	Route 390 south to Exit 7 (Mount Morris, Route 408); turn left (south) on Route 408.
89.6	3.8	Village of Mount Morris; jog in Route 408.
98.4	8.8	Route 408 south to Village of Nunda; stay south on Route 408.
101.5	3.1	End Route 408 south; remain southbound on State Street. Sharp left jog in road on south side of Dalton.
104.0	2.5	State Street ends; stay southbound on Allegany County Road #16.
107.1	3.1	Sharp right jog in county road #16 at intersection with county road #24; bear right on county road #16 south.
123.1	16.0	County Road #16 south to the Village of Angelica. Intersection with Peacock Hill Road; turn left (south) onto Peacock Hill Road.
124.6	1.5	Intersection with Herdman Road (entrance to Hylands Ash Monofill); turn right onto Herdman Road.

STOP 3. PROPOSED HYLANDS ASH MONOFILL

Facility: Hylands Ash Monofill

Location: Town of Angelica, Allegany County (Angelica 7.5" Quadrangle)

Hydrogeologic Setting: Glacial Till Over Bedded Sedimentary Rock.

Site Description: The site contains a broad, southfacing, hanging valley that forms a natural amphitheater. A small intermittent stream is supplied by a bedrock spring along the east side of the site and by overland flow toward the valley center in the southern portion of the hanging valley. Surface water drainage flows into Angelica Creek which, in turn, discharges to the Genesee River.

Surficial Geology: Glacial overburden consists of lodgement till overlain by ablation till. The overburden ranges in thickness between approximately 5 and 115 feet. The till is thinnest on hill crests and thickens rapidly toward the center of the hanging valley. An oxidation front marked by the vertical change from gray, unaltered till to brown, weathered deposits cross-cuts and partially obscures the lodgement till/ablation till contact. The lodgement till is overconsolidated; assuming no net erosional lost of glacial sediments near the valley floor, the brown, oxidized lodgement till is estimated to have been compacted beneath a minimum of 500 ft. of ice and probably considerably more (Michael Mann, personal communication). The ablation till is generally less dense than the lodgement till and contains thin, discontinuous, water-sorted lenses.

High subglacial pore pressures are suggested by the fracturing of bedrock and injection of till seams to depths approaching 200 ft. below ground surface. The upper 20-30 ft. of rock is typically so intensely fractured that the contact between glacial till and bedrock may be completely gradational. In ascending order, the gradational interval consists of 1) in-situ rock mass with till seams; 2) approximately even proportions of in-situ rock mass and till with gravel-sized angular rock fragments; 3) boulder-size, rotated or detached blocks of bedrock in a gravelly clay till matrix (deformation till); and 4) channery lodgement till.

Bedrock Geology: Strata are assigned to the Upper Devonian (Fammenian) Canadaway and Conneault Groups and consist of a series of interbedded, fossiliferous sandstones, siltstones and clay shales (see Fig. 6A). In ascending order the following formations subcrop on site: Machias Shale, Cuba Sandstone and Wellsville Shale. The site is situated on the northwest limb of a northeast-trending anticline. Strata dip to northwest at approximately 1 degree.

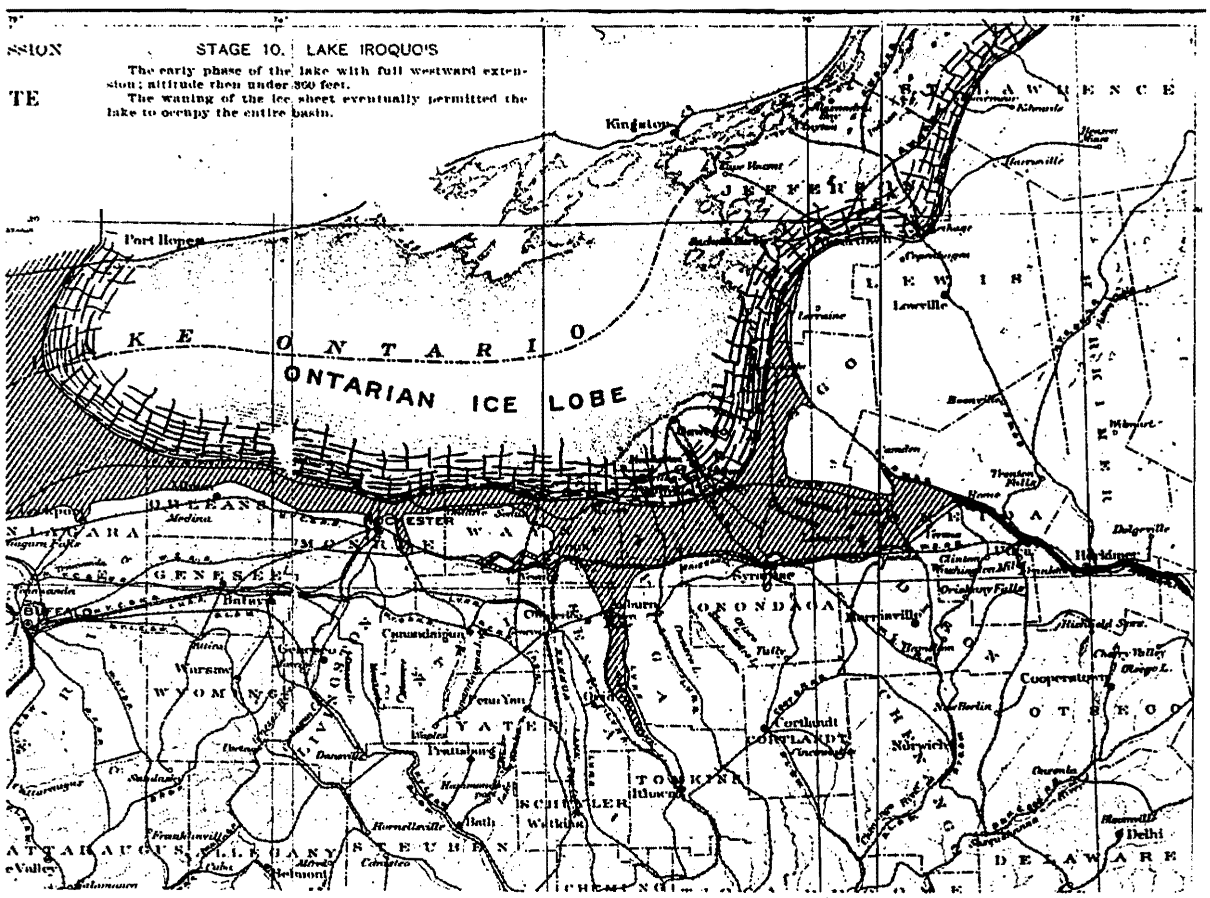
Groundwater Flow: A perched water table exists in the ablation till profile in the valley center. The perched groundwater table is recharged by interflow and overland flow directed toward the valley center from the east, west and north. Along the axis of the valley, groundwater flows southward toward the intermittent stream.

Groundwater flow in bedrock is divided into two flow systems. A high elevation flow system contains a recharge area in open pastures upgradient (east) of the site. The discharge zone occurs at the spring line that marks the base of the Cuba Sandstone subcrop belt along the east side of the site. A second, deeper bedrock flow system occurs in the Machias Shale beneath the landfill footprint. Strong downward gradients are characteristic of the upper portion deeper bedrock system. The horizontal flow vector is directed primarily to the west/southwest. Westward flow on the east side of the till-filled valley is impeded by the gray till aquiclude. Consequently, artesian pressure builds in the deeper bedrock system in the southeast corner of the site. Hydraulic gradients in bedrock beneath the west side of the site are steeply downward. Flow diverges on either side of the topographic divide in the shallow rock profile beneath the western site boundary. Deeper flow in rock trends toward the west/southwest.

124.6	0.0	Exit Herdman Road; Turn left (north) onto Peacock Hill Road.
126.1	1.5	Peacock Hill Road to Junction with Route 16. Turn right (northbound) on Route 16.
145.2	19.1	Route 16 to transition to State Street in Dalton.

147.7	2.5	State Street north to transition to Route 408 north.
163.4	15.7	Route 408 north to Mount Morris; Juntion with Route 390; Take Route 390 north.
199.4	36.0	Route 390 north to Rochester, Scottsville Road Exit; Turn left (north) onto Scottsville Road.
200.4	1.0	Scottsville Road north to Junction with Elmwood Avenue. Take Elmwood Avenue east across Genesee River.
200.9	0.5	Turn left at light onto Wilson Boulevard; Hutchinson Hall is on the right.

END OF TRIP



Map of Lake Iroquois

[From Fairchild, 1909, Glacial waters in central New York.
N.Y.S. Mus. Bulletin 127, Plate 42]

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

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INTRODUCTION

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

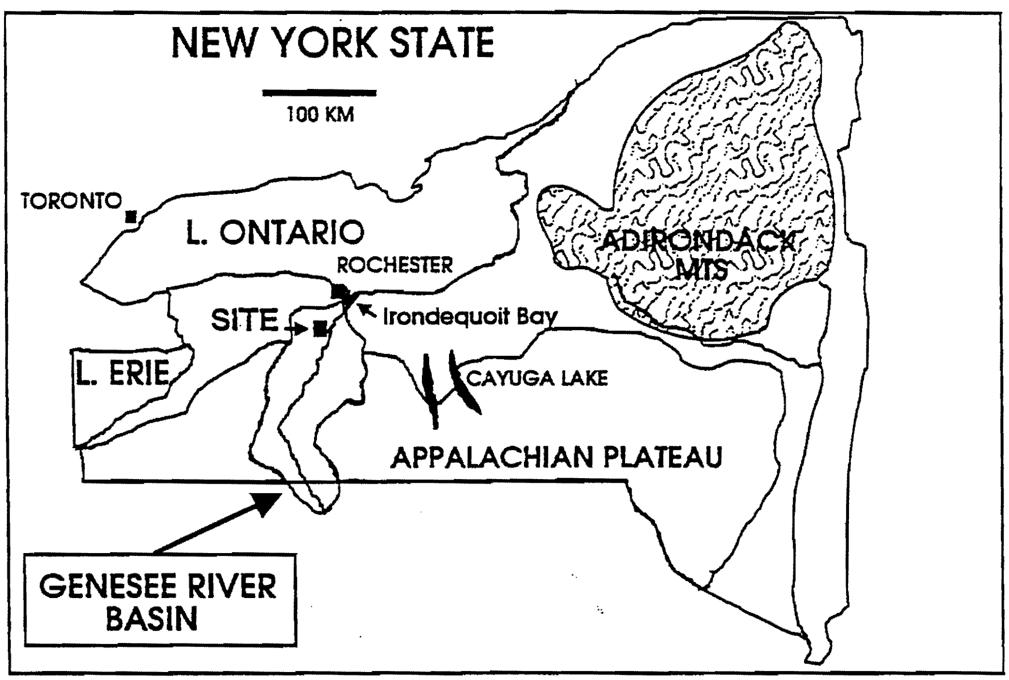

Figure 1: Location of Irondequoit Bay and Middle Wisconsin gravel pit site in Genesee Valley.

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

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

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

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

IRONDEQUOIT BAY SAND BAR SECTION

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

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

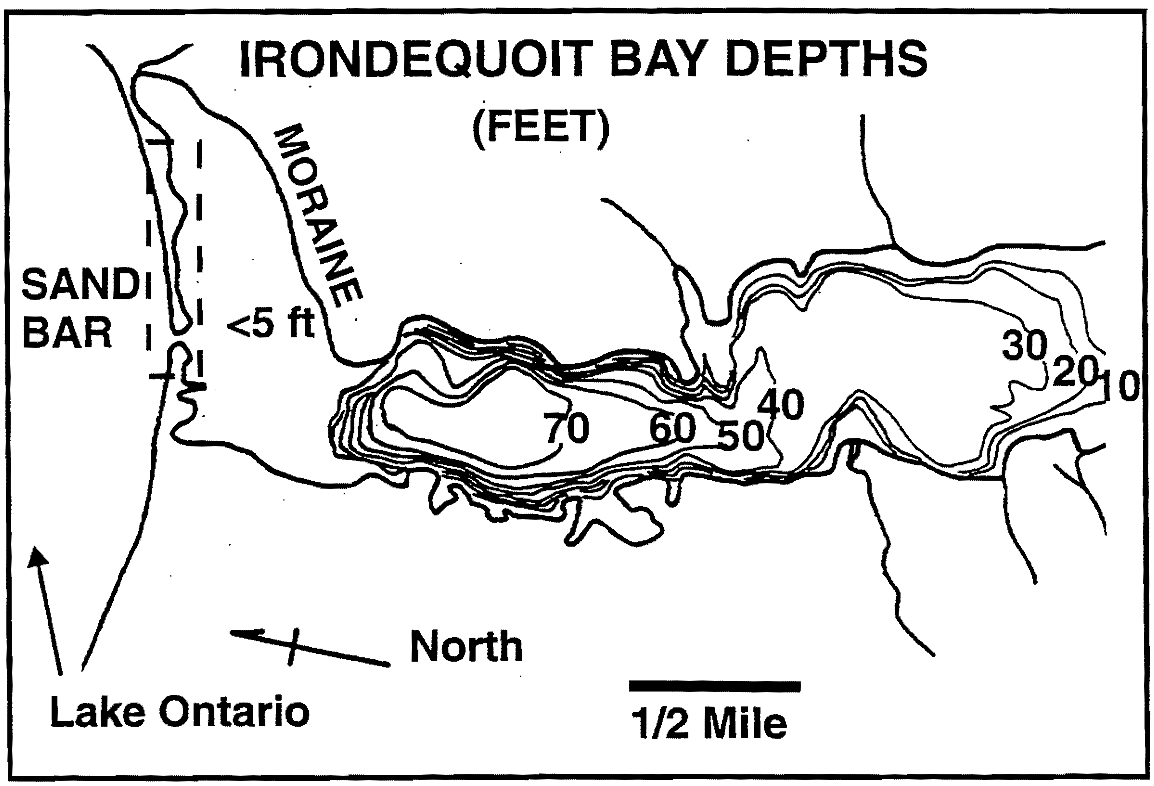

Figure 2: Northern end of Irondequoit Bay with location of sand bar wells (dashed box) and contours showing sediment filling mouth of Bay. Box is also location of Figure 3.

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

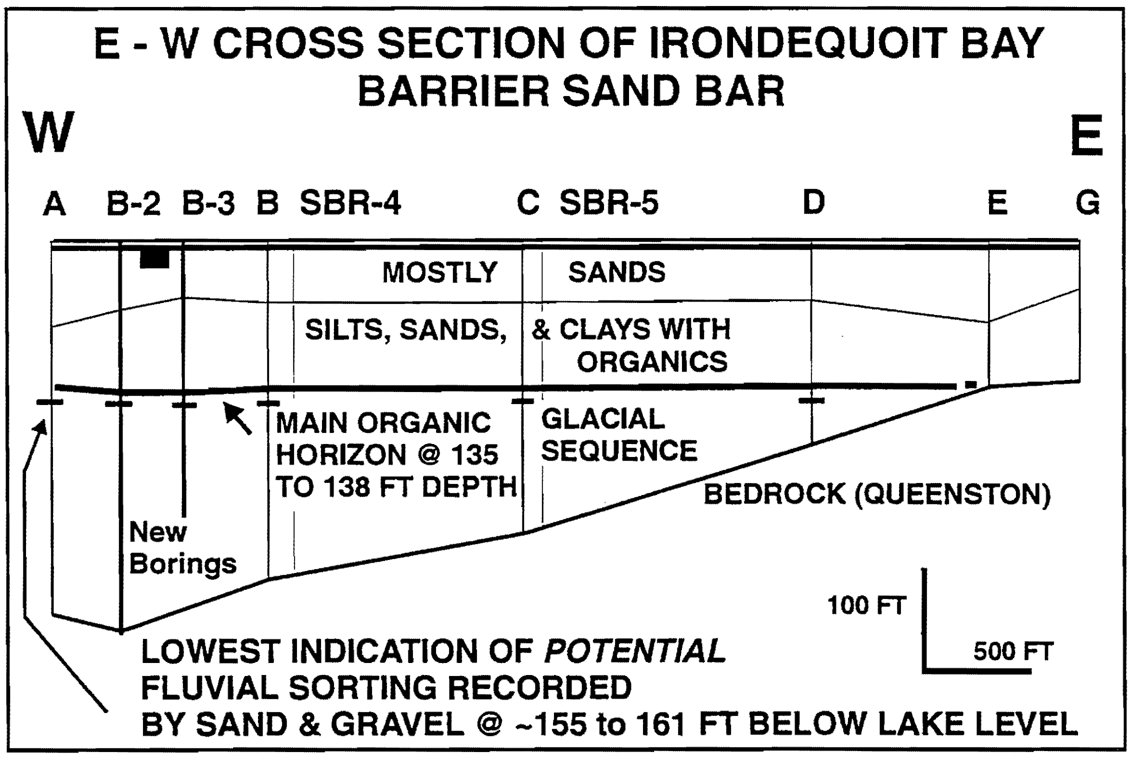

Figure 3: Generalized section through Bar on Figure 2 showing upper, postglacial lacustrine sequence and lower glacial sequence separated by marl and peat horizon near 135 ft depth.

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

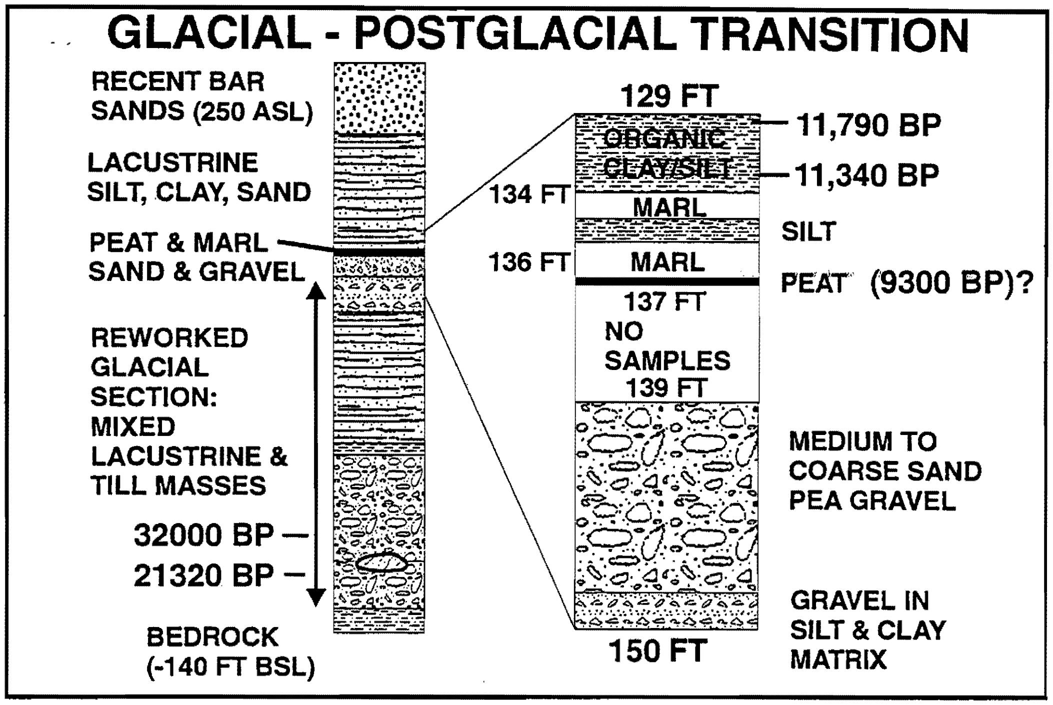

Figure 4: Composite, expanded view of stratigraphy near glacial-postglacial transition in borings B-2 & B-3 (Figure 3) and horizons from which dates were obtained. 9300±BP age is assumed to be in error due to close agreement between two older ages taken from 130 to 134 foot interval. Compare left column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

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

INTERPRETATION

Lake Levels

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

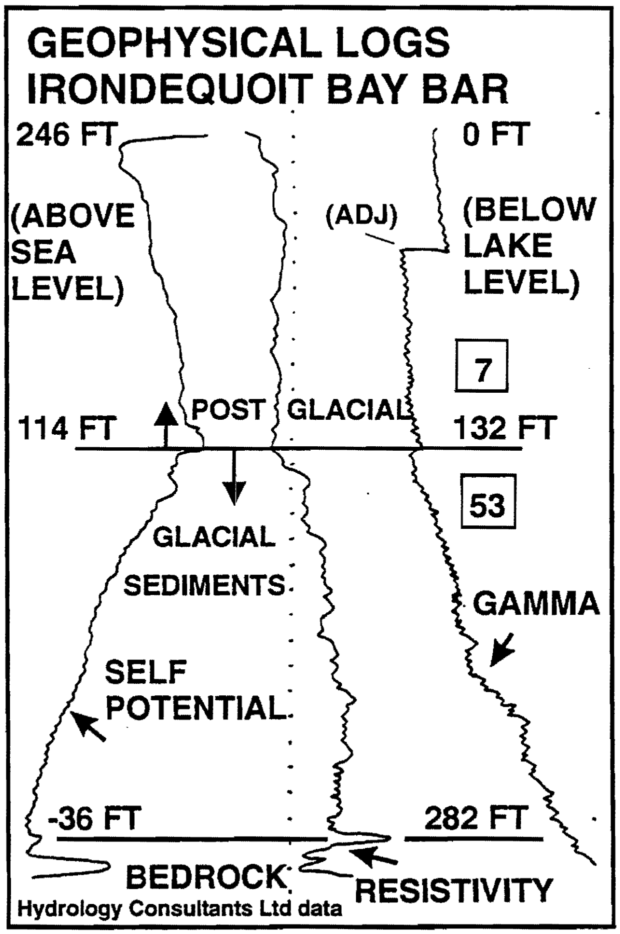

Figure 5: Geophysical logs from boring C (Figure 3) showing contrast between postglacaial lacustrine sequence (top) and denser sediments below. Numbers in boxes indicate average split-spoon blow counts (140 lb hammer) for 6-inch sampler penetration. Change in sediment properties with depth on geophysical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

Pollen Record and Radiocarbon Dates

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

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

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

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

Relation of Lake Stage to Sand Bar

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

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

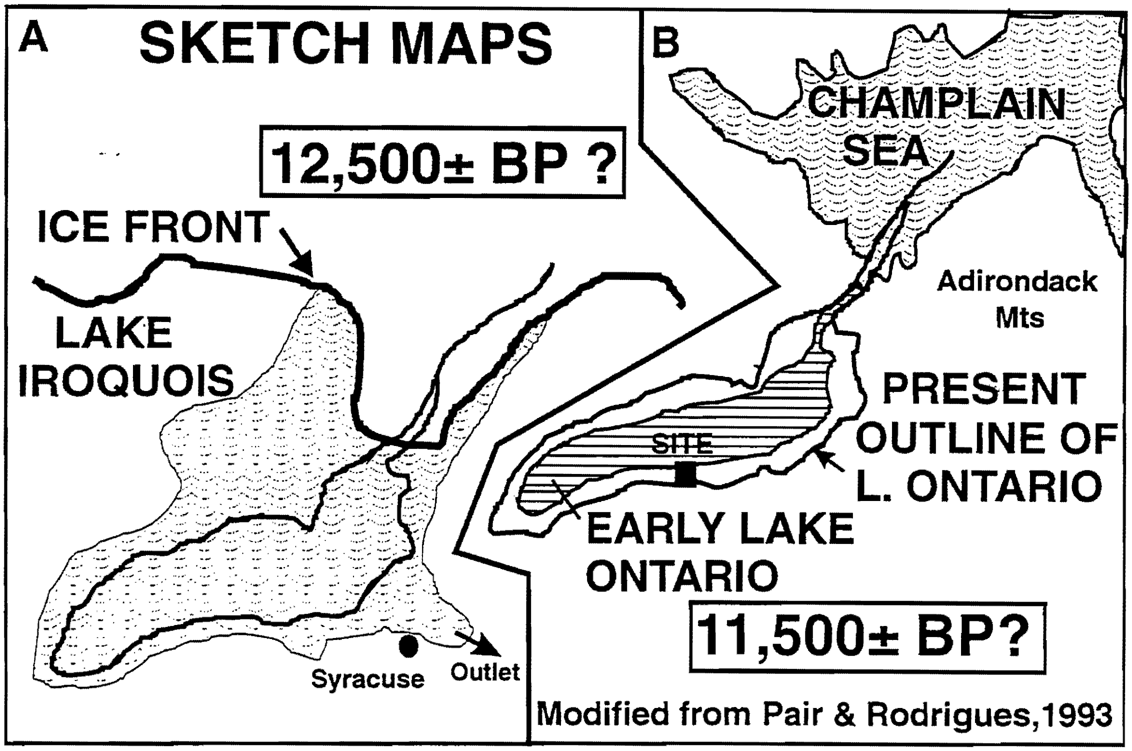

Figure 6: Recession of the ice sheet from the St. Lawrence Valley led to the low Early Lake Ontario stage coincident with the near incursion of the Champlain Sea into the basin while sea level was low and postglacial rebound less complete than at present.

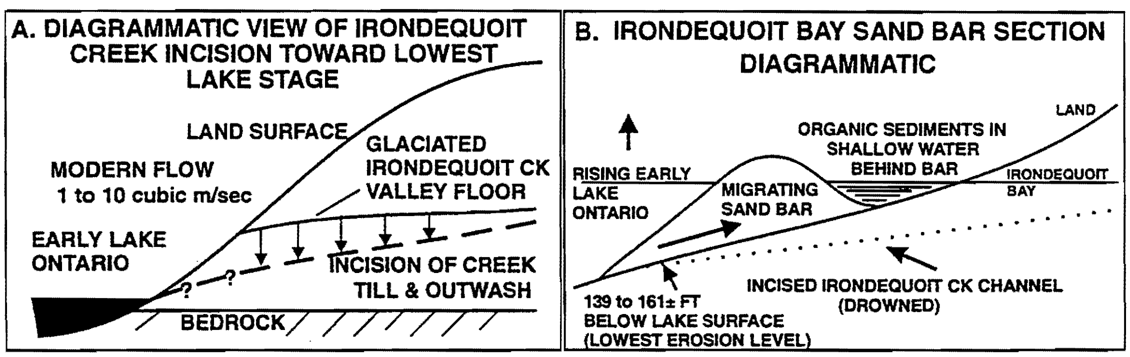

Figure 7A: Diagrammatic cross section of incision of Irondequoit Creek through glacial section following establishment of Early Lake Ontario. 7B: Progradation of sand bar southward with rising lake stages. Organic sediments become buried by bar migration.

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

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

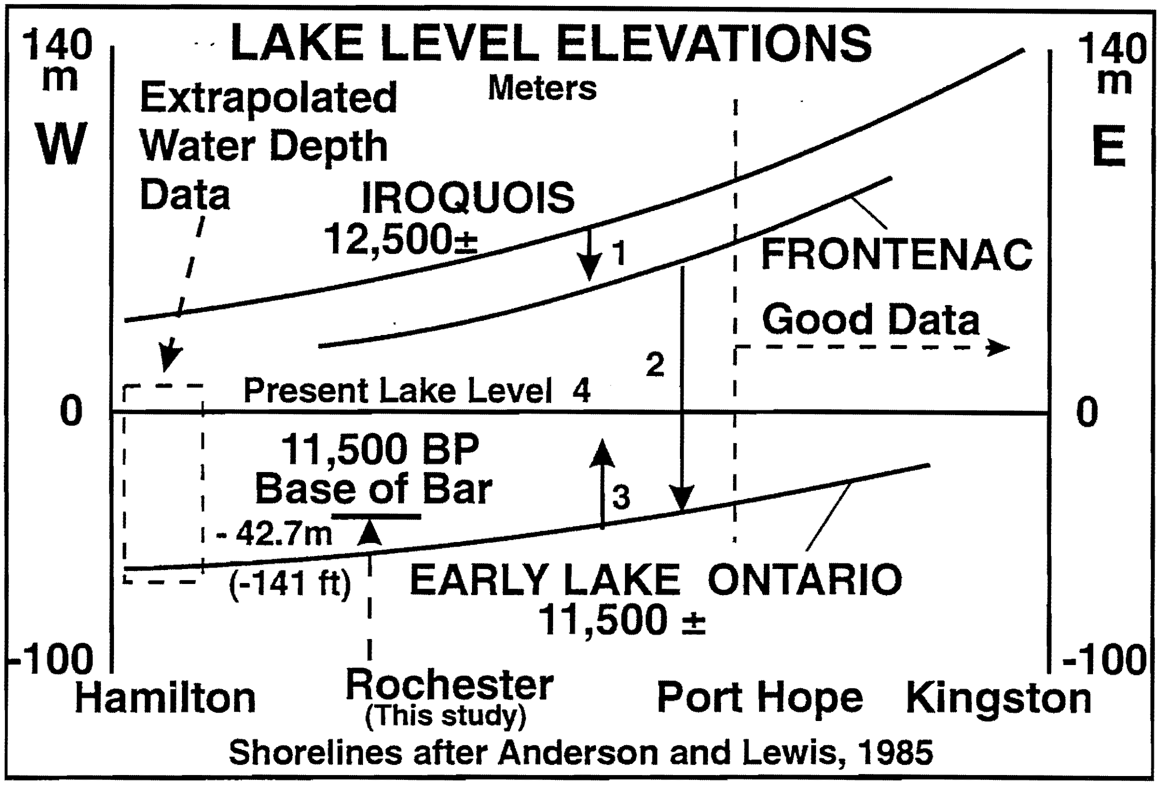

Figure 8: Sequence of falling and rising lake levels (numerical order) illustrating discrepency between earlier work and Rochester data. Dashed box is area of uncertain water depth data.

DISCUSSION

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

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

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

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

MIDDLE WISCONSIN SITE IN NORTHERN LIVINGSTON COUNTY

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

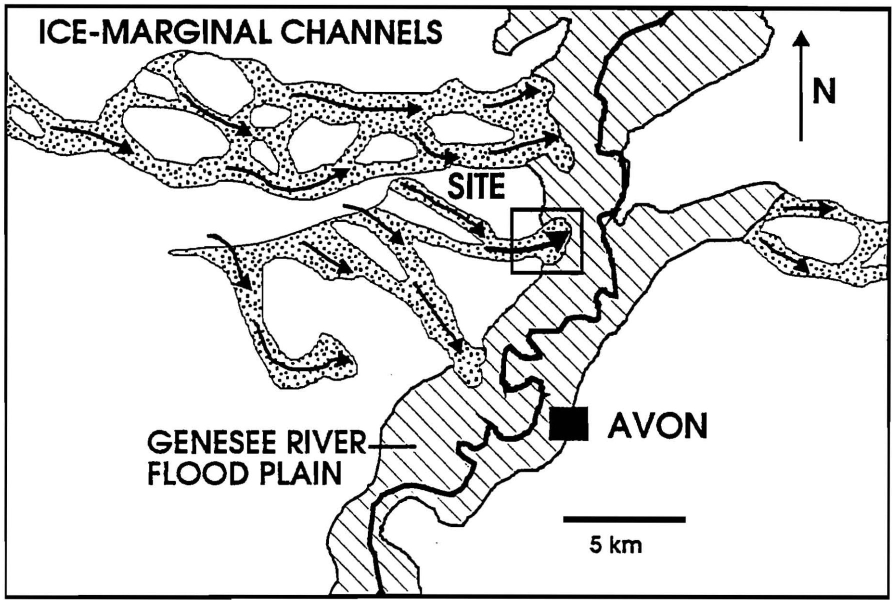

Figure 9. Location of Middle Wisconsin site in northern Livingston County.

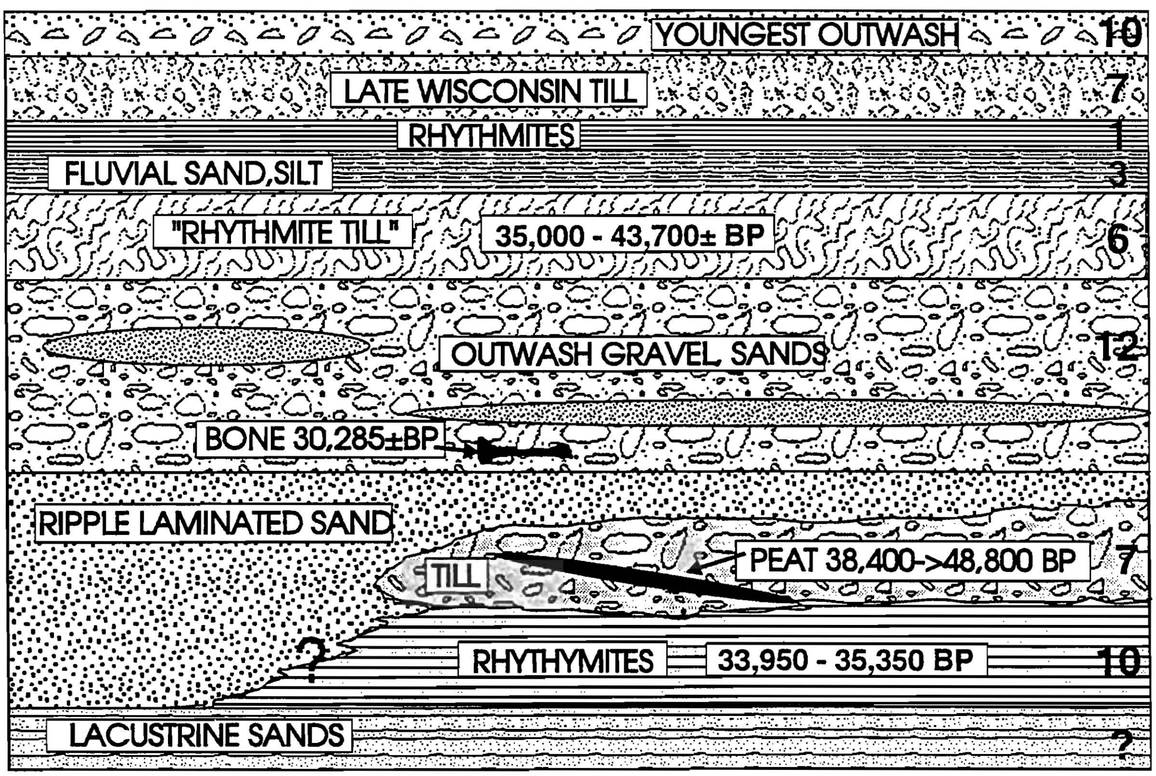

Figure 10. Diagrammatic section of Middle Wisconsin site of Figure 9. Average thicknesses of units at right side in feet. There is no conspicuous evidence of weathering profiles between units except irregular oxidation front (brown) extending from surface(?) down into gray "rhythmite till."

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

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

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

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

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

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

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

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

POLLEN RESULTS

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

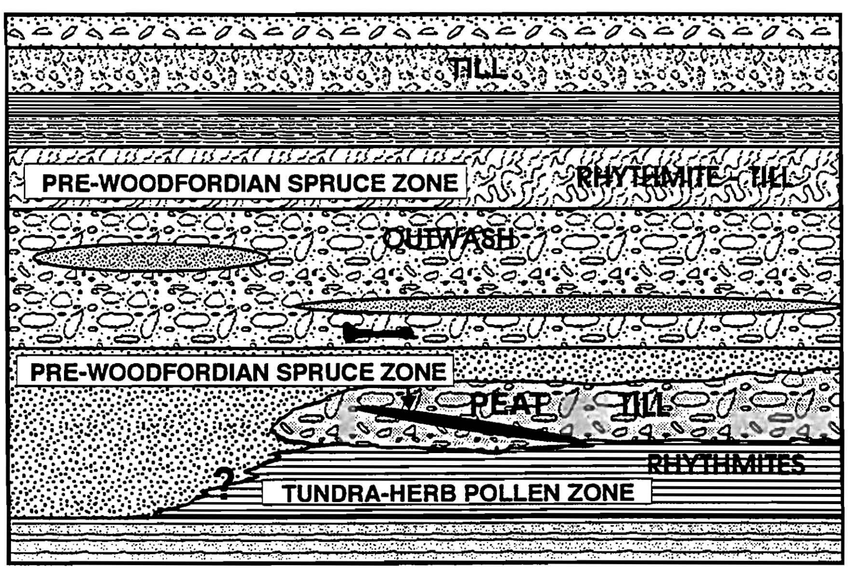

Figure 11. General pollen associations determined by Les Sirkin (See Appendix).

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

AGE DETERMINATIONS

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

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

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

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

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

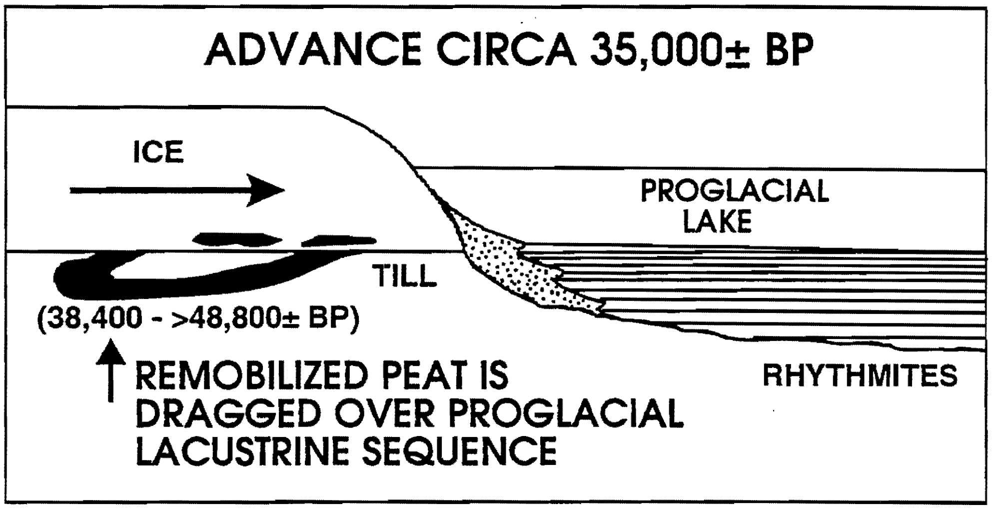

Figure 12. Inversion of radiocarbon age relations by remobilization of peat and redeposition onto younger proglacial lake sequence during advance.

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

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

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

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

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

MECHANISMS AND IMPLICATIONS FOR ICE SHEET TRANSPORT AND EROSION

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

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

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

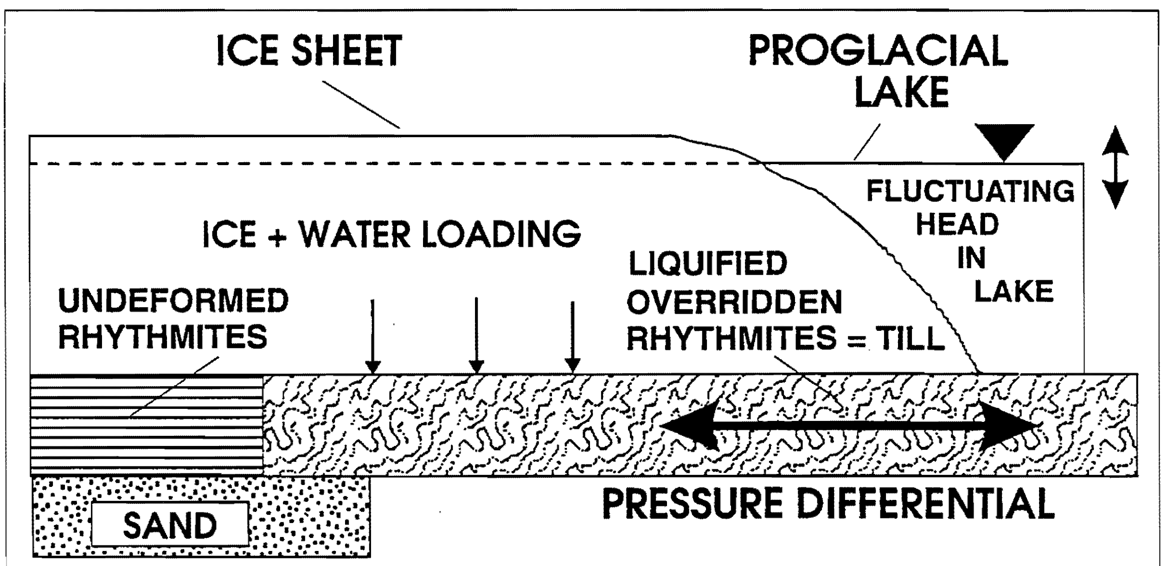

Figure 13: Diagram of hypothetical ice conditions leading to spontaneous liquefaction of rhythmite sediments by ice advance into proglacial lake.

RHYTHMITE LIQUEFACTION TO PSEUDO "TILL" TEXTURE

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

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

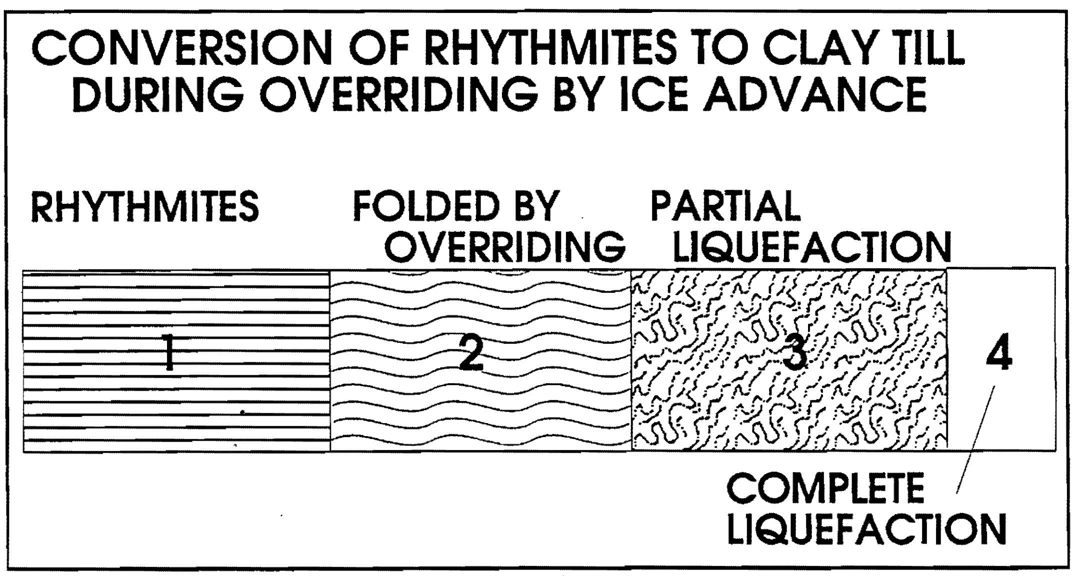

Figure 14. Diagrammatic lateral transition of ryhthmite sequence (1) to massive clay "till" (4) seen in pit exposures

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

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

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

DISCUSSION AND SUMMARY

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

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

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

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

ACKNOWLEDGEMENTS

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

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

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

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

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

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

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APPENDIX
Pollen Histograms

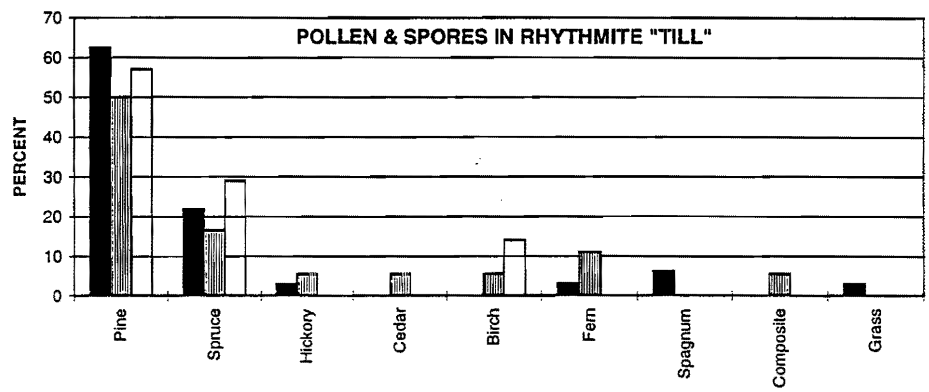


Diagram 1. Pollen and spores from three random samples near top of overridden rhythmite section (Figure 10) or "rhythmite till". Histogram bars are precentages of total sample, based on number of grains counted. Samples: Solid bars (32 grains); Open bars (7 grains); Hatchured bars (18 grains).

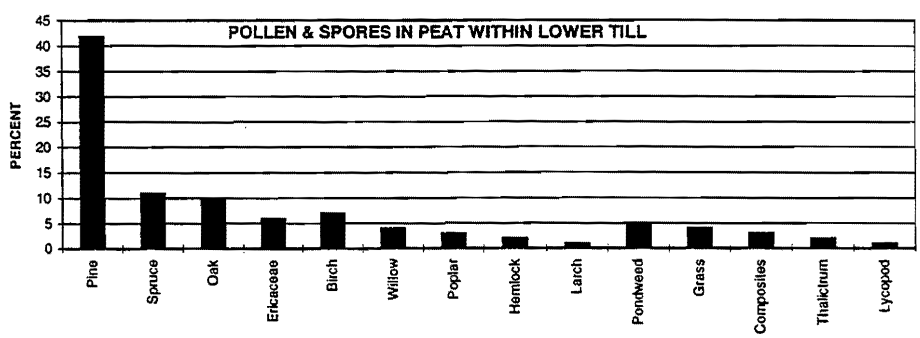


Diagram 2. Pollen and spores from single sample of overridden peat incorporated in lower till (Figure 10). Histogram bars are percentages of total sample, based on numbers of grains counted. Number of grains in sample = 102. Abundant Spagnum spores not included.

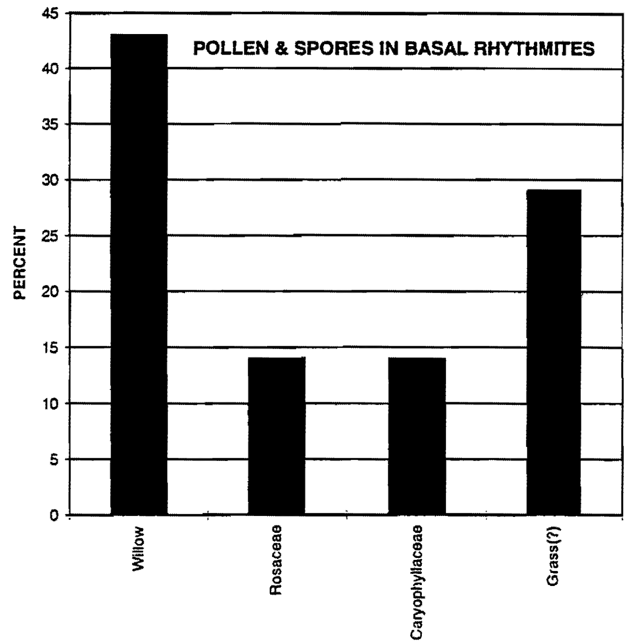


Diagram 3. Pollen and spores from single sample of lower rhythmites below peat-bearing till (Figure 10). Recovery was only seven grains (converted to percent for histogram) plus a few reworked, oxidixed pre-Cenozoic spores (not included in graph).

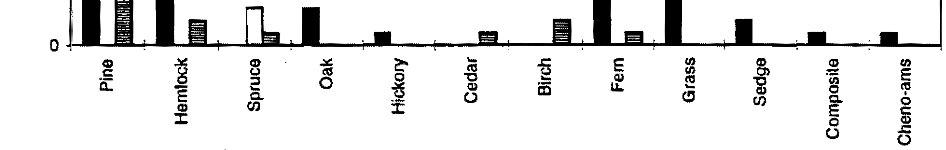


Diagram 4. Pollen and spores from three samples in borings for I-390 in Genesee Valley. Histogram bars are pecentages of total sample, based on number of grains counted. Samples: Solid bars = Boring at Pioneer Rd.and railroad crossing; depth 32 ft., contact of peat over lacustrine clay (44 grains). Hatchured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

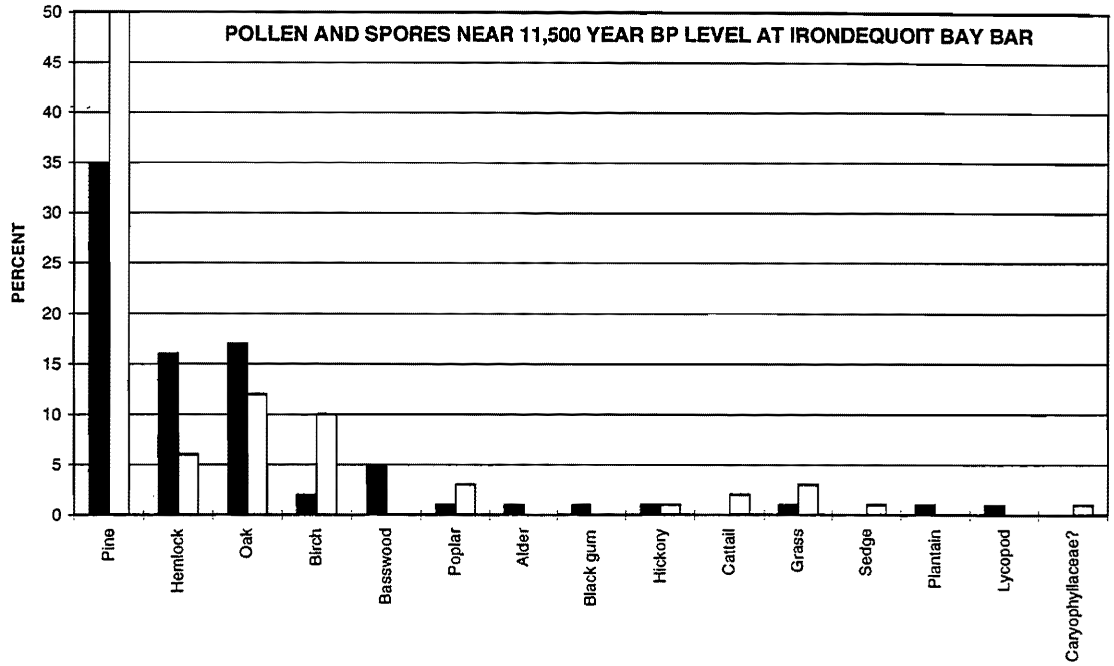


Diagram 5. Pollen and spores from 2 samples near base of postglacial lacustrine sequence at Irondequoit Bay bar. Histogram bars are percent of total sample, based on numbers of grains counted. Samples: Solid bars = depth of 133.5 feet below lake level, boring B-2 (Figures 3, 4). Plot based on 82 grains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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

STOP 4. INFORMAL STOP ALONG ROUTE 39 OVERLOOKING VALLEY

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

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

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

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

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

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

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

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

END OF FORMAL TRIP

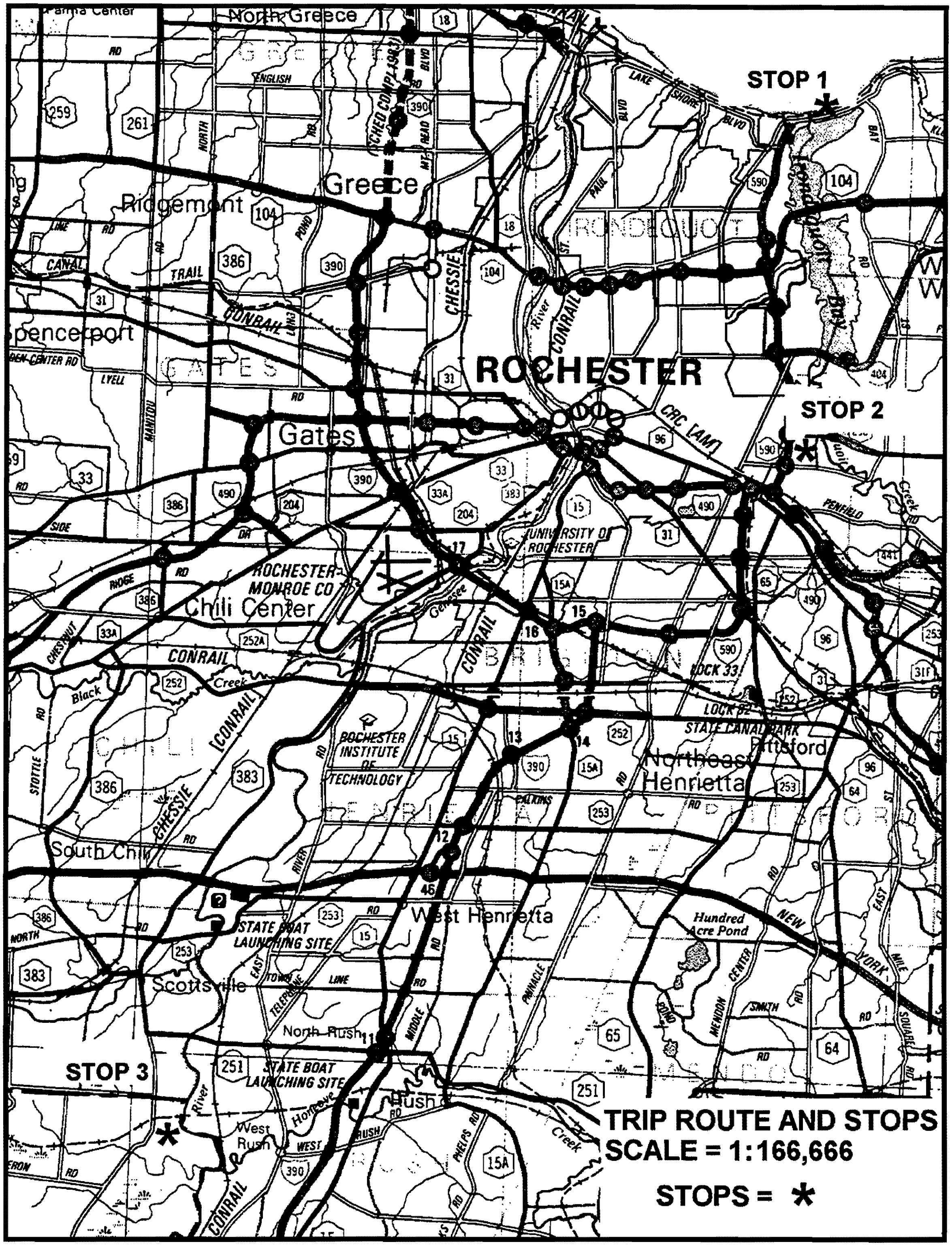
At this point in the trip, depending on the hour, the time spent at the Middle Wisconsin site (STOP 3), and the ongoing developments at the Akzo-Nobel salt mine collapse at Cuylerville (March 1994 to present), the trip will either focus on an informal tour of the salt mine collapse area (with maps and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

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

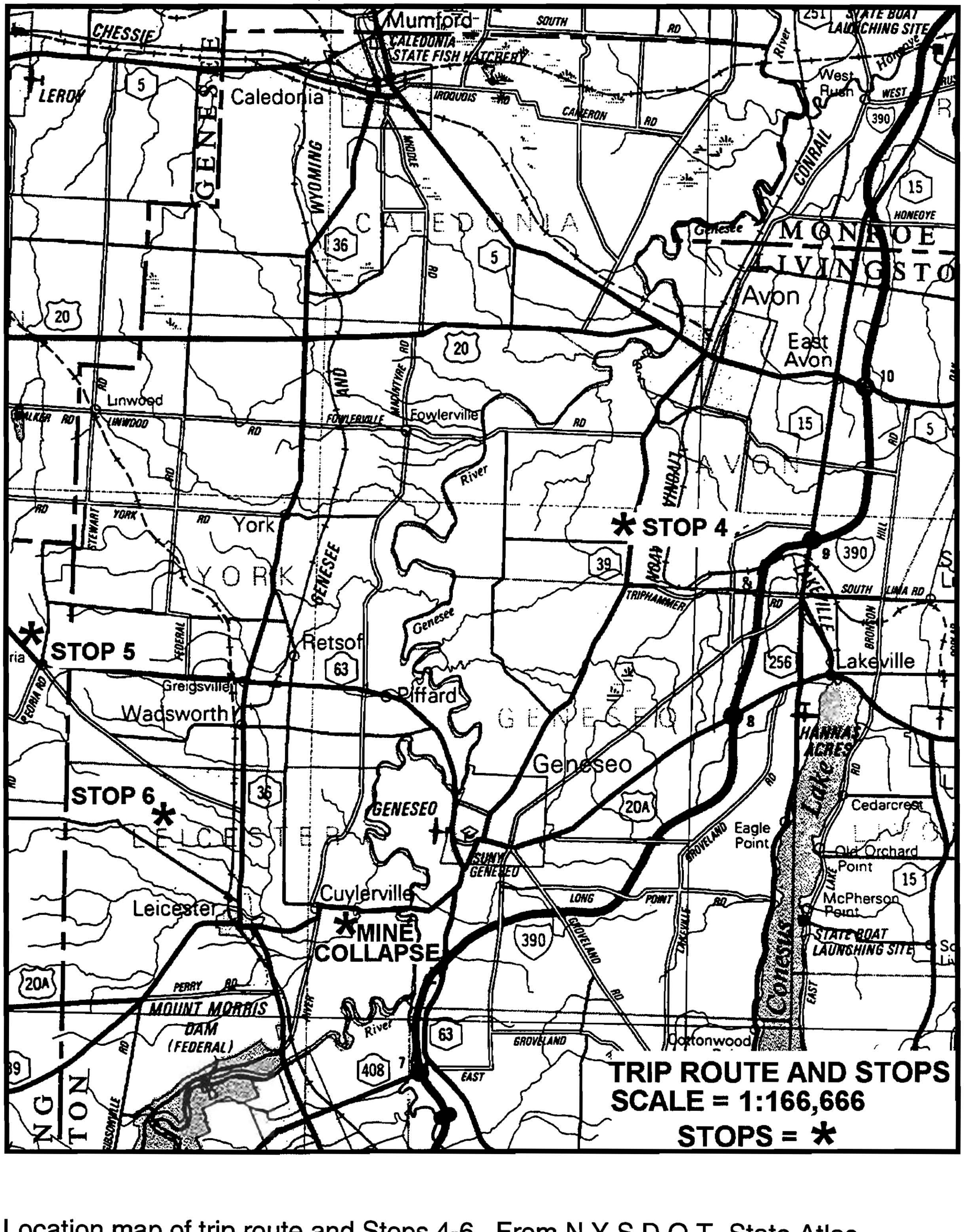
The Valley Heads Moraine complex at the south end of the Canaseraga Valley is a unique, multi-lobate feature, which extended south and west from Dansville in a series of semicircular arcs (not unlike some piedmont lobes in plan view). The lobate extensions of the moraine actually turned <u>northward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

REFERENCES CITED

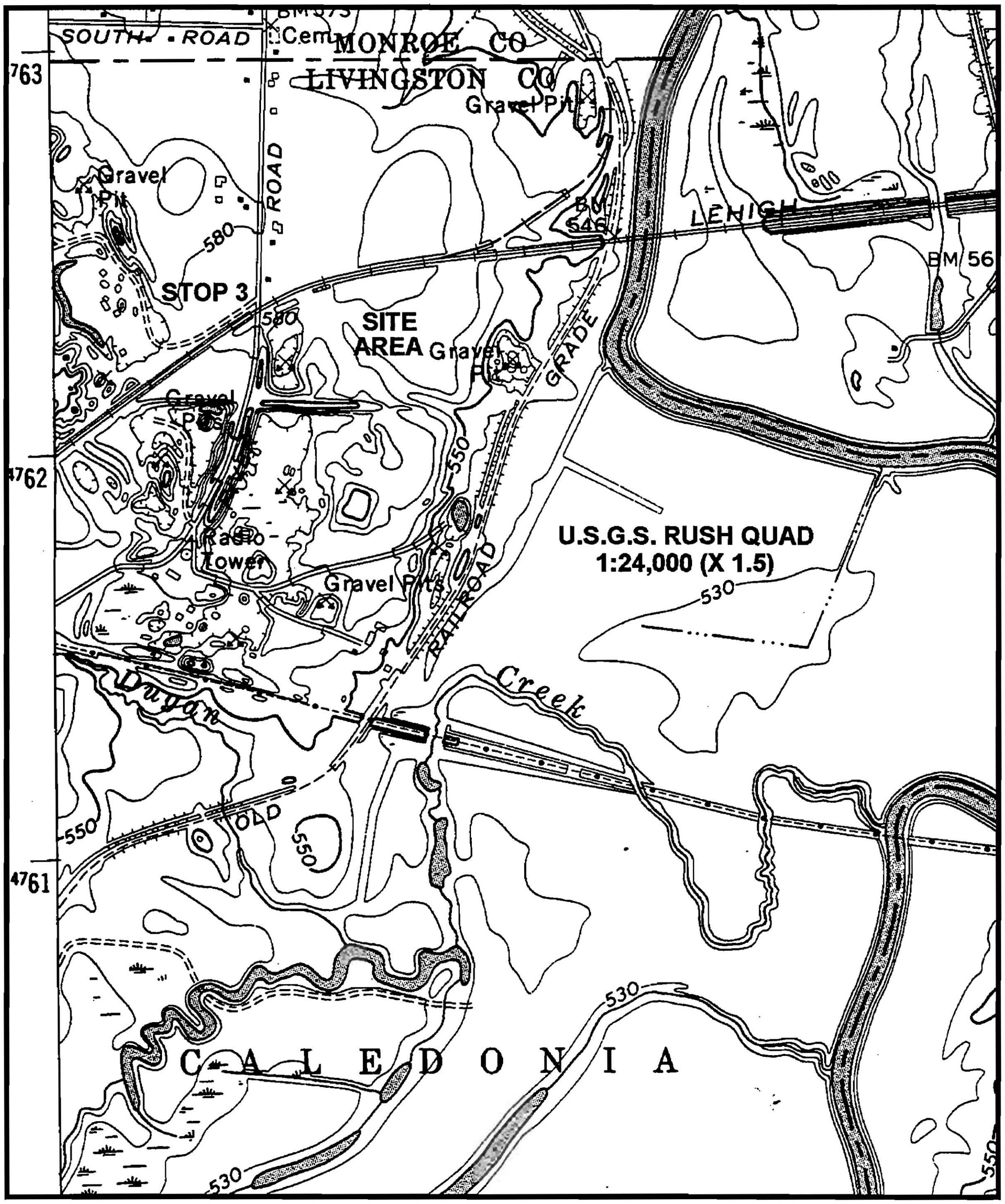
(See accompanying article)



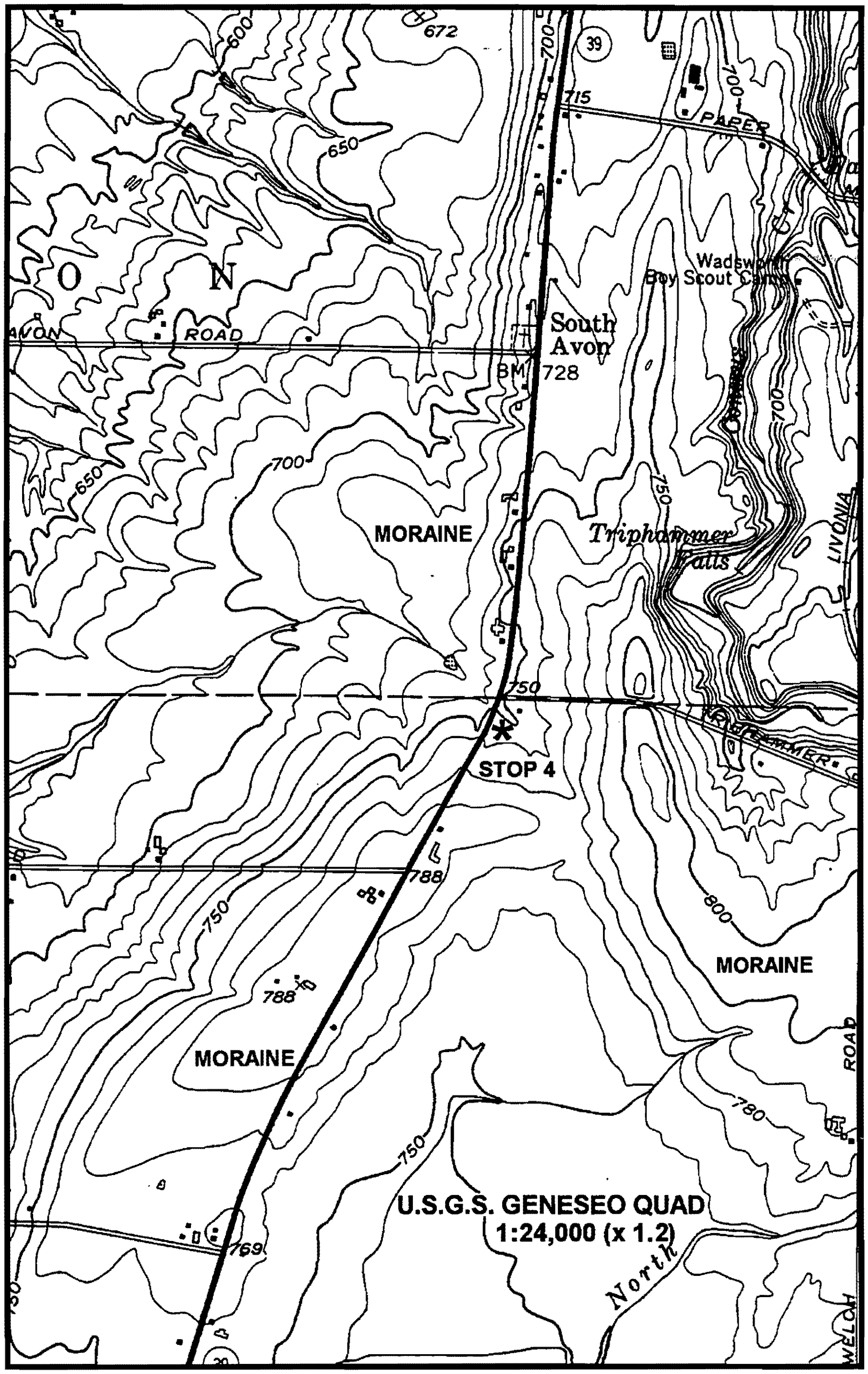
Location map of trip route and Stops 1-3. From N.Y.S.D.O.T. State Atlas.



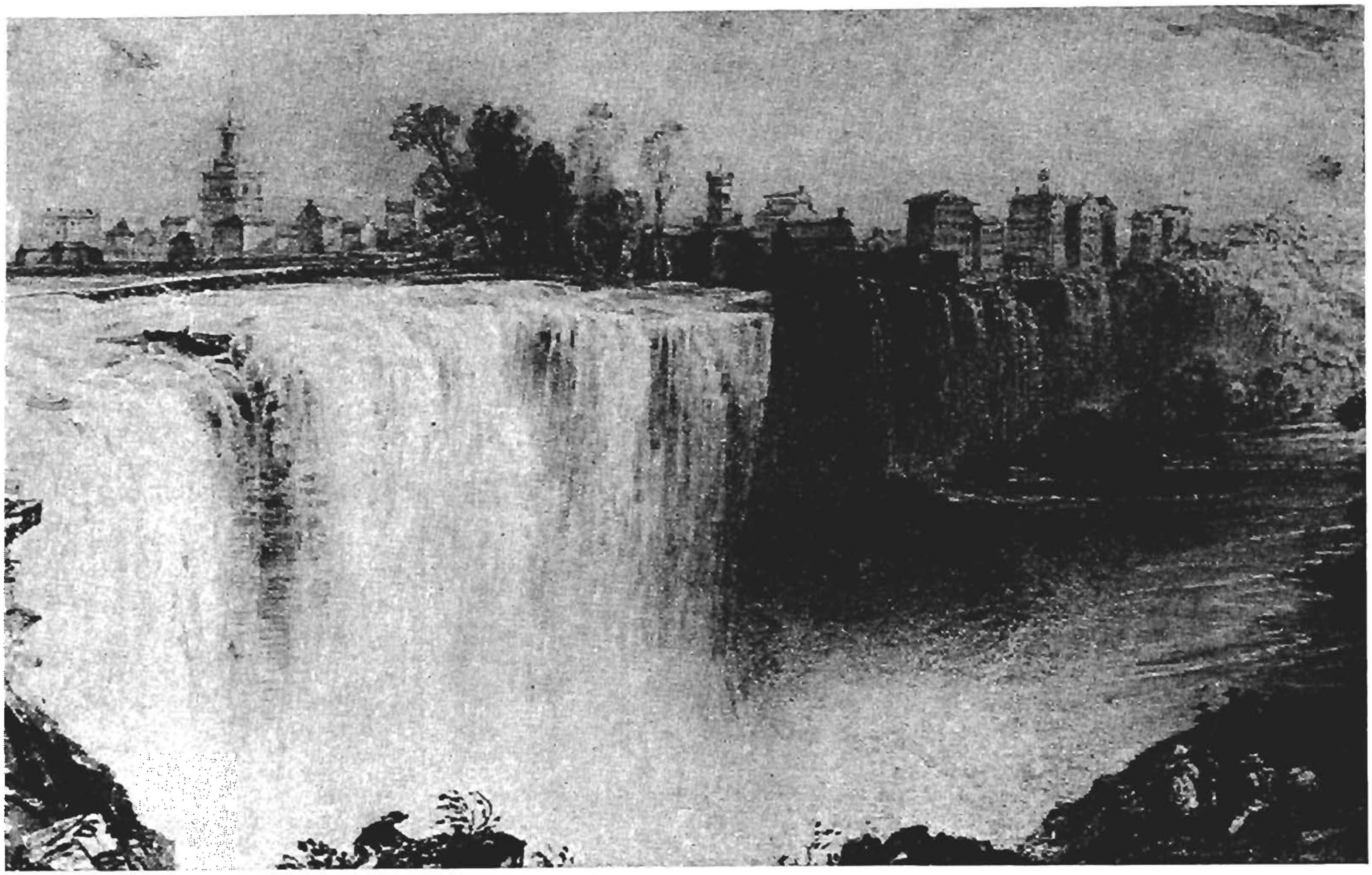
Location map of trip route and Stops 4-6. From N.Y.S.D.O.T. State Atlas.



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



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



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

GEOLOGY OF THE ERIE CANAL, ROCHESTER GORGE, AND EASTERN MONROE COUNTY, NEW YORK STATE: IN THE FOOTSTEPS OF AMOS EATON AND JAMES HALL

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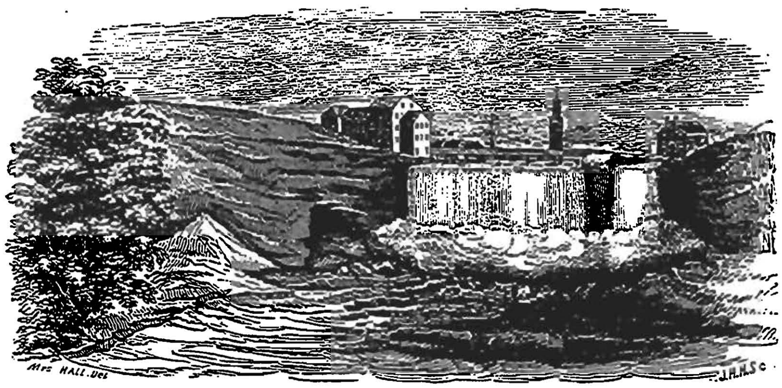

The Upper Falls at Rochester drawn by Mrs. James Hall (Hall, 1843, p. 82).

INTRODUCTION

The Phelps and Gorham Purchase, from the Seneca Nation in 1788, opened the lands west of Seneca Lake for settlement. Westward migration began as a trickle, but with the coming of the Erie Canal to the Genesee Country in 1822-23, a flood of immigration occurred that transformed western New York from dense forest to farmland, villages, and cities. For example, Rochester's first permanent settler did not arrive until 1812 while real growth blossomed in the 1820's. From the colonial period to the opening of the canal the west began at Schenectady.

The growth of geologic thought followed a similar pattern. A few random notices and sketches were made early on, primarily by army officers serving during the colonial wars and the Revolution. The Erie Canal changed all that. The surveys for the proposed project required organized, detailed, observations of the terrain and its underlying geology. Enter now DeWitt Clinton whose 1810 chronicle of a journey across the state, along with other writings, records numerous observations and interpretations of the geology from Albany to Buffalo. However, it wasn't until the seminal work of Amos Eaton (Father of New York Geology) in

1824 (*A Geological and Agricultural Survey of the District Adjoining the Erie Canal in the State of New York*) that a comprehensive and amplified attempt was made to unravel the superposition of western New York's strata. The foundation he laid ultimately led to the founding of the State Geological Survey in 1836. The following year James Hall came to the Genesee Country and examined its fossiliferous strata. Hall's work from 1837 to 1843 finally unraveled the stratigraphic column of western New York as we know it today, save for some modern refinements.

This paper attempts to show the evolution of geological investigations in Eastern Monroe County and Rochester as revealed in the writings and quotations of Amos Eaton and James Hall. The Erie Canal, its history and geology, and the geology of the Rochester Gorge are equally main themes as they provided the framework for the development of geologic thought.

In 1826 Amos Eaton led an expedition by canal boat across New York State. The surviving journals of two of the participants, George W. Clinton and Asa Fitch, provide unique insights into the state of geological knowledge, the nature of the landscape, and human development in the region at that time. These are quoted throughout the text but more liberally so in the roadlog. The writings of James Hall and to a lesser extent Sir Charles Lyell, are included to illustrate the degree of scientific advancement in the score or so years following Eaton's labors.

The paper is divided into five major subdivisions including the roadlog. The life, geological thoughts, and contributions of Amos Eaton is followed by a section on James Hall which contains a brief description of the rise of the New York State Geological Survey. The middle portion is devoted to the history and geology of the Erie Canal and its impact on New York State. With all of the historical perspectives in place the paper proceeds to the site specific, fourth, section - a discussion of Rochester and the Rochester Gorge. Finally the roadlog and boat trip is devoted principally to retracing the route of Amos Eaton's 1826 tour to the Rochester Gorge. The journal entries of the participants illuminate what travel was like at that time as well as highlighting Eaton's personality and ideas. Here Eaton's geology is contrasted with that of James Hall along with today's interpretation of the stratigraphic sequence and glacial topography. Hall's ideas are revealed in his 1837 field notebooks housed in the New York State Library in Albany and of course his monumental 1843 *Survey of the Fourth Geological District.*

AMOS EATON

Amos Eaton (Figure 1) was born in 1776 at New Concord, New York in the Taconic Mountains of Columbia County and died at Troy in 1842. He attended Williams College in western Massachusetts and graduated in 1799. Eaton eventually moved to New York City in 1800 to study law, where he became acquainted with and studied under two of the scientific leaders of the day, David Hosack and Samuel L. Mitchell. Under these men Amos Eaton's fervor for natural science was first nurtured. Details of Amos Eaton's life and his labors on New York geology may be found in McAllister's (1941) comprehensive biography, Wells'

FIGURE 1 Portrait of Amos Eaton (N.Y. State Library)

(1963) treatise on early geological investigations in New York State, and Fisher's (1978) synopsis of early New York geologists.

In 1802 Eaton was admitted to the bar and began working as a land agent for John Livingston, a wealthy land owner of Schoharie and surrounding counties. In 1804 Eaton quit his post with Livingston and moved to Catskill, New York to manage the 5,071 acres he purchased with his father the previous year and to establish his personal land agency. Eaton prospered in Catskill for several years when in September, 1809 he was falsely accused and indicted for forgery. Eaton was then 33 years old. The case against him eventually led to another more damaging suit in 1811 that resulted in bankruptcy that year and a trial on 26 August 1811 in Catskill. The jury returned a guilty verdict and Amos Eaton was sentenced to the state prison at hard labor for the rest of his life.

The state prison at that time was located in what is now Greenwich Village in New York City, on the banks of the Hudson at, ironically, Amos Street (McAllister, 1941 p. 142). While in prison Eaton continued his pursuit of scientific knowledge by studying botany and geology. In addition he met many influential people, not the least among them being DeWitt Clinton the mayor of New York and John Torrey the future eminent botanist. John Torrey, then a teenager, was the son of William Torrey the state prison agent. From his lengthy visits with

Amos Eaton, John Torrey received his early instruction in botany. The younger Torrey, in time, convinced his father to help seek Eaton's release from prison.

After Eaton had languished in prison for four years, his friends finally succeeded in securing his release when Governor Daniel Tompkins granted a pardon on 17 November 1815 on the condition that Eaton leave the state within three months, never to return. Eaton was finally vindicated on 15 September 1817 when Governor DeWitt Clinton granted an unconditional pardon. Amos Eaton persevered through an economic and personal storm that would have crushed a lesser person and at the age of 40 life began anew. In the Spring of 1816 Eaton moved to New Haven, Connecticut and studied at Yale, under Benjamin Silliman. In 1817 he returned to Williams College where he lectured on mineralogy and was conferred a Masters of Arts degree on 3 September 1817. Twelve days later DeWitt Clinton issued his pardon. From 1817 to 1824 Eaton wandered through eastern New York and western New England as an itinerant lecturer, although from 30 April 1819 he made Troy, New York his lifetime residence when he settled on Second Street near Ferry Street (McAllister 1941 p. 193). He soon met and came under the patronage of Steven Van Rensselaer, the last of the great Dutch Patroons. With the Patroon's financial support Eaton began his New York geological studies in earnest and eventually, in 1824, founded the Rensselaer School, now Rensselaer Polytechnic Institute in Troy.

EATON'S GEOLOGY

Before Eaton, not much was known of New York geology. As Wells notes (1963, p. 25):

> "Geologic work in the Northeastern States might soon have reached a dead center of vague generalization and haphazard observation and remained there for years, had it not been for the impact in 1818 of the forceful and, to many, irritating character of Amos Eaton."

Up to this time only William Maclure and Samuel Latham Mitchell had made attempts at unraveling New York's stratigraphic sequence on a grand scale. Maclure introduced the Wernerian classifications to American geology, thereby hindering its progress for many years (Wells, 1963). Maclure, to a lesser degree Mitchell, and others had failed to realize the significance and importance of fossils in establishing chronology. Amos Eaton was no exception. Beginning with his earliest geological publication in 1818 on the geology of the northern states, and culminating in his two monumental works, the 1824 canal survey and 1830 textbook, Eaton never unraveled the regional superposition of New York's formations (Figure 2). This was due to his adherence to Wernerian doctrine and his failure to recognize the general southerly dip of strata striking east-west across central New York State.

James Hall (1843, pp. 5-7) summarized Eaton's labors on the geology of New York as follows:

"The name of the late Stephen Van Rensselaer will always be remembered with reverence by the American student of geology.....Through his munificence, Professor Eaton was enabled to make a very extended and systematic survey of the rocks of New York; [Geological and Agricultural Survey of the District Adjoining the Erie Canal 1824].....if.....somethings are not.....in accordance with recent discoveries.....at that period.....he was in fact describing rocks.....not understood in Europe...... Had he evinced still more independence of European classifications.....pursued the investigation.....to a more thorough detail; published sections illustrating the order of superposition....with their fossils so numerous and characteristic, he would have left an undying fame to himself and his noble patron. We can only regret that this was not done in the most extended and perfect manner.....

"In that work (Survey of the Canal Rocks),.....it is evident that the author was fully aware of the great extent of our undisturbed strata as compared with those of Europe. He remarks, that:

> "Our secondary rocks along the line are several hundred miles in extent, and remarkably uniform in their leading characters.
>
> "After examining our rocks with as much care and accuracy as I am capable of doing, I venture to say, that we have at least five distinct and continuous strata, neither of which can with propriety take any name hitherto given and defined in any European treatise which has reached this country. The late work of Phillips and Conybeare describes many of the beds, and some of the varieties found among the rocks referred to; but the nomenclature of these very able geologists cannot be adopted to our district, without mangling and distorting the unprecedented simplicity of our rock strata (Canal Rocks, pages 7-9).

"I quote this, to show that Mr. Eaton was aware that the names and arrangement adopted in the systems of European authors did not apply to the rocks of New York; and yet, most fatally he attempted to apply that arrangement as far as possible, all the time supposing himself to be investigating rocks of the same age, while in truth they were much older than any described by the authors quoted. This attempt.....arose from the general belief that the older or Transition strata were in a highly disturbed and altered condition.....thus when so great a range of undisturbed strata, abounding with organic remains, was presented, as along the line of the Erie Canal, it was quite natural to refer them to the secondary deposits; ....

> "Having been a pupil of Prof. Eaton in the Rensselaer Institute, and receiving there my first instruction in Geology, it was natural to speak of him and his labors, as a tribute of respect as well to himself as to Mr. Van Rensselaer....."

Amos Eaton's canal survey of 1824 was his major achievement up to that time. The Erie Canal provided ease of travel, new exposures, and ready access to distant localities which were paramount in Eaton's construction of a framework and foundation for his geologic interpretations. In short, the Erie Canal resulted in the first geological survey of the state. This survey produced Eaton's (1824) cross section along the canal route from Boston to Lake Erie, but culminated in his 1830 textbook containing the first geologic map of New York State as a unit. In the 1820's and 1830's, the Erie Canal was a focus for geological work in New York, before the beginning of the New York Natural History Survey in 1836.

Eaton's theory of New York geology gradually emerged to embrace a concept of alternating episodes of non-marine and marine deposits (Figure 2). He divided the New York column into five series, each one bearing a cyclic sedimentary package of isochronous transgressions and regressions, beginning with a carboniferous unit, succeeded by a quartzose, and terminating in a calcareous deposit. According to Eaton, as the seas withdrew, land plants would occupy the emerged terrain. The ocean, when it returned, buried the plants to form his carboniferous formations, usually dark to black shales or perhaps even coal measures, at the base of each series. These units were generally his argillites or slates. The suprajacent deposit of the transgressing ocean would be coarse clastics or his quartzose formations, usually conglomerates and sandstones including red beds. He called most of these sequences graywackes. The continued flooding of the land culminated in the deposition of his calcareous formations such as dolostones and limestones. The seas would withdraw and the cycle would then repeat itself in the next series. Figure 2 clearly reveals Eaton's ignorance of the true superposition, yet he was one of the first to recognize local facies changes when, in 1828, he thought that the Catskill redbeds passed westward into the gray shales and sandstones of central and western New York (Wells, 1963).

Eaton's contributions to New York geologic investigations, although primitive and often erroneous were none-the-less significant. He made basic stratigraphic errors, but we must remember that New York geology at that time was like a cryptogram yet to be translated. Amos Eaton made the initial attempts, and in so doing laid the foundation for the more refined stratigraphic studies which were to follow. His students, such as James Hall, would invest the strata with new meaning and interpretations, carrying on where he left off.

JAMES HALL

James Hall, the "patriarch of American Paleontology, geological organizations, and state surveys" (Fisher, 1978) was born in 1811 at Hingham, Massachusetts, on the South Shore of Boston Bay. This setting influenced his love for natural history as he became very interested in

Series	Formations		Approx. Present Equivalents	Age
Anamolous	Analluvion, Diluvion		Glacial drift	Pleistocene
Tertiary (Fifth Series)	5th calcareous 5th quartzose 5th carboniferous	Shell marl Marine sand marly clay, plastic clay	Coastal Plain Long Island	Cretaceous
Upper Secondary (Fourth Series)	4th calcareous	oolitic oolitic rocks - Ohio	Glacial erratics in Catskills	
	4th quartzose	Third graywacke	Olean Conglomerate Marine Catskill Beds "Chemung" (Canadaway-Conewago Grps.)	Pennsylvanian L. Devonian
	4th carboniferous	Pyritiferous slate	Tioga Coal Genesee to West Falls Grp. U. Hamilton (W.N.Y.S.)	Pennsylvanian L. Devonian M. Devonian
Lower Secondary (Third Series)	3rd quartzose calcareous	Corniferous limerock Geodiferous limerock	Lower Hamilton Grp. Onondaga Ls. Salina Grp. (W.N.Y.) Lockport Grp. Helderberg Grp. (E.N.Y.S.)	M. Devonian M. Devonian L. Silurian L. Silurian E. Devonian
	3rd quartzose subordinates	Lias Ferriferous rock	Salina Grp. (C.-E. N.Y.) Clinton Grp.	L. Silurian E.-L. Silurian
	3rd quartzose	Saliferous rock	Newark Series Catskill Redbeds Oneida Cgl.- Herkimer Ss. Medina Grp. (Grimsby) Queenston Oswego Ss. Upper Lorraine Grp.	L. Triassic- E. Jurassic L. Devonian E.-Silurian- L. Silurian E. Silurian L. Ordovician L. Ordovician L. Ordovician
	3rd carboniferous	Second graywacke	Anthracite- Carbondale Pa. Lower Hamiton-Catskill Front Lower Lorraine Grp. Utica	Pennsylvanian M. Devonian L. Ordovician M. Ordovician
Transition (Second Series)	2nd calcareous	Metalliferous limerock Claciferous sandrock Sparry limerock	Helderberg Grp. Trenton Grp. Black River Grp. Tribes Hill Little Falls Chazy Grp. Beekmantown Grp.	E. Devonian M. Ordovician M. Ordovician E. Ordovician L. Cambrian M. Ordovician E. Ordovician & L.M. & E. Cambrian
	2nd quartzose	First graywacke	Shawanqunk Cgl. Schenectady Snake Hill Taconic sequence	E.-L. Silurian M. Ordovician M. Ordovician
	2nd carboniferous	Argillite	Normanskill Taconic sequence	M. Ordovician Ordovician & Cambrian
Primitive (First Series)	1st calcareous 1st quartzose 1st carboniferous		Green Mts. Manhattan Schist High Taconics Berkshires Adirondacks Highlands of the Hudson	Ordovician Ordoiviian Cambrian Cambrian Precambrian

FIGURE 2 Amos Eaton's Synopsis of New York Rocks, 1830 with approximate present equivalents.

[T.X. Grasso Drafted by: Richard D. Hamell, July 21, 1994]

During his tenure as State Geologist and as State Paleontologist, which lasted until his death in 1898, Hall continued to work on the stratigraphy of Western New York State. Much of this work was done in the "Genesee Country", as the area west of Seneca Lake was known, particularly in Monroe County. Hall named the Lockport Limestone, Rochester Shale, and Medina Sandstone with illustrations to clarify the stratigraphy, including the first vertical section showing relative thickness of formations drawn for the New York State Survey (Aldrich and Leviton, 1987). Another illustration was a map of the Genesee River with formations marked along its banks. This stratigraphy, with formational boundaries clearly marked, became incorporated into the New York System, the North American equivalent to the Silurian and Devonian Systems of Europe. The other State geologists eventually followed Hall's idea of using geographic localities for formation names, rather than the more colorful, lithologic terms of Eaton and others, although the New York System was dropped as correlations with the Silurian and Devonian of Europe became clearer.

Charles Lyell visited New York in 1841. He and Hall traveled the Erie Canal, examining outcrops along the way. Although Hall had little to do with the development of the Erie Canal, he was very critical of the financial waste in construction caused by choosing low quality rocks for building its locks, aqueducts and culverts. They also toured the Genesee River Valley, but little information exists of any impact Lyell had on Hall in this region. However Lyell, became much enamored with Niagara Falls, thereby leading Hall to devote considerable space to them in his final report. Hall also supported Lyell's idea that the surficial deposits of the region were the result of materials dropped in the sea by icebergs and swirling ocean currents. Early New York State geologists did not adopt a glacial theory for these drift deposits although Louis Agassiz had proposed the notion of glaciation (1837) in Europe (Aldrich and Leviton, 1987).

During his tenure as State Geologist and State Paleontologist, James Hall did much to promote the paleontology and stratigraphy of Western New York and laid the groundwork for future research in this region. For a more complete look at the life and contributions of James Hall see J. M. Clarke's biography of his mentor (1921).

ERIE CANAL HISTORY AND GEOLOGY

INTRODUCTION

The old Erie Canal (Figure 5) was not the first canal constructed in North America or even New York State, but it was by far the most successful. This is evidenced by the simple fact that one can still go by canal from Buffalo to Albany, across the state, albeit in a different location for much of the distance. The Erie Canal is still operating. This cannot be said of the Pennsylvania, New Jersey, Virginia and other state canals that attempted to link the interior of the continent with the eastern seaboard. The Erie Canal pierced the wilderness of nineteenth century New York State, transforming a sparsely populated hinterland of impenetrable forests into a line of burgeoning metropolitan complexes. As an alternative to the tortuous overland journey west, it opened the mid-continent of North America to a flood of migration and settlement.

Yankee, not Irish laborers, constructed the original canal, using teams of horses drawing scrapers and scoops, shovels, wheelbarrows and certain makeshift tools invented as the need arose. One such elegant device was a contraption to pull tree stumps out of the ground, as shown in Figure 6. Workers completed the 363 mile long canal in eight years. Syracuse, Buffalo, and many smaller communities either grew dramatically or came into existence as a result of the canal.

The original Erie Canal cost $7,143,789.86, well above the engineers' estimate of $4,881,738 (Whitford, 1906). However, in 10 years, from the tolls generated, it repaid all construction costs, including principal and interest on the loan, paid all maintenance and repair costs, and showed a profit. The State has apparently never surpassed this remarkable undertaking. Shipping costs were cut 80-90 percent, and the trip from New York City to Buffalo was reduced from a month to about two weeks. It was these economic incentives which inspired the novel idea, but also underlie its success.

FIRST CANALS IN NEW YORK STATE

In March 1792, an act of the legislature established two private canal companies -the Western Inland Lock Navigation Company (W.I.L.N.) and the Northern Inland Lock and Navigation Company (N.I.L.N.) - apparently brought to fruition by Elkanah Watson, a former assistant to Benjamin Franklin (Whitford, 1906). Watson was a friend of George Washington, from whom he probably acquired his passion for canals. General Philip Schuyler, a prominent member of the Senate, was instrumental in obtaining the law that created the two lock-navigation companies.

Although both companies were private stock ventures, each was linked with the state through monetary gifts, loans and purchases of stock. The Northern Company, incorporated to facilitate a water communication between the Hudson River and Lake Champlain, accomplished nothing beneficial while somehow expending $200,000. The Western Company, as shown on Figure 7, succeeded in making modest improvements in the Mohawk River Valley by constructing a mile-long canal with five locks around the falls at Little Falls (1795) and another two-lock canal connecting the Mohawk River with Wood Creek at Rome (1797). It also built a short canal around two rapids in the Mohawk River east of Herkimer (1798) and made minor improvements in Wood Creek (1793, 1803). After the year 1800, however, the canals fell into general disuse and after 1803 the company faded into oblivion.

The demise of the private navigation companies stemmed from high tolls, wasteful management, lack of experience and a host of other factors, one of the most significant being that short canals offered only a partial solution to the problems of inland water navigation. In fact, the failures of the lock navigation companies set the canal movement back 15 years as opponents of the Erie pointed out the futility of building a canal 363 miles long through

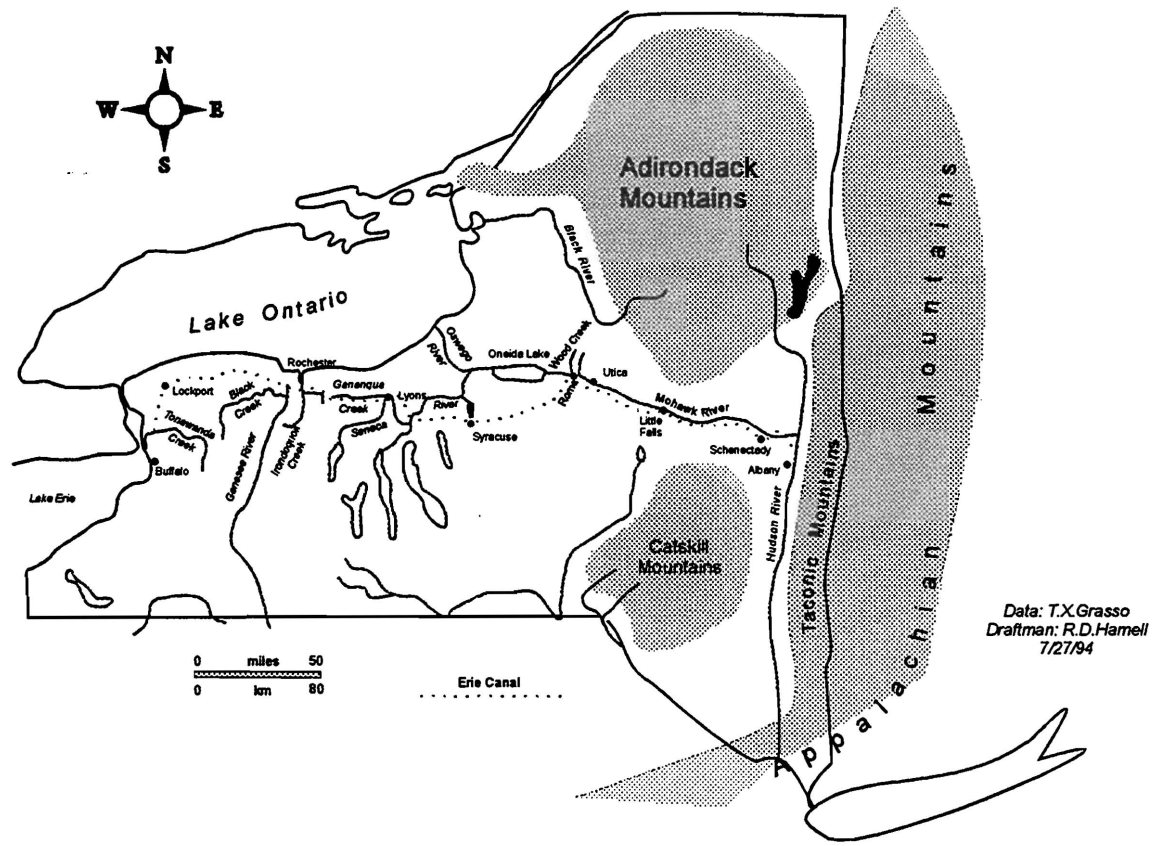

Figure 5 Location map of Erie Canal (dotted line) and neighboring highlands.

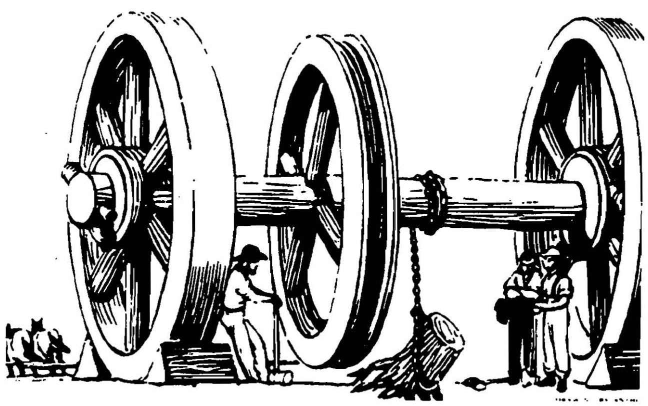

Figure 6 Device to grub stumps (from Andrist, 1964, p. 453; drawing by Anthony Ravielli)

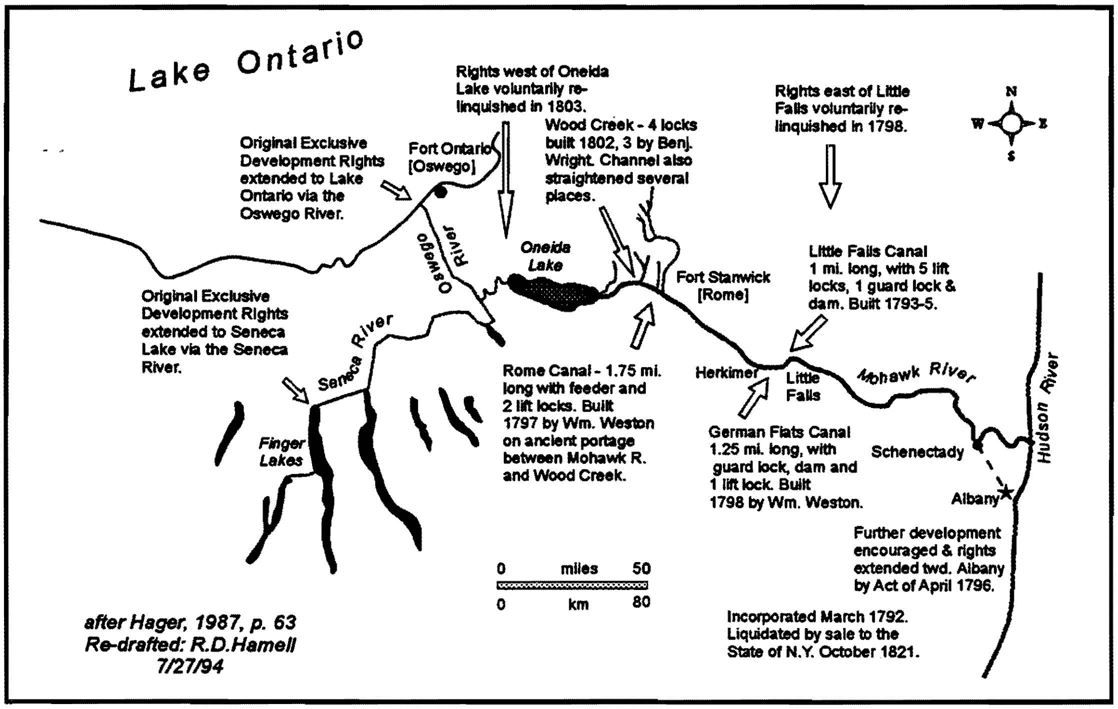

Figure 7 Map showing developments of the Western Inland Lock Navigation Company

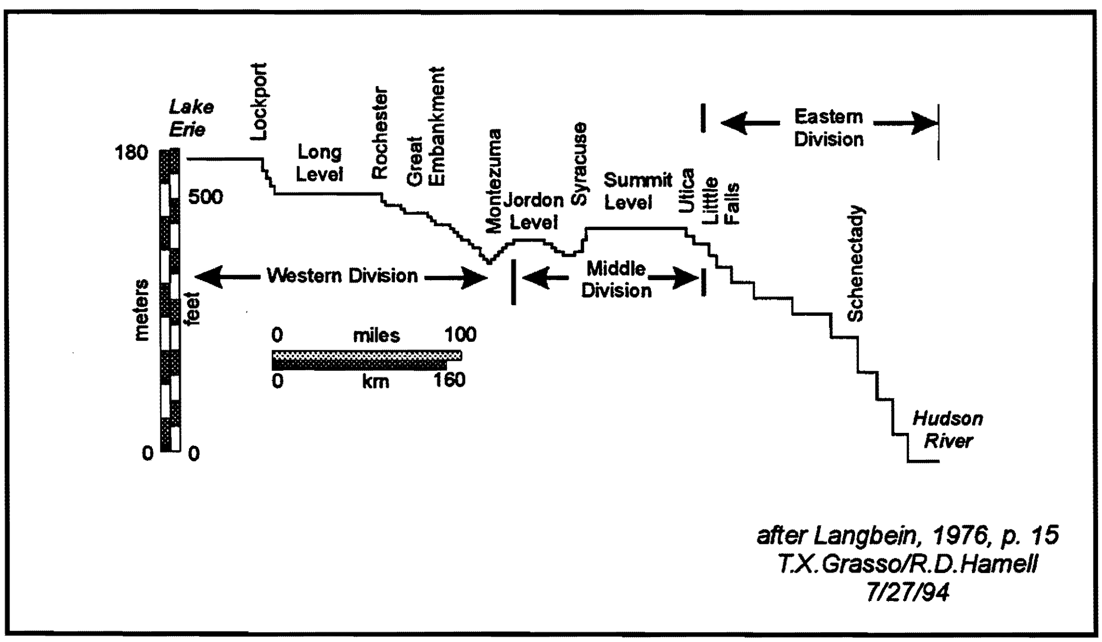

Figure 8 Profile of the Erie Canal

unquestionable authority, that the reports that had been so industriously circulated, respecting the sickness and death of many of the workmen on the Montezuma Marshes are entirely unfounded."

Clinton's Ditch crossed the Seneca River on grade, the towpath being carried across the river on a bridge supported by 130 bents (frames). To obviate the difficult grade crossing in the first enlargement, Van R. Richmond designed and built the great Montezuma Aqueduct. Work began in 1849 and it was brought into use in the spring of 1856. The stone aqueduct rests on a wood foundation floor, covering an area of approximately two acres, supported by 4,464 bearing piles from 15 to 30 feet long (4.5 to 9 m). The aqueduct had 31 towpath arches and was 841 feet long, the second longest aqueduct on the Erie Canal. The central portion was dismantled in the winter of 1917-1918 to free the Seneca River for Barge Canal traffic.

Genesee-Irondequoit Valleys

The route of the canal west from Montezuma lay up the valley of the Clyde River to Lyons. From here it ascended the tributary Ganarqua (Mud) Creek, the shallow draft navigable headwaters of which are found at Palmyra. These are misfit streams that occupy a plexus of abandoned meltwater channels collectively known as the Fairport Channels. The Fairport Channels drained proglacial Lake Dawson in the Rochester region, carrying its waters and all the Great Lakes drainage east to Lyons where it emptied into an early Lake Iroquois. These channels cut a swath through the drumlin belt, permitting a low gradient path for the canal.

The preglacial course of the Genesee River led east from Rush then north through what is now the Irondequoit Valley. Near Rochester, Lake Dawson ranged from 483 feet (147 m) elevation to 462 feet (141 m). Details of the glacial lakes succession in the Genesee Valley region may be found in Muller, et. al (1988). Lake Scottsville occupied the Genesee Valley south to Avon at the same time as Lake Dawson's waters existed over the northern part of Rochester. Lake Scottsville at an elevation of 540 feet (165 m) was dammed at its northern margin by the Pinnacle Hills moraine along the southern margin of Rochester. Lake Scottsville waters drained north across the Pinnacle Hills moraine because the eastward leading leg of the old channel was plugged with glacial deposits. However the north-south portion of the preglacial valley was scoured into a deep U-shaped trough. The Genesee River, now flowing northward in a new postglacial course, began carving the Rochester Gorge when it encountered the Niagara Escarpment. Withdrawal of the ice margin farther north just over 12,000 B.P. permitted Lake Iroquois to expand westward to completely occupy the Lake Ontario basin. Lake Scottsville drained away and Lake Dawson rapidly lowered a total of 120 feet (37 m) to an elevation of 425 feet, the level of Lake Iroquois, the southern beach ridge of which is Ridge Road (NY 104).

The deep U-shaped trough of the preglacial Genesee, the Irondequoit Valley, was to be a dire hurdle. The canal commissioners wanted to maintain a profile that was downhill from Lake Erie to the Genesee River and as far east as Montezuma (Figure 8). Luckily, east of Pittsford near Bushnell's Basin (Figures 10, 29), the valley was occupied by a kame and esker

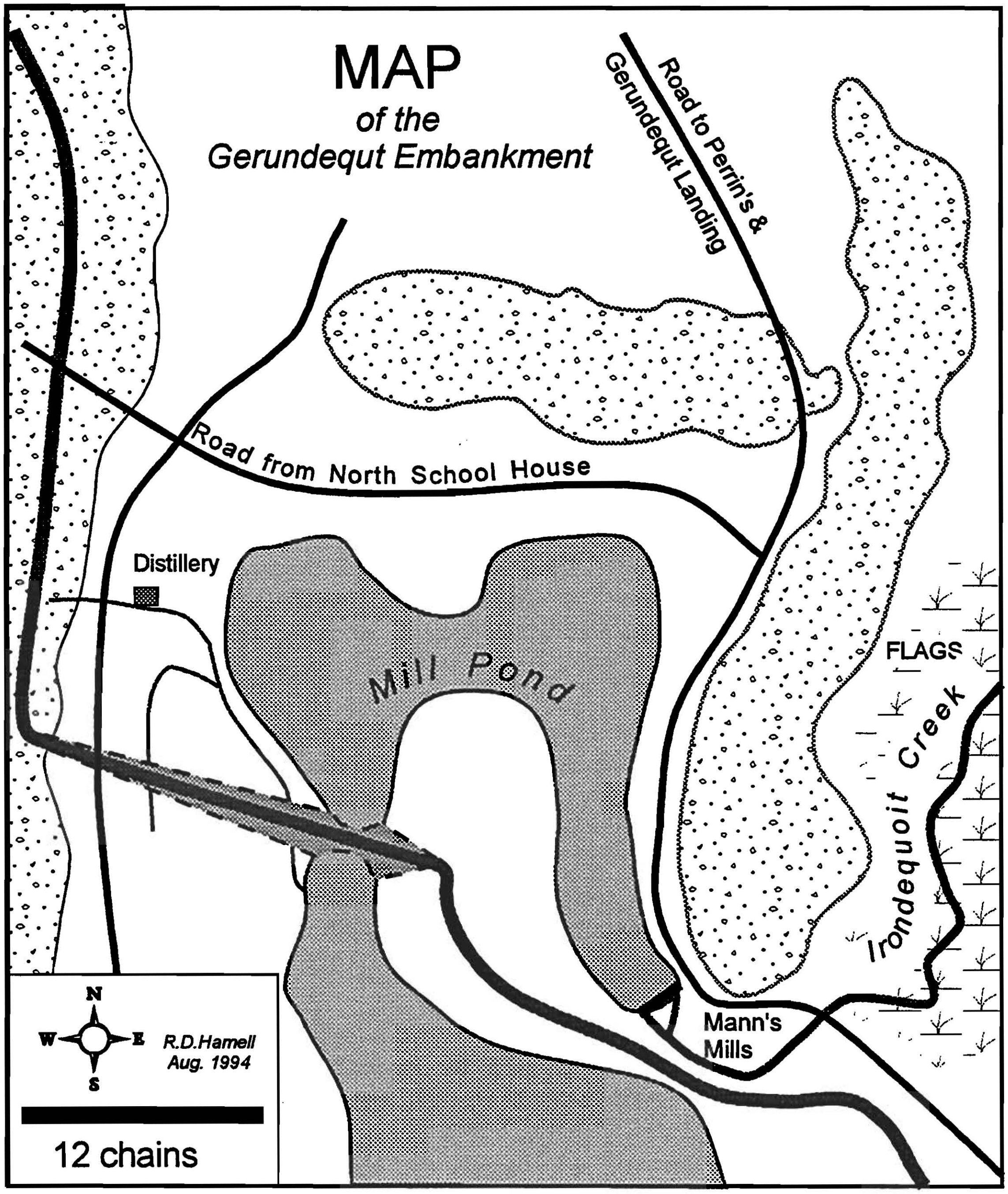

Figure 10

James Geddes' 1808 map of the Irondequoit Valley showing the proposed Great Embankment. The stippled patterns are the "ridges" of Geddes. The ones north of Mann's Mills and the loop in Irondequoit Creek show the Cartersville Esker. The canal eventually crossed the esker north of Mann's Mills.

(Map modified by R. D. Hamell)

complex the crests of which were all at the same elevation. The commissioners approved a scheme to connect the natural deposits with an artificial embankment which would carry the Erie Canal some 60 feet (18 m) over Irondequoit Creek. A culvert 240 feet (72 m) long and 17 feet (5 m) high was constructed to carry Irondequoit Creek beneath the embankment. The structure, which came to be known as the Great Embankment, was completed in 1822 and was heralded far and wide as one of the great engineering marvels of the Erie Canal (Figure 11).

James Geddes in his 1808 survey discovered the esker ("ridges" as he called them) when he wrote (Canal Laws, 1825, p.43):

> "...The passage of the Irondequoit Valley is on a surface not surpassed, perhaps in the world, for singularity. . . Those ridges along the top of which the canal is carried, are in many places of just sufficient height and width for its support, and for 75 chains the canal is held up, in part by them, and in part by artificial ridges, between 40 and 50 feet above the general surface of the earth; the sides of them are in most places remarkably steep, so that when the work is finished, the appearance to a stranger will be that nearly all those natural embankments were artificial works."

At Rochester the canal crossed the Genesee River on a large aqueduct 810 feet (247 m) long (Figure 12), making it the third longest aqueduct on the entire line of the Enlarged Erie. Further, it was the only all stone aqueduct as the others carried the canal in a wooden trunk, supported by stone piers, with towpath arches on one side only. The commissioners went to the added expense, not for the sake of architectural beauty, but because of the severe flooding that long plagued the Genesee River and which was arrested only by the completion of the Mount Morris Dam at Letchworth State Park completed in 1953.

Lockport and the Niagara Escarpment

Canal construction through Lockport proved to be extremely difficult and this was the last section completed in 1825. Two prominent escarpments exist on the Lake Ontario plain in Western New York - the Niagara on the north (once known as the Mountain Ridge) and the Onondaga on the south. Of the two, the Niagara Escarpment is more pronounced. The Lockport Dolostone which crowns it is famous for its crystals of calcite, dolomite, fluorite, gypsum and other sulfates, as well as sulfides.

Beginning in Lake Dawson time and continuing into Lake Iroquois time, a lake was impounded between the Niagara and Onondaga Escarpments which was named Lake Tonawanda by Kindle and Taylor (1913). This lake extended east to near Holley and debauched north across the Niagara Escarpment through several spillways. Of these the spillway at Lockport gained ascendancy over the others, probably due to greater isostatic rebound to the east. This spillway carved a northeastward trending notch in the Niagara

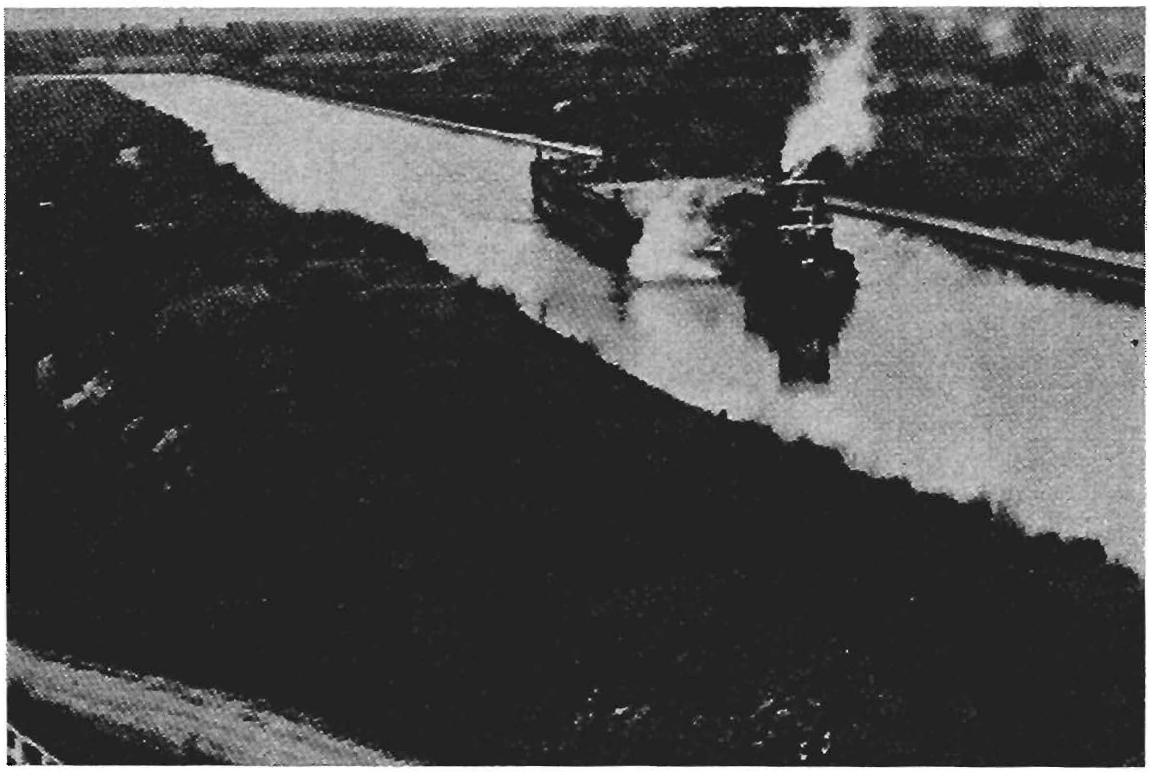

Figure 11 The Great Embankment circa 1921 looking west from Bushnell's Basin. Jefferson Road (NY96) in lower left foreground.

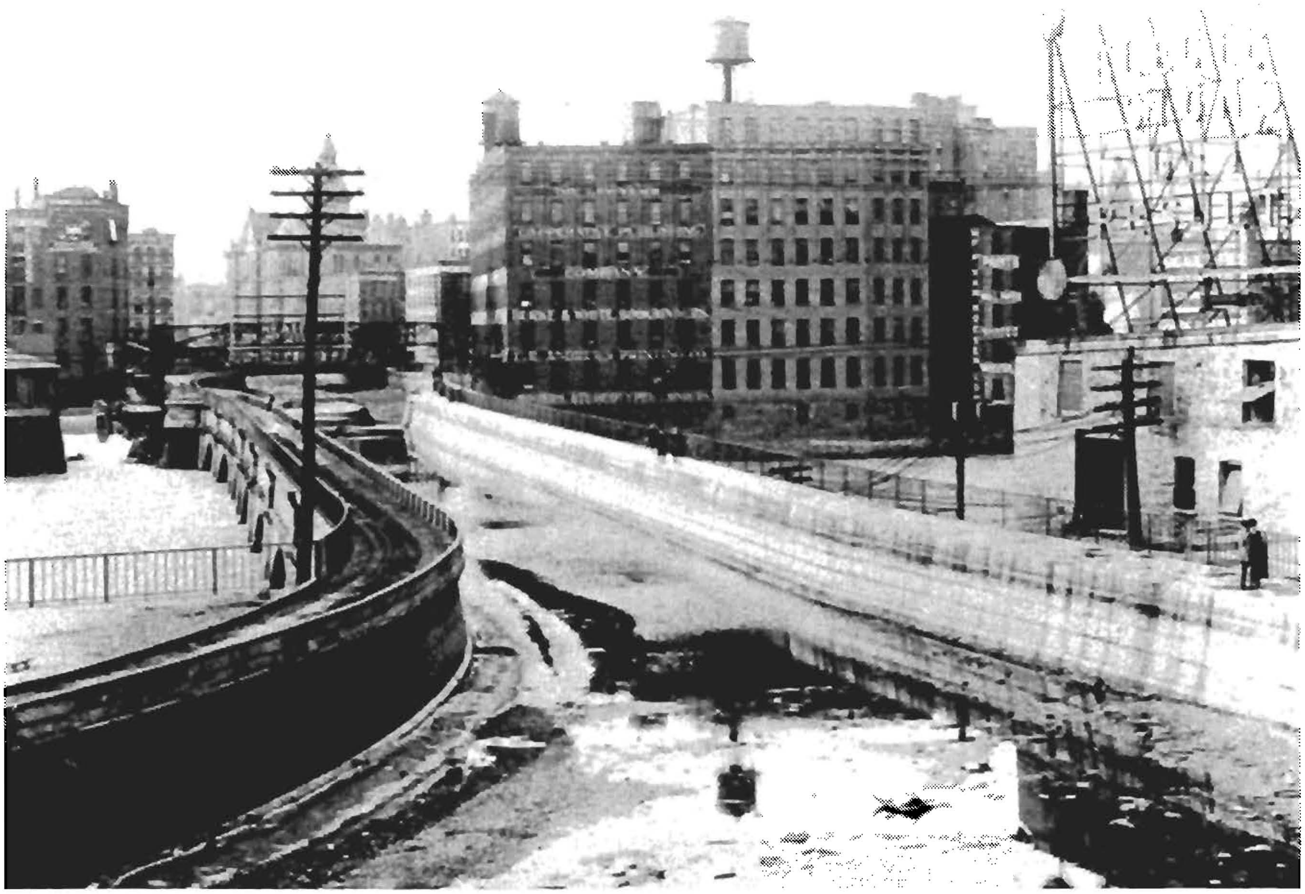

Figure 12 Downtown Rochester Aqueduct looking west, circa 1913 (Stone Collection, Roch. Mus. and Sci. Center)

Figure 13 Lockport flight of combined locks, circa 1880. Note canal boats in lowest and uppermost chambers of the west flight. Rochester Shale capped by Decew Dolostone on right (west) bank.

Escarpment that, before the canal, carried a misfit tributary to Eighteen Mile Creek (not the classic Eighteen Mile Creek of Grabau famous for its fossils - which is south of Buffalo).

In this defile Nathan Roberts completed, in 1825, the famous Lockport flight of five double combined locks that raised or lowered boats fifty-five feet across the face of the escarpment (Figure 13). Furthermore, to maintain a gentle gradient eastward for the flow of Lake Erie water, Roberts' crews had to blast through the dip side of the cuesta, thereby creating a "deep cut" from Pendelton at Tonawanda Creek to Lockport with a maximum depth of forty feet (12 m). The commissioners had once contemplated a route going directly east from Tonawanda Creek, running south of the escarpment instead of turning north at Pendelton. This plan had the advantage of being shorter and cheaper as it would have negated the cost of blasting the deep cut and constructing five combines. However, this would have meant going uphill from Lake Erie, not down, and the water supply from Lake Erie would have been lost. This early plan was therefore abandoned. From the mouth of Tonawanda Creek the canal paralleled the east side of the Niagara River, terminating in the Erie Basin in downtown Buffalo.

GEOLOGY OF ROCHESTER AND THE ROCHESTER GORGE

INTRODUCTION

The Erie Canal came to Rochester in 1822-1823, and the "Young Lion of the West" was born. The appellation was accurate enough as evidenced by the economic explosion and unparalleled physical growth, fostered by the canal, that was this city's hallmark in the first quarter of the 19^{th} century. The original Erie Canal was only 40 feet wide at water surface and just 4 feet deep. Yet this narrow ribbon of water made New York the Empire State, and made Rochester the first "boom town" in America. The canal lowered transportation costs by more than 90% and the rapid growth that resulted is clearly revealed by the fact that 3,130 souls resided in the Village of Rochesterville in 1822, more than double the population only two years before. Further affirmation is found in the 1827 village directory, which boasted:

> "...not one adult person is a native of the village. The oldest person living in the village that was born here is not yet 17 years old."

It is difficult to imagine that in 1810, when DeWitt Clinton and the canal commissioners came here on an exploratory trip for the proposed Erie Canal, there wasn't a single person residing in what would become the Village of Rochesterville - today's downtown Rochester. Later, in 1827, Clinton (found in O'Reilly, 1838, p. 416), wrote to Everard Peck recalling his visit:

> "When I saw your place here in 1810 without a house who would have thought that in 1826 it would be the source of such a work? This is the most striking illustration that can be

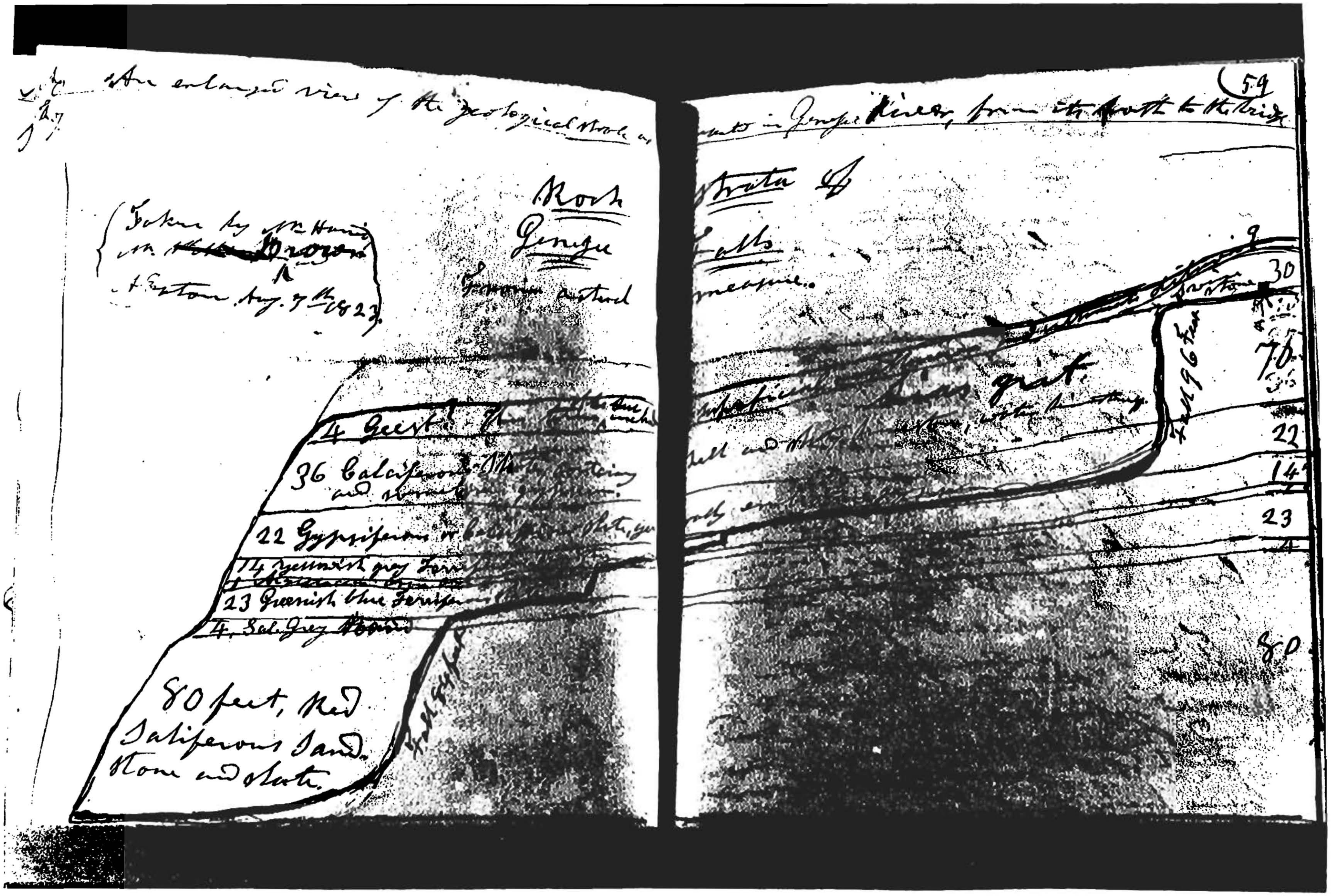

Figure 14 Eaton's 1823 drawing of the stratigraphy and falls in the Rochester Gorge. Note the northerly dip of the strata. (Journal C, Aug. 1823, N.Y. State Library)

> furnished of the extraordinary progress of your region in the career of prosperity."

Not everyone shared the governor's enthusiasm for Rochester. In 1826, Amos Eaton led a geological expedition across the state aboard the canal boat LAFAYETTE. Earlier, in 1823, Eaton made the first detailed geologic study of the Rochester Gorge where he depicted the sequence of formations in a sketch contained in his field notebook (Figure 14). Eaton's understanding of the geology of the gorge was further advanced by the Reverend Chester Dewey. Eaton's work formed the basis for the illustration entitled "Section of Rocks on the Genesee River etc." on page 77 of Henry O'Reilly's 1838 *Sketches of Rochester*, and redrawn here (Figure 15) with annotations.

Returning to Eaton's 1826 field excursion, the group included Rensselaerian School students, faculty, and several dignitaries such as George W. Clinton, the governor's son, Asa Fitch the future state entomologist; and Joseph Henry, the brilliant young physicist who performed some of the seminal experiments on magnetism in this country and thereafter helped found the Smithsonian Institution. The party arrived in Rochester on Sunday, May 14, and Asa Fitch recorded Eaton's impressions of Rochester:

> "This place Prof. Eaton says is a mere mushroom springing upon a moment and is destined to decay and fall away to nothing. He predicts that there will not be a third of the present number of buildings in the lapse of a few years - that he does not believe there is a place on earth so remarkable for its splendor and poverty."

Fifteen years later (August, 1841) Rochester was visited by the eminent Scottish geologist Sir Charles Lyell (1845, p. 23) who recorded in his diary:

> "We explored the picturesque ravine through which the Genesee flows at Rochester, the river descending by a succession of cataracts over the same rocks which are exposed farther westward on the Niagara...The contemplation of so much prosperity, such entire absence of want and poverty, so many school-houses and churches rising everywhere in the woods...fills the traveler with cheering thoughts and sanguine hopes..."

One can only wonder if Eaton and Lyell visited the same municipality. (*All Lyell quotes here and elsewhere in this paper were graciously provided and authored by Gerald M. Friedman Ph.D., D.Sc., Rensselaer Center of Applied Geology, affiliated with Brooklyn College, City University of New York*).

Rochester was preordained for greatness, not so much by its location on the canal--a major east-west artery of commerce--but by its location at the waterfalls and gorge carved by the muddy waters of the Genesee in its head-long dash north to Lake Ontario. The water power, derived from the river's drop, churned water wheels that in turn drove gears, shafts and leather

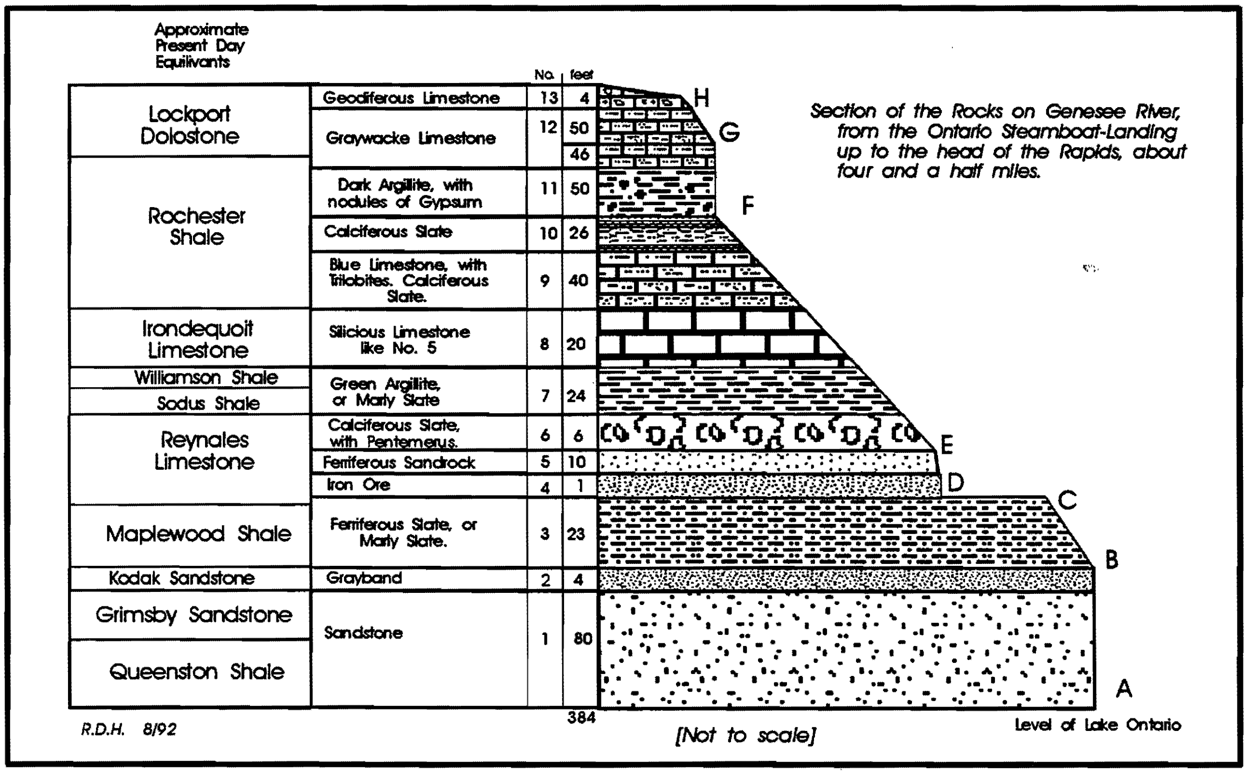

FIGURE 15 Geology of the Rochester Gorge, contrasting authors' rock units with those in use 1823-1838.

A to B - height of the last step (84 feet) of Lower Falls [Lower Falls]*
B to C - ascent to the upper step of those falls [Middle Falls]
D to E - height of upper step of the Lower Falls, about 25 feet [Middle Falls]
E to F - ascent up the river
F to G - height of the Middle Falls, 96 feet [High Falls]
G to H - ascent to the Rapids [old Upper Falls-base of Court Street Dam]

*Brackets are author's for current terminology belts that provided smokeless industrial power. A spectrum of products for domestic, agricultural and industrial use, such as flour, furniture, edge tools, farm machinery, beer, barrels, canal boats, and fire engines were but a few of the products directly or indirectly, turned out by Rochester's many water powered mills.

The primacy of the waterfalls and river gorge in the historical settlement and industrial development of Rochester cannot be understated and therefore geologic and human history are intertwined. Geologic events of the near and distant past laid the foundation that guided the course of human events resulting in the founding and growth of Rochester. In broad strokes, the geological-historical interrelationship can be explained by the gorge's attraction of the early settlers because of the ample water power. The mills that located at the falls produced goods that could not be inexpensively transported to New York City markets. The attraction of great reductions in shipping costs provided the incentive for locating a canal across the Genesee River at Rochester, giving Rochester millers and their products a ready access to eastern markets. This suddenly transformed what before the canal was a struggling enclave at the small Upper Falls of the Genesee, into a thriving, vibrant community after the ditch's completion. Had the Rochester gorge not formed and if it did not contain the rock types necessary for the formation of waterfalls, the canal may have crossed the Genesee River elsewhere--perhaps farther south--and maybe Canandaigua, Avon, or Batavia would have attained Rochester's status as the major city in the region.

THE NIAGARA ESCARPMENT

A prominent rock ridge, approximately 200 feet high, rises abruptly from the lake plain that borders the southern shore of Lake Ontario. This topographic feature was called the Mountain Ridge by the early white settlers, and later the Niagara Escarpment by mid 19^{th} century geologists. Although ill-defined east of Rochester, the escarpment becomes more pronounced westward forming a sharp angular feature in the Niagara River region (Figure 16).

Downtown Rochester and I-490 West, from downtown to the Spencerport exit, lie at or near the crest of the escarpment; numerous rock cuts along the way proclaim its presence. Further evidence of the escarpment's existence in the Rochester area is revealed by the sharp drop in the lay of the land north of Ridgeway Avenue and the view of Lake Ontario that northbound motorists can observe from NY390 just north of Lexington Avenue.

In the Rochester region the Niagara Escarpment and Ridge Road (the southern beach of Lake Iroquois - see Fig. 31, p. 59) were two prominent topographic features that did not escape the keenly watchful eye of DeWitt Clinton during his canal tour of 1810 (Campbell, 1849, p. 114). In his journal entry for Sunday July 29^{th}, he wrote:

> "Shortly after leaving the Genesee River, we entered a remarkable road called the Ridge Road, extending from that river to Lewiston, seventy-

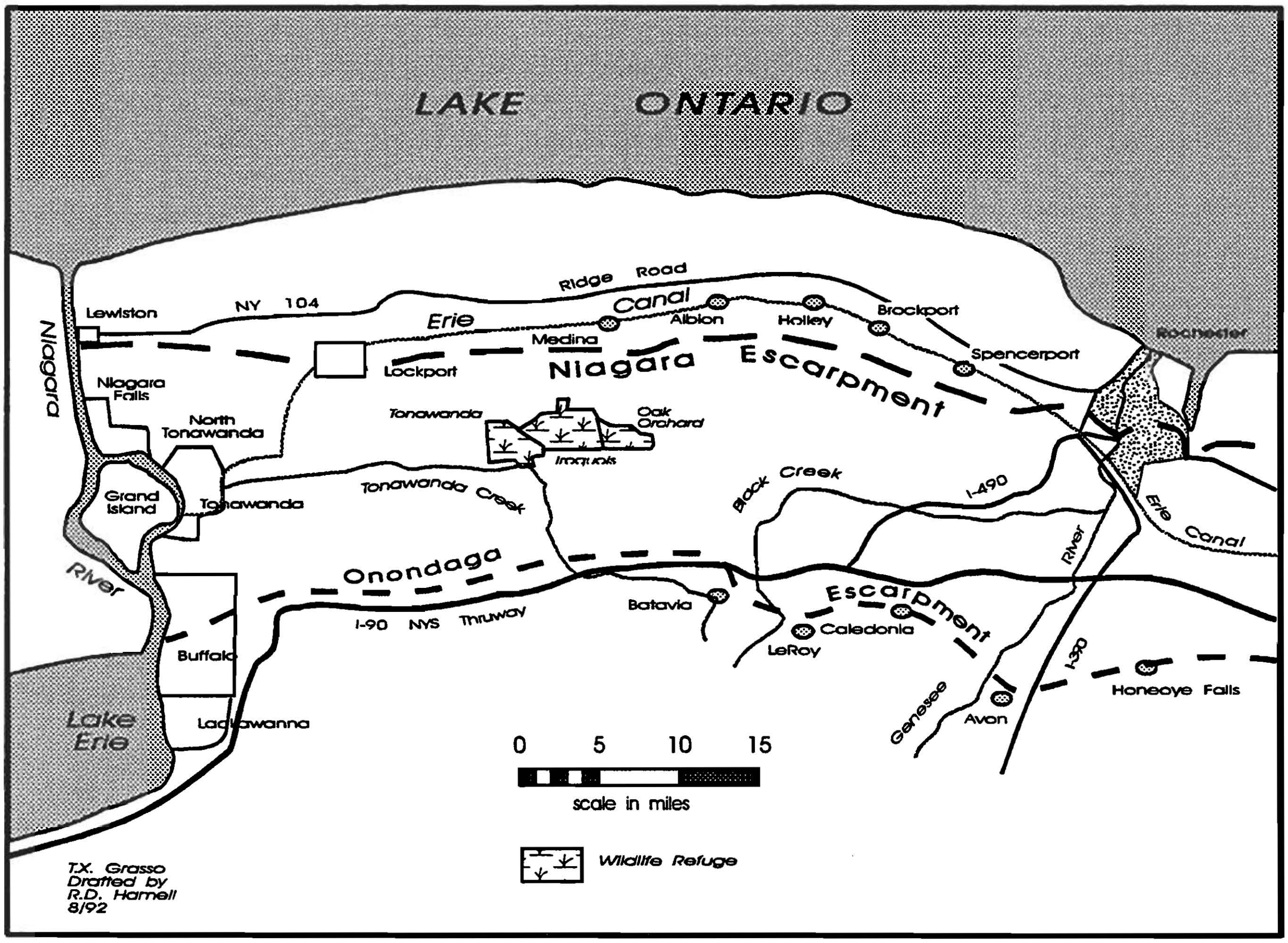

FIGURE 16 Map of the Niagara and Onondaga Escarpments in Western New York.

> eight miles. The general elevation of the ridge is from ten to thirty feet, and its width varies........About from three to half a mile south, and parallel with this ridge, there is a slope or terrace, elevated 200 feet more than the ridge, with a limestone top [Lockport Dolostone], and the base freestone [Queenston Shale/Grimsby Sandstone]. The indications on the ridge show that it was originally the bank of the lake. The rotundity of the stones, the gravel, and c., all demonstrate the agitation of the waters........"

When rivers and streams eventually crossed the Niagara Escarpment and cascaded down its face, waterfalls and rapids formed. These, in time, receded or migrated upstream leaving behind deep bedrock gorges notched into the escarpment. Thus did the Rochester and Niagara Gorges come into being. But before all this could happen, the underlying bedrock units had to be deposited uplifted, tilted to the south $^1/_2°$, and beveled by erosion.

BEDROCK GEOLOGY

The strata exposed in the Rochester Gorge were deposited discontinuously over a span of 20 million years beginning in the Late Ordovician Period and ending in the Middle Silurian Period. The total sedimentary package of primarily shelf strata is about 450 feet (137 m) thick but was deposited in environmental settings ranging from marginally marine intertidal to fully marine, deep basin, anoxic, water. The stratigraphic column is subdivided into four major rock units (from older to younger): Queenston Formation, Medina Group, Clinton Group and Lockport Group. These units, save for the Queenston, can be further subdivided into finer scale units (Figure 17).

The Queenston Formation and Medina Group are well exposed in the lower gorge from the Stutson Street Bridge to the Lower Falls, just south of the Driving Park Bridge (**STOP 4**). The top of the 84 foot (25.6 m) high Lower Falls exposes the top of the Medina Group, namely the Kodak Sandstone, a prominent gray band exposed about half way up the side of the gorge just above the red Medina Group. Early in Rochester's history this falls was known as the Lower Step of the Lower Falls (Figures 18, 19).

At the Lower Falls, above the Kodak Sandstone, the gorge reveals nearly the entire Clinton Group terminating in the lower part of the Rochester Shale. The Reynales and Irondequoit Formations are relatively conspicuous as they jut out from the side of the gorge. They have a brownish color and are made up of thicker beds than the intervening thin bedded shale (Figure 20). The Reynales Formation is made even more prominent by the presence of a thin (1 foot thick) red iron ore bed, a few feet up from the base of the unit. This bed, formerly the Furnaceville Hematite, has recently been renamed the Seneca Park Hematite by LoDuca and Brett (1990). The upper part of the Reynales Limestone caps the Middle Falls which earlier was known as the Upper Step of the Lower Falls. Today RG&E operates a hydroelectric dam on top of the Middle Falls. The remainder of the Clinton Group is exposed more or less

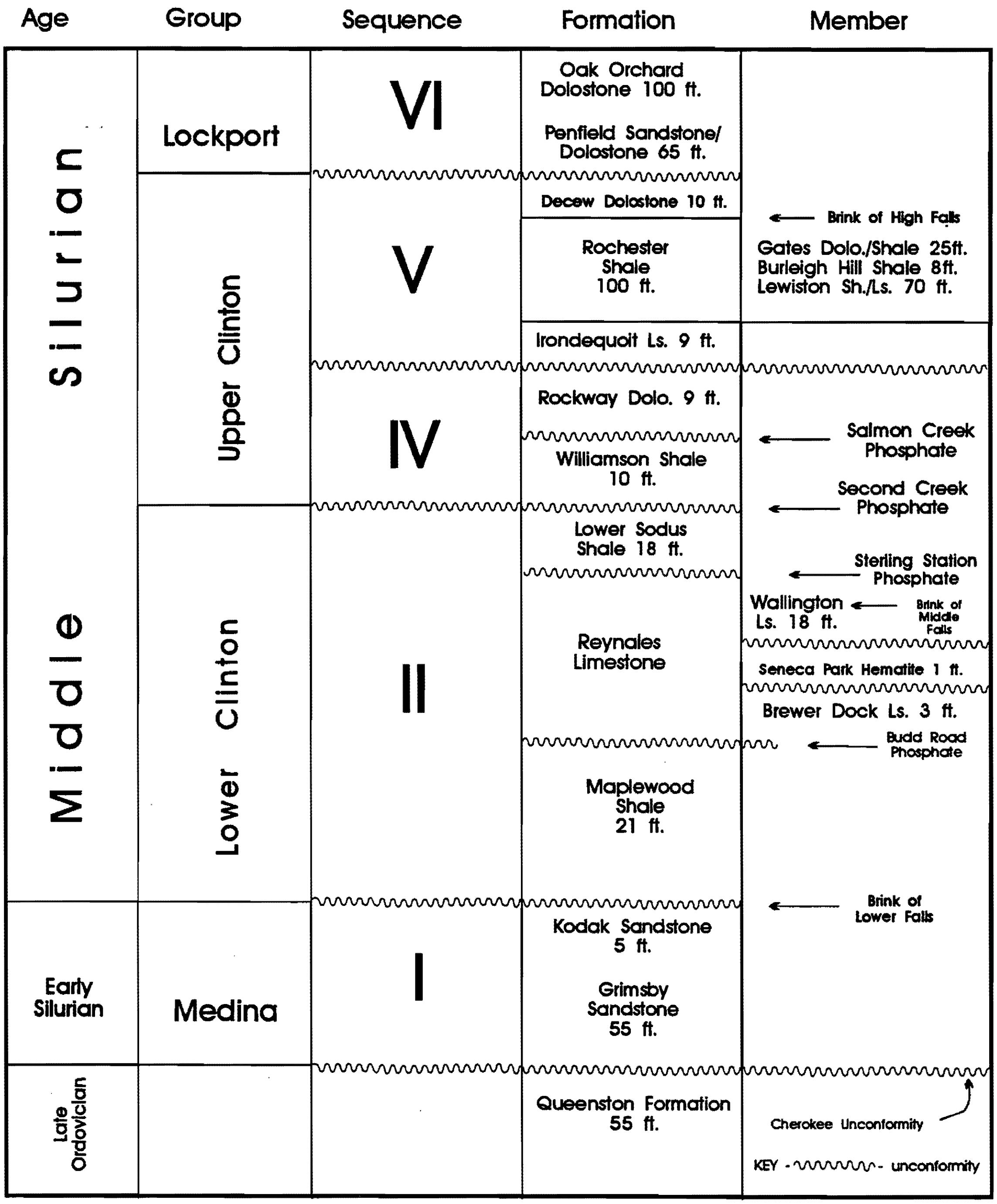

FIGURE 17 Stratigraphy of the Rochester Gorge

[After: C. E. Brett, W. M. Goodman and S. T. Lo Duca. 1990. Figure 4, p. C11. NYS Geol. Assoc. Guidebook, Fredonia State Univ., 62nd Ann. Mtg., modified by T.X. Grasso, drafted by R.D. Hamell, Oct. 1991]

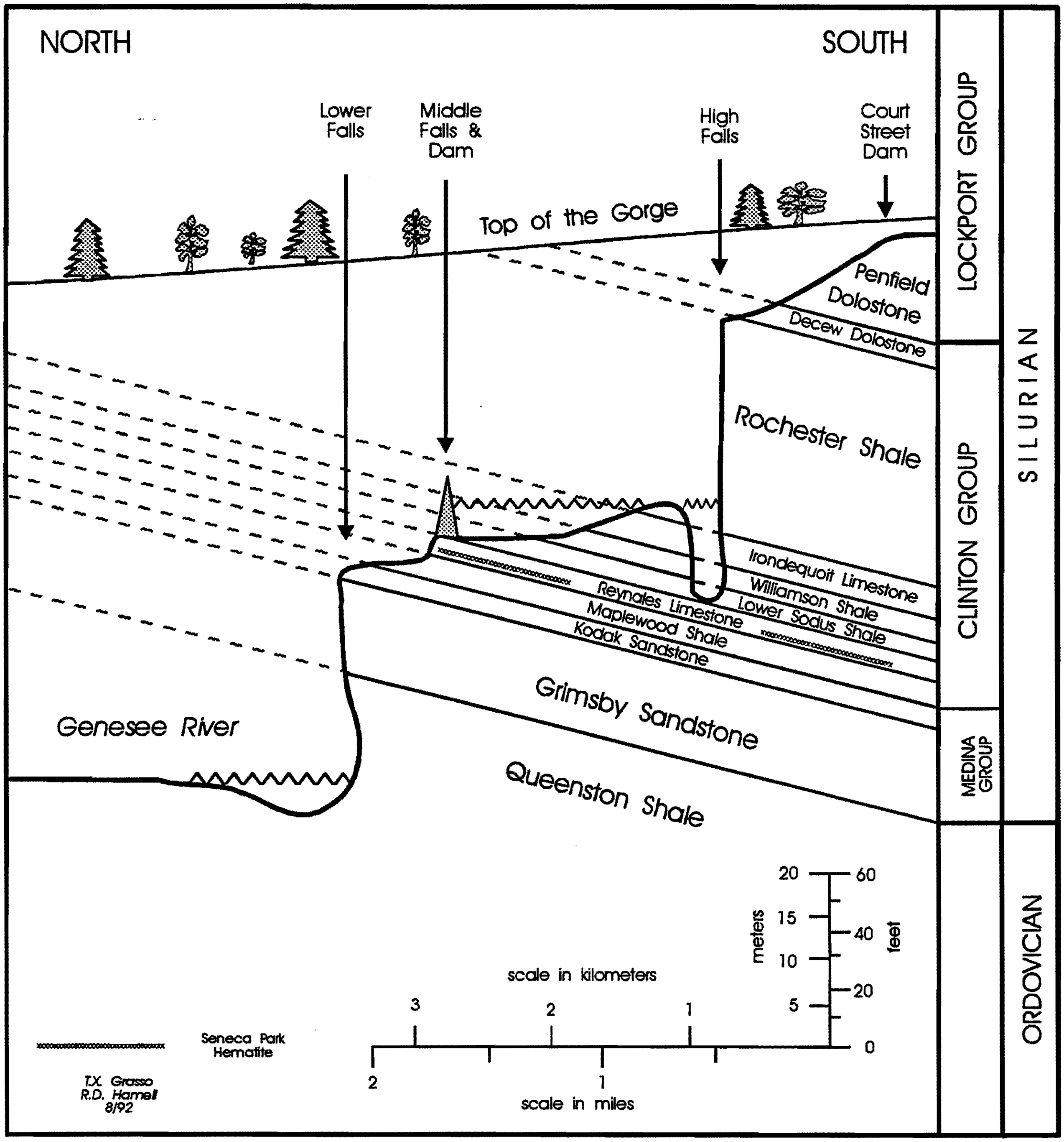

FIGURE 18 Stratigraphic Profile of the Rochester Gorge

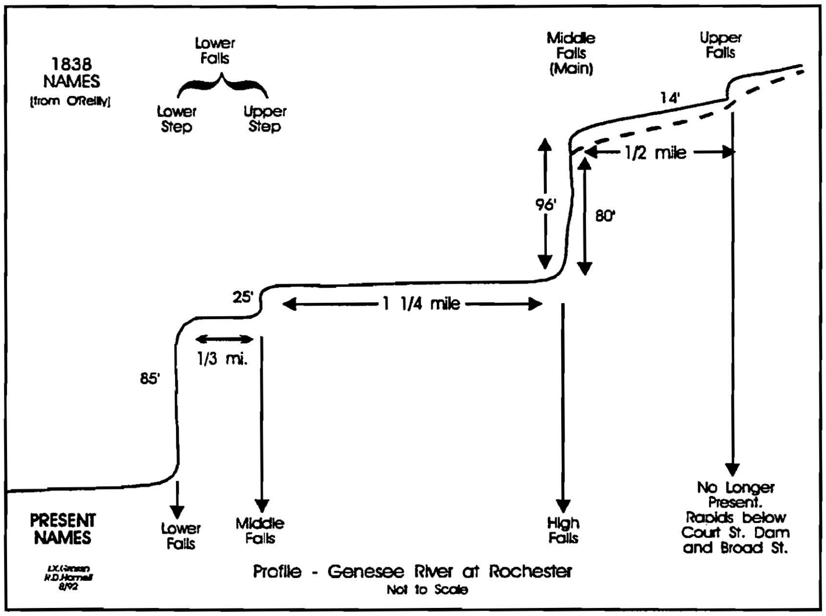

FIGURE 19 Old and current names for the waterfalls in the Rochester Gorge. Dashed line upstream from High Falls is approximate present day profile.

FIGURE 20 East side of gorge just downstream from the Lower Falls. Rock Units: 1=Queenston, 2=Grimbsy, 3=Kodak, 4=Maplewood, 5=Reynales, 6=Lower Sodus, 7=Williamson, 8=Irondequoit, 9=Rochester. (Bill Clar)

continuously to the lip of the High, Main, or Upper Falls just north of the Inner Loop and Conrail bridges and south of the Pont de Rennes Bridge over the river. At one time this falls was called the Middle Falls.

From the brink of the High Falls to the Court Street Dam the Lockport Group is exposed in banks and bed of the river and once formed a small cascade of about 14 feet that stood where the Broad Street Bridge (second Erie Canal aqueduct) is today. This small cataract was, before canal construction, the Upper Falls of the Genesee. Over the years the change in terminology for Rochester's three waterfalls (originally four) led to much confusion. Going downstream from downtown, the original Upper Falls no longer exists. The old Middle Falls is now the High (Main or Upper) Falls, while the Upper Step-Lower Falls is now the Middle Falls and the Lower Step-Lower Falls is currently known as the Lower Falls (Figure 19).

Depositional Models

From the time the strata in the gorge were first examined by Amos Eaton in 1823, numerous models have been advanced to explain how the rocks were deposited. A relatively recent one, is the concept of sequence stratigraphy (see Brett et al., 1990 for a comprehensive discussion). The basic tenet of this model is that the sedimentary rocks in the gorge can be grouped into genetically related packages (called sequences), bounded above and below by unconformities.

A typical sequence begins when sea level is low. Therefore a particular region at that time above sea level is subject to erosion forming an unconformity. Another possibility, at this time, is that nonmarine strata are laid down at the base of a sequence above a regional unconformity. Following this period of sea level lowstand, sea level begins to rise, flooding the region. As the shoreline parades inland, gradually flooding the previously exposed land, the transgressive sequence begins with shallow water (continental shelf) deposits that are subsequently overlain by more offshore deposits, generally limestones. At maximum flooding sediment starvation may result in phosphate nodule deposition above the offshore deposit. The upper part of each sequence typically begins with a deep water, outer shelf or, basin deposit, above the phosphate horizon, which is in turn overlain by successively shallower deposits (a shallowing upward or regressive sequence) as sea level falls. Eventually sea level may fall enough to expose the region to erosion, resulting in the formation of another unconformity. However, erosion may also remove portions of the underlying sequence thereby often resulting in only a partial sequence preserved in the rock record. In summation, sequence stratigraphy presumes to record a rise of sea level at the start of the sequence above an unconformity, or nonmarine strata, followed by a fall of sea level at the end of the sequence. The result is another unconformity at the top of the package followed by the next rise in sea level with its signatory sediments. Sequence stratigraphy is remarkably similar to Amos Eaton's five Series (see Figure 2, p. 7) although on a finer scale. (Geologists are apparently struck from time to time by a recurring illness of seeing regular cycles everywhere in the rock record).

Recent studies of the Silurian of Western and Central New York by Brett, et al. (1990), have revealed six major, unconformity bounded, sequences in the section exposed in the Rochester Gorge (Figure 17). That classification and nomenclature is followed here.

STRATIGRAPHY

Queenston Formation (Grabau, 1908)

During the Middle and Late Ordovician Period, the Taconic Orogeny occurred as eastern North America collided with an island arc (Bronson Hill) in the Atlantic. Eastern New York and western New England, were transformed into Alpine-like mountains as the Taconic Mountains of New York, the Green Mountains of Vermont and Berkshire Hills of Massachusetts were born. Rivers flowing off the newly formed upland eroded the landmass, depositing the sediment in a marine basin to the west. Over time, sediment began to fill the marine basin and a large deltaic complex prograded east to west across the state. The result was the Queenston Formation, 1,000 feet (328 m) of unfossiliferous, fine grained, red, shale, siltstone, and sandstone deposited at the seaward margin of the Queenston Delta. Many of the beds display mud cracks and current ripples, further evidence of its subaerial to intertidal depositional setting. Only the upper 50 to 55 feet (15 to 16.7 m) of the Queenston is exposed in the gorge, comprising the lowest strata below the Driving Park Bridge. Following deposition of the Queenston, an interval of emergence prevailed during which a major erosion surface developed. This erosion surface, widely traceable in eastern North America, was called the Cherokee Unconformity by Dennison and Head (1975).

Sequence I

Medina Group (Luther, 1899)

Grimsby Formation (Williams, 1919)

The Grimsby or Red Medina Sandstone represents the return of terrestrial, deltaic and intertidal deposition to the Rochester area following the erosion that formed the Cherokee Unconformity. This 50 to 55 foot-thick (15-16.7 m) red sandstone sharply overlies the Queenston, its lower portion made even more conspicuous by a 10 foot (3 m) thick bed of massive sandstone, which may equate with the Whirlpool Sandstone further west. The Queenston and Grimsby together make up most of the exposures in the Rochester Gorge north of the Lower Falls, their red color a striking contrast to the gray, green and buff colors of the overlying Clinton Group shales and limestones. The Grimsby, especially in the middle of the unit, contains numerous worm burrows and trails--one a U-shaped tube *(Daedalus),* and the other a corrugated, criss-crossing, trail *(Arthrophycus).* These fossils, along with sedimentary structures such as wave and current induced ripples and mud cracks on some bedding surfaces, indicate a deltaic origin for these strata.

The Grimsby Sandstone is famous as a fine building stone, and was quarried below the Lower Falls, early in the city's history. The stone for the first Erie Canal aqueduct over the Genesee River (1822-1823) came from the Grimsby quarried at Carthage (below the Lower Falls). Nearly all of the red sandstone buildings in Rochester are constructed of the Red Medina save for the present City Hall on Church Street (1885-1889) built of Newark Series sandstone from the Connecticut Valley.

Kodak Sandstone (Chadwick, 1935)

The Kodak Sandstone is currently placed at the top of the Medina Group by Brett, et al. (1990) although older nomenclature places the Kodak at the base of the Clinton Group. This stratum of gray sandstone is about 5 feet (1.6 m) thick, lies directly above the Grimsby, and forms a prominent stripe on the sides of the gorge downstream from the Lower Falls (Figure 21). In Amos Eaton's 1823 field book he names it the Gray Band (1824). The Kodak caps the Lower Falls and represents a near shore deposit, the initial but minor transgression of the late Sequence I sea. The remainder of Sequence I strata were removed by erosion that took place when sea level dropped at the end of Sequence I time.

Sequence II

Lower Clinton Group (Vanuxem, 1839)

Maplewood Shale (Chadwick, 1918)

Sequence II begins with marine sediments laid down in a shallow subtidal setting, now the 21 foot-thick (7 m) green Maplewood Shale. The Maplewood therefore represents the initial flooding of the Sequence II transgressing sea into the Rochester region. It is well exposed along the access road to the RG&E substation, off Seth Green Drive, just west of St. Paul and Norton Streets (**STOP 4**), and splendidly so on Densmore Creek down from Densmore Road north of Norton Street in the town of Irondequoit.

Reynales Limestone (Chadwick, 1918)

The Reynales Limestone earlier designated the "*Pentamerus* Limestone" by Hall (1843), is a slightly deeper, more offshore, deposit than the underlying Maplewood. Therefore the two units are interpreted as a deepening upward sequence that recorded the rise of sea level at the start of Sequence II. The Reynales is subdivided into three members. In ascending order, they are Brewer Dock Limestone (3 feet; 1 m), Seneca Park Hematite (Iron Ore, 1 foot; 0.3 m) and the Wallington Limestone (17 feet; 5 m). The Brewer Dock is characterized by thin beds of fossiliferous limestone alternating with thin, green, shale beds resembling the Maplewood Shale below. The Seneca Park Hematite, previously known as the Furnaceville from exposures in northwestern Wayne County, is at a stratigraphically higher level than the Furnaceville iron ore in Wayne County and therefore cannot be the same unit (Brett et al., 1990). The Seneca Park is found just below the crest of the 25 foot (8 m) high Middle Falls and along both sides of the gorge downstream from this point to near Kodak Park.

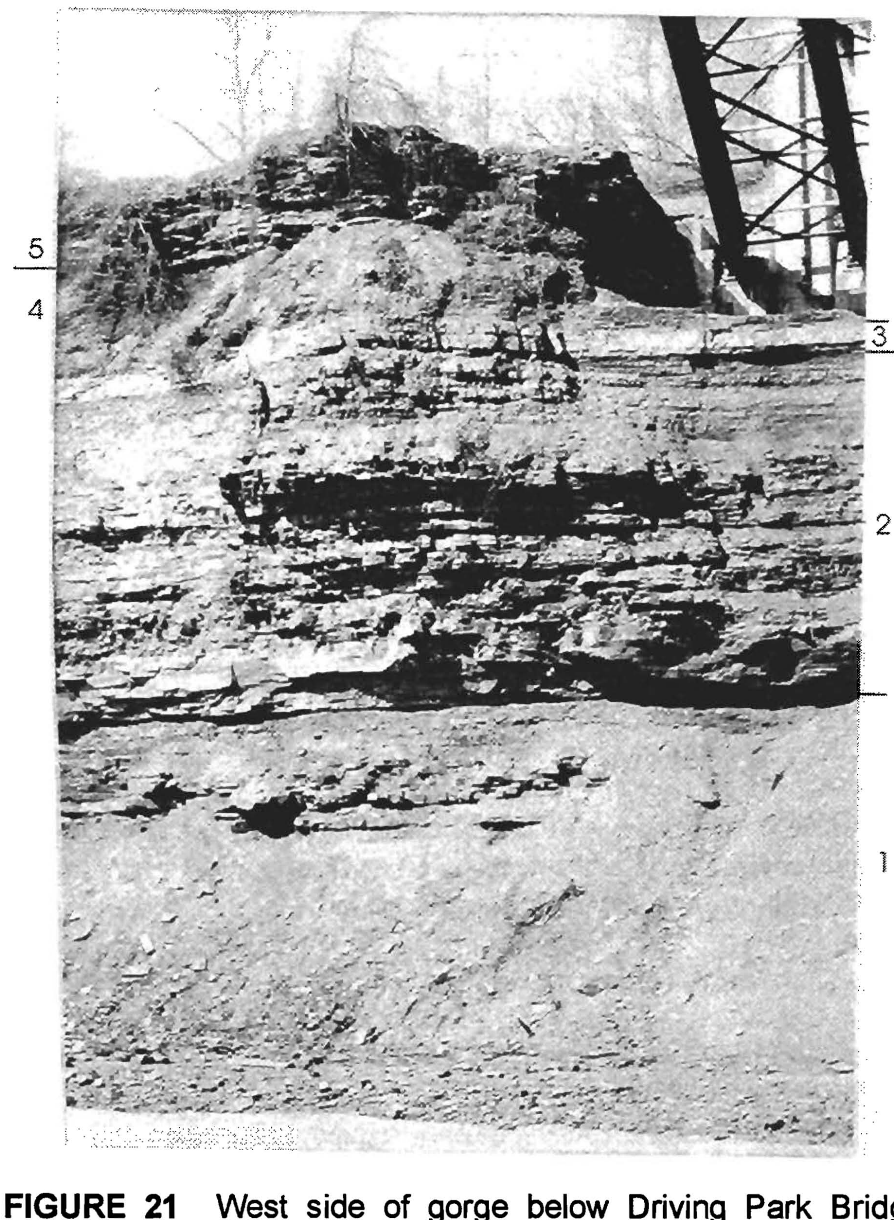

FIGURE 21 West side of gorge below Driving Park Bridge. Rock Units: 1=Queenston, 2=Grimbsy, 3=Kodak, 4=Maplewood, 5=Reynales. (Bill Clar)

The overlying Wallington Limestone was deposited in an open marine, shallow to moderate subtidal, clear, tropical ocean. The evidence for this interpretation is that the rock is coarse grained, some bedding surfaces are ripple marked, but more importantly the unit yields a rich and diverse assemblage of invertebrates, including corals. Some beds, especially in the upper 7 feet (2 m), are packed with masses of the large (2 to 3 inches long; 5 to 7.5 cm), smooth shelled, brachiopod, *Pentamerus* that when alive must have grown like "oyster beds" on the Silurian sea floor. The combination of certain dense strata packed with fossils, interbedded with some containing nodules and stringers of chert (flint) renders the Wallington Member especially resistant to weathering and erosion and, therefore, it forms the cap of the Middle Falls as well as the falls on Densmore Creek down from Densmore Road north of Norton Street (Figures 22, 23).

The area between the Middle and Lower Falls was the site of much industrial activity. At the Middle Falls mills sprouted on each side of the river on mill races beginning above the falls and terminating just below the Lower Falls. Flour, carpet manufacturing, paper and lumber, plus furniture manufacturing were some of the industries that occupied this site from 1817 to 1961.

Upstream from the Middle Falls, near the base of the Bausch Street bridge, several industries occupied the "flats" of the Genesee. Most, but not all, of this acreage was created by fill dumped at the site primarily in the early 20th century. On the east side of the river north of the Bausch Street Bridge, a coal gasification plant was constructed in the late 19th century. It greatly expanded in the early 20th century and eventually extended operations to the west side of the gorge south of Bausch Street. Coal gas was produced by the destructive distillation of bituminous coal for use in heating and perhaps some lighting. The plant ceased production in the 1950's. On the west side "flats" south (upstream) of the Bausch Street Bridge, but north of the gasification plant, the city's waste was incinerated and disposed of at the City Garbage Disposal Plant from the time of World War I to the 1960's. Each of these industries contributed a unique blend of toxic, hazardous, and other waste into the river, that over the years helped to "enrich" its banks not to mention Lake Ontario.

The buried waste from the coal gasification plant that once occupied the Bausch Street site is probably the source of the *creosote* now seeping from bedding planes in the Grimsby Sandstone exposed in the face of the Lower Falls.

Lower Sodus (Gillette, 1940)

Relative sea level dropped after Wallington deposition resulting in the deposition of shallow water, subtidal, green and purple shales of the Lower Sodus Formation. The Lower Sodus, approximately 20 feet (6 m) thick, is particularly distinguished by its "pearly layers", thin (1 to 3 inch thick; 2 to 7.5 cm) beds packed with the lustrous shells of the tiny brachiopod, *Eocoelia*.

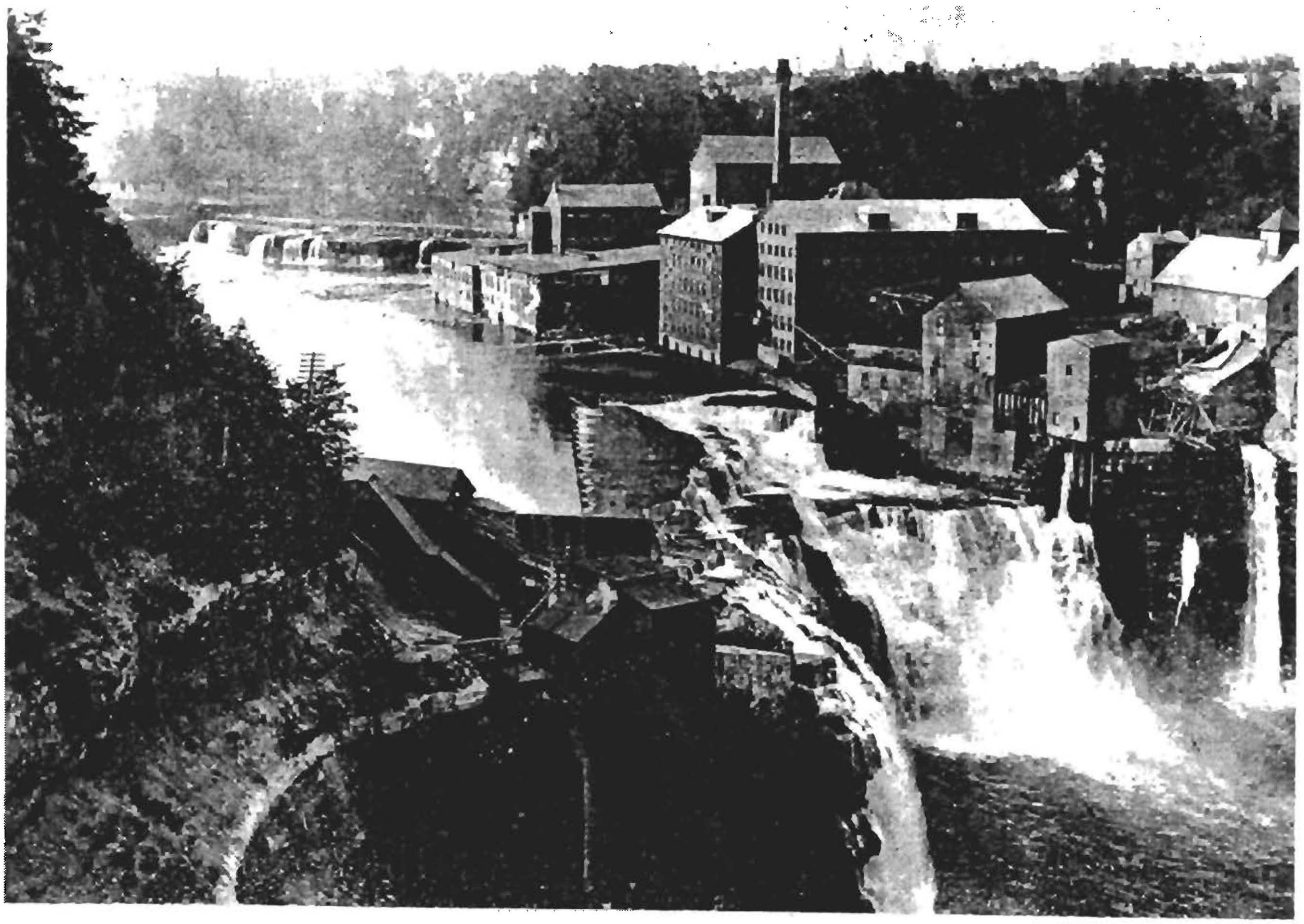

FIGURE 22 Industrial development on the west side of the gorge between the Lower and Middle Falls circa 1880. C.J. Hayden Furniture Co. in foreground, Rochester Paper Co. in middle distance, Middle Falls in distance (Local History Div., Roch. Public Lib.).

FIGURE 23 West side of gorge today, between the Lower and Middle Falls. Compare with Figure 22 (Bill Clar)

Sea level then rose again to deposit other formations above the Lower Sodus Shale at Rochester. However, immediately afterward, a major drop of sea level took place resulting in widespread erosion across western New York thereby bringing Sequence II to an abrupt end. The formations above the Lower Sodus were stripped away at Rochester but are still present in the Sodus Bay region. Therefore the top of the Lower Sodus is marked by a widespread, conspicuous unconformity that not only terminates Sequence II but is a major gap in the rock record where hundreds of feet of strata present in central and eastern New York are absent in western New York.

Sequence III

Middle Clinton Group

The Middle Clinton Group is not present in western New York due to the unconformity mentioned above. However, a narrow seaway persisted in east central and eastern New York where strata of Sequence III were deposited. These rock units will not be described as they are not germane to this paper.

Sequence IV

Upper Clinton Group

Williamson Shale (Hartnagel, 1907)

Overlying the Lower Sodus Shale are 6 to 10 feet (2 to 3 m) of dark, organic rich, black to green, fissile, shales - the Williamson Formation. At the base of the unit is a very thin, phosphatic, pebble horizon named the Second Creek Phosphate Bed marking the regional unconformity. The lithological and biological evidence points to a deep water origin for the Williamson Shale. The Williamson Shale contains few fossils, an indication that environmental conditions were highly stressed Only a sparse bottom dwelling population, if any, could dwell in these deep anoxic waters. Higher in the water column oxygenated conditions prevailed and therefore a thriving community of pelagic organisms, mostly graptolites was able to flourish. When these individuals died they slowly slipped into the murky, black, bottom muds adding their organic matter to what was already there. Graptolite remains blanket some of the bedding plane surfaces in the Williamson Shale.

Therefore, the Williamson represents a rapid rise of sea level commencing with the Second Creek Phosphate Bed. The Williamson accumulated at a time when sea level was at its highest point in the Silurian Period of North America. Toward the end of Williamson time sea level began to fall resulting in a shallowing upward sequence at the top of the unit.

The Williamson and underlying Lower Sodus Shales are exposed on the east side of the gorge north of the Middle Falls and on the west side north of the Rose Garden in Maplewood Park.

Irondequoit Limestone (Hartnagel, 1907)

The Irondequoit Limestone as originally defined encompassed the 18 feet (6 m) of dense, fossiliferous, carbonates that directly overly the Williamson. The lower 9 feet (3 m), now recognized as a separate unit, the Rockway Dolostone (Carlton Brett, personal communication), are fine grained and were deposited in relatively deep water when sea level rose rapidly once again. After this event, sea level fell sharply terminating Sequence IV and producing another regional disconformity.

Sequence V

Sequence V commences with 9 feet (3 m) of Irondequoit Limestone. This unit which sharply overlies the Rockway, is a coarse grained limestone, and contains an abundant and diverse faunal assemblage. The coarse lithology and abundant fossils are consistent with the inference that the Irondequoit was deposited in an inner shelf shallow water environment. This conclusion is further strengthened by the presence of small mound-like reefs or bioherms near or at the top of the unit composed of algae as well as bryozoans. The Irondequoit records the early phase of sea level rise at the start of Sequence V.

The Irondequoit Limestone is exposed in the upper part of the gorge walls near the Lower Falls. In addition it is found on the west side of the gorge, just above the river, upstream from the RG&E Dam on the Middle Falls, and on the east side of the gorge north of the Middle Falls.

Rochester Shale (Hall, 1839)

The Rochester Shale is approximately 90 feet (30 m) thick and makes up nearly all of the gorge walls from the level of the river, above the Middle Falls, to the top of the High (Main) Falls. It comprises all of the rock visible in the gorge below Brown's Race at the newly established High Falls Historic District Park (**STOP 5**). On the gorge's east side the Rochester Shale is well exposed in the banks from river level to a few feet below Upper Falls Park. The brownish colored strata at the very top of the gorge belong to the overlying Decew Formation of the Lockport Group according to Zenger (1965) and Rickard (1975), but uppermost Clinton Group in Brett's (1990) usage (Figure 24).

The lower 20 or so feet (6 to 7m) of the Rochester Shale is highly fossiliferous containing a rich diversity of marine invertebrates including corals, brachiopods, bivalves, snails, cephalopods, trilobites, and echinoderms such as the crinoid *Eucalyptocrinites coelatus* and the rare cystoid *Caryocrinites ornatus*.

FIGURE 24 East side of gorge below Upper Falls. Genesee Brewery in center and left of photo. Rock Units: 1=Lewiston Mbr.-Rochester, 2=Gates Mbr.-Rochester, 3=Decew Fm. (Bill Clar)

The hundreds of invertebrate species found in the lower Rochester Shale combined with the fine grained shale and thin limestone beds in which they are found, indicate deposition in warm, well oxygenated, normal marine waters of intermediate depth. The lower Rochester is, therefore, transitional in bathymetry from the shallow water upper Irondequoit Limestone below to the deep water deposits of the middle Rochester Shale above.

In the past the classic locality for study of the lower Rochester Shale was along Densmore Creek, upstream from Norton Street, east of Culver Road. The unit was exposed above a small cascade over the Irondequoit Limestone. However, a concrete trough or flume was constructed on top of the Rochester Shale thereby rendering it inaccessible. Small exposures of the lower Rochester Shale may be seen along the south side of the Keeler Street Expressway (NY104), just west of the Portland Avenue bridge, and along NY390 at the Ridgeway Avenue bridge.

The succeeding 20 feet (6 m) of the Rochester Shale consists of dark, organic rich, mudstones and shales that are nearly devoid of fossils and suggest deposition in deep, outer shelf water, but not quite as deep as those that prevailed in Williamson time.

The ocean gradually shallowed again as the dark, low faunal diversity mudstone and shale discussed above, passes upward into lighter colored more fossiliferous shale and thin limestone beds of the top of the Lewiston Shale Member (Brett, 1983). These beds form a conspicuous bulge in the middle portion of the east wall of the gorge just downstream from the High Falls (Figure 24). A second deepening event is recorded by the uppermost 33 feet (10 m) of the Rochester Shale. This sedimentary package is characterized by uniform, thin (1 to 2 inch thick; 2.5-5 cm) beds of dolostone and shale of the Burleigh Hill Shale and Gates Members. The dolostone beds have an internal structure that suggests deposition from storm generated events such as hurricanes. The Gates is characteristically even bedded and almost banded. This feature is well displayed in the upper part of the gorge at the High Falls, and also in the rock cut for the entrance ramp to I-390 South from Lyell Avenue (NY31) westbound. The brink of the High Falls today is at the top of the Gates Member.

Decew Dolostone (Williams, 1914)

The Decew Dolostone, was probably deposited in shallow water, therefore, it represents the top of the shallowing upward sequence that began in the underlying upper Gates Dolostone Member. Sea level continued to fall after the Decew was deposited exposing the region once again to erosion. Thus an unconformity was produced on top of the Decew, terminating Sequence V. The Decew is characterized by a distinctive pattern of convoluted and contorted bedding that gives the unit a "ball and pillow" or rounded, concretionary appearance especially on well weathered, vertical surfaces. This distinctive bedding is called enterolithic structure and can be traced west to the Niagara Gorge and into Canada. This internal structure was probably caused by flowing and slumping of the still unlithified Decew sediment, perhaps triggered by a large earthquake that shook the area one day during the Middle Silurian Period 420 million years ago or a meteorite impact!

FIGURE 25 High Falls before upper ledges were removed by blasting. Looking west. Gorsline Building in distance. Nov. 1914. (Stone Collection, Roch. Mus. and Sci. Center)

FIGURE 26 High Falls showing rock debris from blasting. Looking west. Gorsline Building in distance. Nov. 1914. (Stone Collection, Roch. Mus. and Sci. Center)

The Decew weathers to a distinctive tan or buff color which can be seen in the uppermost 6 to 10 feet (2 to 3 m) of strata in the gorge above and just downstream from the Main Falls (Figure 24). It is also is well exposed in cuts on the Inner Loop, along I-490 west from downtown, and at the Lyell Avenue (NY31) interchange with I-390.

The Decew once formed the brink of the High Falls. However, between 1913 and 1919, under a P.W.A. (Public Works Administration) project for flood abatement and protection, the river bed was lowered from a point upstream near Broad Street to the present brink of the High Falls. The blasting lowered the High Falls approximately 10 feet (3 m) and removed the Decew Dolostone caprock from the lip of the falls (Figures 25, 26). A second P.W.A. project from 1936 to 1938 deepened the river once again in order to install a new dam at the Central Avenue Bridge which today can be seen at the upstream side of the Inner Loop Bridge over the river (Howe, 1936). Today the High Falls is approximately 80 feet (24 m) high according to direct measurement and according to elevation data supplied by RG&E, not 96 feet (29 m) as shown in Eaton's 1823 Journal C sketch (Figure 14) and stated by Henry O'Reilly in his 1838 *Sketches of Rochester* plus countless publications since. Compounding the confusion of the falls' height is the questionable accuracy of Eaton's and O'Reilly's original measurement.

Brown's Race was blasted through the Decew Dolostone and upper Gates Member of the Rochester Shale late in the Fall of 1817 by Francis and Matthew Brown. By 1818 several mills were in operation on Brown's Race, and with the opening of the Erie Canal to the west side of the Genesee River in 1823, Brown's Race enjoyed the fruits of the initial boom years. Many industries flourished here and by 1879 Brown's Race generated 3,760 H.P. Francis and Matthew Brown rebuilt, in 1818, the first and original Harford Mill which stood on the tract in 1807, renaming it the Phoenix Mill. One third of the building still stands, formerly the Lost and Found Tavern, later Whispers and now the Public House. For details of Brown's Race mills and industries see Rosenberg-Naparsteck (1988).

Sequence VI

LOCKPORT GROUP (Hall, 1839)

Penfield Dolostone (Zenger, 1962)

The initial deposit of Sequence VI, the Penfield Dolostone, was laid down, when the sea once again transgressed Western New York submerging the sequence bounding unconformity. This unit is a coarse, sandy dolostone containing wave induced ripples. Some beds within the Penfield display crossbedded laminae. The Penfield Formation was most likely a shallow water deposit as, in addition to its coarse grained sediment and sedimentary structures many of the fossils, mostly echinoderm stems, are broken and fragmented. The Penfield Dolostone is exposed in the bed and banks of the river upstream from the High (Main) Falls, to the Court Street Dam (**STOPS 2 and 3**), and in the Barge Canal cut on the west side of the city.

The upper part of the Penfield once formed a small cascade of 14 feet (4 m) between what are now the Court Street and Main Street bridges. This cascade, as described earlier on page 166 and Figure 19, was the original Upper Falls of the Genesee at Rochester. It was first blasted away to make room for the foundation of the first Erie Canal (Clinton's Ditch) Aqueduct over the river, completed in 1823, but was further altered by the two subsequent P.W.A. deepenings.

In 1789, at the original Upper Falls of the Genesee, Ebenezer (Indian) Allen, the first white settler in what is now downtown Rochester, established a saw and grist mill on the west side of the river. The site is now occupied by the Lawyer's Cooperative Publishing Company and Aqueduct Park (Figure 27). The mills of Allen had not prospered, probably because there weren't enough settlers in the region to support them. By 1807 little remained of the first mills at the Upper Falls. The Genesee crossing had not yet come of age. Eventually, in 1803, Nathaniel Rochester, William Fitzhugh, and Charles Carroll purchased Allen's 100 Acre Tract site, which had by now gone through several hands. In 1811, Rochester surveyed the tract and subdivided it into mill and residential lots which he then put up for sale.

Eventually two mill races were blasted out, one on each side of the river. The one on the west side came to be known as the Rochester, Fitzhugh, and Caroll Race (hereafter, in text and figures, referred to as the Rochester Race) and it extended north to Main Street. By 1817 several mills were in operation on it such as the Red Mill erected in 1814 near Main Street. The one on the east side is called the Johnson and Seymour Mill Race as it was blasted by Elisha Johnson and Orson Seymour on July 4, 1817 while ground breaking ceremonies for the Erie Canal were taking place in Rome, New York. Water still cascades from this race to the river beneath Rundel Library. The Johnson and Seymour Race once continued north beneath the present Convention Center, Holiday Inn and Old Rochesterville to the Main Falls.

Both mill races provided power for many industries on both sides of the river north and south of the canal, but primarily north. Flour, farm implements, lumber, machine shops and foundries were some of the industries that flourished here in the 19th century.

There were five water power sites and mill races at Rochester (from downtown going north): the Rochester Race, Johnson and Seymour Race, Brown's Race, Third Water Power, and Upper Step-Lower Falls. Of the five the Brown's, Johnson and Seymour, and Rochester Races dominated the flour and industrial output of the city, as they were located closest to the Erie Canal. Collectively these sites formed the industrial and manufacturing core of Rochester thereby providing the economic underpinning of the city's rapid growth (Figure 28).

Wheat, floated down the Genesee River, supplied the grist for Rochester's mill stones. Early in the canal era, new mills were in operation on this city's three main races and in the first 10 days of the 1823 canal season, 10,450 barrels of flour were shipped eastward (McKelvey, 1949). In addition, 58 boats departed Rochester in this interval while 45 boats arrived, unloading, among other significant items, 4,000 gallons of beer and 2,300 gallons of whiskey at Gilberts Basin alone (behind the Old Stone Warehouse at South and Mt. Hope Avenues).

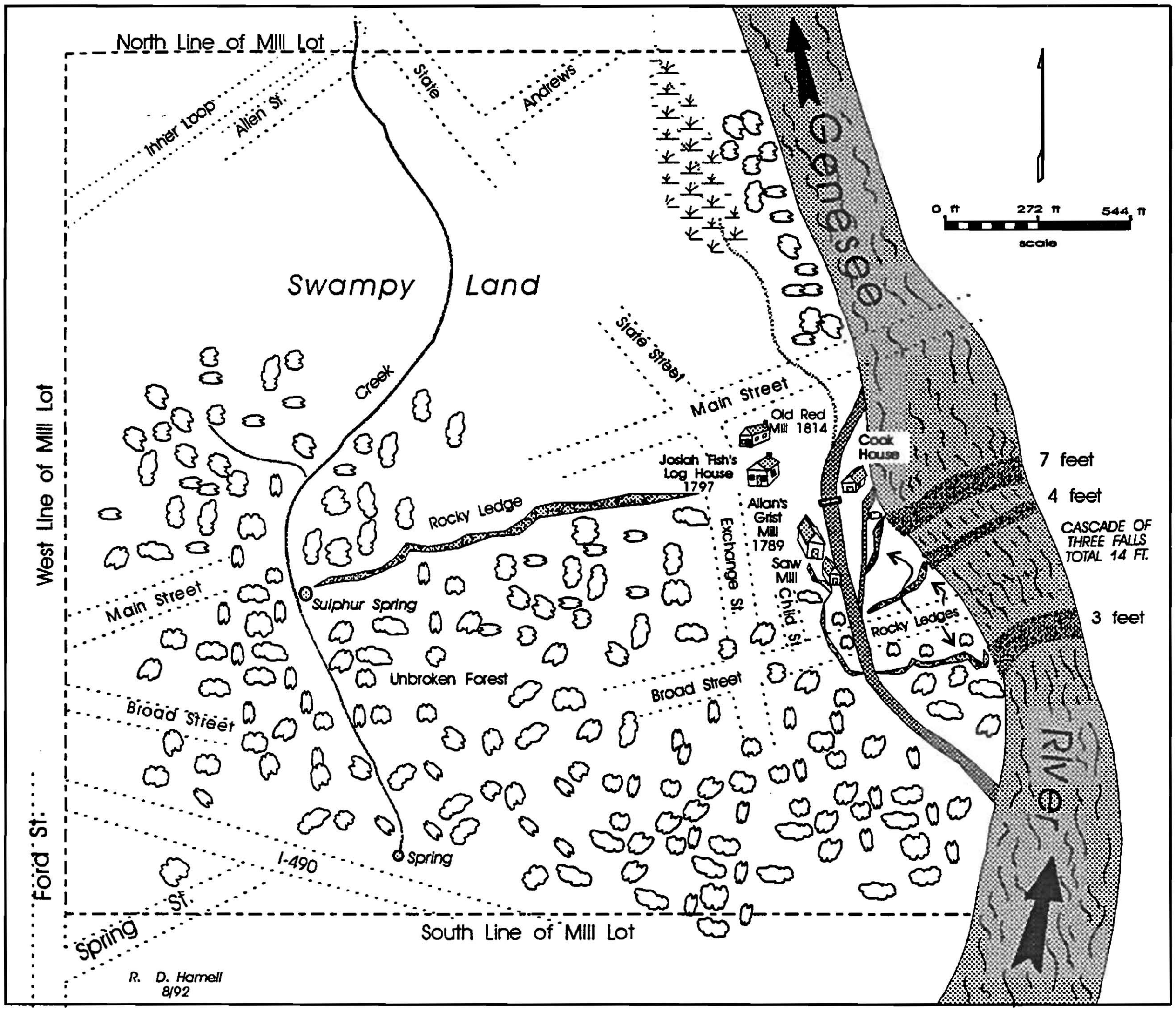

FIGURE 27 An Early Map of the 100 Acre Tract-Unsettled Rochesterville, circa 1789. [Redrawn from McKelvey, Blake, 1979, *Panoramic History of Rochester and Monroe County, New York,* p. 19]

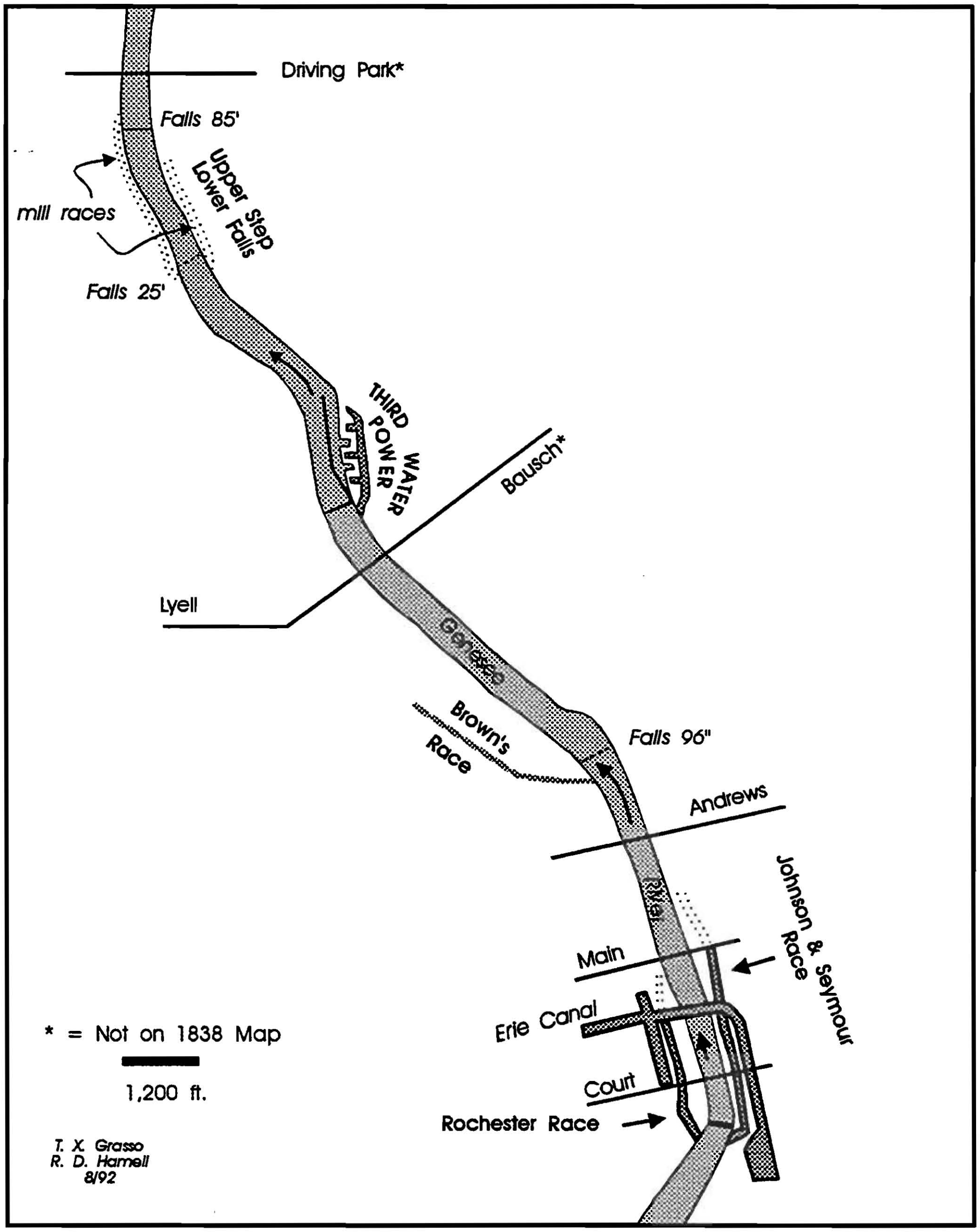

FIGURE 28 Rochester's mill races. Modified from S. Cornell, 1838: O'Reilly's *Sketches of Rochester.*

One can perhaps safely conclude that the rapid growth and industrial boom of the early years fueled an analogous but nearly unquenchable thirst for ardent spirits!

Oak Orchard Dolostone (Howell and Sanford, 1947)

Upstream and south of downtown the Oak Orchard Dolostone (Guelph), the uppermost formation of the Lockport Group, was once exposed. It formed a small rapids in the river and was a common fording point for travelers during low water. The rapids were located just north of the Interfaith Chapel on the University of Rochester campus. From the west side of the river the rapids were located opposite the point where Brooks Avenue joins Plymouth Avenue. The settlement that grew up at the intersection of Brooks Avenue and Genesee Street was known originally as Castletown or "The Rapids" or the "Genesee Rapids" (Rosenberg-Naparsteck, 1992).

By 1822 the State erected a dam at the rapids and constructed the Genesee Feeder on the east side that joined the river, just above the dam, with the Erie canal behind the Old Stone Warehouse (South and Mt. Hope Avenues). This feeder canal conveyed Genesee River water to the Erie Canal, thereby augmenting and replacing the water supply coming from Lake Erie, as most of the Lake Erie water, by the time it reached Rochester, was lost to evaporation and seepage. By 1918 Barge Canal construction had obliterated the rapids by blasting and by raising the level of the river when the Court Street Dam was put in. The pool thus created provides water for the Barge Canal east of the Genesee River.

The Oak Orchard Dolostone contains numerous vugs lined and/or filled with minerals and crystals such as gypsum, selenite, calcite, dolomite, sphalerite and fluorite. It also possesses a fairly high diversity assemblage of corals and other invertebrates although the preservation is often poor. Its inferred environment of deposition was shallow, subtidal marine but deeper than the underlying Penfield Formation.

The Oak Orchard is exposed in road cuts on I-390 north of the airport, I-490 near the NY 531 (Spencerport) interchange, I-490 and I-590 junction, I-490 near the Penfield Road interchange, and in the Barge Canal cut on the west side of the city which also exposes the underlying Penfield, Decew, and Rochester Shale in that order going northwest to Long Pond Road.

SALINA GROUP (Dana, 1863)

Above the Lockport Group is the Salina Group of late Silurian age. It is composed of shallow water, hypersaline, lagoonal deposits of dolomite, salt, and gypsum combined with nonmarine and occasional marine shales. Sparse and widely scattered exposures of the Salina Group may be found south of the city line to Honeoye Falls, Avon, and Caledonia and along the Erie Canal east of Rochester to Lyons. The mines at Garbutt, southwest of Scottsville, once extracted gypsum from the upper part of the Salina Group.

Pittsford Shale (Sarle, 1903)

The lower 20 to 40 feet (7 to 12 m) of the Salina Group, resting on the Lockport in the Rochester vicinity, is the Pittsford Shale. The Pittsford is a sequence of soft green, purple, red, and black shales, one or several feet thick, interbedded with thin fine grained dolostone beds, a few inches to several inches thick. Sarle (1903) first described the Pittsford Shale and its eurypterid fauna from exposures along the old Erie Canal north of Old Lock 62 and south of the Spring House Restaurant when the canal was deepened to 9 feet (3 m) in the winter of 1897-1898. Ciurca (1990) has redefined the Pittsford Shale restricting the name to just a bed of black shale a few feet thick that contains eurypterids (see description in Roadlog; **STOP 1,** p. 79). At the time of its discovery the Pittsford Shale's eurypterid fauna was the earliest one then known and contained 6 species belonging to the genera: *Hughmilleria, Eurypterus, Pterygotus,* and *Mixopterus.* The remainder of the fauna consisted of a rare graptolite, brachiopod (*Lingula*), bivalve (*Pterinia*), Ostracod (*Leperditia*), xiphosurans (pseudoniscids), and phyllocarids.

The Pittsford Shale is a nearshore more shallow water environment than the underlying Oak Orchard Formation probably representing an estuarine, or bay environment. The eurypterids were possibly gently rafted in from a fresh water source most likely to the north.

Overlying the Pittsford Shale is the remainder of the Salina Group namely the Vernon, Syracuse, and Camillus Formations all poorly exposed in southern Monroe County. They are blanketed by Pleistocene glacial deposits.

LATE PALEOZOIC

The Acadian and Alleghenyan Orogenys of Late Paleozoic age produced the Appalachian Mountains. Relative to the Rochester region, these orogenic forces had their greatest impact in New England, eastern most New York State, and Pennsylvania from Williamsport south. In the Genesee Valley the previously deposited strata were given a gentle dip of $1/2^{o}$ south.

MESOZOIC-CENOZOIC ERAS

Throughout the Mesozoic and Cenozoic Eras the Genesee region was undergoing erosion and the Genesee River gradually came into existence. The details of how the Genesee River came to be may be found in the countless publications of Herman Leroy Fairchild but particularly his 1928 *Geologic Story of the Genesee Valley and Western New York.*

Just before the Pleistocene Epoch, nearly two million years ago, the Genesee River flowed north from Avon turning east at Rush nearly to Honeoye Falls where it continued its northward flow through what is now the Irondequoit Valley as shown on Figure 29. At this time the Genesee River joined the Ontario River somewhere near the center of what is now the Lake Ontario Basin. Had the Genesee River remained in its preglacial course human events of the 18th and 19th century would have taken a drastically different direction. The glaciers of the ice age were about to make some dramatic changes in the landscape of the Rochester region.

GLACIAL HISTORY

Pleistocene Epoch

During the Pleistocene Epoch, numerous "ice ages" affected the Northern hemisphere, separated by intervals of milder, ice free, interglacial episodes. The last glacial event, called the Wisconsinan Stage, was the one that left the most indelible imprint on the face of New York State. Although there is documented evidence of earlier glacial stages, the bulk of the depositional record is from the Wisconsinan time. This late Wisconsinan ice sheet was at its maximum extent 19,000 to 22,000 years ago and covered nearly all of New York State, except for a small triangular area in southwestern New York near Salamanca-Olean called the Salamanca Re-entrant. The present course of the Allegheny River in this area was controlled by the ice margin.

Glacial Moraines

At the south edge of the city stretches a southwardly convex, linear, lobe of hills and knolls approximately 4 miles long called the Pinnacle Hills Moraine. From west to east the main components of the range are Oak Hill, now the University of Rochester campus, Mount Hope Cemetery, Highland Park, Colgate-Rochester Divinity School campus, Pinnacle Hill, Cobbs Hill and several smaller knolls extending east to Winton Road (Figure 30). Pinnacle Hill, (750 feet; [27 m] above sea level) is the highest hill in the complex and rises 230 feet (70 m) above the city plain below. The Pinnacle Hills Moraine is part of the Albion-Rochester Moraine deposited nearly 13,000 years BP, and is the best example of a kame moraine north of the south end of the Finger Lakes.

The Pinnacle Hills Moraine formed when the melting edge of the glacier retreated north from the Finger Lakes region then readvanced south to a position that is now the north slope of the Pinnacle Hills Moraine. When this occurred the rock debris in the ice, ranging from fine clay, silt, and sand to boulders, was dumped at the melting edge as an irregular subaqueous ridge (kame moraine), deposited in a proglacial lake parallel to the ice front. The moraine today is a series of irregularly shaped mounds crudely resembling large "eggs in a basket": Mount Hope Cemetery and Highland Park are good examples of this topography. The Pinnacle Hills Moraine is one of several other major recessional moraines in New York State marking successive, positions of the ice margin.

Recessional moraines may be deposited directly by the ice, in which case they are crudely stratified and composed of glacial till, a poorly stratified mixture of boulders, cobbles, sand and clay. If a moraine is constructed by meltwater flowing off the glacier, it tends to be made up of well-sorted, stratified sand and/or gravel and is known as a kame moraine. The term

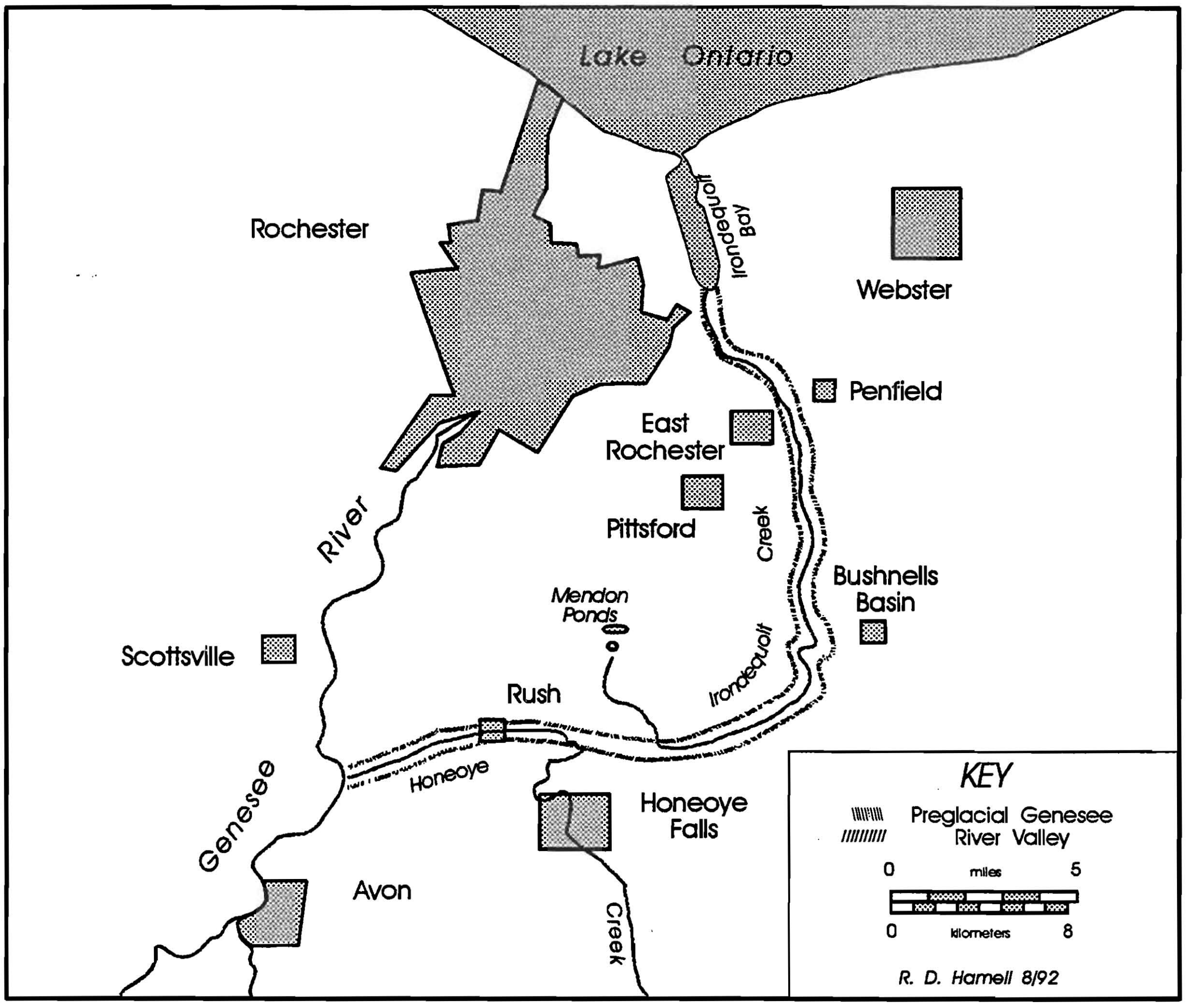

FIGURE 29 Preglacial Genesee River near Rochester

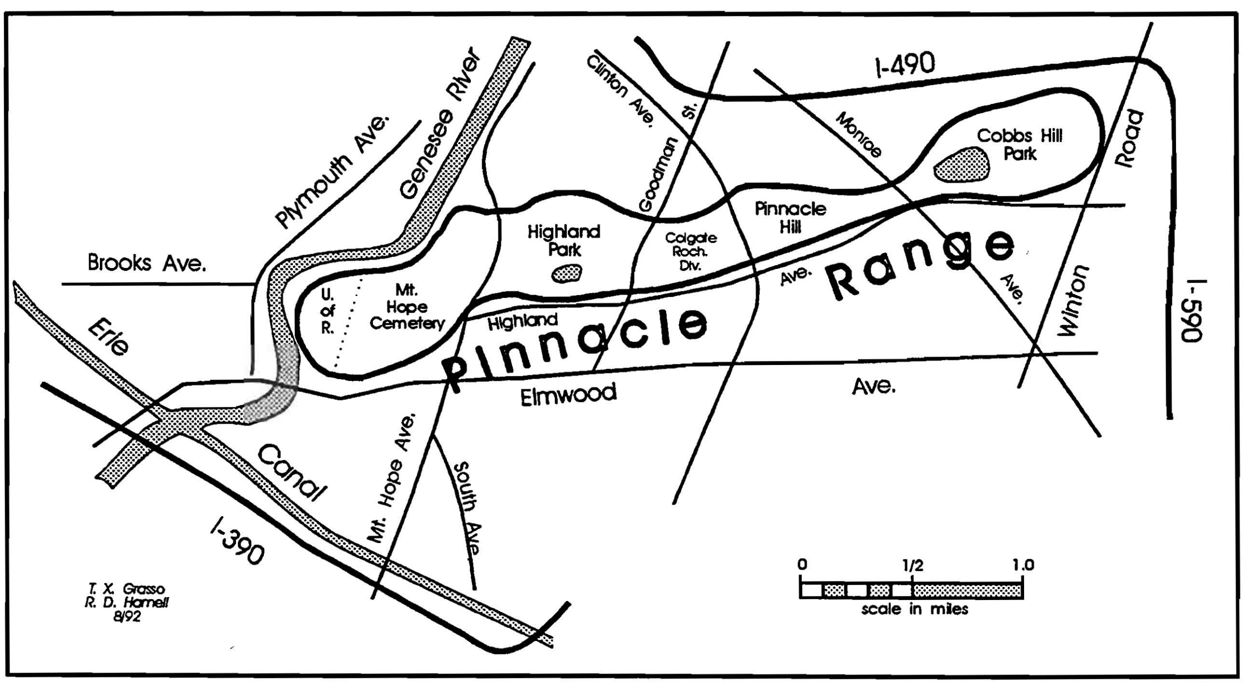

FIGURE 30 Map of the Pinnacle Range

kame has a Scottish origin meaning mound or knoll. The Pinnacle Hills Moraine is primarily a kame moraine, although thin glacial till deposits cap most of the hills in the range. This attests to a minor readvance of the ice which overrode the moraine. The moraine forced the ice to flow over its crest leaving the till behind when the glacier melted. The contorted and faulted strata revealed by 19th century quarries in the Pinnacle Hills Moraine were deformed when the ice advanced over the moraine (Fairchild, 1923).

Proglacial Lakes and the Rerouting of the Genesee River

Proglacial lakes formed when the glacial ice sheet acted as a dam because the terrain in central and western New York slopes to the north. Therefore, as long as the ice occupied the Lake Ontario basin and blocked the Mohawk Valley, the St. Lawrence, and western outlets, waters from northward flowing streams and glacial meltwater were impounded in front of the ice as proglacial lakes. The southern border of the lakes was the high land to the south, while the lake waters to the north impinged against the ice. The glacial lakes found outlets to the ocean either west past Chicago and down the Mississippi or east past Syracuse and Rome and out the Mohawk-Hudson Valleys. In the Genesee Valley fifteen proglacial lakes, at various elevations, existed from 19,000 to 9,000 years BP (Muller et al., 1988).

One of these proglacial lakes was Lake Dawson (13,000 years ago) at approximately 480 feet above sea level, and over 200 feet higher than Lake Ontario. It flooded most of the Rochester area north and east of downtown. The relict shoreline of this lake extends from just north of Ridgeway Avenue, northwest of Rochester, to I-590 just south of the Monroe Avenue interchange, southeast of the city, as shown on Figure 31. Twelve Corners at an elevation of 491 feet above sea level, was less than 11 feet above Lake Dawson's waters (Figure 31).

Contemporaneous with Lake Dawson, Lake Scottsville existed in the Genesee Valley, dammed at its northern end by the Pinnacle Hills Moraine. This lake, at an elevation of 540 feet (165 m) above sea level, was the immediate precursor of the Genesee River, and extended up the valley to Avon. Previous to Lake Dawson time the eastward trending portion of the ancestral Genesee Valley, past Rush and Honeoye Falls, had been filled and buried by glacial deposits making this outlet unavailable for Lake Scottsville. Lake Scottsville's waters drained northward across a sag or saddle in the Pinnacle Moraine at the University of Rochester and did not reoccupy the older valley. The Genesee River flowing due north in its new postglacial course, emptied into Lake Dawson about one mile north of the present High Falls, which then came into existence near the Bausch Street Bridge as the river tumbled over the edge of the Niagara Escarpment forming a series of rapids (Figure 31).

Continued withdrawal of the ice margin farther north exposed a lower outlet than Lake Dawson. The result was a 45 foot (14 m) drop of Lake Dawson's water to form Lake Iroquois at an elevation of 435 feet (133 m) above sea level. As Lake Scottsville completely drained, the mouth of the river shifted north one mile (approximately two miles north

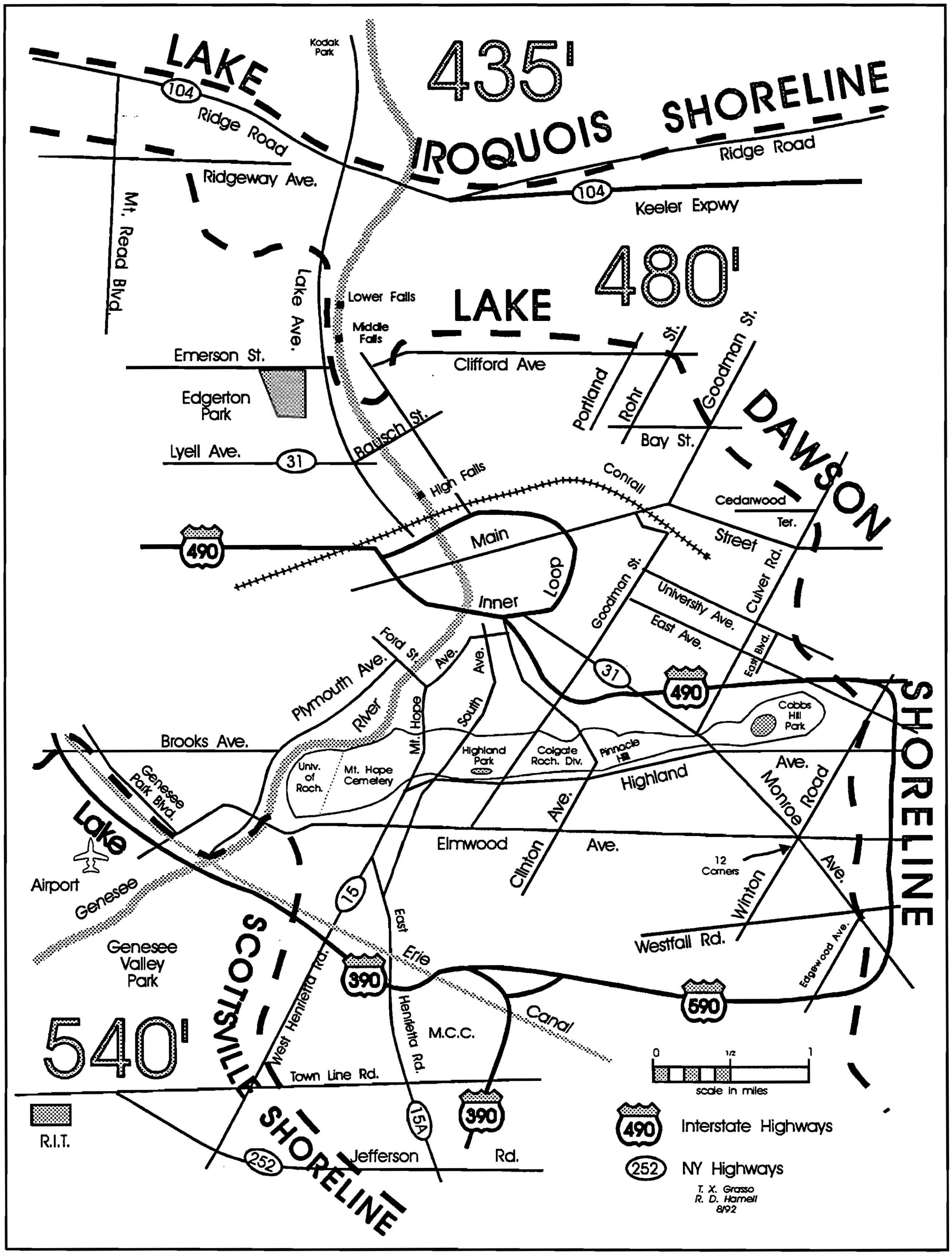

FIGURE 31 Proglacial Lakes in the Rochester Vicinity

of the High Falls) to a point just north of the Veterans Memorial Bridge near Kodak. The High Falls became a cascade just north of its present location and was only 40 feet (12 m) high, compared to nearly 80 feet (24 m) today. Contemporaneously the Middle and Lower Falls emerged as a series of small rapids with the Middle Falls near the Veterans Memorial Bridge and the Lower Falls a short distance north. Once the ice pulled back from Covey Hill in the St. Lawrence Valley, Lake Iroquois lowered to form Early Lake Ontario 50 feet (15 m) above sea level approximately 10,000 years ago. The region which had been depressed earlier by the ice's weight, permitted the existence of a lake approximately 200 feet (61 m) below the present level of Lake Ontario. At this time the mouth of the river extended another nine miles north from its previous location to a position seven miles north of the present southern shore of Lake Ontario at Charlotte, whereupon all three waterfalls were etched into relief and retreated upstream (south) nearly to their present positions. At the same time the lower portion of the gorge, near to and south of Stutson Street, was incised by river erosion. After this time, the region continued to rise due to isostatic rebound, which was greater in the north than in the south. Consequently modern Lake Ontario rose in elevation from the Early Lake Ontario stage and spilled over on its south shore to form Irondequoit Bay, Long Pond and the other bays and inlets along the south shore of Lake Ontario.

In summation the Rochester Gorge formed by river erosion carving its way through solid bedrock as lakes first fell then rose beginning approximately 12,600 years ago. Rochester's most remarkable and certainly its most significant feature had come into existence.

SUMMARY AND CONCLUSIONS

As we have seen, Rochester's preeminence as a water powered industrial giant did not begin with Indian Allen or the purchase of the 100 Acre Tract by Rochester, Fitzhugh, and Carroll. It was preordained for greatness by its long geological history. Back through the arch of time, one can glimpse the seeds of its growth in the deposition of the resistant Lockport Dolostone and its subjacent Clinton and Medina Group strata. The Silurian seas that rose and fell resulted in a depositional package of sediment that contained rock units of varying resistance to the destructive force of erosion. The Grimsby-Kodak Sandstones, Reynales Limestone, and Lockport Dolostone are resistant, while the Queenston, Maplewood, Sodus-Williamson, and Rochester shales are relatively weak. All were uplifted near the end of the Paleozoic Era imparting a southerly dip to the strata and after millions of years the Niagara Escarpment was etched into relief.

Leaping forward to the Pleistocene Epoch, the Pinnacle Hills Moraine temporarily dammed proglacial Lake Scottsville which was prevented from escaping eastward past Rush because the old channel was choked with glacial debris. As Lake Dawson dropped to the Lake Iroquois level, Lake Scottsville drained north across the Pinnacle Hills Moraine forming the Rochester Canyon; first as a series of rapids north of the Bausch Street bridge and then later as the well defined Upper Falls. In a similar fashion, but later in time, a series of rapids farther north would eventually evolve into the Middle and Lower Falls. Finally, the ice melted back and Lake Iroquois fell to form Early Lake Ontario below the present level of Lake Ontario. The Rochester Gorge became deeply incised through its entire length. As Early Lake Ontario waters rose to form modern Lake Ontario the lower gorge from Charlotte to the base of the Lower Falls was flooded.

And so, the rapids and waterfalls came into existence; namely "The Rapids" at the University of Rochester campus at the top of the Niagara Escarpment, the original Upper Falls downtown near Broad Street over the Penfield Formation, the High (Main) Falls over the Gates Member of the Rochester Shale (Figure 32), the Middle Falls over the Reynales Limestone, and the Lower Falls over the Kodak-Grimsby Sandstones. These majestic cascades and waterfalls lay waiting thousands of years for people to one day take up their inheritance. Though humans have nearly spoiled it through neglect, abuse, oversight and overuse, we have finally begun to dimly comprehend how precious, how significant, how important and how beautiful the river and its canyon are for us today. Let us hope that we never forget!

ACKNOWLEDGMENTS

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historical significance. We will attempt to recreate that portion of the trip through western Wayne and Eastern Monroe counties to Rochester. In addition, Sir Charles Lyell traveled through New York State in 1841 (Lyell, 1845), visiting both Amos Eaton and James Hall. Our itinerary therefore incorporates some of Lyell observations.

Eaton's Trip - Syracuse to Rochester 1826

The canal route west from Syracuse follows a series of swampy meltwater channels through Camillus, Jordan, and Port Byron to Montezuma. The melancholy travelers described the "tedious period" as they were "plunged into a swamp of white cedar, pine, hemlock, etc." and there was "nothing cheerful or amusing in it. . . spirits are uncommonly dejected."

The Eaton party arrived at Jordan on Thursday, May 11, and noted that the Saliferous rock [Salina Group] is found nearby and that it persists nearly to Rochester. Also at Jordan there was a Clinton's ditch lock, with an eleven foot lift to the east, located just west of the aqueduct over Skaneateles Creek. George Clinton (p. 287) noted that "the lock and aqueduct are made of a very coarse grained limestone containing terebratulites." However on the return trip Thursday, June 1 he wrote (p.300):

> "In going up I was told by Doctor Eights that it was a coarse grained limestone he having dissolved it entirely. Seeing him so confident and not having an acid by me, I was fool enough to take it upon trust, maugre the evidence of the senses. It is nothing more or less than a sandstone....." [Oriskany Sandstone with *Hipparionyx*]

Today some of these original Clinton's Ditch building stones line the walls of the realigned Enlarged Erie (1845), just west of the enlarged aqueduct and now a village park in Jordan.

A short distance north of the Enlarged Erie aqueduct at Montezuma, the line of Clinton's Ditch crossed the Seneca River on grade, the horses towed the boats across the river on a long towpath bridge. Some poles of this towpath bridge yet remain. Eaton's class crossed the Seneca River on Thursday, May 11, and Asa Fitch described how "the towpath is built on 130 bents . . . in a state of rapid decay and will need rebuilding in a few years." Rezneck (1959) interpreted the poorly handwritten word "bents" as "boats" and thus thought the Seneca crossing to be a pontoon or float bridge.

West from Montezuma the canal follows the bed of Lake Iroquois and, as stated earlier, from Lyons to Fairport the Fairport meltwater channels. George Clinton (p. 287) wrote:

> "We put up for the night at a lock about 6 miles beyond Montezuma [Lockpit]. During our journey we observed several water-snakes, one of which had a small catfish in his mouth, and although chased about for some time preserved his hold until being knocked on the head by a pole, he sunk......we have had recourse to fishing in order

to pass away pleasantly this tedious period. The thing once commenced, of course it was not long confined to such narrow limits......the fish heretofore caught consisted of suckers, dace, and a small fish exactly resembling........the one called *shiner* by the Albanians. This day (Friday, 12th) I caught a yellow perch. We stopped for the night at Palmyra. During the day we observed large numbers of the *Anadonta marginata* [bivalvia] floating on the canal......

Saturday 13. Stopped at Bloss's within 5 miles of Rochester" [Brighton at Jct. of I-490 and I-590].

Asa Fitch recorded the same section of the trip as follows:

"Thursday May 11.....The river [Seneca] has scarcely any current and is not over 4 feet deep.

From here all the way to Buffalo mile boards are put up. From here to Lockpit 6 miles west-the canal is at present on a level with the surface of the river. Alluvial formations are marly clay and black marl. Had to stay on board again at night no tavern for 5 miles of Lockpit.

Friday May 12.....Left boat took a walk collected botanical specimens and thought of home......Ate dinner at Lyons.....considerable business is done here. Messers Hulbert and Root remain to give lectures on Botany and Chemistry. The former is clever fellow rather green in his manners having had but little acquaintance I suppose in company. But he learns with eagerness.....Took supper at Newark. Stayed for the night at Palmyra-slept on shore.

Saturday May 13 Fine weather......finally have gotten through immense swamps of the canal. Beautiful vallies [*sic*] and gently sloping hills [meltwater channel and drumlins]......Pleasant ride to Bushnell's Basin. Few buildings here. Temperature at 2:30 o'clock 86° in a cool shade. High southwest wind blows sand in our eyes [kames and esker] and on our clothes. This deprived us of a good view of the Great Embankment. Put up opposite Bloss's Canal Hotel about 3 miles East of Rochester. I am the daily assistant tomorrow of which I have to bring in my bottle for the boatmen which is the practice of the assistants each day."

Let Lyell's diary speak for him (1845, p.19):

"In the course of this short tour, I became convinced that we must turn to the New World if we wish to see in perfection the oldest monuments of the earth's history, so far at least as relates to its earliest inhabitants. Certainly in no other country are these ancient strata developed on a grander scale, or more plentifully charged with fossils; and, as they are nearly horizontal, the order of their relative position is always clear and unequivocal."

Lyell, jointly with James Hall, examined the drumlins in the Rochester area. Lyell had seen similar features in Sweden, and he wrote (1845, p.24):

"Geologists are all agreed that these and other similar ridges [the drumlins] surrounding the great Canadian lakes, and occurring at different heights above them, were once lines of beach surrounding great bodies of water. Whether these consisted of lakes or seas, -- how the water came to stand at so many different levels, and whether some of the ridges were not originally banks and bars of sand formed under water, are points to be explored."

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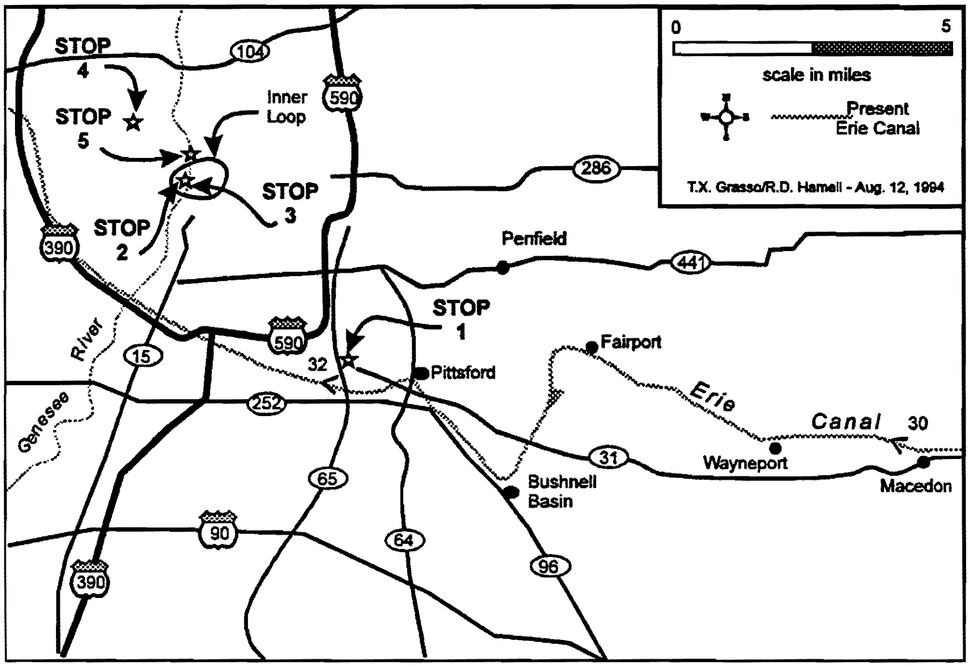
September. 1843.

[From D. Veeder, 1968, *The Original Erie Canal at Fort Hunter*; Fort Hunter Canal Soc., NY, p. 10]

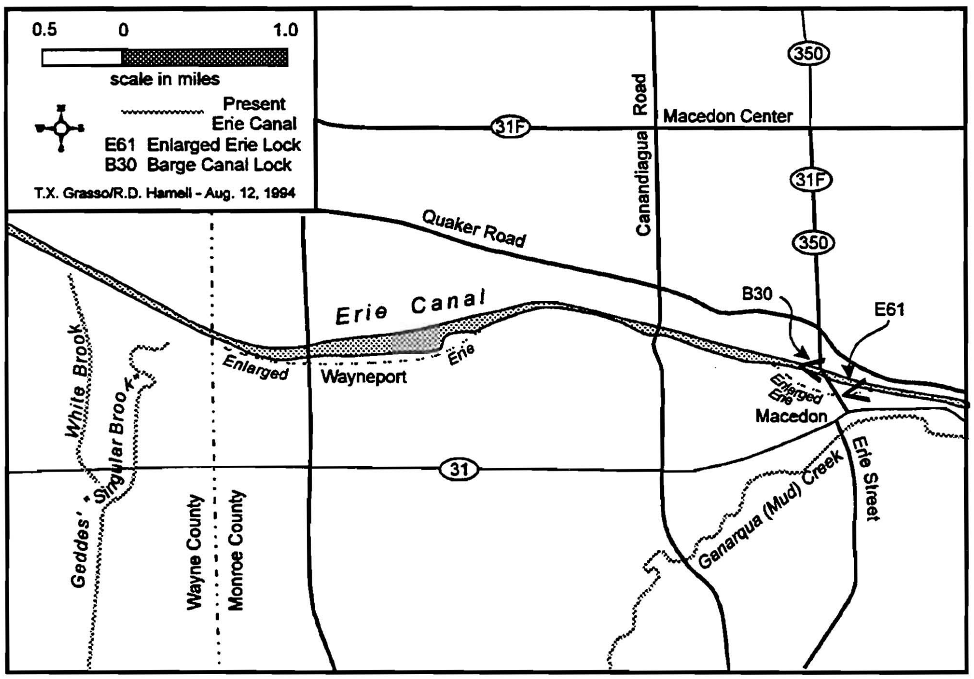
ROADLOG



MAP 1 Field Trip Stops

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
		Begins at intersection of Wilson Blvd. and Elmwood Ave.
0.00	0.00	Proceed east on Elmwood to I-490
2.00	2.00	Pinnacle Hill with TV transmitters on top on left at intersection of Clinton Ave.
4.00	2.00	Jct. I-590 proceed east on Elmwood. I-590 is on alignment of Old Erie Canal.
5.3	1.3	Jct. East Ave (NY96) proceed east on Elmwood.
5.5	0.2	Jct. I-490 turn right (SE) onto I-490.
10.0	4.5	Exit I-490 at Pittsford-Palmyra (Exit 26) interchange. Proceed east on NY31 to Macedon.
11.4	1.4	Cross Erie Canal.

12.6	1.2	Jct. NY250. Proceed east on NY31. Route East to Macedon offers many fine vistas of drumlins and other glacial features.
16.4	3.8	Enter Wayne County - Town of Macedon.
19.4	3.0	Enter Village of Macedon.
20.2	0.8	Jct. NY31F and NY350. Turn left (N) on these routes.
20.4	0.2	Cross alignment of Old Erie Canal. Old Lock 61. Turn left (W) at Lock 30 (Macedon Lock) on the Barge Canal. Lock 30 has a lift of 17 ft. (5 m) West and was completed in 1916.



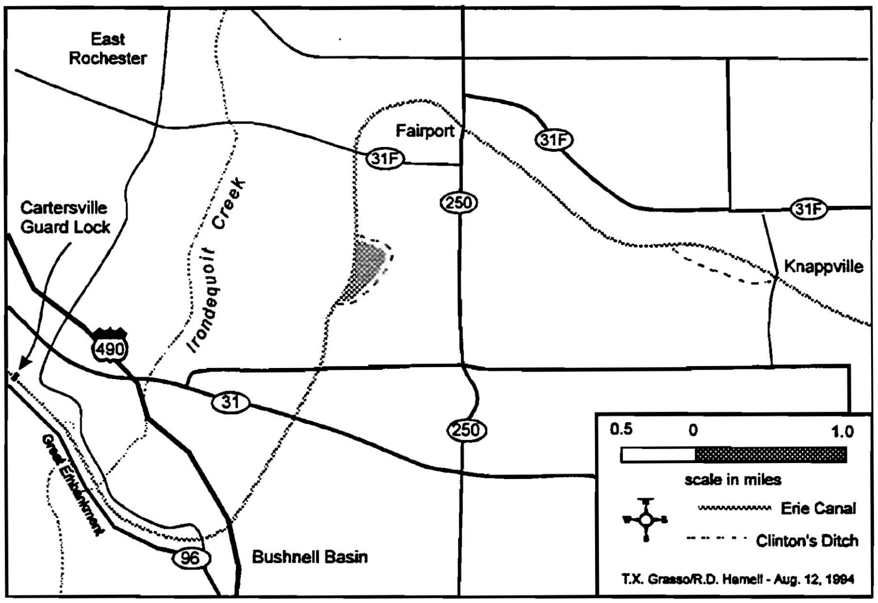
MAP 2 Erie Canal Macedon to White Brook

Board canal boat SAM PATCH for Lock 32 (Pittsford Lock) west of Pittsford. (There is no Lock 31).

<u>The distances listed below are approximate miles from the head of Lock 30.</u>

0.0	0.0	Head of Lock 30. The Erie Canal here partly restored the flow of the ancient Great Lakes (and proglacial) waters through the Fairport meltwater channels. From here to Fairport the canal route follows the main meltwater channel.

1.0	1.0	Bridge over canal.
1.9	0.9	Old Erie Canal channel veers left (S) of the present canal through the hamlet of Wayneport. Start of widewaters created by Barge Canal construction which allowed water of present canal to back up against old canal bed.
3.1	1.2	Wayneport Bridge.
3.4	0.3	Old canal loop enters present alignment - west end of widewaters.



MAP 3 Erie Canal White Brook to Cartersville

3.6	0.2	Monroe County-Wayne County line.
5.3	1.7	Lynden Road Bridge - this site was known as Knappville - a sleepy community in old canal days.
7.0	1.7	Enter village of Fairport.
7.6	0.6	Fairport (Main Street) lift bridge (1914-1915), one of sixteen between here and Lockport and the first one west of the Hudson River on the Erie Canal. Daniel and Henry Deland established the Deland soda and saleratus (baking soda) chemical works on north bank of the canal just east of this

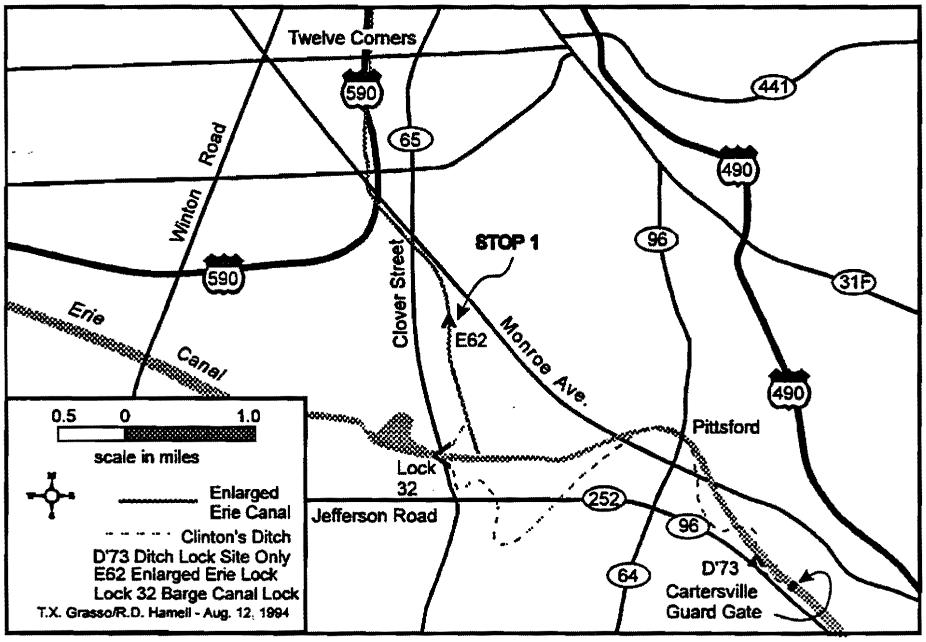
		bridge in 1852. It burned in 1893. The Box Factory and Village Cafe occupy the site.
8.5	0.9	NY31 bridge over canal. This site was known as Fullamtown in Amos Eaton's time as Elisha Fullam, in 1822, built a tavern and warehouse here along with other buildings. Travelers proceeding west to Rochester often transferred to stagecoach at Fullamtown as the overland route was 8.5 miles (13.7 km) shorter.
9.15	0.65	Start of the Oxbow on left (E) - a large loop of Clinton's Ditch that was cutoff when the enlarged Erie was constructed through here circa 1855.
10.3	1.15	NY31 bridge.
11.1	0.8	I-490 bridge.
11.2	0.1	Site of Bushnell's Basin canal break in November 1974. Canal washed out due to collapse of a sewer tunnel being bored beneath the canal. Navigation was not restored until late 1975 - nearly a year later.
11.4	0.2	Marsh Road bridge - Bushnell's Basin.
12.0	0.6	East (SE) end of the Great Embankment. The Erie Canal vaults Irondequoit Valley nearly 60 feet (18 m) above the valley's floor. A culvert nearly 500 feet (152 m) long conveys Irondequoit Creek beneath the embankment. The passage high above the surrounding terrain was sharply curved at its western end (NW) as it followed the serpentine crest of the Cartersville Esker. Dryer (1890, p. 203) described it as a "*kame which extends from Cartersville on the west to and beyond Bushnell's Basin on the east. The north end is a sharp ridge of very coarse gravel, fifty feet in hight* [*sic*], *one mile long and in shape like a rude fish-hook.....Irondequoit river.....has cut the kame completely in two. In the southern portion the gravel is overlain by fifty feet of fine sand which spreads out toward the southeast.....The Erie Canal avails itself of this kame to cross the valley and by a fifty foot embankment restores what probably once existed as a natural feature.*" Upham (1893, p. 190-191) described the feature as part of an esker and kame belt "*.....from a point about a mile east-southeast of Pittsford village to its termination about a mile southeast and south of*

the village of Bushnell's Basin....." Fairchild (1896, p. 135-136) described it fully beginning with ......"*nearly opposite Cartersville and north of Bushnell's Basin is a conspicuous esker.....*" Giles (1918, p. 208-211) named the Cartersville Esker and included a contour map of its northwest portion near Pittsford.

The "rude fish-hook" part of the embankment was deemed too sharp for tugs and barges during Barge construction. Therefore, a new great embankment was built cutting off the esker loop and straightening the canal's alignment. The old loop is still marked by a line of trees north of the present canal.

The new embankment was begun in 1908 and completed in the spring of 1912. The original part of the older embankment over Irondequoit Creek was also modified at this time. In October, 1912, the Great Embankment over Irondequoit Creek collapsed down to the level of the creek in the worst canal breach of its history in this district. Water leaking through the new concrete floor eventually weakened the Irondequoit Creek culvert until it collapsed. Within a few weeks the breach was dammed at each end and a temporary wooden flume erected on piles 60 feet (18 m) high to convey canal boats across the break until the embankment and culvert could be rebuilt. This was accomplished in 1918. The south (SW) wall of the current trough bows outward where the flume used to be as the engineers poured the new wall around the flume. They removed it in the winter of 1917-1918 to pour the new concrete floor. Small tunnels (galleries) were built into the new structure to channel off any leaks and to permit visual inspection. This part of the embankments has held ever since

12.75	0.75	West (NW) end of Great Embankment. When Eaton's party crossed the embankment in 1826 blowing sand prevented them from having a good view of the Great Embankment
13.65	0.9	Mitchell Road bridge. The Clinton's Ditch Pittsford Lock (Lock 73) was located just southeast of this bridge. The LAFAYETTE locked up through here on Saturday May 13, 1826.



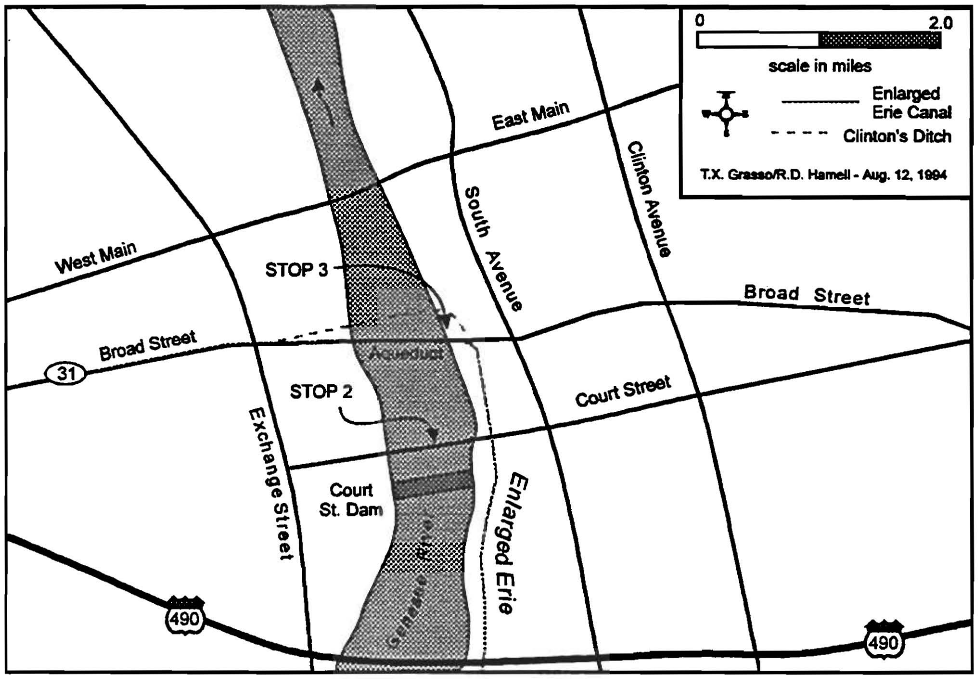
MAP 4 Erie Canal Cartersville to Old Erie Lock 62 (STOP 1)

14.15	0.5	NY31 bridge - Enter Pittsford Between this bridge and the Main Street bridge is the Schoen Place district. In the later part of the 19th century and into the 20th century lumber yards, a malt house, and warehouses lined the south side of the canal while coal sheds, a flour mill, and fruit evaporators were located on the northside. A grain and bean processing mill and Pittsford Lumber Company still survive on the north side amid the rows of boutiques, shops, and restaurants.
14.4	0.25	Main Street bridge
14.6	0.2	Conrail bridge.

The south abutment of the bridge rests on the Barge Canal Bed of the Vernon Formation of Ciurca (1990). It is the youngest of four eurypterid bearing horizons in the lower 100 feet (30 m) of the Vernon Formation. When the canal is drained in the winter the exposure reveals several feet of black shale grading upward into greenish shale. The fauna consists of *Eurypterus pittsfordensis* and rare *Mixopterus* remains.

The other horizons, in order of superposition are the Harris Hill Bed, Monroeav Bed and the Pittsford Bed. Ciurca equates the

"Gananda Bed" of Hamell (1978) with the Barge Canal Bed although the two are 10 miles (16 km) apart.



MAP 4A Map of STOP 1

14.8 | 0.2 | NY31-Monroe Avenue bridge.

The state maintenance shops near the west end of the bridge were built around 1928. Clinton's Ditch ran on the south side of the canal from here to Lock 32.

15.6 | 0.8 | The Old Erie Canal veers north towards Monroe Avenue and downtown Rochester. The present channel from here to Greece was opened in 1918.

15.8 | 0.2 | Cross alignment of Clinton's Ditch in its northward course to downtown Rochester.

15.9 | 0.1 | Clover Street (NY65) bridge and Lock 32.

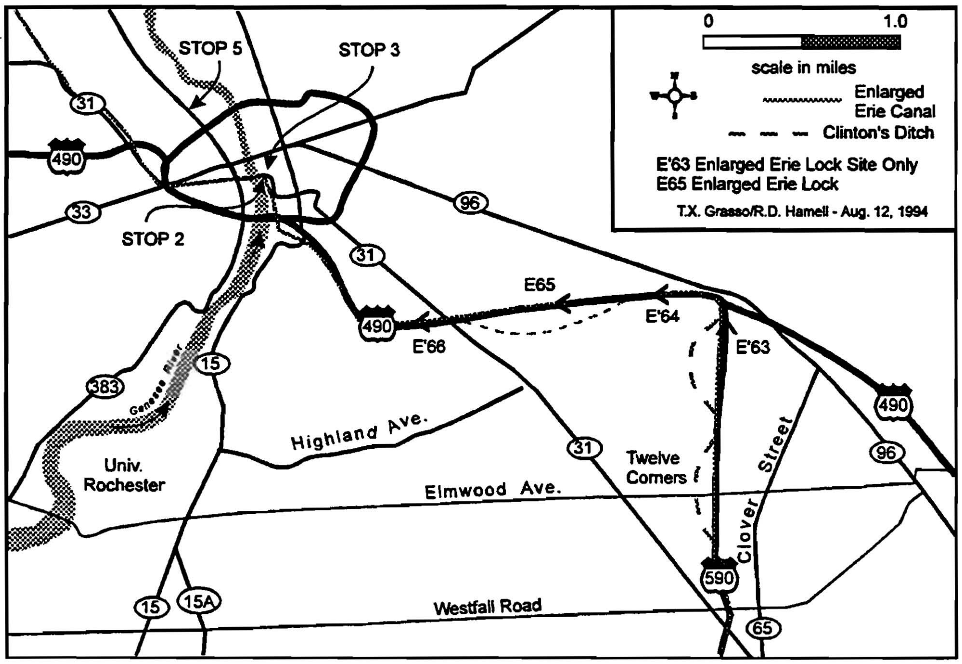
Lock 32 (1917) is the third Pittsford Lock. The first, Lock 73 (1822), was located near the Mitchell Road Bridge; the second, 0.9 mile (1.5 km) north of here is the Enlarged Erie Lock 62 (**STOP 1**). Lock 32 has a lift of 25 feet (7.5 m).

		Board bus - roadlog begins at Clover Street (NY65) and entrance to Lock 32 Canal Park.
0.0	0.0	Proceed north on NY65.
0.6	0.6	Jct. French Road - turn right (E) on French Road.
0.9	0.3	Cross alignment of Old Erie Canal.
		In World War II landing craft were built on the southside of French Road in the old canal.
1.2	0.3	Jct. NY31 (Monroe Avenue) - turn left (NW) on Monroe Ave. Pittsford Plaza on left.
1.7	0.5	Turn left between the second and third signal light opposite the Marine Midland Bank. Proceed straight to the back of the plaza between the Firestone store (left) and First Federal Bank (right).
1.8	0.1	**STOP 1 - Enlarged Erie Pittsford Lock 62 and Pittsford Shale.**

Lock 62 had a lift of 9.2 feet (2.8 m), going west, and was completed as a single chamber, enlarged lock (110 feet by 18 feet; 33.5 m by 5.5 m) by Jesse Petersen in 1855 for $26,260,77. It was doubled by James Wiltsey in 1873 for $34,269.19. At the upper end of this second chamber a drop gate was installed instead of the typical vertically hinged miter gates. This new type of lock gate was horizontally hinged and operated like a draw bridge so that when it was open it rested upstream on the floor of the canal. It was the invention of George Heath and was called Heath's Tumble Gate. They were installed on many Enlarged Erie Locks. The lock was lengthened in 1887-1888 to permit the passage of double headers (two boats lashed together) by B. T. Smith for $25,021.26. The final improvement came in 1890 when a water powered turbine, propelled by the overflow water, was installed at the base of the vertical shaft at the head of the lock. The turbine was in turn connected to geared shafts that operated rope cables wound around capstans and winches which then pulled boats in and out of the locks. These appliances were installed by Nicholas and William Wewple for $1,100. Lock 62 was abandoned at the end of the 1920 season.

Walk along the drive a few hundred feet north of Lock 62 at the base of the canal embankment behind Wegmans.

Pittsford Shale - Several feet of black shale and thin dolostones are exposed here. Ciurca (1990) has redefined the Pittsford Shale restricting it to only the eurypterid-bearing black shale and associated dolostones and renaming it the Pittsford Bed. The Pittsford Bed bears enormous numbers of the eurypterid *Hughmilleria socialis* along with *Eurypterus pittsfordensis*, *Pterygotus monroensis*, and several other non eurypterid species.

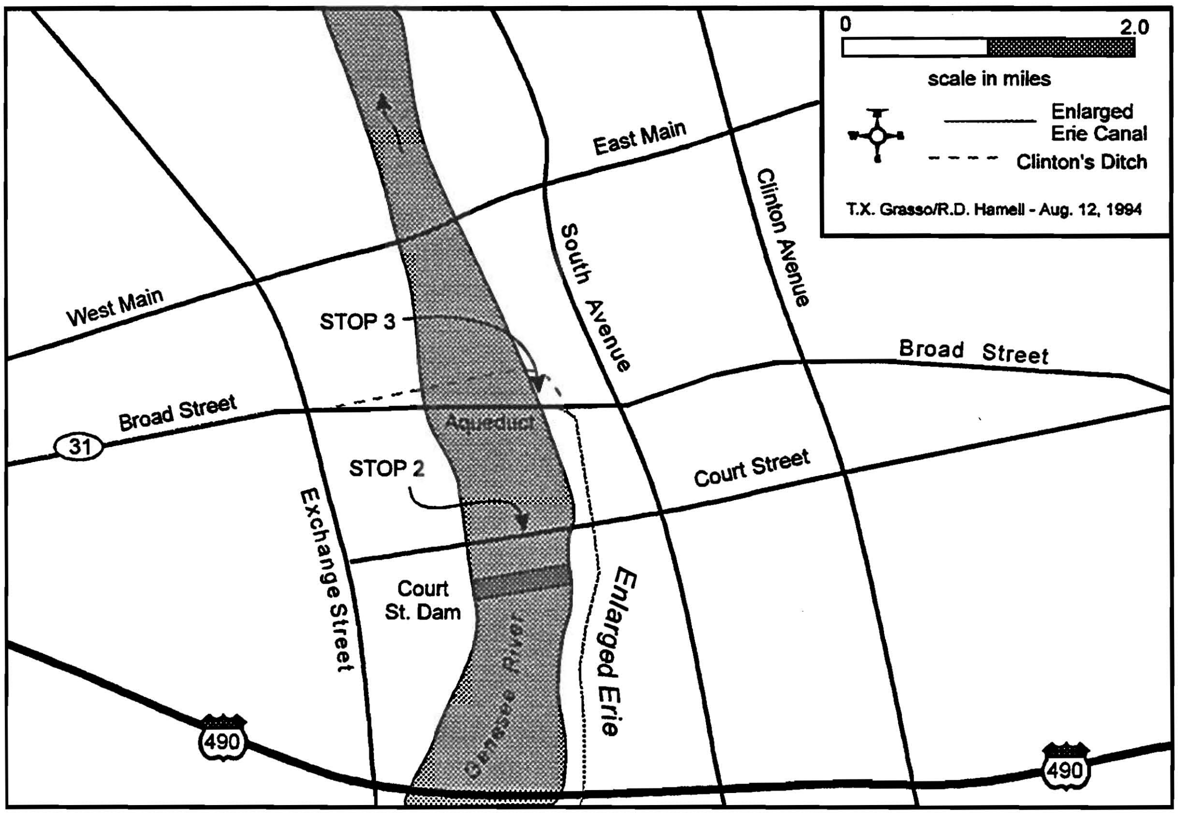


MAP 5 Route to Downtown Rochester

2.0	0.2	Return to NY31-Monroe Ave. Turn left. (NW)
2.7	0.7	Jct. I-590 Rochester interchange. Turn right (N) on I-590 which is built on the alignment of the Old Erie Canal.
4.5	1.8	Jct. I-490. Proceed west on I-490 to Downtown Rochester.
5.6	1.1	Pinnacle Range (Cobbs Hill) on left.
5.8	0.2	Old Erie Canal Lock 65 (Reservoir Lock) on right. It had a lift of 8.8 feet (2.7 m) going west. This was one of the five locks (Locks 62 through 66) east of Rochester that locked westbound boats up to the aqueduct level. Lock 66 was just west of Monroe Ave. and was the last lock west toward Lockport. Locks 32 and 33 on the Barge Canal accomplish the same task today. Clinton's Ditch took a large loop to the south here hugging the base of Cobbs Hill. Enlarged Erie construction cut off the loop thereby creating large "widewaters" between the new towpath and old canal bed. Lock 65 was at the east end

of this "reservoir". Lake Riley in Cobbs Hill Park is all that remains today.

7.6	1.8	Exit at Number 16 Clinton Ave. Downtown.
8.2	0.6	Turn left (W) on Court Street.



MAP 5A Downtown Rochester

8.4	0.2	**STOP 2 Genesee River, Aqueduct, Court Street Dam.**

Just downstream from the Court Street bridge was the head of the 14 foot (4.3 m) cascade that was the original Upper Falls of the Genesee River at Rochester. They were blasted to make way for the first Erie Canal Aqueduct.

Farther downstream is the Broad Street bridge resting on the Enlarged Erie Canal Aqueduct made of Onondaga Limestone and built from 1836 to 1842 by Josiah Bissell for $445,347. When it was completed it had an interior width of 43 feet (13 m), a length of 800 feet (244 m), and was the third longest of the 32 aqueducts on the entire line of the canal. This structure replaced the Clinton's Ditch aqueduct that was just to the north. The original aqueduct was completed in 1823, had an interior width of only 17 feet (5 m), and was constructed of

Grimsby Sandstone. This is worth noting because all other Old Erie Canal aqueducts had wooden trunks supported by stone piers, with arches only on the towpath side for its support. The engineers opted for the stone design due to the severe floods on the Genesee River before the Mount Morris Dam was completed in 1953 (Figure 12). The deck that supports Broad St. was built in 1924 and rebuilt as it stands today in 1974.

Upstream from the Court Street bridge is the Court Street Dam, called a sector dam because each gate is a pie shaped section of a circle. Gates are hinged at the base of the downstream side of the structure and float in a chamber notched into the bed of the river. The level of the water in this submerged chamber can be raised or lowered thereby controlling the gates and the pool above. The dam was completed in 1917, as part of the Barge Canal construction for the canal's Genesee crossing south of here at Genesee Valley Park. RG&E generates hydroelectric power at the west end of these structures.

		Proceed west on Court Street.
8.5	0.1	Jct. Exchange Street. Turn right (N).
8.6	0.1	Cross Broad Street. The 1826 travelers put in at Child's Basin, located where the bank is today at the northeast corner of Broad and Exchange Streets.

Eaton's party arrived at Rochester on Sunday, May 14th. George Clinton walked back on the canal about one mile to examine some "lime rock" [Lockport] along the banks, finding fluorspar in it. Clinton wrote:

> "In the afternoon visited the falls and procured specimens of the saliferous rock [Queenston, Grimsby], grayband [Kodak], ferriferous slate, [Maplewood], iron rock [Seneca Park], ferriferous sandrock [Reynales]. . . At this place . . . the geodiferous limerock [Lockport] and the calciferous slate [Rochester] meet, the latter becomes geodiferous."

8.7	0.1	Jct. Main Street. Turn right (E).
8.8	0.1	Cross Genesee River
8.9	0.1	Jct. South Avenue. Turn right (S)

9.0	0.1	**STOP 3 Penfield Formation, Lockport Group in the Genesee River**

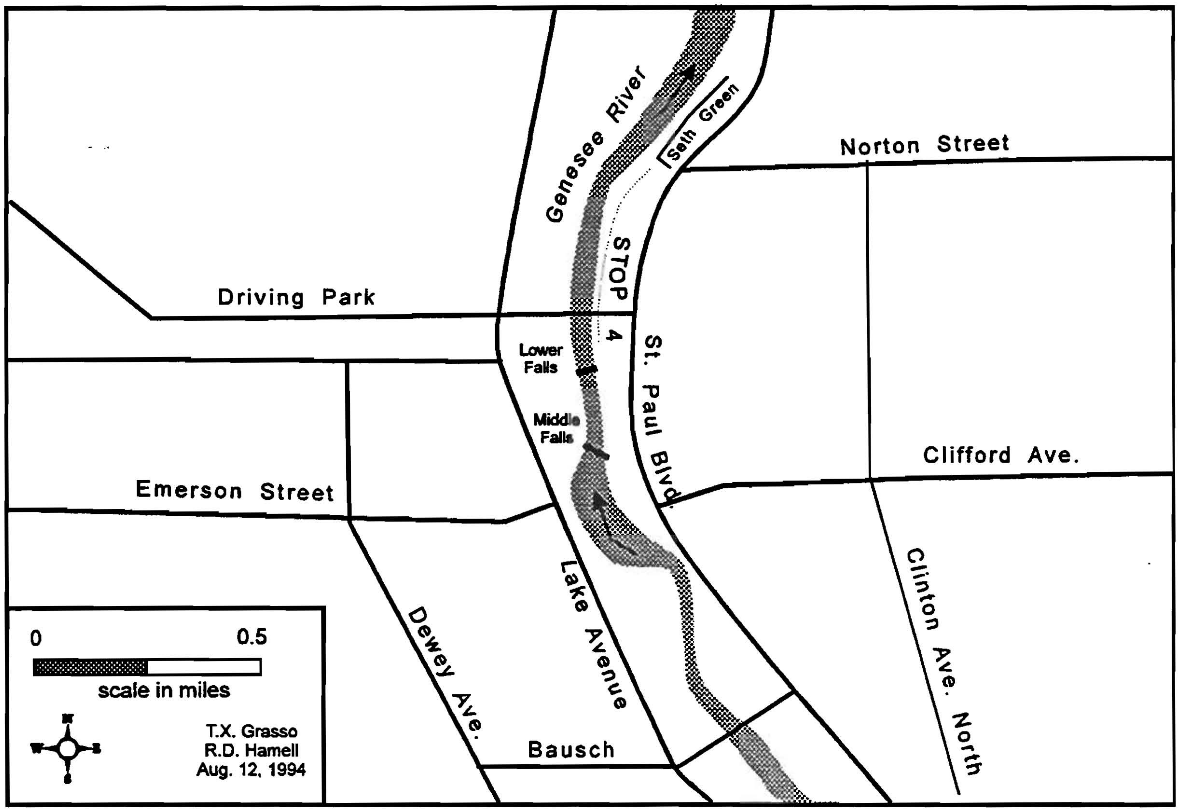
Stop at drive leading down to the back of the Convention Center and RG&E Water Street Substation just north of Broad Street. Follow the path, through the gate between the aqueduct and the south end of the RG&E brick building.

At the river level several feet of crinoid rich, coarse grained, sandy dolostone of the Penfield Formation may be examined. Ripple marks on the tops of some beds. The Penfield grades into the crinoidal Gasport Limestone and Goat Island Dolostone of the Niagara Region. These ledges are all that remain of the original Upper Falls.

The RG&E building rests on the original 1827 foundation of Hervey Ely's flour mill. Across the river in the park was the original site of Indian Allen's 1789 mill. Upstream, through the arches of the aqueduct, the Court Street dam can be viewed.

		Proceed south on South Avenue to Court Street.
9.1	0.1	Turn left (E) on Court Street.
9.2	0.1	Jct. Clinton Avenue - turn left (N) on Clinton Avenue.
9.6	0.4	Cross Main Street.
9.8	0.2	Jct. Andrews Street. Turn left on Andrews Street then immediate right (N) on Bittner Street.
10.0	0.2	Jct. Saint Paul Blvd. Proceed straight (N) on St. Paul.
11.7	1.7	Jct. Avenue E - Driving Park bridge (left). Proceed north on St. Paul.
12.1	0.4	Jct. Seth Green Drive and Norton Street. Turn left (W) on Seth Green.

Before the Days of Rapid Transit [Edward L. Henry, 1900]
[From *The Course of Empire: The Erie Canal and the New York Landscape*, 1984, Memorial Art Gallery of the University of Rochester, Rochester, New York, p. 32]



MAP 6 Rochester Gorge North of Downtown

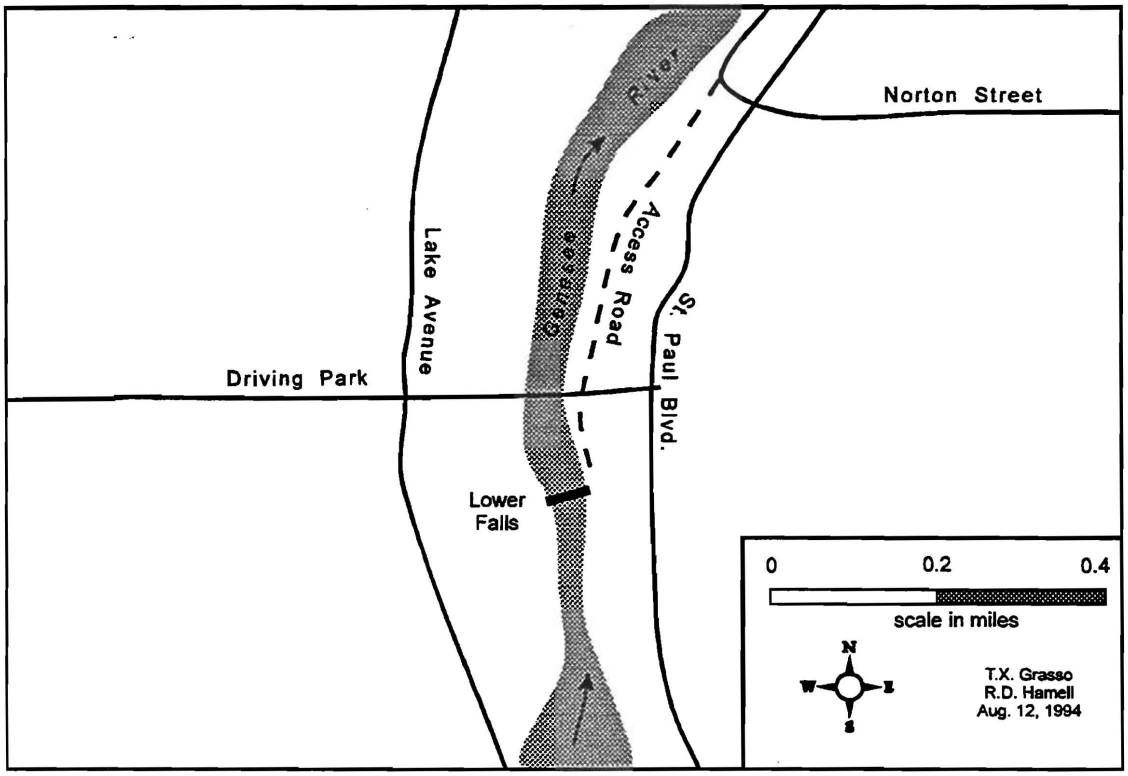
12.2 0.1 **STOP 4 Lower Falls**

Eaton's entourage arrived at the west side of the Lower Falls on Monday May 15, 1826. Asa Fitch recorded that they stopped at the Upper Falls on route but could only see a small portion of them from the west side. Arriving at the Lower Falls at 10:00AM they stayed until 2:00PM. On the west bank Amos Eaton, upon his arrival, gave a lecture on the strata ".....could see the rocks clearly. They are 1st Saliferous rock [Queenston, Grimsby], 2nd Grayband [Kodak], 3rd Ferriferous slate [Maplewood], 4th Ferriferous Sandstone [Reynales]....."

"Swam across Genesee River twice about 1/2 mile above falls 1/4 mile wide....."

George Clinton noted (p. 289):

"Is not Mr. Eaton's ferriferous sandstone rather poorly characterized. Some of the layers resembled shell limestone, contain petrifactions, and effervescing strongly with acids; others exactly resembled in external appearance the calciferous sandrock [Little Falls Dolostone, Tribes Hill Formation] obtained at Nose Hill [the Noses east of Canajoharie] and contained the same imbedded minerals, viz. hornstone, chalcedony, semiopal and agate. Is he right in considering shell-limestone, vermicular limestone and gypsum as mere beds in calciferous slate?"



MAP 6A Rochester Gorge and Lower Falls (STOP 4)

The Genesee River, diverted from its preglacial outlet through nearby Irondequoit Bay, has carved a post-glacial gorge approximately two hundred feet (62 m) deep during the past twelve thousand years. The strata revealed in the Rochester Gorge at this section are Upper Ordovician to Middle Silurian, overlain by Pleistocene till. James Hall, along with A.W. Grabau, named and described several of the formations in the gorge including the Rochester Shale, the first formally designated stratigraphic unit in North America (Brett, 1983).

The oldest unit in the gorge is the Upper Ordovician Queenston Formation, a series of brick-red, thin bedded shales and siltstones with several sandstone layers near the top (Figure 21). Although only the upper 50 to 60 feet (15-18 m) are exposed here, the Queenston Formation is more than 1,000 feet (300 m) thick and underlies much of the Lake Ontario Basin. The Queenston Formation represents extensive nearshore deltaic deposits, which covered much of the northern portion of the Appalachian Basin during Upper Ordovician time. This paleoenvironmental determination is based on well preserved sedimentary features including cross-stratification, ripple marks of both current and oscillation types, desiccation features including mud cracks, mud-chip pebble conglomerates, and sole marks (Liebe and Grasso, 1988). Following deposition of the Queenston, an interval of emergence prevailed, and a major erosion surface called the Cherokee Unconformity (Figure 17), widely traceable in eastern North America, was formed.

The overlying Grimsby Formation (also known as the "Red Medina") is a series of red-green mottled sandstones and silty shales. In his 1837 field notebook Hall originally described this unit (and the Queenston) as "sandstone, variegated, red and green, with fucoids." The base of the formation is marked by a massive 10 foot (3 m) coarse grained sandstone interpreted by Martini (1972) as a "low energy beach and/or bar(?)" deposit. This unit is overlain by a "Highly burrowed sequence of high tidal flat-flood plain or lagoonal environment which is cut through at times by shallow distributary channels and/or tidal channels" (Martini, 1972). Intraformational conglomerates and mud cracks are common. Fossils are rare in the formation, but specimens of the inarticulate brachiopod *Lingula* have been found elsewhere, and some of the massive sandstones near the middle of the unit contain excellent casts of the trace fossils *Arthrophycus alleghaniensis* and *Daedalus archimedes* (Hall's fucoids). The conspicuous bioturbated (burrowed) zone, 12 feet (4 m) below the Kodak Sandstone apparently correlates with the Thorold Sandstone of the Niagara Gorge (Brett et al., 1990). The origin of the prominent color mottling is not clearly understood; however, the green splotches are generally thought to represent the irregular alteration of ferric iron to the ferrous state by either downward percolating ground water or the presence of disseminated organic material (Liebe and Grasso, 1988).

The Grimsby Formation grades into the Kodak Sandstone, a 6 foot (2 m) thick, massive, fine grained, bioturbated gray sandstone. Hall, in his initial work in the Rochester Gorge, followed Eaton and described this unit as the "grayband", a coarse, gray sandstone which was harder than the underlying sandstones and less destructible. At one time the Kodak Sandstone was thought to be correlative with the Thorold Sandstone in the Niagara Gorge. Work by Brett et al. (1990) recognized that these units are two distinct sandstones separated by 10 feet (2-3 m) of red and green shale, informally known as the Cambria Member of the Grimsby Formation. This coincides with the work of Chadwick (1935) and Fisher (1966) who used the name Kodak for the prominent formation that forms the caprock of the Lower Falls.

The basal unit of Sequence II is the Maplewood Shale, a smooth, platy, in-part calcareous shale that unconformably overlies the Kodak Sandstone. This unit was described by Hall in his field book (1837) as a "ferriferous slate, soft and green, contains a few shells, and near its limit below, a very thin stratum, sometimes two, composed almost wholly of shells" In addition to these thin limestone layers, phosphatic pebbles are common, especially in the basal portion. The Maplewood is approximately 21 feet (7 m) thick here but becomes much thinner to the west where it is correlative with the Neagha Formation in the Niagara Gorge. Both formations are thought to represent either quiet, offshore marine or lagoonal deposits as part of the initial flooding of the transgressing Sequence II sea into the Rochester area.

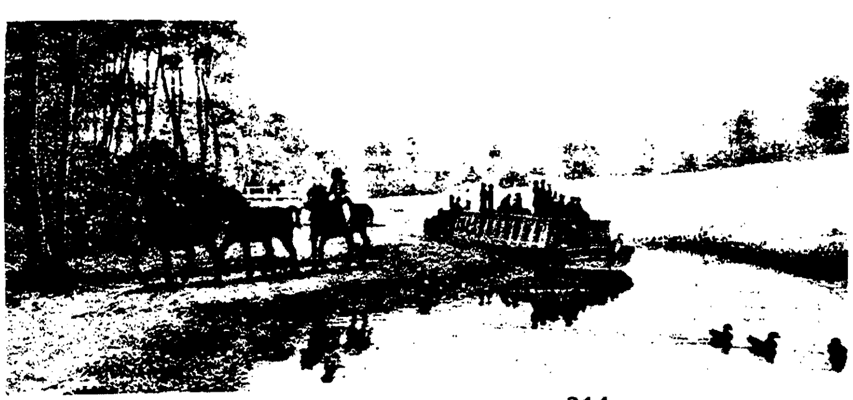
The youngest unit exposed along the access road to the Power Station is the Reynales Formation, although Hall (1843) first designated this unit the "Pentamerus Limestone" for the numerous pentamerid brachiopods present. The Reynales Formation consists mainly of 21 feet

(7 m) of gray, crystalline limestone or dolomitic limestone with numerous shale partings and is divided into three members. The lower member, the Brewer Dock Limestone 3 feet (1 m), is separated from the upper Wallington Limestone Member by the well known Seneca Park Hematite (1 foot; 0.3 m). The Seneca Park was previously known as the Furnaceville, but recent mapping by Brett et al. (1990) revealed that the iron ore in the Genesee Gorge is actually at a higher stratigraphic horizon than the type Furnaceville in Wayne County and, therefore, cannot be the same unit. The Seneca Park Member was originally deposited as a fossiliferous limestone. The hematite has since replaced most of the calcium carbonate; however, samples will still effervesce when acid is applied, indicating much of the original carbonate material remains. The mechanism and time of replacement are much debated, but consensus seems to favor a penecontemporaneous alteration of the original fossiliferous limestone by iron-rich solutions from streams descending the iron-rich Taconic landmass located in eastern New York State at that time.

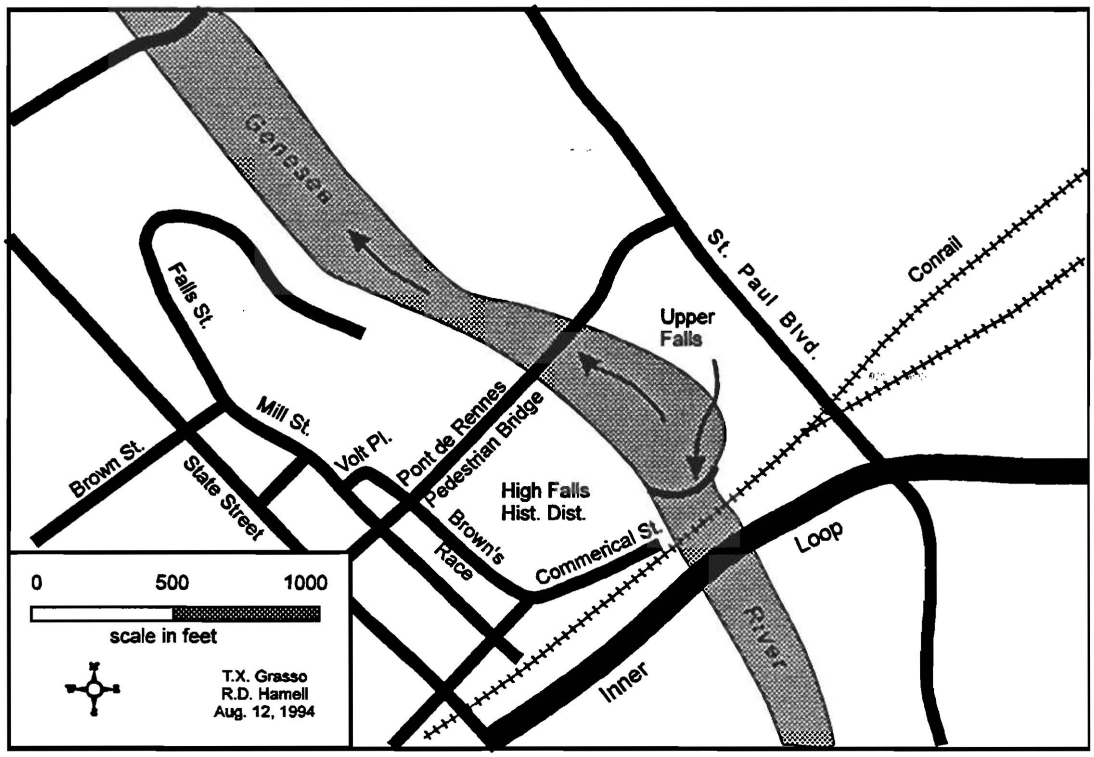
The Wallington Member (17 feet; 5 m) is a coarse grained, fossiliferous limestone, with beds of the smooth shelled brachiopod *Pentamerus*, as well as corals and bryozoans. The combination of dense strata packed with fossils and the replacement of some of the fossils and beds with chert make this unit particularly resistant, well suited to form the cap rock of the Middle Falls. The Reynales Formation then represents a shallow, clear water, offshore marine deposit of Sequence II.

Visible along the walls of the Rochester Gorge, but not easily accessible at this locality, is the upper member of Sequence II, the Sodus Shale, as well as the Williamson Shale and Irondequoit Fm. (Sequence VI) and the Rochester Shale (Sequence V). The latter unit, as well as the overlying Decew and Lockport Dolostone (Sequence VI) will be seen at **STOP 5.**

12.3	0.1	Return to St. Paul Blvd. Turn right (S).
12.7	0.4	Jct. Avenue E-Driving Park Avenue. Turn right (W). Cross Genesee River. Lower and Middle Falls on left.
13.0	0.3	Jct. Lake Avenue. Turn left (S)
14.7	1.7	Jct. Platt Street. turn left (E). Enter High Falls Historic District.
14.8	0.1	At end of Platt Street turn right on Browns Race.



Pittsford on the Erie Canal by George Harvey, ca. 1840. [From *The Course of Empire: The Erie Canal and the New York Landscape*, 1984, Memorial Art Gallery of the University of Rochester, Rochester, New York, p. 26]



MAP 7 Brown's Race Historic District and High (Upper) Falls (STOP 5)

STOP 5 High Falls (Upper Falls), Browns Race

Although visually attractive to visitors, the High Falls could rarely be seen as industrial development along Matthew and Francis Brown's mill race blocked the view. It was sometimes said that Rochester had developed the river, for commercial purposes, to such a degree that the best view of the river and its falls could only be had from the back door of a factory or mill.

The upper part of the Lewiston Member is exposed in the lowest part of the gorge's east wall below the Upper Falls while the Burleigh Hill Member of the Rochester Shale forms the conspicuous bulge in the middle portion (Figure 24). The overlying 25 feet (8 m) of the Gates Member makes up the upper portion of the gorge walls downstream from the Upper Falls while its uppermost beds form the brink.

The Decew Formation is exposed upstream from the lip of the falls, in the uppermost 6 to 10 feet (2 to 3 m) of the east side of the gorge just downstream from the falls and on the west side at the base of the stone wall, where the enterolithic structure is clearly discernible, (north wall of the Kidd Iron Works) east of the water wheel. Another exposure is at the top of Falls Street just north of its intersection with Brown Street.

Figure 32 High Falls looking south about 1910. Gorsline Building on west side, Rochester Railway and Light Company on east. Decew Formation forms small step just upstream from brink. (Stone Collection, Rochester Mus. and Sci. Center)

Walk along Brown's Race Street to view Brown's Race at Commercial Street. From here walk east on Commercial Street and opposite the High Falls Parking lot climb the embankment to the railroad. Walking east cross the river on the north side of the railroad bridge above the lip of the falls. Excellent vistas of the gorge and the historic district are obtained from this bridge. Continue walking east and then north along the railroad spur for Genesee Brewery to the Platt Street (Pont de Rennes) bridge (see Figure 32 for a 1910 view taken from this bridge) crossing back to the west side and returning to the starting point.

END OF TRIP - Return to U. of R.

The evening of May 15, 1826 was disastrous for Amos Eaton as at the end of the day he was struck by a fit of delirium. Asa Fitch wrote:

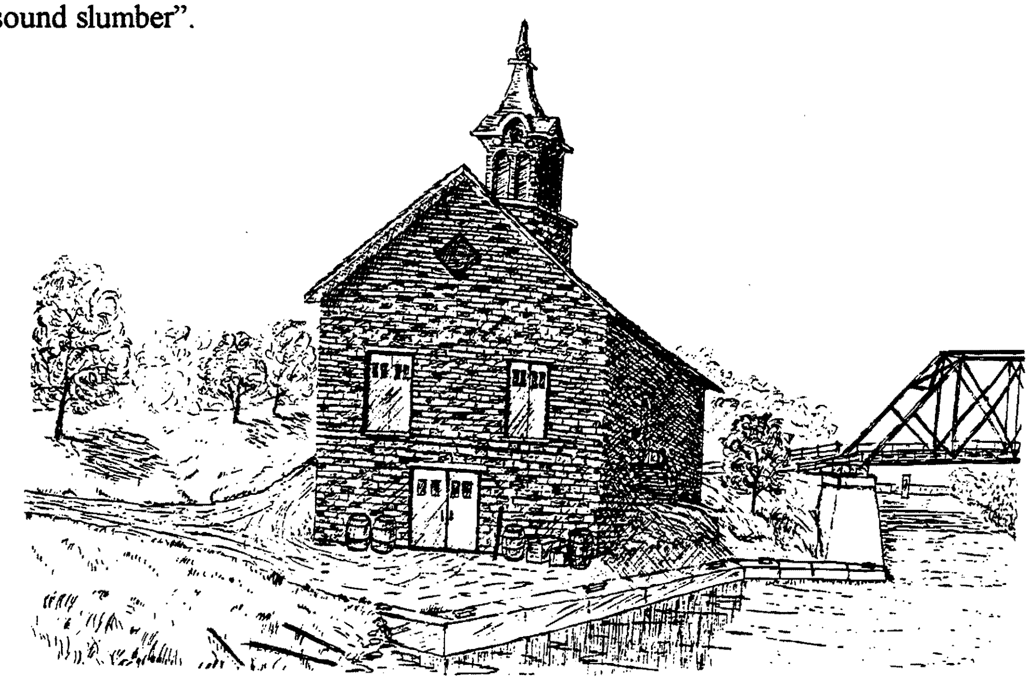
> "......met Prof. Eaton who was taken with a fainting turn. He leaned on my arm till we came to one end of the dock at the other end of which the boat

> lay. Here he got so weak he could not proceed.....We took him into a store a few steps off and in a few minutes he became completely crazy.....We got him to the boat and for about a hour he continued as crazy as any body I ever saw. He thought he was among savages.....He thought the medicine Dr. Marvin gave him was poison and was uneasy about it. Said that the canal survey was not yet finished we might poison him if we wished that he would then drink with pleasure. After an emetic had operated his senses immediately came to him. He was afraid he had said or done something disgusting during his fit."

George Clinton described the event (p. 288) in more general terms:

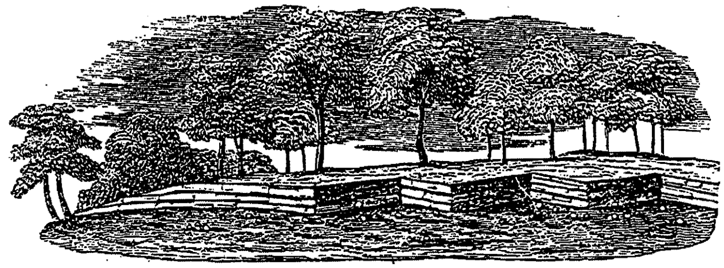
> "In the evening Professor Eaton was seized with a fainting fit brought on probably by fatiguing himself so much during the day. He was delerious for nearly 1 hour, during which time the soundness of his remarks proved that his mind although uncontrollable, was by no means defective in strength. The physicians called in administered repeated doses of sulphate of zinc and ipecacuanha [a tropical plant the dried roots of which are used to make a preparation that induces vomiting]. As soon as their operation had ceased, his reason returned and he is now (10PM) enjoying an apparently sound slumber."

Hopefully we will not need similarly strong medication after the banquet following our trip to enjoy a "sound slumber".



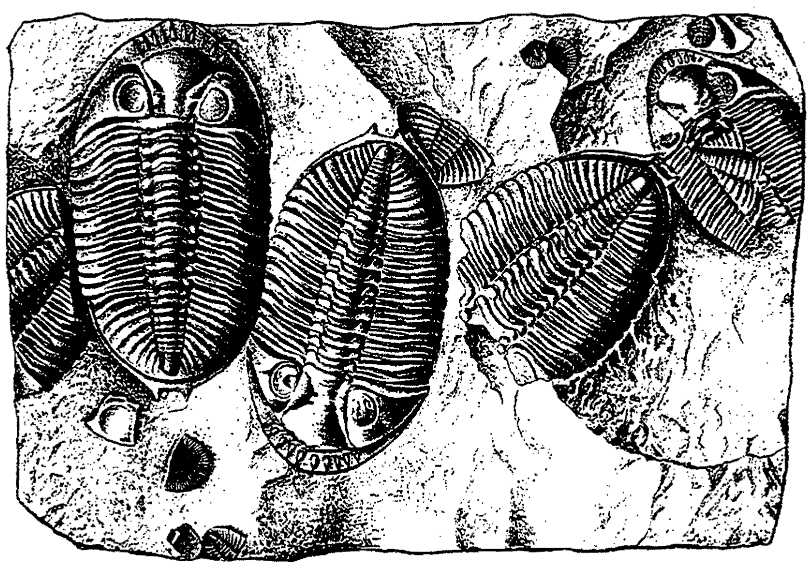
The Scheaffer Cold Storage House on the canal near Gasport, N.Y., built in 1870 and destroyed by fire in 1967. Drawing by E. Mayes.

[From R. Garrity, 1977, *Canal Boatman: My Life on Upstate Waterways*, Syracuse Univ. Press, Syracuse, N.Y., p. 98.]



CORNIFEROUS LIMESTONE.

Upper part of the Corniferous Limerock of EATON. *Seneca Limestone of the Annual Reports.*



TOP: From Hall, 1843, p. 161.

BOTTOM: *Odontocephalus selenurus*, Onondaga Limestone, Seneca County, New York. From Hall and Clarke, 1888, Plate 12.

STRATIGRAPHY AND FACIES RELATIONSHIPS OF THE EIFELIAN ONONDAGA LIMESTONE (MIDDLE DEVONIAN) IN WESTERN AND WEST CENTRAL NEW YORK STATE

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INTRODUCTION

The Middle Devonian Onondaga Formation of New York generally represents broad, carbonate platform facies that were deposited in the northern part of the Appalachian Basin during early to middle Eifelian time. Finer-grained, more basinal facies in central New York shallow laterally to the east and west into more biologically productive platform settings toward the Buffalo and Albany areas. Numerous biostromal to biohermal buildups, which include pinnacle reefs in the subsurface of southern New York, are rooted in the shallowest-water basal Onondaga facies.

Carbonates of the Onondaga Formation are characterized by calcarenitic to cherty to argillaceous limestones and minor shales deposited in a shallow epicontinental sea. The formation comprises a generally deepening-upwards succession, but apparently represents two major (third order) transgressive-regressive cycles. It is generally subdivided into four members across New York State (Edgecliff, Nedrow, Moorehouse, and Seneca, from the base upwards; Oliver, 1954; Rickard, 1975). A fifth member (Clarence) has been recognized only in western New York (Ozol, 1964; Oliver, 1966b); we herein designate the Clarence as a local facies of the Edgecliff Member.

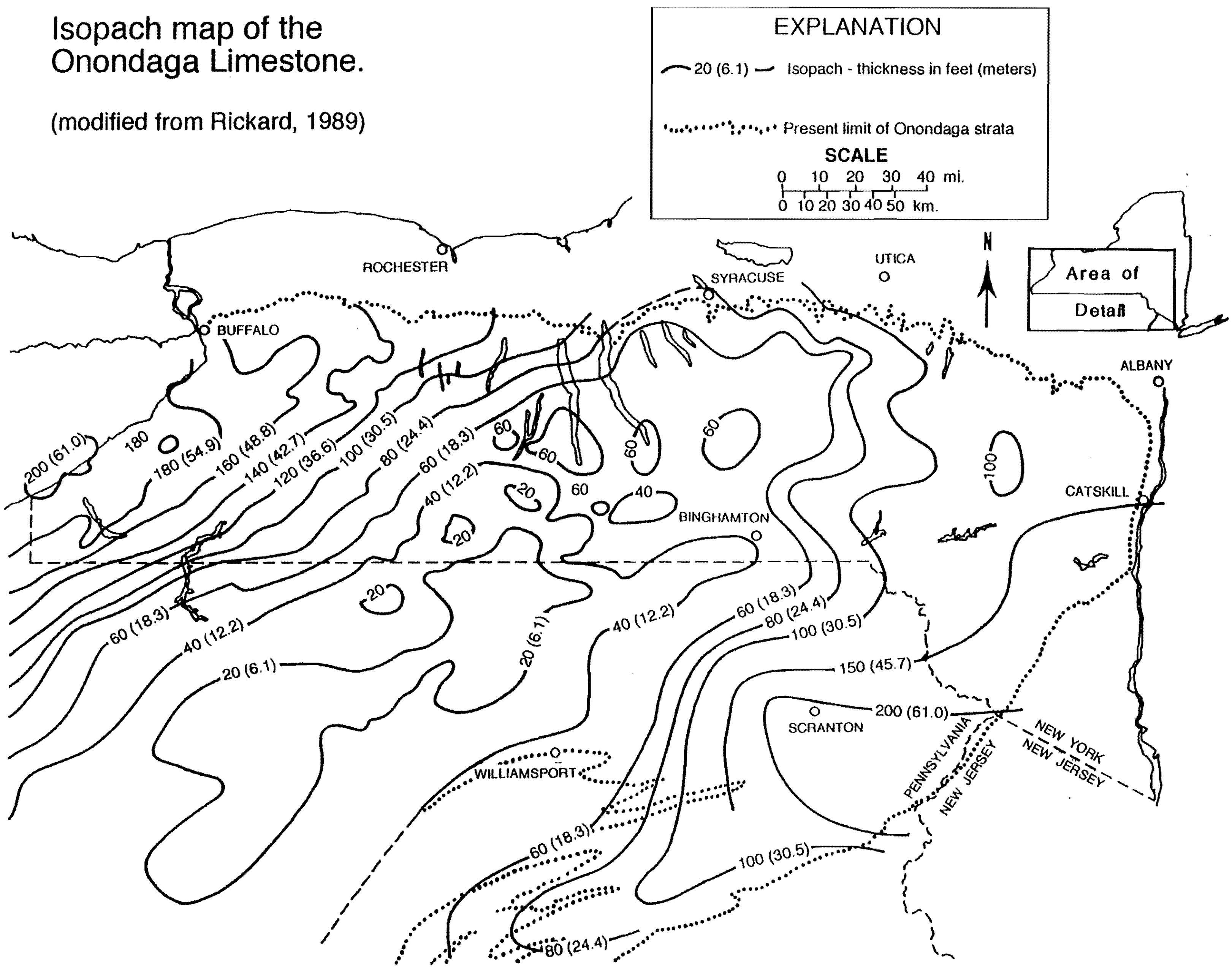
The basal contact of the Onondaga Limestone is conformable in parts of eastern New York, but westward becomes unconformable and overlies increasingly older Devonian to Silurian strata. Throughout western and central New York State Eifelian-age strata are underlain by a major sub-Onondaga unconformity, commonly referred to as the Wallbridge Unconformity. The Wallbridge was recognized long ago by Sloss (1963), as one of the major sequence-bounding erosion surfaces in the Phanerozoic of North America. He used it to define the boundary between his Tippecanoe and Kaskaskia Megasequences. In actuality, the sub-Onondaga unconformity in Western New York State is a composite of two to three distinct erosion surfaces which are locally merged (Oliver, 1966c; see below). The nature of the upper contact has long been debated; it appears to represent a westward-younging, sediment-starved, submarine unconformity below overlying black shales at the base of the Middle Devonian Hamilton Group.

Underlying the Onondaga Formation in some parts of west-central to western New York are quartz arenites and minor carbonates of the upper Lower Devonian Tristates Group. These rocks, which include strata equivalent to the Oriskany Sandstone and/or the Carlisle Center and Schoharie Formations of eastern New York, occur sporadically across the region. In the Buffalo region a westwardly-thickening wedge of cherty limestone (Bois Blanc Formation) appears and thickens into southern Ontario.

Limestones of the Onondaga Formation (Figure 1) represent relatively tabular, clean, dominantly biogenically-derived sediments deposited in the northern part of the Appalachian Basin during a period of relative tectonic quiescence. Underlying and overlying clastics (Lower Devonian Tristates Group and Middle Devonian Hamilton Group, respectively) represent terrigenous sediments shed from a rising mountain belt in New England during two separate active "tectophases" of the Devonian Acadian Orogeny (Ettensohn, 1985). The concentration of altered volcanogenic strata (Tioga Bentonites/Ash Beds of Dennison and Textoris, 1970, 1978, 1987) in the upper part of the Onondaga Limestone heralds the

Isopach map of the Onondaga Limestone.

(modified from Rickard, 1989)



progradation of clastic sediments into the Appalachian foreland basin.

Interestingly, the area of maximum erosion and, presumably greatest uplift, during pre-Onondaga Emsian time experienced an inversion of topography in the early Eifelian. The main basin axis during deposition of the Onondaga Formation was a northeast-southwest oriented trough of active subsidence that intersects the New York outcrop belt approximately at the location of the central Finger Lakes (e.g., Seneca Stone quarry between Seneca and Cayuga Lakes = Stop 6 of this trip), but which migrated eastward through the Eifelian.

In this paper we will examine the following: 1) a new detailed stratigraphy for the Middle Devonian Onondaga Formation and associated underlying strata in western to west-central New York; 2) the occurrence of the Tioga Bentonites cluster in the Onondaga Formation; 3) apparent lateral shifts in more basinal facies through Onondaga time; 4) a sequence stratigraphic interpretation of the Onondaga and underlying strata of the upper Lower Devonian Tristates Group; and 5) a new regional synthesis of the Onondaga Formation and equivalent strata in the northern and central parts of the Appalachian Basin (New York and Pennsylvania).

GEOLOGICAL BACKGROUND

GEOLOGICAL SETTING

The Onondaga Formation of New York was deposited in the northern part of the Appalachian Basin during a time of relative tectonic quiescence in the early Middle Devonian. Basin topography across New York through much of Onondaga time consisted of three relatively shallow, gently dipping ramps, two in eastern New York that sloped to the southwest into Pennsylvania (Lindemann and Feldman, 1993) and westward into central New York, respectively, and a third that sloped eastward from western New York-Ontario into central New York. The latter two ramps gently dipped toward a central deeper, more basinal northern tongue of the central trough of the Appalachian foreland basin. The main basin axis during Onondaga deposition was a northeast-southwest trending area of more active subsidence that intersects the New York outcrop belt approximately at the location of the central Finger Lakes. Dennison (1985) likened the shape of the Appalachian Basin during much of the Devonian Period to the keel of a boat. The deepest portions of this basin lay south of the New York outcrop belt area, and are probably recorded near Altoona (west-central Pennsylvania) and in the subsurface to the west.

The eastern interior seaway during the earliest Middle Devonian was not restricted to the Appalachian Basin, however; carbonate-dominated strata equivalent to the Onondaga Limestone are widely distributed across eastern North America (Figure 2), from the James Bay region of northern Ontario to southeastern Quebec and Maine to southwestern Virginia to Illinois, and include such units as the Columbus Limestone of Ohio and the Jeffersonville Limestone at the classic Falls of the Ohio River (Koch, 1981).

PREVIOUS WORK

In the American Journal of Science in 1828, Amos Eaton (Eaton, 1828) first reported the "Cornitiferous Limerock" that constitutes the modern day Onondaga Formation. Subsequent work by the first New York Geological Survey (Vanuxem, 1839, 1840, 1842; Conrad, 1837; Hall, 1841, 1843; Mather, 1843) recognized two to four divisions of Eaton's Cornitiferous across New York State. In their final reports they noted three main

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Figure 1. Isopach and outcrop map of the Onondaga Limestone and equivalent strata in the Northern Appalachian Basin (modified after Rickard, 1989).

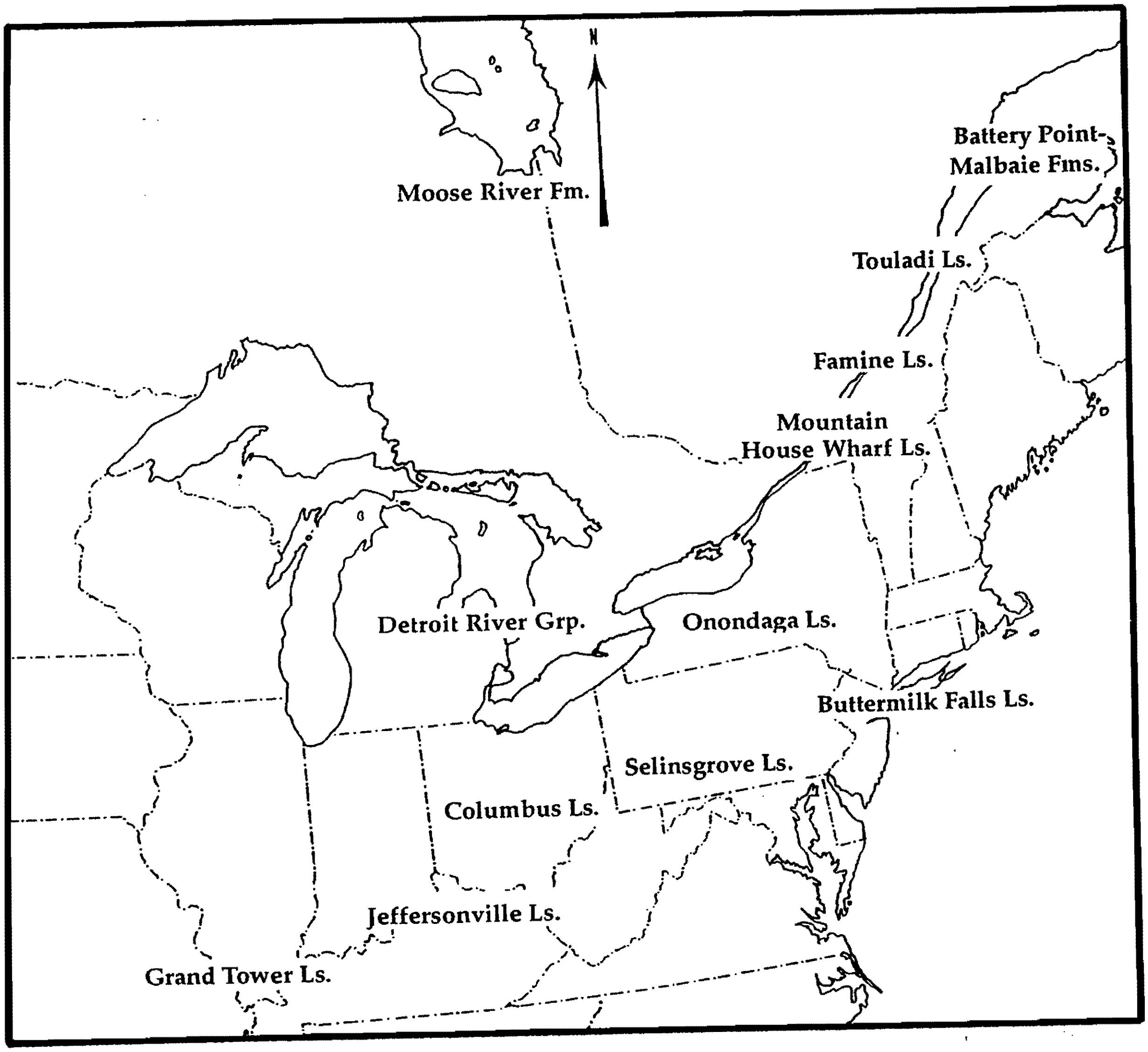

Figure 2. Map of Onondaga-equivalent strata across eastern North America (source=Koch, 1981).

units: the "Onondaga Limestone" (coral-bearing, crinoidal limestone), the "Corniferous Limestone" (lower shaly and upper cherty units), and the "Seneca Limestone" (with relatively little chert and common chonetid brachiopods). The unit as a whole came to be termed the Onondaga near the turn of the century (for further discussion see Oliver, 1954; Chadwick, 1944).

The first modern stratigraphic and paleontologic study of the Onondaga Formation, upon which all recent study is based, was by Oliver (1954, 1956a; see Figure 3). Oliver subdivided the formation into four members that roughly correspond to the "Onondaga" (= Edgecliff Member), the two subdivisions of the "Corniferous" (=Nedrow and Moorehouse Members) and "Seneca" (= Seneca Member) Limestones of the first New York Survey.

James Hall's classic series, *Natural History of New York: Paleontology* (e.g., Hall, 1867, 1877) included the first comprehensive systematic study of the fauna of the Onondaga

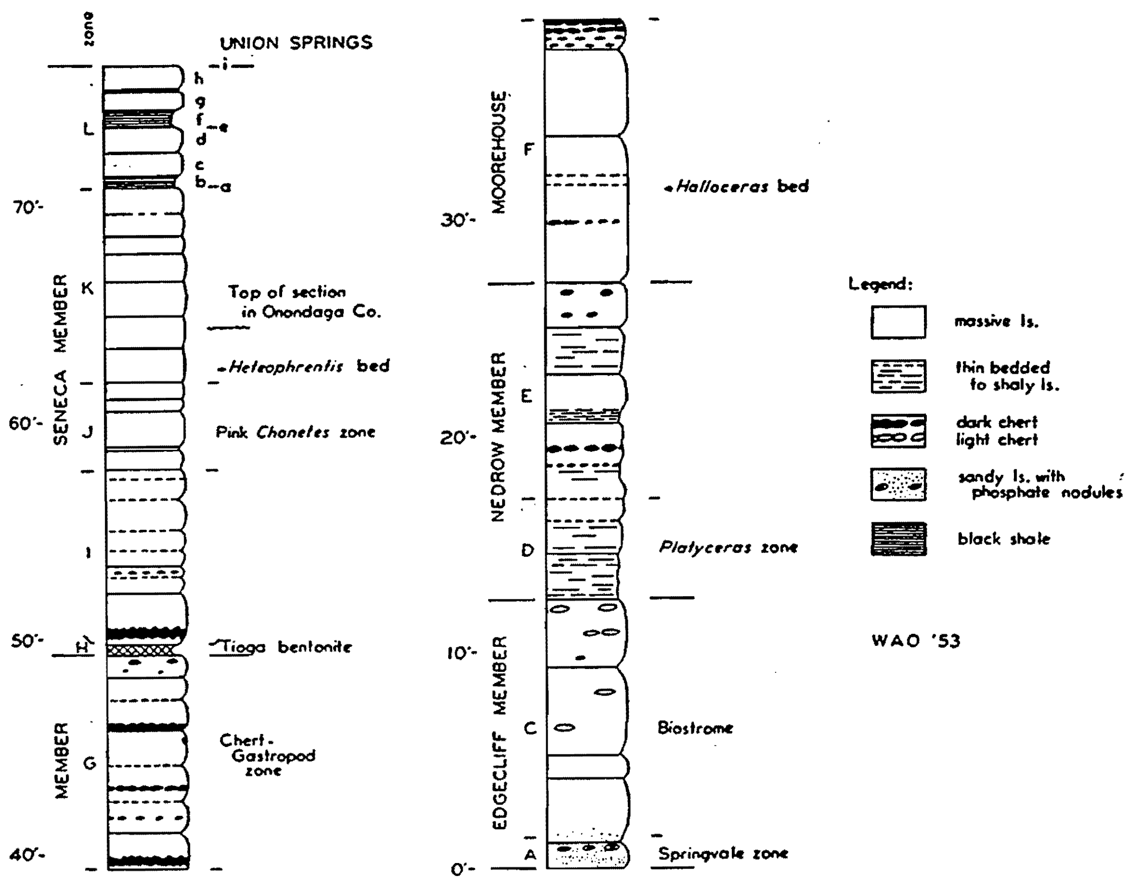

Figure 3. Oliver's (1954) composite section of the Onondaga Formation in the type area in central New York. Measurements in feet.

Limestone. Recent studies since Oliver's classic papers have chiefly focused on the corals (Oliver, 1960, 1976), brachiopods and brachiopod paleoecology (Feldman, 1980, 1985, 1994; Koch, 1981), and dacryoconariids and styliolinids (Lindemann and Yochelson, 1984). Lindemann (1980), Lindemann and Feldman (1981), and Feldman and Lindemann (1986) also examined the paleoecology of the Onondaga across New York and outlined a series of faunal communities.

Coral bioherms have been the focus of much study, in part due to the economic significance of pinnacle reefs in the subsurface of southern New York. Oliver's (1956b) report on bioherms and biostromes in eastern New York State was the first of many papers, most notably by Wolosz (1988, 1992), Wolosz and Paquette (1988), Kissling (1981), Kissling and Coughlin (1979), and Cassa and Kissling (1982); in addition, local bioherm buildups along the New York outcrop have been the subject of numerous master's theses since the 1960's (e.g., Coughlin, 1981; Poore, 1969).

Carbonate lithologies and detailed facies analysis of the Onondaga have been addressed by Lindemann (1980) and Lindemann and Feldman (1981). Additional petrologic study of the formation has focused chiefly on limestone (Lindholm, 1967, 1969a) and dolomite (Lindholm, 1969b; Selleck, 1985), chert (Ozol, 1964; Selleck, 1985; Maliva and Siever, 1989; Moyer, 1956), and clay mineralogy (April et al., 1984).

BIOSTRATIGRAPHY AND AGE

The Lower-Middle Devonian (Emsian-Eifelian) boundary was long placed at the base of the Onondaga Limestone in New York. Recent definition of the global Emsian-Eifelian stage boundary by the Subcommission On Devonian Stratigraphy, however, raises questions on its placement in New York. The latest statement, from Kirchgasser and Oliver (1993), places the boundary at or near the base of the formation, although they state that it could lie as high as the base of the Nedrow Member (the tenuous placement of the boundary is due to a lack of diagnostic conodonts in the Edgecliff Member).

Biostratigraphy of the Onondaga Formation is discussed in a series of papers in Oliver and Klapper (1981) and is summarized here. Goniatite biostratigraphy is poorly restrained for the Onondaga Limestone, with only one species reported. The entire formation is placed within House's (1981) goniatite Fauna #3. Klapper (1971, 1981) reports two conodont zones within the Onondaga, the "*patulus*" and "*costatus costatus*" zones. The former occurs within the lower part of the formation, from an unknown base (noted above) through the lower part of the Nedrow Member; the latter includes the upper part of the Nedrow Member through the top of the Onondaga. Klapper states that a fauna from low in the overlying Union Springs could be in either the *costatus costatus* or the younger *australis* zone.

The Edgecliff, Nedrow, and lower part of the Moorehouse Members occur in the upper part of the *Amphigenia* Assemblage Zone, in the upper of two subzones, the *Fimbrispirifer divaricatus* subzone (=large *Amphigenia* Zone; Dutro, 1981). Upper Moorehouse and Seneca strata are within the overlying *Paraspirifer acuminatus* zone. Oliver and Sorauf (1981) summarize the rugose coral biostratigraphy for the Onondaga Formation, and report two Assemblage Zones (*Acinophyllum segregatum* zone and an unamed zone). The first is broken into two subzones which include the Edgecliff (*Synaptophyllum arundinaceum* subzone) and the Nedrow-Moorehouse Members (*Eridophyllum seriale* subzone).

PALEONTOLOGY AND FAUNA

Brachiopods, corals, bryozoans, trilobites, and gastropods are more common elements of the fauna of the Onondaga Limestone. Their distribution varies throughout the formation with changes in lithofacies. Pelmatozoans, generally represented by disarticulated crinoid material, are common through parts of the formation, most notably in coarser facies where crinoid ossicles may be 1-2 cm in diameter and form crinoidal grainstones. Styliolinids are common in finer-grained facies, and may compose up to 95% of the rock (Lindemann, 1980).

Lindemann (1980) analyzed the fauna of the Onondaga and recognized nine communities within the formation across New York. Of these, two are dominated by corals, two by bryozoans, three by brachiopods, and one by trilobites; the additional community is split between a diverse brachiopod fauna and numerous corals. Lindemann also noted that the ichnofauna and degree of burrow-mottling varies throughout the formation. Coarser-facies communities feature occasional vertical burrows, while finer-grained, calcisiltite-rich assemblages may feature a diverse and abundant ichnofauna, generally dominated by *Chondrites.* Intense bioturbation occurs in some facies, and the ichnofauna is indistinguishable. Additional community analysis by Feldman (1980; Lindemann and Feldman, 1987) focuses specifically on the brachiopod communities of the Onondaga across New York.

Changes in communities through the Onondaga, as discussed above, are dominantly associated with tracking specific facies. Overall, the fauna of the Onondaga is very stable throughout the late Emsian to early Eifelian; Feldman (1994), in a recent monograph on the brachiopod fauna, states, "...taxa found in the Bois Blanc-Onondaga interval indicate a high degree of stasis." Dutro (1981) also recognizes, "the general unity of the Schoharie and Onondaga" brachiopod faunas, a fact he states was previously recognized by Cooper. Reportedly more than 80% of Schoharie brachiopod species persist into the Onondaga despite more marked changes in other taxa such as trilobites and cephalopods. A much smaller proportion of the brachiopods (no more than 10% of species) and very few of the trilobites, mollusks, or echinoderms of the Onondaga Formation, however, persist into the overlying Hamilton Group.

Several authors in recent years have interpreted a cooler, warm temperate setting for the Appalachian Basin during Oriskany to Onondaga time (Koch and Boucot, 1982; Boucot, 1990). Based on the lack of abundant stromatoporoids, stromatoporoid-dominated bioherms, oolites, dasycladacean or udoteacean algae, and a lower diversity gastropod fauna, Blodgett et al. (1988) infer a probable warm temperate environment for the Northern Appalachian Basin during the early Eifelian Stage. Koch and Boucot (1982) reach similar conclusions based on the paucity of gypidulinid brachiopod-dominated communities in the Oriskany to Onondaga Formations of the Appalachian Basin. Furthermore, in examination of Edgecliff reefs across New York and Ontario, Wolosz (1990a,b) reports a westward trend of increased size and abundance of stromatoporoids in the direction of the Devonian paleoequator.

Faunas across the Appalachian Basin from the Oriskany to Onondaga Formations show more affinity with colder, temperate Malvinokaffric Realm faunas than with warmer subtropical to tropical faunas of the more equatorward Old World Realm (Boucot, 1975, p. 327-331). This affinity breaks down, however, in the overlying Marcellus succession, and warmer subtropical faunas immigrate into the Appalachian Basin (see discussion in Ver Straeten et al., this volume)

THE LOWER DEVONIAN TRISTATES GROUP IN CENTRAL TO WESTERN NEW YORK STATE

Pragian- to Emsian-age strata of the upper Lower Devonian Tristates Group are poorly represented in west-central to western New York. In eastern New York these strata are chiefly represented by the Oriskany-Glenerie, Esopus, Carlisle Center, and Schoharie Formations and range up to 300 m in thickness (Rickard, 1989). The Tristates Group is thin to locally absent across the central to western part of the state and is generally represented by sand-dominated facies, with a wedge of limestone facies appearing in the Buffalo area that thicken westward into Ontario.

Quartz arenites of the Pragian-age Oriskany Formation and equivalent strata occur widely across eastern North America and comprise the basal sandstone of Sloss' (1963) Kaskaskia Megasequence. Large, robust brachiopods and other forms characterize the Oriskany Formation. Erosional remnants of the Oriskany Sandstone occur sporadically across west-central to western New York. Across much of the area, however, post-Oriskany processes removed and reworked the quartz sand during Emsian and/or earliest Eifelian time.

Younger, thin, upper Tristates sandstones, sometimes termed "Springvale Sandstone," also occur sporadically across the region. These are better exposed in the Syracuse region where they are represented by approximately 2-3 m of glauconitic and phosphatic to clean quartz arenites, which may locally be hematitic. Thin sandstones also occur locally at the base of the Onondaga Formation in the region (for more discussion, see below).

A thin wedge of upper Tristates limestone, the Bois Blanc Formation, occurs locally in western New York (Oliver, 1966a&c, 1967). The Bois Blanc is a dark gray, brachiopod-rich limestone (Boucot and Johnson, 1968) with numerous corals. It is finer grained than the overlying Edgecliff Member of the Onondaga Limestone. The Bois Blanc ranges from 0-1.3 m across western New York; it thicken westward into Ontario, and is 30 m in thickness at Woodstock, Ontario. The Bois Blanc Formation is equivalent to the Schoharie Formation of eastern New York.

THE ONONDAGA FORMATION IN WESTERN TO CENTRAL NEW YORK STATE

PHYSICAL STRATIGRAPHIC SUBDIVISIONS OF THE ONONDAGA LIMESTONE

Introduction

The subdivision of the Onondaga Limestone into members was completed largely as a result of the detailed studies of Oliver (1954; see Figure 3). Part of our work is focused upon the Nedrow-Seneca interval of the Onondaga in an attempt to correlate it westward from the type area south of Syracuse into western New York. This has resulted in discovery of several key marker beds, most notably a pair of dark shales and a fossil-rich bed overlain by thin clay-rich notches that may be K-bentonites. These markers have been correlated, not only into western New York, but subsequently into eastern New York State and into the Selinsgrove Limestone of central Pennsylvania. In the following sections, the various members of the Onondaga will be described in some detail, followed by a sequence stratigraphic interpretation of Onondaga stratigraphy.

Onondaga Formation - Overall Thickness And Facies

The Onondaga Formation in western and west central New York State (Figure 4) consists of approximately 18 to 50 m (Rickard, 1989; Figure 1) of light to medium gray-weathering and commonly cherty limestone. The lower Edgecliff Member and portions of the Moorehouse Member are characterized by a crinoidal pack- and grain-stone lithology with local micritic bioherms or patch reefs developed in the Edgecliff Member and biostromes of fasciculate corals, particularly *Eridophyllum* and *Synaptophyllum*, in the Moorehouse Member. These coarser-grained facies are typically non-cherty, or contain only small, gray chert nodules.

The most chert-rich portion of the Onondaga Limestone consists of up to 10 meters of micritic argillaceous wackestone (calcisiltite) interbedded with medium to dark gray cherts. This facies is particularly well developed within the upper part of the Edgecliff Member (Clarence facies) where chert nodule-rich beds and nearly continuous chert bands may constitute nearly 50% of the volume of the rock. Similar facies occur in the Moorehouse Member in western New York.

Lindholm (1967) recognized two fine-grained subfacies within the Moorehouse and Seneca Members based upon the percentage of clay; these were the fossiliferous calcisiltite with about 5% clay and biocalcisiltite with a smaller amount of clay and about 10 to 50% fossils. The latter facies appear to become dominant upward in the Moorehouse and Seneca Members of Western New York State, as well as in the upper Moorehouse of the Albany region. Finally, in the Nedrow Member, Lindholm recognized a fossiliferous calcisiltite with about 25% clay and less than 10% fossils.

Isopach maps for the Onondaga (Mesolella, 1978: Rickard, 1989; see Figure 1) demonstrate that the basin axis trend was associated with areas of minimal thickness for the Onondaga in south-central New York State. The outcrop expression of these more basinal facies indicates that they are composed mainly of argillaceous lime mudstones

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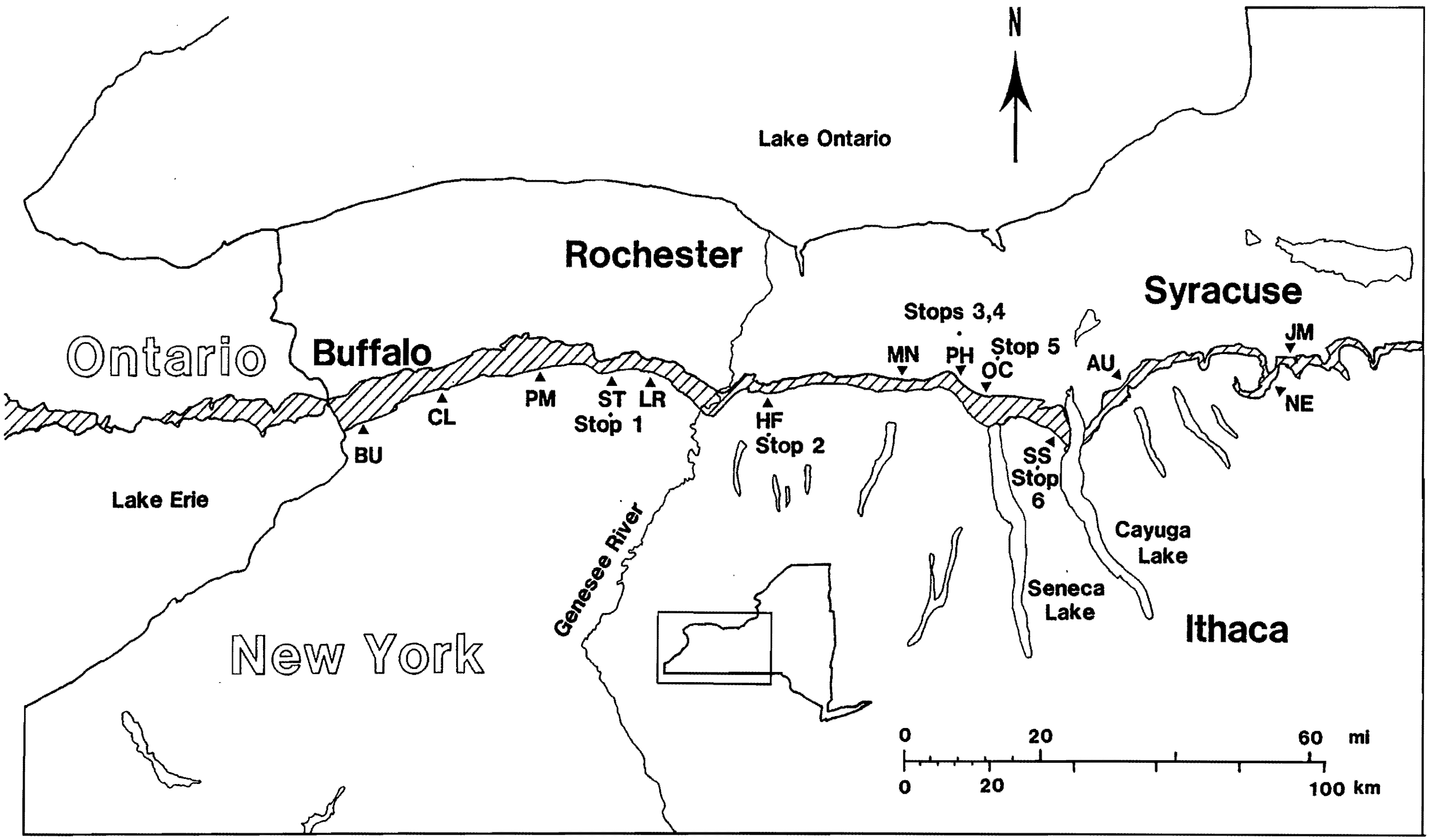
Figure 4. Outcrop map of the Onondaga Formation in west-central to western New York (modified after Rogers et al., 1990). Localities are as follows: BU=Buffalo, CL=Clarence, PM=Pembroke, ST=Stafford (Stop 1 of fieldtrip), LR=LeRoy, HF=Honeoye Falls (Stop 2 of fieldtrip), MN=Manchester, PH=Phelps (Optional Stops 3&4 of fieldtrip), OC=Oaks Corners (Stop 5 of fieldtrip), SS=Seneca Stone quarry (Stop 6 of fieldtrip), AU=Auburn, NE=Nedrow & Onondaga Indian Nation, JM=Jamesville.


N
Lake Ontario
Rochester
Syracuse
Ontario
Buffalo
Stops 3,4
Stop 5
MN
PH
OC
AU
JM
NE
BU
CL
PM
ST
LR
Stop 1
HF
Stop 2
SS
Stop
6
Lake Erie
Genesee River
Cayuga
Lake
Seneca
Lake
Ithaca
New York
0
20
60
mi
0
20
100
km

(biocalcisiltites or wackestones) interbedded with thin shales, particularly in the central portion of the Onondaga Formation. These light-weathering, micritic limestones are only slightly cherty relative to facies both east and west of the basin center area and carry a sparse fauna comprised mainly of dalmanitid trilobites and small brachiopods. Litho- and bio-facies of these more argillaceous rocks closely resemble the Onondaga-equivalent Selinsgrove Limestone (member of the Needmore Formation) exposed in cuts along the Susquehanna River at Selinsgrove Junction in central Pennsylvania, essentially due south of the Seneca Stone quarry. Indeed, detailed correlations (Ver Straeten, unpublished) reveal a precise correlation between meter-scale subdivisions of the Onondaga Limestone of the central Finger Lakes and those of the type Selinsgrove Member in central Pennsylvania. However, sections west of the type area, near the Allegheny Front, reveal that the Onondaga becomes even thinner towards the actual deepest portion of the sedimentary basin and is composed here of a larger portion of dark gray to nearly black shale, particularly in the upper beds (see below).

In the following section, we outline some of the new stratigraphic refinements to each unit of the Onondaga that have resulted from our recent field work. We emphasize those marker beds which can be correlated over considerable distances and which have utility for very detailed event and cycle stratigraphy within the Onondaga Formation.

Edgecliff Member

The Edgecliff Member is widely recognized as relatively coarse, coral to crinoidal-rich, non- to sparsely cherty strata in the lower part of the of the Onondaga Formation. Recent study across New York, and specifically in west-central to western New York, permits a new view of these strata. Our recent work demonstrates that much or all of the Clarence Member of the Onondaga Limestone (Ozol, 1964; Oliver, 1966b) in its type area of Erie County grades laterally into the middle and upper portions of the Edgecliff Member in its type area. Correlation of distinctive marker beds within the Nedrow Member across central to western New York demonstrate that the Nedrow overlies the Clarence throughout the region. Rickard (1975, 1989) originally noted this relationship based on subsurface log analysis. We have observed lateral gradations from non-cherty grainstone ("Edgecliff") to cherty packstone ("Clarence") facies within the wall of a single quarry (east wall of LeRoy Bioherm quarry; see Wolosz, this volume). Clarence facies is present and relatively thick (8.0 to 14 m) from Oaks Corners (Stop 4) westward to Buffalo. It is a very minor component of the Edgecliff interval from Seneca Falls (Seneca Stone quarry, Stop 6) eastward to Albany. However, thick, light-weathering, cherty, micritic facies reappear within the Edgecliff Member to the south of Albany along the Hudson Valley and into the Buttermilk Falls Formation in northeastern Pennsylvania. Because of the interfingering and lateral gradation of cherty, fine-grained limestone ("Clarence") and non-cherty crinoidal packstone, we choose to retain the term Edgecliff Member for the entire interval between the basal contact of the Onondaga Formation and the lowest shaly to argillaceous marker beds of the Nedrow Member. As such we retain the term "Clarence" as an informal name for the cherty, micritic facies that predominate in western <u>and</u> southeastern New York; we also propose the informal term "Jamesville Quarry facies" for the sparsely- to non-cherty crinoidal pack- and grainstones of central to eastern New York. A basal bed or "tongue" of the Jamesville Quarry facies is present in nearly all "Clarence facies"-dominated sections of the Edgecliff Member in New York and equivalent strata in northeastern Pennsylvania.

Near its type locality in the Syracuse area of Onondaga County (e.g., Jamesville quarry), the Edgecliff Member attains thickness of approximately 5.7 m. The member in its type area consists dominantly of chert-poor, crinoidal pack- and grainstone which we informally term the "Jamesville Quarry facies" of the Edgecliff Member. Chert occurs at two intervals locally, as a meter-thick package a short distance above the contact and as distinctive, yellowish-weathering, blue-gray chert nodules approximately 5 meters above the base of the member. The basal contact of the unit is sharp, but appears gradational because of the similar lithology of underlying uppermost Tristates Group strata. These upper 40 cm at the top of 2.2 m of post-Oriskany, Emsian-age strata (=Esopus?, Carlisle Center, and Schoharie Formations?) consist of silty to sandy packstone rich in large brachiopods such as *Amphigenia*. There is no basal Edgecliff fine-grained unit as appears in eastern New York sections (see Oliver, 1956a). The basal Edgecliff at Jamesville is rich in large crinoid columnals and also features a diverse rugose and tabulate coral assemblage. The upper boundary of the Edgecliff is sharply defined by a greenish-gray, silty, calcareous shale, which is rich in small corals, brachiopods, platyceratid gastropods, and other fossils. A similar thin shale layer occurs about 30 cm below the highest typical Edgecliff limestone, which suggests some interfingering of these two lithologies. A cherty interval near the top of the Edgecliff in the Syracuse area may correlate to a thin cherty band observed at the Seneca Stone quarry (see below).

To the west, the Edgecliff Member thins to a minimum of approximately 2.5 m at Seneca Stone quarry (Stop 6; see Figure 5). Here the unit is notably finer-grained, composed mainly of non-cherty, rather massively-bedded, wacke- to packstone. The Edgecliff, however, still carries large crinoid columnals near the base, and again near the top. The basal contact is clearly unconformable, although it appears gradational as a result of stratomictic processes. Lowest Edgecliff strata at Seneca Stone consist of a sandy conglomerate that includes clasts eroded from the Manlius Limestone and, more abundantly, dark, phosphatic sandstone cobbles derived from erosion of underlying Tristates Group strata seen preserved at nearby localities to the east. A thin biostrome just above the conglomerate in the basal portion of the unit is exceptionally rich in solitary and colonial rugose corals and small *Favosites*.

A distinctive 0.5 m-thick band of cherty calcisiltite (micritic limestone with black chert nodules) occurs 1.8 m above the base of the Edgecliff at the Seneca Stone quarry. This is the only trace of Clarence-like, cherty lithology at this section. The uppermost portion of the Edgecliff, which sharply overlies this cherty band, consists of crinoidal wacke- to packstone. The upper contact is sharply defined by an abrupt change to sparsely fossiliferous, medium dark gray, very shaly limestone of the Nedrow Member.

A most dramatic facies change within the Edgecliff Member occurs between the Seneca Stone quarry and the next major outcrop 21 km to the northwest near Phelps (Oaks Corners quarry, Stop 5). At this section (see Figure 5) the Edgecliff interval is much thicker, and

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Figure 5a. Four quarry sections of the lower part of the Onondaga Limestone and associated strata in west-central to western New York. Datum = Nedrow-Moorehouse Members contact. See Figure 4 for quarry section localities. Onondaga Formation lithology = limestone except where otherwise shown. Bold lines = datum or formational boundary; thin lines = member boundaries; dashed lines = various correlated beds. Key: 1=black shale, 2=dark, generally calcareous shale, 3=interval of pyrite nodules, 4=limestone nodules, 5=reworked phosphatic sandstone clasts, 6=nodular to bedded chert, 7=covered interval.

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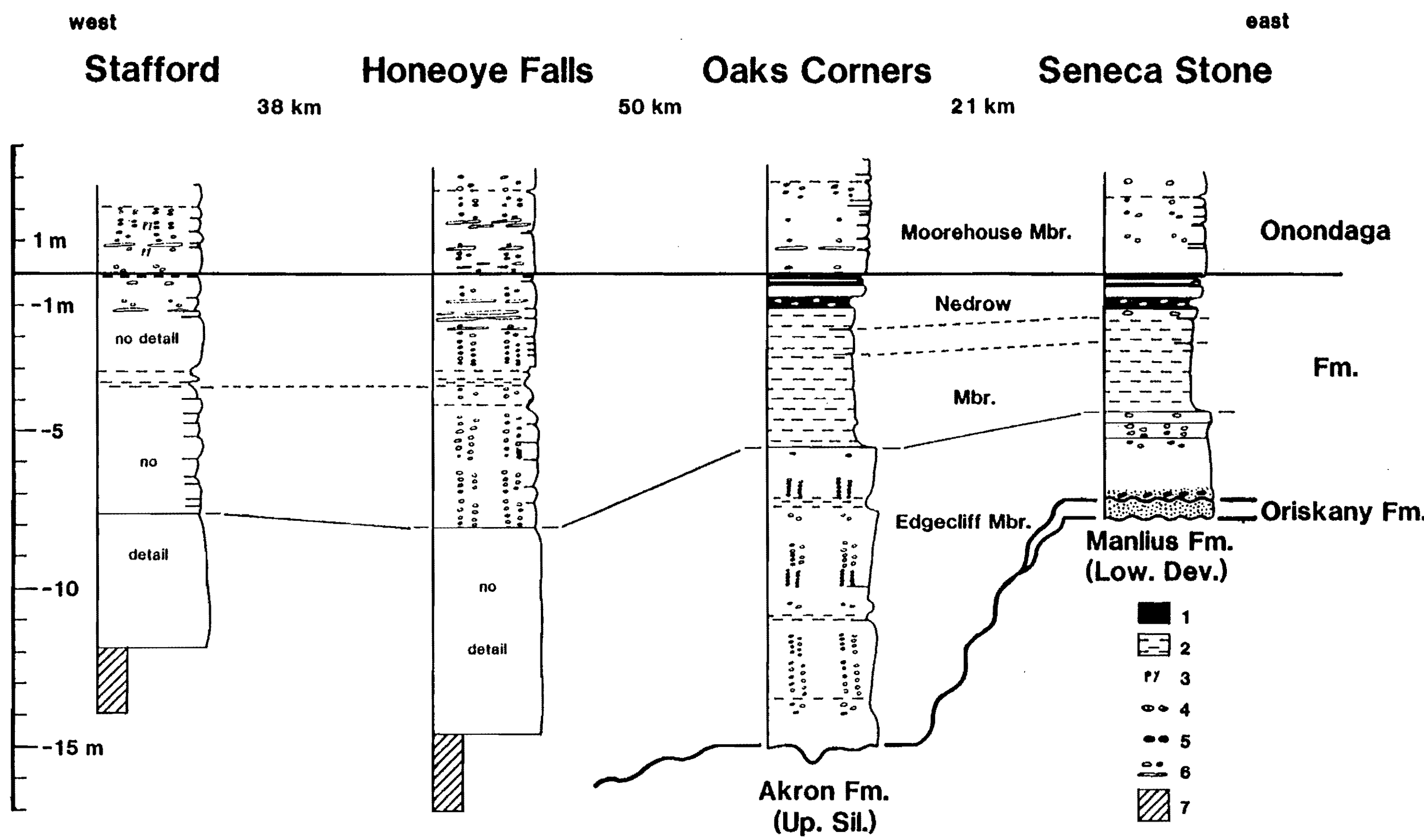
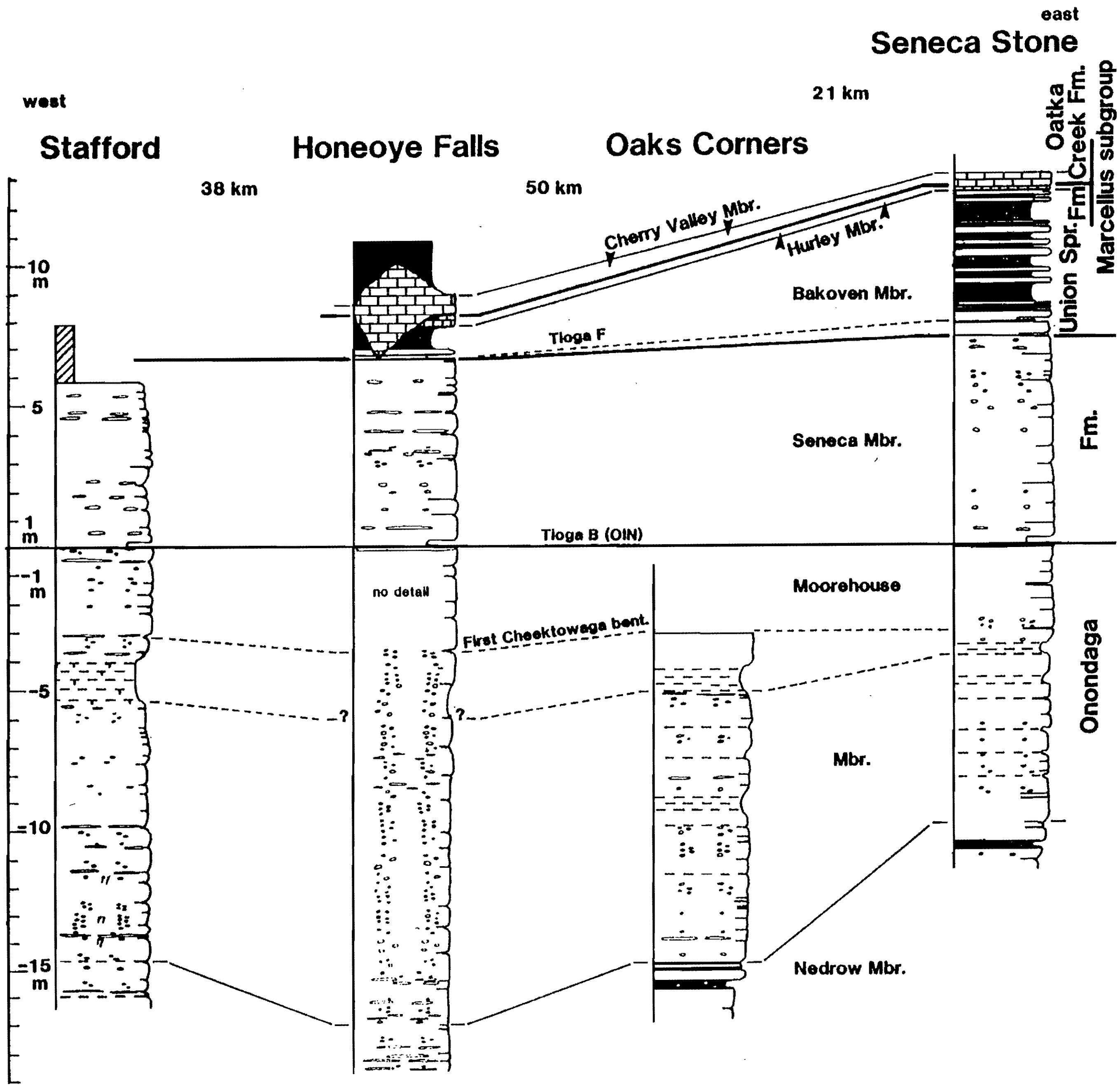
Figure 5b. Four quarry sections of the upper part of the Onondaga Limestone and lower part of the overlying Marcellus Shale in west-central to western New York. Datum = Tioga B (=OIN) Bentonite. Not all chert bands shown for Stafford quarry. Informal revised stratigraphy of the "Marcellus subgroup" after Ver Straeten et al., this volume. For Key see Figure 5a; arrows point to Tioga bentonite beds; q=shale bed with scattered quartz granules to pebbles at Seneca Stone quarry.



west
east
Stafford
Honeoye Falls
Oaks Corners
Seneca Stone
38 km
50 km
21 km
1 m
-1 m
-5
-10
-15 m
no detail
no
detail
no
detail
Moorehouse Mbr.
Onondaga
Nedrow
Fm.
Mbr.
Edgecliff Mbr.
Oriskany Fm.
Manlius Fm.
(Low. Dev.)
Akron Fm.
(Up. Sil.)
1
2
3
4
5
6
7



west
east
Stafford
Honeoye Falls
Oaks Corners
Seneca Stone
38 km
50 km
21 km
Cherry Valley Mbr.
Hurley Mbr.
Bakoven Mbr.
Tioga F
Seneca Mbr.
Tioga B (OIN)
Moorehouse
First Cheektowaga bent.
Mbr.
Nedrow Mbr.
no detail
Oatka Creek Fm.
Union Spr. Fm
Marcellus subgroup
Onondaga Fm.
10 m
5
1 m
-1 m
-5
-10
-15 m

is comprised primarily of sparsely fossiliferous calcisiltites with abundant bands of dark gray to black chert, closely resembling the typical western New York Clarence facies. The total thickness of the Edgecliff Member ranges from about 8.5 to 9.2 m. This variability is a result of prominent channeling at the base of the unit. Channels with a relief of up to 70 centimeters are cut into the underlying Akron Dolostone. These low spots are infilled with more typical Edgecliff lithology which consists of crinoidal wackestone to packstone up to almost a meter in thickness, but in places as little as 30 cm thick. These are overlain by about a half meter of transitional stylonodular limestone with small, dark chert nodules. The next 3.5 m consist of distinctly cherty, micritic limestone which is capped by about 30 cm of light-weathering, crinoidal packstone which forms a distinct pale gray band in the quarry walls. This crinoidal packstone is abruptly overlain by slightly argillaceous, cherty micritic limestone, resembling that immediately below, which persists through another 2.1 m and is capped in turn by a second light-gray weathering crinoidal packstone that features very large crinoid columnals. This third coarser crinoidal interval is, in turn, overlain by about 0.5 m of dark, gray, non-cherty, argillaceous calcisiltite which closely resembles the typical Nedrow lithology seen higher in the section. This shaly interval weathers to form a distinct notch in the quarry face and is separated from the typical Nedrow by about 1.2 meters of additional Clarence cherty, micritic lithology capped by a sparsely fossiliferous crinoid-bearing wackestone. As at the Seneca Stone quarry, there is an abrupt change from this top bed of the Edgecliff Member into overlying nodular, non-cherty argillaceous Nedrow facies.

The details of the Edgecliff interval are not as clearly shown in quarries to the west. The interval is not exposed in a quarry near Manchester (13 km west of Oaks Corners quarry), and is poorly exposed near the base of the Honeoye Falls quarry south of Rochester. However, in the latter a somewhat comparable thickness of typical cherty Clarence facies has been observed overlying non-cherty Jamesville Quarry facies lithology.

Near Clarence, New York, in the Buffalo area, the Edgecliff Member is approximately 12 to 14 m-thick (Oliver, 1966b) and dominantly consists of the chert-rich Clarence facies. However, previous estimates of the thickness of the Edgecliff in western New York may have included a somewhat cherty, argillaceous interval at the top which we recognize as equivalent to the Nedrow Member to the east. Hence, the Edgecliff which appears abruptly between the Seneca Stone and the Oak Corners quarries, thickens some to the west, and then appears to maintain a more nearly uniform thickness from LeRoy to the area to Erie County . At present, it is not known whether the three small-scale cycles of the Edgecliff clearly recognizable at the Oaks Corners quarry, can be correlated over this region.

Nedrow Member

The Nedrow Member was defined by Oliver (1954) as a thin interval (approximately 4.5 m-thick by original definition) of argillaceous and typically rather fossiliferous limestone, characterized by platyceratid gastropods. The Onondaga Indian Reservation quarry in the Town of Nedrow, south of Syracuse was designated as the type section. The interval was correlated by Oliver, both east and west of this area; however, stratigraphers always encountered difficulty in tracing the unit into western (and eastern) New York. As previously stated, the Nedrow had been interpreted by some authors as laterally correlative with the "Clarence Member" (Clarence facies herein) in western New York (see Buehler and Tesmer, 1963, Oliver, 1966b.) However, later workers recognized that in west-central New York the Nedrow Member in places overlies cherty facies, comparable to Clarence (Rickard, 1975; Lindemann and Feldman, 1981). In most areas, from eastern New York at least to the Seneca Stone quarry (Stop 6), the thin Nedrow argillaceous limestone beds directly overlie pack- to wackestone deposits of the Edgecliff Member with a sharp, but conformable contact. Recently, we have recognized several marker beds within the Nedrow Member that facilitate detailed correlations into western New York sections, where they prove that the Nedrow indeed is a separate stratigraphic unit distinct from the underlying chert-rich strata (Clarence facies) of the Edgecliff Member. The same marker beds also permit correlation into the central part of the Appalachian Basin in Pennsylvania (Ver Straeten and Brett, 1994; see below).

Oliver's original description of the Nedrow-Moorehouse contact in the type area placed the member contact within an interval of interbedded shales and limestones. We have found that the specific bed at the contact is locally correlatable through a detailed bed-by-bed examination, but it is difficult to correlate very far. An informal criterion used to recognize the Nedrow-Moorehouse Member contact outside of the type area has been the position of the lowest dark chert above argillaceous strata; the lowest chert changes position, however, across facies changes, which makes it difficult to recognize a time-equivalent member boundary. The authors have recently recognized a pair of black to dark gray shale-dominated beds within the Nedrow Member that are widely correlatable across a large area of the northern and central parts of the Appalachian Basin (see Pennsylvania discussion below). Due to the ease of recognition and the widespread mappability of the two black shales, we functionally use them as the upper boundary of the Nedrow Member. We do note, however, that Oliver's original position of the Nedrow-Moorehouse boundary may be of significance; that bed has been located at the Seneca Stone quarry (Stop 6), where uncommon quartz pebbles and granules occur within a shaly layer.

Sections at the Oaks Corners and Seneca Stone quarries (Stops 5 & 6; Figure 5) display the Nedrow Member in its apparently most basinal facies along the New York outcrop. At the Seneca Stone quarry, the Nedrow is approximately 4.2 meters thick. It displays a relatively abrupt basal contact with the underlying upper wackestone unit of the Edgecliff Member. The basal shales are typically medium dark gray, but contain scattered fossils, particularly the brachiopods *Pseudoatrypa*, *Leptaena,* and small rugose corals, as well as pyritic burrows. The lower three meters of the Nedrow at this location consist of alternations of medium gray, sparsely fossiliferous, calcareous shales, and gray, somewhat nodular, highly argillaceous and noncherty limestones. Several distinctive crevices within the interval may represent K-bentonite beds.

About three to four meters above of the base of the unit at Seneca Stone, the Nedrow displays four thin (<5 cm-thick) bands of very dark gray shale. A thicker, nearly black calcareous shale band approximately 30 cm-thick forms a prominent marker bed 3.4 m above the base of the Nedrow. The black band locally contains two or more levels of small, ellipsoidal, black chert nodules and a thin middle limestone. This black marker unit, one of the most important regionally correlative beds within the Onondaga Formation, is overlain by a second important marker, a ledge-forming, light gray-weathering, somewhat fossiliferous wackestone 33 cm thick. This band, which is equally traceable, is overlain by an additional distinctive dark gray shale interval, up to about 20 cm in thickness. In detail this thin interval consists of two dark shales separated by a thin limestone bed. Unlike the lower dark to black shaly beds, this unit contains relatively abundant, small- to medium-sized brachiopods, particularly the orthid *Schizophoria*. This unit, informally referred to as the "*Schizophoria*" bed, has been located at least as far east as Syracuse and westward at least to the Manchester quarry. A similar, dark gray, relatively fossiliferous shaly band occurs to the west, and forms an easily recognized upper boundary for the Nedrow Member as defined herein. The triplet of the thick lower black shale, the light-colored middle micritic limestone, and the overlying thin, dark gray shale has recently been correlated throughout much of the northern to central Appalachian Basin (Ver Straeten and Brett, 1994; see below). It is identifiable in the Nedrow of the Syracuse region, in the middle portion of the Selinsgrove Limestone of central Pennsylvania, and in the Onondaga Limestone of the mid-Hudson Valley area of New York (Kingston region).

The Oaks Corners and Seneca Stone quarries (ca. 21 km apart), display very similar Nedrow sections. At Oaks Corners, however, the member overlies Clarence cherty facies as noted above. This quarry section is very significant in that it displays the complete Nedrow interval in context with the Clarence facies, and clearly demonstrates that the Nedrow is a younger stratigraphic unit, while "Clarence" strata represent Edgecliff-equivalent, chert-rich facies. At Oaks Corners, the Nedrow is somewhat thicker (ca. 5.4 to 5.5 m-thick) than in the section at Seneca Falls. The member displays the same distinctive marker beds seen at the Seneca Stone quarry. As previously noted, the basal shales abruptly overlie a crinoidal wackestone at the top of the Edgecliff Member. The Nedrow fauna at Oaks Corners is dominated by atrypid brachiopods, *Leptaena*, and small rugose corals. The lower four meters of the Nedrow consists of rhythmically interbedded calcareous shales and marly nodular limestones. These occur in clusters that appear to define four small scale (ca. 1.1 m-thick) shallowing-upward cycles. Each cycle contains four to five thin nodular limestones with interbedded medium to dark gray calcareous shales. In the upper part of the member there are about three thin, darker gray shale bands that are overlain by approximately 30 centimeters of black, chert-bearing shale, equivalent to the lower black shale marker bed seen at Seneca Stone quarry and elsewhere. The black bed is overlain by a prominent, light-weathering, noncherty micritic (calcisiltite) ledge which in turn underlies the upper marker unit of the Nedrow Member as defined herein, i.e., the *Schizophoria*-bearing dark shale unit. As at Seneca Stone quarry, this unit is characterized by two thin shales surround a middle, 10 cm-thick, argillaceous limestone bed. This horizon forms a prominent parting, and has been used as a lift level in the quarry which provides good exposures for examination of this shale and its small brachiopod fauna.

Excellent exposures of the Nedrow Member occur along Oak Orchard Creek immediately south of Route 96, just west of the Town of Phelps. These display the interval of the lower rhythmically-bedded calcareous shales and nodules (see description of optional Stop 4). The lower shaly beds along the creek contain abundant, well-preserved brachiopods and scattered small rugose corals. Among the prominent faunal elements are robust, articulated specimens of *Pseudoatrypa*, *Leptaena*, and *Pentagonia*. Upper thin argillaceous limestones and calcareous shales are only sparsely fossiliferous. The upper part of the Nedrow Member, including the black beds, is not exposed at this outcrop.

A comparable succession has been measured at the Manchester quarry, 13 km west of Phelps. Here, the prominent lower dark shale bed displays rusty weathering, somewhat larger, black chert nodules. Again, a shaly parting about 65 cm above displays abundant *Schizophoria*, and other small brachiopods.

To the west, correlation of the Nedrow Member becomes more difficult as the result of rather prominent facies change within the interval. However, at sections in the Honeoye Falls quarry (see Stop 1 description), and in quarries to the west near LeRoy and Stafford (see Figure 5), the interval is still recognizable. It is best displayed in the old Gulf Road quarries approximately 3 km northeast of LeRoy. Here the Nedrow Member is 7.7 m-thick and is distinctly cyclic in appearance. The lower part (ca. 6.0 m) consists of four alternations, which may correspond, in part, to the four small cycles observed at the Oaks Corners quarry. At LeRoy, the cyclic interval consists of fine, sparsely fossiliferous, medium gray calcisiltite with abundant dark gray to black chert nodules that alternate with lighter gray, crinoid- and coral-rich packstones with light gray chert. The base of the unit displays a relatively sharp contact with upper bed of the underlying Edgecliff Member, which carries an abundance of large, chert-replaced tabulate and rugose corals. The basal 20 cm of the Nedrow that overlies this coral bed is a dark gray, nearly barren argillaceous limestone, or very calcareous shale. A thin fossil-rich bed occurs at approximately one meter above the base of the Nedrow, and a second at about 2.5 meters above the base. A third, distinct, 20 to 40 cm thick, light-colored, very fossiliferous coral-rich limestone is overlain by an approximately 70 cm-thick, greenish, more argillaceous unit. This unit can be subdivided into a thin, coral-rich shale bed, a 10 cm nodular limestone and an upper, thin, less fossiliferous shale. Dark gray and somewhat shaly nodular chert layers and an 8-

cm-thick shale about 1.4-1.8 m above the green argillaceous unit are considered to be the probable equivalent of the lower black shale marker bed at the Seneca Stone, Oaks Corners, and Manchester quarries. The shale is overlain by a distinctly light buff-weathering, sparsely cherty limestone (ca. 90 cm-thick at LeRoy and ca. 1.0 m-thick at Stafford). This buff band, which may be subdivided into three intervals, contains scattered corals in a light micritic matrix. It is overlain, in turn, by another thin dark gray shale which appears to correlate with the *Schizophoria* bed at the top of the Nedrow Member at Seneca Stone quarry and other sites.

At present, the Nedrow Member has not been correlated west of the LeRoy-Stafford area. However, gamma ray logs correlated by L.V. Rickard (1989) show a similar signature at the top of the Edgecliff Member in the Buffalo area to that displayed in the known Nedrow section of the LeRoy to the Manchester area.

Moorehouse Member

The Moorehouse Member was named by Oliver (1954) for Moorehouse Flats near the Onondaga prison quarry southeast of Syracuse, New York. At that location, the Moorehouse consists of approximately 12.5 m of medium to dark gray, cherty, fine-grained limestone with thin, very calcareous shale beds and a thicker, more prominent shale about eight meters above the base of the unit. The thick shaly zone is recognized by the authors across New York State and in the Selinsgrove Limestone of central Pennsylvania and the Buttermilk Falls Limestone of eastern Pennsylvania (Ver Straeten and Brett, 1994; see below).

In the study area of west-central New York, the Moorehouse Member ranges from approximately 9.8 m-thick at the Seneca Stone quarry, one of its thinnest sections, to about 14.9 m in the Stafford-LeRoy area (see Figure 5). At the Seneca Stone quarry, the lower 6.4 m of the unit is composed primarily of burrowed, sparsely fossiliferous calcisiltite with a relatively small percentage of black chert. These are bounded by intervals of highly argillaceous limestone or very calcareous shale that seem to define at least six to seven small-scale cycles within the lower Moorehouse. The prominent shaly bands occur at about 1.5 to 1.7 m, 2.5 m, 3.3 m, 4,0 m, 5.0 m, and 6.0 m above the base of the unit. Notch-forming thin clays, which may represent K-bentonites, occur at several levels, notably at 0.95 m and 1.05 m above the base of the member. The thin shales tend to be relatively fossiliferous and contain skeletal debris of small crinoids, trilobites, and brachiopods. Rarely, articulated specimens of the trilobite *Odontocephalus* occur at those bedding planes. In some cases, these thin, fossil hash-bearing shales overlie irregularly burrowed surfaces that represent firmgrounds or possible hard grounds.

Chert becomes more prominent in the upper third of the Moorehouse, overlying the uppermost, prominent calcareous shale bed at 5.9-6.4 m above the base. This shaly unit, which we loosely term the "false Nedrow" shale, is medium to dark gray and somewhat more argillaceous than the remainder of the Moorehouse. In weathered outcrops, the "false Nedrow" appears shaly and is rich in pyritic burrows and small brachiopods. It appears to become somewhat more fossil-rich and more calcareous toward the top. The unit is widely traceable across the northern and central parts of the Appalachian Basin.

Two prominent notches that feature biotite-rich claystones occur at 3.0 m and l.3 m below the upper contact of the Moorehouse. These appear to represent the First and Second Cheektowaga Bentonites of Conkin and Conkin (1979, 1984; Conkin, 1987); the upper bed is the equivalent of the Tioga A Bed of Pennsylvania workers (Way et al., 1986). The highest beds of the Moorehouse Member, immediately below the Onondaga Indian Nation bentonite (OIN; = Tioga B bentonite of Way et al., 1986), contain specimens of the high-spired gastropod *Palaeozygopleura* and the enigmatic tubular fossil *Coleolus*.

At the Seneca Stone quarry, both the upper and lower contacts of the Moorehouse are sharply defined. The base can be drawn unambiguously at the top of the dark gray *Schizophoria*-bearing shale bed which is herein defined as the upper boundary of the Nedrow

Member. As previously stated, this definition of the Moorehouse base differs slightly from that used in the past. A thin shale approximately 2.5 m above the base, within which we have noted several quartz granules and pebbles, appears to correspond to the Nedrow-Moorehouse contact as originally defined by Oliver and shown in Figure 3. The upper boundary of the Moorehouse Member is readily identifiable here, as in most sections, as it is overlain by the 15 cm-thick Onondaga Indian Nation ("Tioga B") K-bentonite. This interval and an immediately underlying black, pyrite-rich, rusty-weathering chert bed form a distinctive marker in the wall of the Seneca Stone quarry. The OIN ash bed forms a prominent notch in the wall, and forms a clear separation from the overlying Seneca Member.

The section at Oaks Corners quarry displays an incomplete section of the Moorehouse Member. Again, its base is sharply defined at the *Schizophoria*-bearing shale beds of the top Nedrow as defined herein. Approximately 6-7 cyclic intervals, each capped by a prominent shale bed, can be recognized at this location and these appear to be correlative with cycles observed in the Moorehouse at the Seneca Stone quarry. The basal Moorehouse unit is a prominent massive limestone interval about one meter thick, overlain by twin beds in thin notches that represent possible K-bentonite beds. Prominent shale beds occur again about 2 and 2.5 meters above the base of the Moorehouse. The upper shale has yielded specimens of the stemless crinoid *Edriocrinus* and well-preserved bryozoans and sponges. A prominent, notch-forming calcareous shale interval about 80 cm-thick occurs at just over four meters above the base of the Moorehouse at this locality. The underlying and overlying beds are massive, light gray-weathering, fine-grained limestone (calcisiltite) with very prominent dark gray to brownish-gray chert nodules generally surrounded by buff weathering, dolomitic rims. A bed 1.4 m below the 80 cm shale carries abundant specimens of the large, coiled, frilled nautiloid *Gyroceras*.

Higher beds in the quarry, approximately 6 meters above the base of the Moorehouse, carry a diverse brachiopod fauna within argillaceous or shaly partings; particularly notable here are abundant *Atrypa*, *Leptaena*, and *Megakozlowskiella*. Trilobites, particularly *Phacops* and *Odontocephalus* are common in this horizon. This assemblage resembles the *Leptaena-Megakozlowskiella* community of Feldmann (1980). The widely recognizable Nedrow-like shale in the upper Moorehouse occurs approximately 10.5-11.5 m above the member base; the uppermost 1.2 m of cherty, fossiliferous limestone in the quarry represent the lower part of the coarser upper part of the Moorehouse Member.

Sections of the Moorehouse Member to the west, at Honeoye Falls and the LeRoy-Stafford area, not only display a thickening of the interval, but also an increased proportion of dark gray, chert-bearing beds, and several levels of crinoidal packstone or grainstone. Complete sections are displayed at both Honeoye Falls and Stafford quarries. At these sites again, the Moorehouse is delineated at its base by fossiliferous shales that correlate with the *Schizophoria* bed at the top of the Nedrow Member further east. The upper boundary is, as at all localities, sharply drawn at the base of the Tioga B-OIN bentonite. In these western localities, the Moorehouse clearly appears divisible into two submembers, separated by the very shaly "false Nedrow" limestone generally about one meter in thickness. The lower interval contains marker beds that appear to be correlative into the shaly caps of the six to seven cycles seen at Oaks Corners and Seneca Stone quarries. The first thin shale bed of the Moorehouse, about 90 cm above the top of the Nedrow Member, contains a diverse fossil assemblage. In the north quarry of the LeRoy Stone Company adjacent to Perry Road, this three cm-thick bed of dark gray, calcareous shale, has yielded an abundant and diverse fauna of brachiopods, including *Elytha, Schizophoria, Megakozlowskiella*, and others, as well as an abundance of small *in situ* stalked rhenopyrgid edrioasteroids (G.C. McIntosh, personal communication, 1990). The paired, thin, possible K-bentonite beds a short distance above the base in other sections may correspond to notches 2.3 and 2.6 m above the Moorehouse-Nedrow contact at the LeRoy quarries.

A prominent marker bed in the LeRoy and Stafford quarries, which occurs at 4.5 m above the base of the Moorehouse Member, is a recessive shaly and hackly-weathering dark chert bed. This may correspond to the interval of dark gray *Pacificocoelia*-bearing shales noted at the Oaks Corners quarry. The interval overlying this chert bed displays a distinctive coarsening-upward facies. About 2 m above the chert layer, in quarries at Stafford and LeRoy, is another very distinctive marker interval, consisting of biostromes of the fasciculate rugose coral *Synaptophyllum* and/or *Eridophyllum*. These thicket-like colonies occur in two beds over an interval of about 0.5 m. The beds are distinctive in that corals colonies are partially incorporated into dark chert nodules, in which the individual corallites stand out because of the contrast of their very light coloration to that of that surrounding chert. This marker bed was previously recognized by Oliver (1966c) and referred to as the *Eridophyllum* beds at Stafford. An interval overlying the coral biostrome bed, about l.25 to 1.5 meters thick, consists of crinoidal pack- and grainstones which are highly fossiliferous. The upper 30 cm of this interval is somewhat more argillaceous and has yielded abundant intact specimens of stalked rhenopyrgid edrioasteroids as well as relatively well preserved crinoids, primarily *Arachnocrinus*, *Tripleurocrinus* and *Schultzicrinus*.. These beds are overlain by argillaceous, *Zoophycos*-bioturbated limestone. This shaly, fine-grained, brachiopod-rich interval correlates with the prominent, thick, widespread, "false Nedrow" shale. Overlying this meter-thick interval of bioturbated, argillaceous limestone are about four to five meters of highest Moorehouse strata that feature chert nodules at several levels and, toward the top, intervals of crinoidal pack- and grainstone. At the Stafford quarry, the First and Second Cheektowaga Bentonites occur 3.6 and 0.5 meters below the top of the Moorehouse Member. As at more easterly localities, an abundance of gastropods and coleolids has been recognized in the highest beds of the member.

Seneca Member

The Seneca Member comprises the uppermost strata of the Onondaga Formation in New York. The name was first applied by Vanuxem (1839) for darker limestones with abundant chonetid brachiopods above the heavily chert-rich "Corniferous Limestone" of the old terminology. The Seneca Member as now defined (Oliver, 1954) is part of a general fining-upward trend that extends from the underlying upper part of the Moorehouse Member into the overlying Marcellus Formation. Smaller scale cyclicity is superimposed on this general deepening-upwards pattern, however. Lateral lithologic trends again show a general coarsening-outward east and west of the central New York trough. The Seneca Member is thickest in the central Finger Lakes Region. Regional trends laterally across the state show a slight thinning westward and a distinct bevelling and cutout of the Seneca Member eastward; the Seneca Member is reportedly absent from the Albany area of eastern New York, but pinches back in southward through the Hudson Valley and southwestward into northeastern Pennsylvania (Rickard, 1989).

The Seneca Member is less well exposed than the underlying Edgecliff-Clarence, Nedrow, and Moorehouse Members. The unit features a relatively low diversity fauna, especially toward the central New York trough. Chonetid and atrypid brachiopods may be abundant, along with less common *Leptaena* and other forms. Rugose corals are generally uncommon. Ostracod/minute brachiopod hash beds occur in some basinward sections. Fine to medium-grained wacke- to packstones dominate the Seneca Member, becoming finer-grained mudstones in parts of more basinal sections. In outcrop the member appears as thick-bedded to massive limestones separated by thin shaly to bentonitic partings. Light-weathering chert is common to rare in parts of the member, and disappears toward easternmost outcrops. Thin dark shales increase in number in the higher part of the unit.

The lower contact of the Seneca Member is sharply defined by the base of the prominent, 15-20 cm-thick bentonitic clay layer termed the Onondaga Indian Nation Ash (OIN; Conkin and Conkin, 1979, 1984; Conkin, 1987; = Tioga B bed of Pennsylvania, Way et al., 1986). It should be noted, however, that Conkin (1987) proposed a placement of the base of the Seneca approximately 0.47 to 1.25 m above the base of the OIN bentonite across New York at the position of a paracontinuity surface and biostratigraphic boundary.

The section at Seneca Stone quarry (see Figure 5) exposes the complete Seneca Member, which may be 7.1 or 8.4 m-thick dependent on the placement of the upper contact with the Union Springs Shale. The 20 cm-thick OIN bentonite lies at the base, overlain by two meters of argillaceous wackestones with a sparse fauna and scattered nodular chert. An overlying interval of upward-coarsening diminutive brachiopod or ostracod shell hash is capped by additional fine-grained limestones. Chonetid brachiopod coquinite beds of the "*Hallinetes* Zone" (Racheboeuf and Feldman, 1990; ="Pink *Chonetes* Zone" of Oliver, 1954) and an interval of small rugose corals that occur about 3-4 m above the base are widely correlatable through central New York. Thin dark shale interbeds begin to appear high in the section, above another coarsening-up package of limestone beds.

The position of the Onondaga-Marcellus contact in the type area of the Seneca Member is treated differently by different authors. The contact is well exposed at the Seneca Stone quarry (Stop 6), 8.5 km northwest the type section of both the Seneca Member and Union Springs Shale immediately south of the village of Union Springs. The debate centers around an interval of transitional strata between the two units which comprise Oliver's (1954) Zone L.

At 7.15 m above the base of the Seneca an apparent discontinuity occurs (base of Oliver's 1954 zone L). The surface of the bed is marked by a small-scale irregular topography and pieces of pinkish weathering, possibly hematitic limestone; overlying dark shale adhering to the surface exhibits scattered phosphatic fish bone fragments. The surface is overlain by fine-grained, dark, styliolinid-rich limestones and thin, interbedded black shales and bentonites that cap the carbonate succession; this post-discontinuity succession (Zone L) was assigned by Oliver (1954) to the Seneca Member. This transitional interval forms the top of the carbonate-dominated part of the section in the high walls of the Seneca Stone quarry. The beds were placed in the Seneca based on faunas found in apparently correlative strata 32 km to the east. Other authors have placed the contact lower in the section either at the 16 cm-thick parting associated with the upper, 12 cm-thick, Tioga "F" bentonite (Clarke, 1901; Cooper, 1930) or at the previously noted irregular burrowed surface 7.15 m above the base (Conkin and Conkin, 1979, 1984; Conkin, 1987). Conkin and Conkin (1979, 1984) also note a widespread bone bed (bone bed #7; Onondaga-Union Springs contact bone bed of Ver Straeten et al., this volume) and discontinuity on this same surface. The authors tentatively follow the usage of Conkin and Conkin, but more attention is needed to resolve the problem.

Two prominent Tioga K-bentonite beds occur: 1) at the base of the Seneca (OIN Ash = Tioga B of Pennsylvania workers); and 2) within the previously discussed transitional interval (7.6 m above OIN) at Seneca Stone quarry. Additional, thinner K-bentonites occur within the member also, chiefly concentrated in its upper part. As many as eight Tioga Ash Beds may occur in the Seneca Member at Seneca Stone quarry.

A complete section of the Seneca Member is also exposed at the Honeoye Falls quarry (Stop 2, ca. 75 km west of Seneca Stone quarry; see Figure 5). The Seneca is thinner at this locality (6.65 m-thick) and has become coarser-grained, composed of crinoidal wacke- to pack-stones; nodular to bedded cherts occur throughout the section. The fauna is more varied, but diversity still is relatively low. Chonetid brachiopods, characteristic of the member in central New York (including the "Pink *Hallinetes* Zone") have not been noted. More typical brachiopods of the Genesee Valley region include atrypids and *Leptaena*. A thin interval of corals occurs in the upper part of the Seneca Member at Honeoye Falls, closely associated with a five cm-thick bentonite bed approximately 1.3 meters below the top. Notably, the top of the Seneca at this locality is relatively coarse, and consists of crinoidal grainstones with numerous *Leptaena* brachiopods.

Approximately 6.65 m of wacke- to packstones of Onondaga-like lithology are capped by a discontinuity surface similar to that seen at Seneca Stone quarry. Resting on that surface in places is a thin veneer of dark shale overlain by approximately 15 cm of coarse, biotitic tuff to soapy, yellow-brown claystone and a thin (ca. 18 cm-thick) interval of styliolinid limestones. A prominent phosphatic lag deposit with scattered fish bone material occurs within the styliolinid limestones interval (discussed in Ver Straeten et al., this volume).

As at Seneca Stone quarry, as many as eight Tioga bentonites may be associated with the Seneca Member at the Honeoye Falls quarry, including the prominent OIN Ash bed and the previously mentioned 15 cm-thick biotite-rich bed in the base of the overlying Marcellus Shale. Notably, coarser-grained parts of the section appear, in some cases, to be concentrated about some K-bentonite beds. Several of the more notable thin K-bentonite beds in the Seneca at Honeoye Falls are discussed elsewhere in this paper with the Tioga Bentonites cluster.

Regional study of the Seneca Member across New York indicates a more complex picture than that of the underlying members. Of five complete sections studied between Honeoye Falls and Cherry Valley (ca. 235 km, west to east), the Seneca Member appears thickest at the Seneca Stone quarry in the central Finger Lakes area (7.15 m-thick). Westward the member thins somewhat to 6.65 m south of Rochester (Honeoye Falls quarry), but there does not appear to be any significant difference in the successions at each quarry. East of Seneca Stone quarry, however, the top of the member is progressively beveled off, from 5.0 m at Jamesville (south of Syracuse) to 3.2 m at Oriskany Falls to 2.0 m at the classic Cherry Valley exposures. No Seneca Member is reported east of Cherry Valley; Rickard (1989, Plate 31) projects a total absence of the unit in the Helderberg-Albany area, and its reappearance southward, where it thickens towards northeast Pennsylvania. Uppermost Onondaga strata are exposed at several localities in the Helderberg-Hudson Valley region in the east, but the prominent OIN bentonite has not been reported.

The junior author has recently found and positively identified several K-bentonite layers in the upper part of the Onondaga in eastern New York. In the Helderberg region two notable bentonites, approximately 8-10 cm-thick, occurs approximately 2-3 m below the top of the Onondaga and at the Onondaga-Union Springs contact. Either bed could potentially be the OIN-Tioga B ash; however, the OIN is generally on the order of 15-25 cm in thickness. On the other hand, all other K-bentonite beds in the Onondaga of New York are thin relative to these beds, rarely up to five cm-thick. At this time, no other marker beds that are associated with the OIN farther to the west have been identified in the Helderbergs. Further work is needed, but if the lower bed does prove to be the OIN, then it is the first report of the Seneca Member east of Cherry Valley.

The eastward cutout of the Seneca Member appears to be associated with a regional submarine unconformity, as stated by Rickard (1984). The progressive eastward loss of upper Seneca strata is confirmed by the disappearance of upper beds of the Tioga interval east of Seneca Stone quarry. It may be notable, however, that black shales of the Union Springs Shale overlie progressively coarser facies in an eastward direction, an indication that shallower platform conditions persisted toward the east through deposition of the lower part of the Seneca Member. The fact that fine-grained black shales overlie relatively coarse deposits in the east, that a phosphatic pebble-bone bed occurs at the contact, and that no bentonite layers have as yet been found in the overlying black shale succession seems to support the idea of a regional unconformity at the top of the Onondaga. The Onondaga-Marcellus contact is discussed in more detail in the accompanying paper by Ver Straeten et al. (this volume).

Reports of the thick, upper Tioga "F" (Tioga "restricted" of Conkin and Conkin, 1979, 1984; Conkin, 1987) immediately above the Onondaga as far east as Catskill are apparently erroneous; the bed appears to be cut out by the unconformity as far west as Jamesville, approximately 65 km east of Seneca Stone quarry.

TIOGA BENTONITES CLUSTER

Geologists' understanding of the package of volcanogenic strata collectively termed the "Tioga Ash Beds" or "Tioga Bentonites" has evolved rapidly over the last 15 to 20 years. Early reports and correlation of a single bentonite layer have now been superceded by recognition of a cluster of K-bentonite layers concentrated within, but not exclusive to, the upper part of the Onondaga Formation and equivalent strata across the Appalachian Basin and eastern North America.

Altered volcanic ashfall strata of Paleozoic age are generally termed K-bentonites (or metabentonites) due to the greater alteration of the original volcanic glass fragments to potassium-rich clays (mixed illite-smectite to illite). K-bentonite beds, generally considered to represent single volcanic ashfall events deposited on the order of hours to days, are ideal isochronous layers important to the stratigrapher. Their use in construction of a high resolution stratigraphy yields crucial data for a detailed basin analysis and for the timing of geologic and biologic events. Furthermore, bentonite layers in foreland basins adjacent to deeply-eroded mountain belts generally preserve the best record of paleovolcanic activity during an orogenic episode.

The prominent Tioga B-OIN bentonite bed that marks the Moorehouse-Seneca member contact in New York was first noted by James Hall (1843) who reported an "unctuous" clay layer in the upper part of the Onondaga Limestone. A possible volcanic origin for the bed was first suggested by Luther (1894). The term "Tioga Bentonite" was first used by Fettke (Ebright et al., 1949; Fettke, 1952) who recognized biotite-rich bentonite in well cuttings, which he used for subsurface correlation in Pennsylvania. Oliver (1954, 1956a) first recognized the bed now termed the OIN Ash (=Tioga B of Pennsylvania workers) in outcrops in New York and used it to mark the base of the Seneca Member.

Dennison was the first worker to specifically focus attention on the Tioga Ash Beds, initially in his Ph.D. work (1960, 1961) and subsequently in a series of papers that address such issues as distribution and isopach thicknesses of the Tioga Bentonites, potential volcanic sources, and even Devonian paleowind directions based on the Tioga ashfall record (Dennison, 1986; Dennison and Textoris, 1970, 1978, 1987). Conkin and Conkin (1979, 1984) have also made major contributions to Tioga studies, chiefly through recognition of numerous additional bentonites in the Tioga interval and their distribution from New York cratonward into Ohio, Indiana, and other areas.

The most detailed analysis of the cluster of Tioga Bentonites within the Appalachian Basin to date is from central Pennsylvania (Smith and Way, 1983; Way et al., 1986). Way et al. (1986) present a microstratigraphic correlation of seven Tioga Ash Beds (Beds A-G) across 280 km of the Valley and Ridge Province in central Pennsylvania (Figure 6). Rickard (1984) presented another correlation scheme for four bentonite beds in the Tioga Interval (Beds A-D) based on subsurface gamma ray log correlation (Rickard, 1984, p. 824, states that his beds B and C merge toward central New York. Field study shows this not

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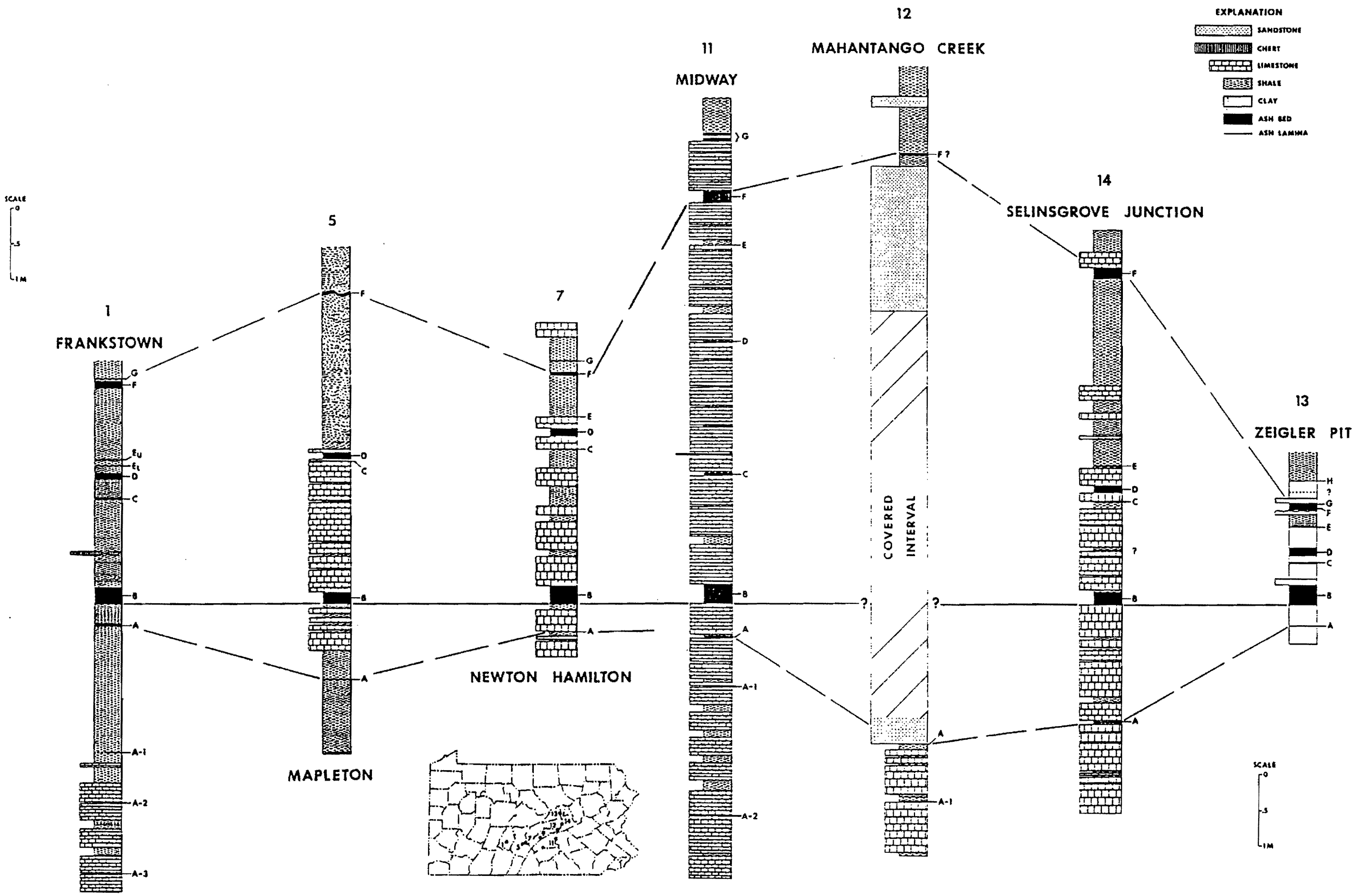
➤ Figure 6. Cross-section of the Tioga Bentonites Cluster across central Pennsylvania (modified from Way et al., 1986). Localities 5, 7, 11, and 12 correspond to Localities MP, MI, MH, and SJ of this paper (Figure 7), respectively.


EXPLANATION
SANDSTONE
CHERT
LIMESTONE
SHALE
CLAY
ASH BED
ASH LAMINA
SCALE
1
FRANKSTOWN
5
MAPLETON
7
NEWTON HAMILTON
11
MIDWAY
12
MAHANTANGO CREEK
COVERED INTERVAL
14
SELINSGROVE JUNCTION
13
ZEIGLER PIT

to be true). Way et al.'s (1986) terminology of Tioga beds A-G is utilized in this paper and is recommended for standard usage in favor of Rickard's scheme due to: a) the problems of translating well log thicknesses to cm-scale outcrop study; and b) the highly detailed resolution of the Pennsylvania workers scheme. The authors also recognize the usefulness of specific names for very distinctive, key beds (e.g., Onondaga Indian Nation (OIN) Ash of Conkin and Conkin, 1979, 1984; Conkin, 1987). The use of the name "Tioga" for a specific bed within the interval (e.g., "Tioga (restricted) Ash" of Conkin and Conkin, 1979, 1984; Conkin, 1987), however, should be suppressed.

Lithology of the Tioga bentonites ranges from coarse biotite crystal tuffs to tuffaceous clays to claystones that generally appear in outcrop as recesses between more resistant carbonate beds. Beds may appear dark to tan to gray or yellow, and leave a slippery, soapy feel when rubbed between the fingers. Biotite, which may appear bleached of color, is commonly recognizable in the field in some beds. Volcanogenic zircons, apatites, and beta quartz fragments are common accessory minerals that aid in positive identification of beds as altered volcanic ashfall layers. Preserved pumice may also be found. Some beds may show distinct layering, as if formed by several different ashfall events. Geochemical analysis of pristine glass inclusions that are found inside of volcanic quartz from the beds indicate a high-K rhyolitic composition of the source magmas (Schirnick and Delano, 1990, 1991).

A key aspect of recent Tioga studies is the recognition of a multiplicity of K-bentonite layers in Onondaga and equivalent strata in New York and across eastern North America. As previously noted, Conkin and Conkin (1979, 1984) and Rickard (1984) report four or more K-bentonites from the Onondaga Formation in New York. The junior author recognizes a number of additional possible K-bentonites in the Onondaga across New York (we state "possible" as some beds appear very similar to other known bentonite layers, but have not as yet been positively identified as such). Approximately eight possible Tioga bentonite beds occur in the Seneca Member and the base of the overlying Union Springs Shale at the Honeoye Falls quarry (Stop 2) and at the Seneca Stone quarry (Stop 6). The OIN bed and the thick bed in the base of the Union Springs (= Tioga "restricted" bed of Conkin and Conkin, 1979, 1984; Conkin, 1987) we correlate with the Tioga B and Tioga F Beds of Way et al. (1986; see Figure 6). Correlations of other beds in the Seneca between New York and Pennsylvania are tentative at this time.

Additional beds occur in underlying Moorehouse, Nedrow, and possibly the Edgecliff Members at these localities, including the First and Second Cheektowaga Bentonites of Conkin and Conkin (1979, 1984; Conkin, 1987) which occur in the upper part of the Moorehouse Member across New York, above the "false Nedrow" shaly interval in the middle of the member. Detailed work by the authors shows that the Second Cheektowaga Bentonite is the equivalent to the Tioga A of Way et al. (1986). Apparent K-bentonite layers in the lower three members of the Onondaga Formation tend to be relatively thin, generally on the order of 1-2 cm or less, rarely up to 5 cm in thickness.

In addition, one of the authors (CAV) has found several previously unreported bentonite beds in the upper part of the Onondaga Limestone of the Helderberg-Hudson Valley region of eastern New York (two of these are discussed with the Seneca Member above). These eastern beds, however, cannot at this time be correlated into central to western New York sections.

The Eifelian-age Tioga Ash Beds represent the youngest of three known Devonian ashfall-rich clusters in the Appalachian Basin. Lower Devonian bentonite-rich strata occur in the Lockhovian-age Kalkberg Formation (Helderberg Group; Bald Hill Bentonites of Smith and Way, 1988) and the late Pragian- or Emsian-age lower part of the Esopus Formation (Tristates Group; Sprout Brook Bentonites of Ver Straeten, 1992, ms. submitted; Ver Straeten et al., 1993) and equivalent strata across the Appalachian Basin. Notably, the Tioga and the Sprout Brook bentonite intervals mark major transitions between platform orthoquartzite-carbonate suites and overlying basinal dark gray to black shales (Onondaga

Limestone to Marcellus Shale and Oriskany Sandstone-Glenerie Limestone to Esopus Shale, respectively). In both cases deposition of ash-rich strata associated with an apparent increase in volcanic activity coincided with subsidence of the foreland basin (Ettensohn, 1985) and a eustatic rise in sea level (Johnson et al., 1985). These events were, in part, concurrent with a flush of fine-grained siliciclastics into the basin during the onset of two separate tectophases of the Acadian Orogeny (Ettensohn, 1985).

Rhyolitic volcanic rocks biostratigraphically equivalent to the Lower Devonian Bald Hill and Sprout Brook Bentonite intervals occur in the Northern Appalachians. However, no volcanic rocks equivalent to the Middle Devonian Tioga bentonites are reported from the Appalachians. Dennison and Textoris (1970, 1978), based on regional isopachs, project a potential source area for the Tioga Ash Beds in northeastern Virginia.

DISCUSSION

ONONDAGA FACIES GRADIENTS AND DEPOSITIONAL ENVIRONMENTS

Dark gray, sparsely fossiliferous, calcareous shales to very argillaceous limestones (Nedrow Member-type facies) appear to have accumulated toward the basin center. These were relatively starved of carbonate input. The source of the siliciclastics remains uncertain. However, the relative increase of shaly strata southward nearer to the main trough of the Appalachian foreland basin in central Pennsylvania suggests a possible southeastward source area. Indeed, the entire Onondaga interval appears to pass into shaly facies in northern Virginia and West Virginia (Dennison, 1960, 1961). The most distal Nedrow-type facies are represented by very dark gray to black, laminated calcareous shales. These facies carry a very sparse fauna of brachiopods (e.g., ambocoeliids), and styliolinids. Slightly lighter gray facies display abundant burrows of *Chondrites.* In places these calcareous shales contain a low diversity of brachiopod-dominated fauna with occasional small, solitary rugose corals, such as *Amplexiphyllum* and small *Heterophrentis.* These biofacies were referred to as the *Amplexiphyllum -chonetes* and *Amplexiphyllum-Odontocephalus* communities by Lindemann (1980); the brachiopod components are assignable to the low diversity *Atrypa* community and rarely to the *Pacificocoelia* community of Feldmann (1980).

The shaly Nedrow facies are interbedded with and grade laterally into argillaceous lime mudstones (calcisiltites) with sparse, typically fragmentary fossil assemblages, including minor crinoid debris, atrypid brachiopods, and, particularly the dalmanitid trilobite *Odontocephalus.* Such argillaceous limestones are typically non-cherty and form blocky tabular to slightly nodular bands, interbedded with calcareous gray mudstones. The sharp contrast in resistance between beds due to varying siliciclastic content has probably been diagenetically enhanced by early cementation of the more carbonate-rich intervals (for similar examples see Eder, 1982; Hallam, 1986; Ricken, 1991). These facies are typical of the Selinsgrove Limestone of central Pennsylvania (Inners, 1975) as well as parts of the Nedrow and Moorehouse members in portions of central New York State.

These fine-grained argillaceous limestones appear to grade laterally in some cases into sparsely fossiliferous lime mudstones which are less argillaceous and display layers and pods of gray chert. The latter typically appears to outline relatively large, branching burrow galleries of *Thallasinoides* type.

The Clarence facies of the Edgecliff Member is typically quite sparsely fossiliferous, although it locally contains medium-sized rugose corals, particularly *Heterophrentis,* fenestellid bryozoans, and brachiopods, such as *Atrypa.* Locally, these cherty facies appear to interfinger with and grade laterally in upramp directions into skeletal wacke- to packstones that are particularly rich in medium to large camerate crinoid columnals (Jamesville Quarry facies of the Edgecliff Member). Biostromes of fasciculate rugose and tabulate corals, such as *Syringopora, Acinophyllum, Synaptophyllum, Cylindrophyllum,* and *Eridophyllum* occur at some levels. These slightly muddy facies commonly display graded beds of skeletal debris passing upward into calcisiltites that suggest storm winnowing and deposition. Hence, these facies appear to have been deposited in areas below normal wave base, but above storm-wave base.

Finally, biostromal wackestone facies pass apparently upramp into coarse, typically winnowed and, in some cases, cross-laminated crinoidal pack- and grainstones of the typical Edgecliff lithology. These facies were deposited near normal wave base as implied by evidence of winnowing. Corals are abundant in these facies and include solitary rugosans such as *Heliophyllum, Siphonophrentis,* and large heads of tabulates such as *Favosites, Lecfedites,* and *Emmonsia.*

The spectrum of facies seen in the Onondaga Limestone from non-cherty calcareous shales and argillaceous nodular limestones to heavily cherty calcisiltites and fossiliferous calcarenites (wacke-, pack-, and grainstones) is repeated in a number of other units. For example, a similar sedimentary facies transition is observable in the Silurian Lockport Group and in the classic Lower Devonian Helderberg Group carbonates in the Hudson Valley. The classic Coeymans-Kalkberg-New Scotland transgressive succession documented by Rickard (1962) and Laporte (1969) displays the progression of crinoid pack- and grainstones to cherty micritic limestones, nodular non-cherty limestones, and, finally, highly argillaceous non-cherty limestones and calcareous mudstones.

Clearly, in part, the facies spectrum records decreasing environmental energy. Current- and wave-winnowing processes dominated in near-wave base facies (represented by grainstones) while muddy offshore facies were affected only by occasional intermittent storm action and, finally, deeper water shaly facies deposited below the effects of all but the most severe storms. A gradient from highly oxygenated to dysoxic environments is also suggested.

The most problematic feature of this carbonate facies spectrum is the presence of chert in a restricted facies belt between the high energy crinoidal shoal deposits and the most offshore muddy limestones that are typically non-cherty (see Maliva and Siever, 1989). Several key questions remain to be answered about the chert. First of all, what was the source and nature of the silica? Why is chert restricted to fine-grained but low argillaceous-content facies? Selleck (1985) provided an important working model to understand the occurrence of cherts within the Onondaga Formation. The source of silica, according to Selleck, was finely particulate silica, probably largely in the form of biogenic deposits, such as the opaline of sponge spicules. Such fine silica-rich sediment was winnowed from shallow shelf areas and, hence, typically transported into more offshore regions. Dissolution of the unstable opaline silica phases led to an enrichment of dissolved silica.

Perhaps the most enigmatic part of the story is the reprecipitation of this silica as cryptocrystalline quartz or chert (in particular, what caused reaggregation of this silica in local areas). The presence of chert bands suggests that enrichment of silica occurred intermittently or possibly periodically during deposition of parts of the Onondaga facies. Potentially, these cherts mirror certain surfaces of sediment starvation associated with minor climatic oscillations which temporarily reduced the amount of carbonate being produced within the system.

Another alternative is that carbonate within the cherty facies was deposited episodically as thin, storm-derived calcisiltite layers and that these depositional events were separated by much longer intervals of non-deposition in which the silica built up within near-surface pore waters. The common outlining of burrow galleries by chert suggests that special properties of the burrow fillings, including the increased porosity of the sediment filling, as well as possible geochemical alterations associated with decay of organic burrow linings, may have contributed to local aggregation of silica. The volume of silica at some levels, particularly in the Clarence facies of the Edgecliff Member, however, demands further consideration. One possibility is that the sediments became enriched with finely particulate silica as a result of volcanic ash input. Certainly, thin bentonites are known in the upper portions of the Onondaga and possibly as low as parts of the Nedrow and Edgecliff Members. However, any such possible ash beds are extremely thin within the Edgecliff Member, which is richest in chert. Note that this does not preclude the possibility of a general "background" input of finely particulate windblown volcanogenic silica. In the Moorehouse Member, there appears to be an association of thick, dark chert layers immediately below at least the thickest of the ash beds. However, again, other chert-rich intervals, particularly those of the Silurian Lockport Group, do not appear to be associated with times of volcanism.

Another very tentative possibility is that silica was locally concentrated by certain organisms, primarily siliceous sponges, whose spicules dissolved within the sediment to produce pore water solutions rich in dissolved silica or silicic acid (Maliva and Siever, 1989). Chert nodules in the Devonian Onondaga and Kalkberg, and the Silurian Lockport chert-rich intervals commonly display small spheroidal siliceous sponges (*Hindia*). External molds of spicules have also been observed in certain chert beds. These observations demonstrate at least that sponges were modestly common in areas where chert tended to form. However, such sponges are volumetrically relatively insignificant. Hence, if the chert-rich facies do represent belts of environments highly dominated by sponges, one must assume that the vast majority of sponge remains were completely dissolved and therefore that their fossil record has been almost eradicated. The absence of chert in the more argillaceous facies may reflect an absence of an appropriate source of opaline silica or may reflect the inhibitory effect of clay minerals in the reaggregation and precipitation of local masses of silica (see Selleck, 1985). In any event, our understanding of the depositional setting of these chert-rich facies remains very incomplete and will require much further study.

SEQUENCE STRATIGRAPHIC INTERPRETATION: TRISTATES GROUP AND ONONDAGA FORMATION

Overview

From the standpoint of sequence stratigraphy, the Tristates Group, Onondaga Formation and overlying Hamilton Group in central to western New York constitute a single large subdivision or "holostrome" of Sloss's (1963) Kaskaskia megasequence. This early phase or Tristates-Onondaga-Hamilton phase of the Kaskaskia is bounded by major unconformities. At the base, the sub-Oriskany Wallbridge Unconformity and a composite Wallbridge-upper Tristates-sub-Onondaga unconformity represent the major sequence boundary; at the top the sub-Tully or "Taghanic" unconformity bevels some portions of the upper Hamilton Group and forms the upper major sequence boundary for the holostrome. This is a second order depositional sequence in the terminology of Vail et. al. (1991). Detailed stratigraphic work permits resolution of a number of finer subdivisions within this major package that correspond approximately to third order depositional sequences as recognized by sequence stratigraphers (Vail et. al., 1977, 1991). Sequence stratigraphic study of the Tristates Group and Onondaga Formation is not as fully developed as that of the overlying Hamilton Group at this time, but some generalizations can be offered on the basis of detailed regional stratigraphy that is now nearing completion (Ver Straeten and Brett, 1994; Ver Straeten et al., this volume) In general, it appears that the Tristates-Onondaga interval can be subdivided into at least four third order sequences, each one comprising one to three million years in duration. However, the sequence boundaries between these stratal packages are not everywhere as clear cut as are found in other strata studied in the Appalachian Basin (e.g., Lower Silurian, Brett et al., 1990; Middle Devonian Hamilton and Tully Groups, Brett and Baird, in press).

Basal Megasequence Boundary

The Wallbridge Unconformity is one of five major Phanerozoic unconformities that define the boundaries of Sloss' (1963) "megasequences." It marks a major second order sequence boundary, associated with a major fall in relative sea level. The Wallbridge underlies basal quartz arenites/orthoquartzites of the upper Lower Devonian Oriskany Formation or younger strata across much of eastern North America. In western to west-central New York the Wallbridge Unconformity occurs as an irregular, karstified erosion surface at the top of the late Silurian Bertie or Akron Formations, or the overlying earliest Devonian carbonates of the Manlius Formation. This unconformity, which displays a relief of a few meters in some outcrops, is overlain locally by lenses of Oriskany Sandstone, by thin, sand- or limestone-rich strata equivalent to the Schoharie Formation of eastern New York, or directly by the lower Middle Devonian Onondaga Limestone.

Collectively, the overlying Tristates, Onondaga, and Hamilton strata, up to the major Taghanic Unconformity (below the Tully Limestone), constitute the first holostrome of the Kaskaskia Megasequence. This holostrome is divisible into a series of smaller-scale sequence-like units. Between these two scales, however, are two packages of strata, each bounded by unconformities.

The first succession is marked at its base by the Wallbridge Unconformity proper, which is overlain by transgressive Oriskany quartz arenites. Sandstones of the Oriskany Formation are composed of relatively clean, well rounded and frosted quartz sand grains, and commonly feature a fauna of large brachiopods and other forms. While the Oriskany Sandstone is missing at most localities in the west-central New York region, it does occur as a very thin quartz arenite in outcrops in the central Finger Lakes area, particularly at the Seneca Stone quarry (Stop 3), where the Lower Devonian Manlius Limestone is overlain by a 0-60 cm-thick lense of Oriskany Sandstone (Oliver and Hecht, 1994). The latter is rich in robust, thick-shelled brachiopods and favositid corals. Sands of the Oriskany may also locally be represented by thin neptunian dykes and fissure fillings on the Wallbridge karst surface. In places Oriskany sands may extend more than two meters downward into cracks within the upper Akron or Bertie dolostones. In the Hudson Valley, the Oriskany Sandstone grades laterally into sandy, cherty limestone of the Glenerie Formation. Together, the Oriskany and Glenerie Formations constitute an orthoquartzite-carbonate suite of sedimentary rocks that characterize not only the initial transgression of the Kaskaskia sea, but also an interval of tectonic quiescence prior to an episode of abrupt foreland basin subsidence and influx of a thick wedge of siliciclastic sediments.

The Oriskany Sandstone or Glenerie Limestone is overlain in eastern to east-central New York by dark gray to black silty and generally highly bioturbated Esopus Formation mudstone to sandstone. The latter is up to 100 m-thick in the Kingston area of the Hudson Valley but thins markedly to the northwest.

An erosional contact at the base of the Carlisle Center Formation appears to truncate upper beds of the Esopus Formation in a northwestward direction. West of Cherry Valley, New York, the Carlisle Center equivalent comes to rest directly on Oriskany Sandstone or other units below the Tristates Group and, consequently, below the Wallbridge Unconformity. Hence, the base of the Carlisle Center itself is a third-order sequence boundary. Again, a deepening upward succession (in eastern New York) from glauconite- and phosphate-bearing sandstones of the lower Carlisle Center and overlying calcareous, dark gray silty shales appears to record either another lesser tectonic pulse or an interval of eustatic deepening. However, this apparent deepening upward is reversed in the upper portion of the Carlisle Center which apparently grades upward into the overlying increasingly calcareous nodular to bedded shaly limestones of the Schoharie Formation. The upper Carlisle Center and Schoharie together appear to represent the highstand deposits of a second sequence developed above the Kaskaskia megasequence erosional boundary.

In the Hudson Valley, south of the Catskill area, the contact between the Schoharie and the overlying Onondaga formation appears to be conformable and perhaps gradational. However, as this boundary is traced northward into the region of Clarksville and westward beyond Cherry Valley, the basal Onondaga contact becomes distinctly sharp and the Schoharie Formation is progressively truncated. At Cherry Valley the basal Edgecliff Member of the Onondaga is comprised of some phosphatic and glauconitic material, as well as possible erosional clasts and sand derived from the underlying units.

Thin remnants of the underlying Carlisle Center and/or Schoharie Formations persist westward, at least to the Syracuse-Auburn area of central/west-central New York State. In these regions, the thin Tristates Group equivalent characteristically occurs as glauconitic silt- to sandstone (Carlisle Center Formation?) overlain by sandstone with large (up to 30 centimeters in diameter) dark gray to black phosphate-cemented sandy concretions (Mesolella, 1966; Schoharie Formation?). A 4-5 cm-thick bed of dark, non-calcareous shales is found locally at the base that may represent a thin remnant of the Esopus Shale. The upper bed of the Tristates immediately below the Onondaga carries a diverse and rich brachiopod and small coral fauna in a sandy limestone. This interval is sharply set off from the overlying Edgecliff Limestone. To the west, in the region near Seneca and Cayuga lakes (e.g., in the Seneca Stone quarry, Stop 5), the basal Onondaga unconformity becomes marked and the Edgecliff Member comes to rest on the Oriskany Sandstone with the underlying Tristates Group removed and/or cannibalized into the basal Edgecliff to form a sandy phosphatic conglomerate. At the Seneca Stone quarry the large phosphatic nodules from upper Tristates strata and pieces of phosphatized Oriskany Sandstone and Manlius Limestone have been reworked and accumulated as an erosion lag along the basal unconformity of the Onondaga Limestone. In west-central New York, the sub-Edgecliff unconformity locally merges with the Wallbridge to form one major disconformity, in which the Edgecliff rests directly upon the upper Silurian units. This composite hiatus ranges up to 20 million years in some areas where the Edgecliff Member may rest unconformably on the channeled upper surface of beds low in the Bertie Dolostone of Late Silurian (Pridolian age). However, still farther west, in Erie County and in the adjacent Niagara peninsula of Ontario, the Oriskany Sandstone and lateral equivalents of the Schoharie Formation reappear (Bois Blanc Formation) although they are still locally overlain with sharp contact by the Edgecliff Member. Therefore, as many as three unconformity surfaces may locally occur across New York; the sub-Oriskany Wallbridge Unconformity, a second pre-upper Tristates (Carlisle Center to Bois Blanc) unconformity, and a third sub-Onondaga unconformable surface.

The composite unconformity below the Onondaga Formation clearly represents a prolonged period of relative sea level lowstand, exposure, erosion, and karstification over much of New York State west of the Hudson Valley. However, the sub-Onondaga unconformity itself remains much more enigmatic. It is evidently not simply a case of non-deposition of the Tristates Group, as there is evidence for erosional truncation of some of the beds, including the phosphatic nodule-bearing interval in the Seneca Stone quarry. Furthermore, the contact between the Onondaga and the upper part of the Tristates Group (Schoharie Limestone in eastern New York and Bois Blanc Limestone in Ontario) appears to be conformable to nearly conformable. In those areas skeletal crinoidal grainstones of the Edgecliff Member appear to cap a shallowing upward cycle that commences within the underlying Schoharie-age interval. Yet, in much of western and west-central New York State, the base of the Edgecliff Member is very sharply defined and appears to truncate different levels within the Schoharie-equivalent, Emsian-age beds below. Indeed, over much of this region there is no vestige of the Tristates Group whatsoever, except for the occasional fillings of Oriskany sands into fissures of the underlying Silurian dolostones. Thus, there is tentative evidence that the central portion of New York State was subjected to uplift and erosion, perhaps under subaerial conditions, during the later part of the Emsian Stage. Simultaneously, marine deposition of silty limestone continued both to the east and west of this area. This suggests an intriguing pattern in which the central portion of New York may have been upwarped in a broad arch, possibly a forebulge produced as an isostatic response to crustal loading in the Acadian orogenic thrust belt in New England. These events would be associated with the development of an actively subsiding, narrow, Emsian-age foreland basin to the east and southeast in which a thick succession of dark gray shales of the Esopus Formation and the overlying, increasingly carbonate-rich Carlisle Center and Schoharie Formations accumulated during Acadian Tectophase I of Ettensohn (1985).

Edgecliff Member Sequence Stratigraphy

The Edgecliff Member is sharply bounded both at its base and top in nearly all localities in west-central to western New York. The basal erosional unconformity may or may not have developed under subaerial conditions, at least in the central portion of New York State. The sub-Edgecliff unconformity is interpreted as a third order sequence boundary. However, as discussed in the preceding section, the sub-Edgecliff unconformity may or may not have been produced by eustatic sea level drop.

In western and west-central New York State, the basal Onondaga Edgecliff Member clearly onlaps an irregular erosion surface in many places. For example, at the Oaks Corners quarry (Stop 2), there is nearly a meter of relief on the basal unconformity. Channel-like areas, no more than a few tens of meters across, are infilled with differentially-thickened pack- and grainstone deposits of the basal Edgecliff unit. Laterally, these units thin to nearly zero in areas between the channels.

This phenomenon also may occur on a larger scale. Apparently, the basal Onondaga deposits may thicken and thin more radically in association with topographic highs on the onlap surface. For example, the Edgecliff interval displays marked thickening between the Seneca Stone quarry and the Oak Corners quarry, approximately 21 km to the northwest (see Figure 5). This thickening is also associated with a relatively dramatic facies change, from dominantly crinoidal wacke- to packstone at Seneca Stone quarry into dominantly fine-grained, sparsely fossiliferous, and highly chert-rich calcisiltite of the Clarence facies. As noted in preceding sections, the Edgecliff displays a prominent cyclicity at both Seneca Stone quarry and Oaks Corners quarry. However, only two major cycles can be recognized at the former locality, whereas in the latter, much thicker section to the northwest, a total of four cycles, each one to three meters in thickness, can be discerned. Moreover, the cycles display distinctly different motifs. At Seneca Stone quarry, the cycles are dominated by sparsely fossiliferous, non-cherty, crinoid-bearing wackestone, capped by thin, crinoidal packstone beds. The dark, chert-rich, micritic bed is thought to represent the deepest water facies within the Edgecliff Member at the Seneca Stone locality. It closely resembles the majority of the Edgecliff interval (Clarence facies) in the adjacent Oaks Corners quarry. This cherty bed, in turn, passes abruptly upward into approximately 50 cm of crinoidal wackestone immediately underlying Nedrow Member calcareous shales. This latter clearly appears to represent a major marine flooding surface.

The two Edgecliff cycles at Seneca Stone quarry resemble PACs, (Punctuated Aggregational Cycles; Goodwin and Anderson 1985) and are interpreted in the light of sequence stratigraphy as parasequences. That is, each represents a more or less asymmetrical, shallowing upward cycle capped by crinoid-rich facies and abruptly overlain by deeper water cherty calcisiltites or calcareous shales.

By contrast, the four cycles in the Oaks Corners quarry, except for the lowest, are dominated by cherty, nearly barren limestones which closely resemble the cherty marker band at Seneca Stone quarry. Only the relatively thin caps of these cycles feature more fossiliferous crinoidal wackestones, which are more sparsely fossiliferous than the cycle caps at the Seneca Stone quarry and are similar to the main bodies of these cycles there. Cherty beds that immediately overlie the cycle caps are somewhat argillaceous; in the case of the fourth cycle, a dark gray, non-cherty calcareous shale that closely resembles the overlying Nedrow Member abruptly overlies the sparse crinoidal wackestone of the underlying cycle cap.

The overlying Nedrow Member displays a lesser degree of facies change between the Seneca Stone and Oaks Corners quarries and its internal stratigraphic subdivisions are readily correlated between the two sites. Therefore, in attempting to correlate the Edgecliff cycles, we have worked downward from the abrupt contact with the Nedrow. Using this approach, the upper cycle is only slightly thicker at the more westerly Oaks Corners locality than at Seneca Stone. However, its basal interval displays an apparent downslope facies transition from the cherty micrites at Seneca Stone quarry to the dark gray, Nedrow-like facies tongue at Oaks Corners, whereas the upper portion of the cycle becomes more distinctly chert-rich and much less fossiliferous at the Oaks Corners section. Assuming that this cycle is correctly correlated, the lower or first cycle at the Seneca Stone quarry represents only the third of four cycles at Oaks Corners. It, too, displays marked facies change, having become non-cherty in the distance between the two quarries. The coral-rich bed at the base of Seneca Stone quarry may correlate with the relatively fossiliferous cap of the second cycle at Oaks Corners. This latter interval is notable in carrying relatively large crinoid columnals and some favositid corals, even at Oaks Corners. This correlation scenario would suggest that the lower half of the Edgecliff interval exposed at Oaks Corners quarry, including the one-meter crinoidal lower cycle and bulk of the chert-rich second cycle, is completely absent at the Seneca Stone quarry. This pattern of apparent depositional pinch-out of units, together with the evidence of some thinning in the overlying beds and the above-noted facies changes, suggest that during Edgecliff deposition the area of Seneca Stone quarry lay in an up-ramp position with respect to sections to the west as in Oaks Corners. To the east, in and near its type section in the Syracuse area, the Edgecliff becomes increasingly fossiliferous and is composed of primarily pack- to grainstones. Although it again thickens in this direction, there is reason to suspect that this thickening is in a continued upramp direction.

The absence of basal Onondaga cycles I and II in the vicinity of Cayuga Lake may represent a lack of accommodation space in this region. As noted, tongues or pods of the Bois Blanc Formation of the Tristates Group occur eastward to the vicinity of Phelps, a few miles west of Oaks Corners quarry. And again, Tristates Group equivalents appear in the Auburn area, approximately 25 km northeast of Seneca Stone quarry. The vicinity of Seneca Stone appears to have experienced the greatest amount of removal of Tristates group in the sub-Onondaga unconformity. This fact, in turn, suggests that this particular area represented the topographically highest region of the sea floor during initial Onondaga transgression. If the basal Edgecliff units have any correlatives in this area, they lie within the reworked phosphatic conglomerates that sharply overlie the Oriskany at Seneca Stone quarry. In any event, it appears that the region to the west near Oaks Corners was much more actively subsiding in at least lower Edgecliff time. The Clarence cherty facies actually thickens somewhat from the Oaks Corners quarry westward at least to LeRoy but may thin again toward Buffalo where the lower Edgecliff again is represented by crinoidal pack- and grainstone facies. These facies are virtually absent at the Oaks Corners quarry, except for fillings of erosional channels at the very base of the Edgecliff Member.

Thus, a pattern of facies change, at least within the Edgecliff Member, indicates a local basin center in the Phelps to Manchester area (Ontario County, New York). This region appears to have been bordered both to east and west by shallow carbonate ramp transitional to shallow shelf, crinoidal shoal facies, both in the extreme west and in central portions of the state. The meter-scale parasequences observed within the Edgecliff member may represent minor eustatic sea level oscillations (see, for example, Goodwin and Anderson 1985), or minor pulses of subsidence, followed by progradational buildup of the seafloor by a few meters of carbonate deposition. However, the thinness of the cycles, together with evidence for abrupt shifts not only to deeper, but also to shallower facies (see, for example, the boundaries of the chert-rich marker bed at Seneca Stone quarry) suggest that progradation was probably not the chief mechanism for shallowing.

Overall, the three to four small-scale cycles or parasequences that make up the Edgecliff Member locally display an overall pattern of upward deepening or retrogradation. Thus, while each parasequence is a rather asymmetrical, generally upward-coarsening and -shallowing succession, the facies which overlie flooding surfaces at the tops of these small-scale cycles display a progressive trend from shallower to deeper water. This is particularly well seen in the four cycles of the Edgecliff Member at Oaks Corners quarry. The strata overlying the top of the basal meter-thick cycle are nodular, somewhat fossiliferous, and sparsely cherty. Those overlying the cap of the second cycle are thin-bedded, slightly argillaceous cherts. Strata overlying the third cycle cap consist of the Nedrow-like non-cherty argillaceous facies. Finally, the top of the fourth cycle is a major marine flooding surface at the contact between the Edgecliff Member and the overlying shaly Nedrow Member. This appears to represent the strongest deepening event of the four.

Overall, this retrogradational pattern is characteristic of a transgressive systems tract, as outlined by Van Wagoner et al. (1988) and Vail et al. (1991). It clearly appears that the Edgecliff was being deposited at a time of overall deepening of the facies and this pattern is seen at virtually all outcrops. We do note however, as discussed elsewhere in this paper, that the central New York region was also undergoing more rapid subsidence than other parts of the state during Edgecliff time.

Nedrow Member Sequence Stratigraphy

The Edgecliff-Nedrow Member contact is an extremely widespread, apparent deepening event. In many localities throughout western and central New York, it is marked by a dark gray to greenish-gray thin fossil hash-rich calcareous bed. Many of the fossils within this level appear heavily corroded and pitted, as though they had been exposed for prolonged periods on the seafloor. The greenish coloration of the shale suggests trace quantities of glauconite, which is commonly associated with sediment-starved intervals. In typical Nedrow outcrops, such as in the central Finger Lakes area, bundling of calcareous shales and thin to somewhat thicker, tabular, argillaceous micrite bands define at least five small-scale, slightly asymmetrical cycles within this unit. Again, these cycles appear to present a continued retrogradational pattern, at least up to the level of the distinctive upper Nedrow black shale marker bed. As noted, this black, laminated shale interval is extremely widespread throughout much of central New York, the south-central portion of the Hudson Valley, and into central Pennsylvania. We suggest that this episode of deposition of high dysoxic to anoxic laminated muds within the central portions of the Appalachian basin may represent one of the strongest pulses of deepening within the Onondaga succession.

Thus, while the Edgecliff to Nedrow contact is interpreted as a surface of sediment starvation, the thin shaly to nodular cycles of the Nedrow appear to constitute a condensed interval and the black shale band may represent a maximum flooding event. This sequence is somewhat unusual in that the maximum flooding surface and surface of maximum starvation appear to be offset from one another by at least four minor parasequences. Overall, the Nedrow is interpreted tentatively as an early highstand facies. The extremely widespread nature of the Nedrow black shale suggests that this may well have been an eustatic event, although it is not recognized as such on the well-known Devonian sea level curve of Johnson et al. (1985).

Some previous workers have interpreted the Nedrow as a shallow water facies. For example, Feldman (1980) argued that the *Pacificocoelia* community of the Nedrow Member in central New York represents one of the shallow water assemblages in the unit. However, this assemblage appears to be displaced in an upramp direction by more diverse brachiopod associations. Furthermore, *Pacificocoelia* appears in deeper water assemblages in other portions of the stratigraphic column. Moreover, the Nedrow Member is known to contain some of the more diverse offshore *Polygnathus* conodont elements that permit recognition of the *patulus-costatus costatus* zonal boundary (see Klapper, 1981). This contrasts with the typical shallow water *Icriodus* conodont biofacies represented in much of the rest of the Onondaga Limestone. The regional geometry of the Nedrow Member, with dark shales being confined to the more central portions of the Onondaga basin, also strongly suggests that the Nedrow represents one of the deepest intervals during Onondaga deposition.

In sections upramp from the basin center, the Nedrow interval becomes increasingly rich in limestone. Again, the more shaly intervals appear to pass laterally into light gray and dark chert-rich bands which display distinct cyclicity in quarries in the LeRoy to Stafford area (see Figure 5). In contrast, the nodular or tabular marly limestone beds of the cycle caps in the Nedrow of the basin center appear to pass upramp into fossiliferous wacke- to packstone lithologies, which bear increasingly diverse faunas of brachiopods, gastropods, and corals. The enrichment in fossils at the tops of some of these beds is spectacular and may reflect minor intervals of sediment starvation.

Understanding of facies trends in the Nedrow Member between the Seneca Stone quarry and the Oaks Corners quarry are somewhat ambiguous at this time. The sections are similar, although the Nedrow at Oaks Corners quarry is somewhat thicker than that seen at Seneca Stone quarry (see Figure 5). The dark marker bed is nearly black shale in both sections although thin beds of dark gray shale occur more prominently below the bed at the Seneca Stone quarry than at Oaks Corners quarry and the lower Nedrow at least appears to be slightly more fossiliferous in the latter than at the former. Certainly, facies trends in the overlying transitional Nedrow to Moorehouse interval strongly suggest a down-to-the-east ramp extending from Oaks Corners quarry to the Seneca Stone quarry. There is no doubt that the Nedrow undergoes a substantial lateral upslope change westward from the Oaks Corners quarry.

These lines of evidence suggest that the ramp deepened gradually from the area west of LeRoy through the area of Phelps to a deepest basinal position somewhere in the vicinity of the Seneca Stone quarry during Nedrow deposition. Certainly the higher beds of the Nedrow and the overlying Moorehouse strongly support this contention. While the Moorehouse is moderately fossiliferous and highly cherty at Oaks Corners quarry, it is thinner, much more sparsely fossiliferous and contains much less chert at the Seneca Stone quarry. Thus, in the time interval within lower Edgecliff deposition and the onset of Nedrow deposition, the basin center appears to have shifted slightly, by about 20 km, to the southeast from approximately the Phelps area to the Cayuga Lake region. This southeastward drift of the basin center may have continued into the later Eifelian during deposition of the higher Moorehouse and Seneca Members.

Moorehouse Member Sequence Stratigraphy

As defined herein, the base of the Moorehouse Member is drawn at the upper of the two dark shale marker beds ("*Schizophoria* bed"). Basal Moorehouse strata are among the most sparsely fossiliferous in most Onondaga sections, and consist of sparse echinoderm skeletal wackestone or lime mudstone with very scattered fossil debris. Weathered surfaces of beds low in the Moorehouse display markings diagnostic of the trace fossil *Zoophycos*. Hence, these rocks can be interpreted as bioturbated carbonate silt deposits that may have accumulated relatively rapidly and possibly under high turbidity conditions that were unfavorable to many benthic organisms. The appearance of abundant dark brownish gray chert nodules within the Moorehouse parallels situations seen in the Clarence facies of the Edgecliff Member and demonstrates the enrichment of silica, possibly from biogenic sources. Conceivably, these represent a sponge-rich facies which is nonetheless rather poor in brachiopods, corals, and other organisms. In contrast, the thin shaly intervals within the Moorehouse, which appear to record minor reversions to condensed Nedrow-like facies, display more diverse fossil assemblages with small rugose corals and relatively rich, brachiopod-dominated benthic communities.

A generally shallowing upward succession is seen in a series of approximately six to seven small-scale cycles in the lower to middle Moorehouse Member, that have been assigned by different authors to the upper part of the Nedrow Member (in part) or to the lower part of the Moorehouse Member. These cycles are represented by intervals of non-cherty to chert-bearing limestones capped by thin, dark gray to greenish-gray brachiopod-rich shales. The shaly beds are enriched in heavily fragmented or corroded fossil debris, suggesting sediment starvation associated with marine flooding events. The lower shale partings carry a modest diversity of *Pacificocoelia* and *Schizophoria* brachiopod assemblages in some areas. In contrast, some of the higher capping shales are richer in brachiopods and show a much higher diversity assemblage assignable to Feldman's (1980) *Leptaena-Megakozlowskiella* community.

In western sections this apparent shallowing-upward trend in the lower part of the Moorehouse culminates in approximately one to two meters of coral-rich, crinoidal packstone. Locally, in the Stafford-LeRoy area, this interval features one or more distinctive coral biostromes. In other units (e.g., Middle Silurian Lockport Group) similar biostromes commonly mark the caps of shallowing cycles, although they appear slightly below the coarsest beds in the middle Moorehouse Member.

The coral biostromal layers and associated crinoidal packstone interval is abruptly overlain by the "false Nedrow" calcareous shale to shaly limestone. Indeed, this argillaceous interval has been mistaken in some areas for the Nedrow Member (see Conkin and Conkin, 1979, 1984; Conkin, 1987). This shaly interval appears to reflect a marine flooding event or deepening that took place midway through deposition of the Moorehouse Member. This cycle has been overlooked by most previous workers on the Onondaga, but is recognizable not only in western New York, but also in central and eastern portions of the New York outcrop and southward into eastern and central Pennsylvania where it is commonly represented by a very dark gray shale interval that may resemble the Nedrow "black beds."

A fossil-rich crinoidal hash bed that occurs beneath this important marker may signify a condensed interval associated with the abrupt deepening pulse. However, in many sections it is abruptly overlain by coarser strata of the upper part of the Moorehouse Member. These strata comprise 3-4 m of crinoid-rich pack- and grainstone both in western and eastern New York State. In the central basin region (e.g., Seneca Stone quarry, Stop 6) the interval is a more chert rich and less argillaceous micritic limestone. It is typically rich in fossil mollusks, particularly high-spired gastropods.

These more fossiliferous facies of the upper Moorehouse Member extend upward to the Tioga B-OIN Bentonite that demarcates the top of the Moorehouse. Very similar facies occur in the lower part of the Seneca Member immediately above the Tioga B in most sections. In some sections (e.g., Seneca Stone quarry, Stop 6) the Tioga B bed is overlain by a thin interval of dark gray shale that grades upward through argillaceous limestone into wacke- or packstone of the lower Seneca. Hence, at least a small scale deepening-shallowing cycle appears to be associated with the bentonite. This and other evidence suggests that the Tioga B may represent a type of condensed bed and flooding surface. On the other hand there is no major shift in facies at this juncture; as stated, lowest Seneca strata immediately above the Tioga B are similar to upper Moorehouse strata.

Seneca Member Sequence Stratigraphy

Contrary to previous suggestions, the Seneca Member does not form a simple fining-up/deepening-up succession. Details of the member at Seneca Stone quarry (see Figure 5), where it appears most complete, show two larger scale cycles within the Seneca. An apparent deepening is recorded approximately one meter above the base at a transition from fossiliferous wackestone to finer-grained micritic limestone. Coquinas of *Hallinetes* (= "pink chonetid") brachiopods signal shallowing that culminates in a bed with small rugose corals 4.0 m above the base of the Seneca. The recurrence of finer-grained *Hallinetes*-rich limestones above mark a second deepening interval, but these in turn pass into somewhat more fossiliferous beds with small rugosans and a slightly more diverse brachiopod assemblage. The uppermost 0.5 m of Seneca marks a return to more finer-grained, sparsely fossiliferous facies. The upper contact of the Seneca Member as defined in this paper is marked by a thin, apparently widespread lag bed with fish bone material. This bed and the transition upward through the lower part of the Bakoven Member (Union Springs Formation) is discussed further in Ver Straeten et al. (this volume. Although correlative beds are uniformly coarser and more fossiliferous to the west, these overall patterns of change through the Seneca Member are reiterated at western New York localities (e.g., Honeoye Falls and Stafford quarries, Stops 2 & 1, respectively).

SYNTHESIS OF LOWER MIDDLE DEVONIAN CARBONATES, NORTHERN AND CENTRAL APPALACHIAN BASIN

Middle Devonian carbonates equivalent to the Onondaga Limestone of New York State occur widely across eastern North America (Koch, 1981). As previously noted, they range from the James Bay region of northern Ontario to southeast Quebec and Maine to Georgia and Illinois (Figure 2), and include the Columbus Limestone of Ohio and the Jeffersonville Limestone of the classic Falls of the Ohio River at Louisville, Kentucky. Within the Appalachian Basin, however, the relations between the Onondaga Limestone of New York and correlative strata has been somewhat obscure. The faunally diverse, cherty, carbonate platform-setting of the northern part of the Appalachian Basin (New York) in the lower Middle Devonian gives way southward to thinner, increasingly argillaceous strata (upper part of the Needmore Formation) across much of the rest of the basin. Poor biostratigraphic and lithostratigraphic resolution of more mud-dominated facies across central Pennsylvania into the Virginias has hindered detailed correlation and understanding of time-rock relationships between the Onondaga Limestone and its equivalents across the Appalachian Basin (Figure 7).

In the course of fieldwork during 1993, a series of distinctive, correlatable marker beds were found within the lower Middle Devonian Selinsgrove Limestone of central Pennsylvania. This microstratigraphic framework can be correlated across a minimum of 400 km of the Appalachian Basin into the Onondaga Limestone of central New York. As a result, we recognize that: 1) the Selinsgrove Member of the Needmore Formation in central Pennsylvania is the direct equivalent of the Onondaga Formation of New York; and 2) four recognizable subdivisions of the Selinsgrove Member are directly equivalent to the Edgecliff, Nedrow, Moorehouse, and Seneca Members of the Onondaga Formation of New York.

The Selinsgrove Member of central Pennsylvania is the uppermost of three members of the Needmore Formation (Inners, 1975). It is comprised of relatively thin, non-cherty,

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Figure 7. Map of the Onondaga Limestone in New York and equivalent Buttermilk Falls and Selinsgrove Limestones in Pennsylvania (modified after Inners, 1975). Key localities are as follows: [New York] CL=Clarence, HF=Honeoye Falls, SS=Seneca Stone quarry, AU=Auburn, JM=Jamesville, CV=Cherry Valley, CL=Clarksville, CA=Catskill, KI=Kingston; [Pennsylvania] ST=Stroudsburg, WB=West Bowmans, SW=Swatara Gap, MI=Midway, DL=Dalmatia, SJ=Selinsgrove Junction, NH= Newton Hamilton, MP=Mapleton.

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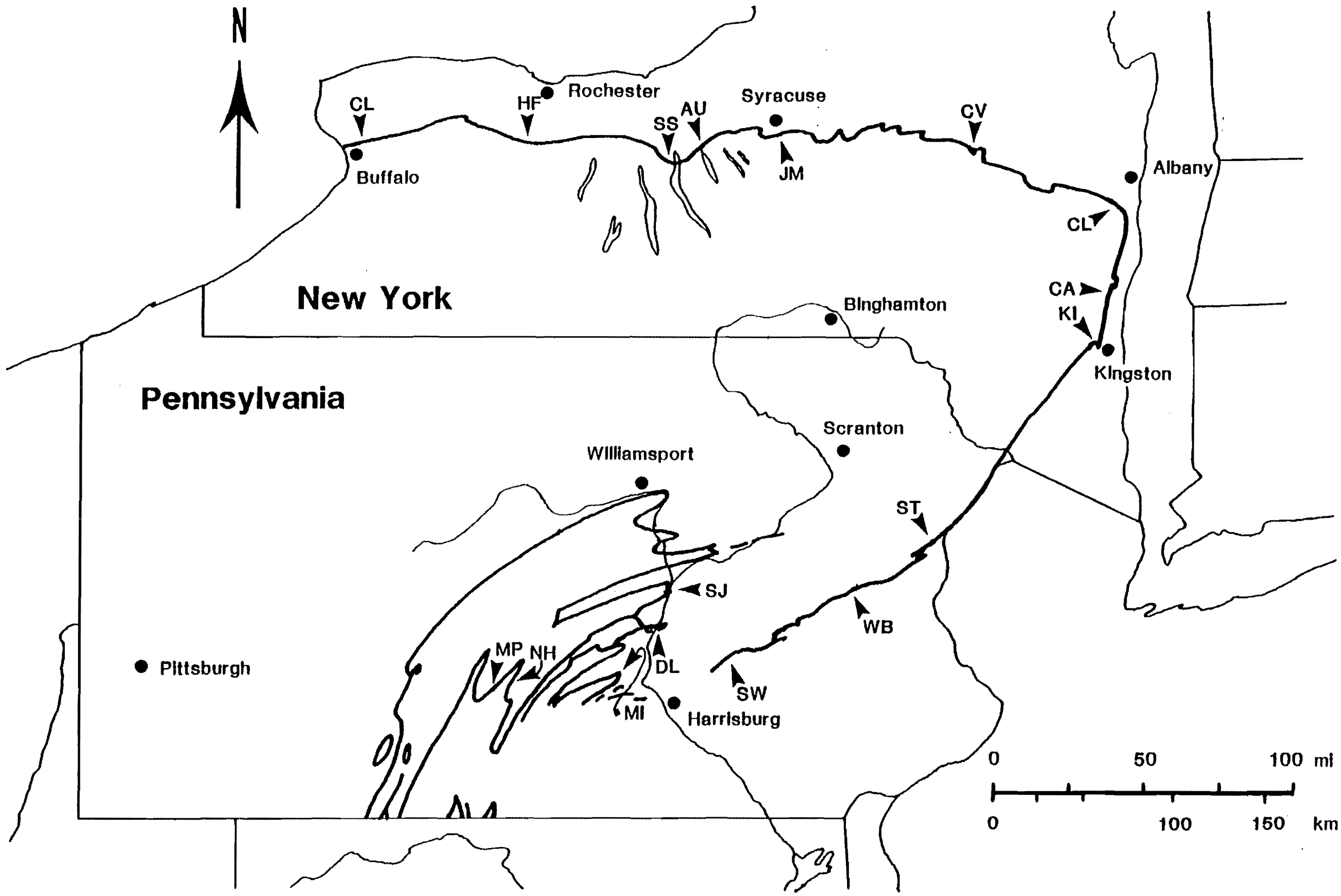
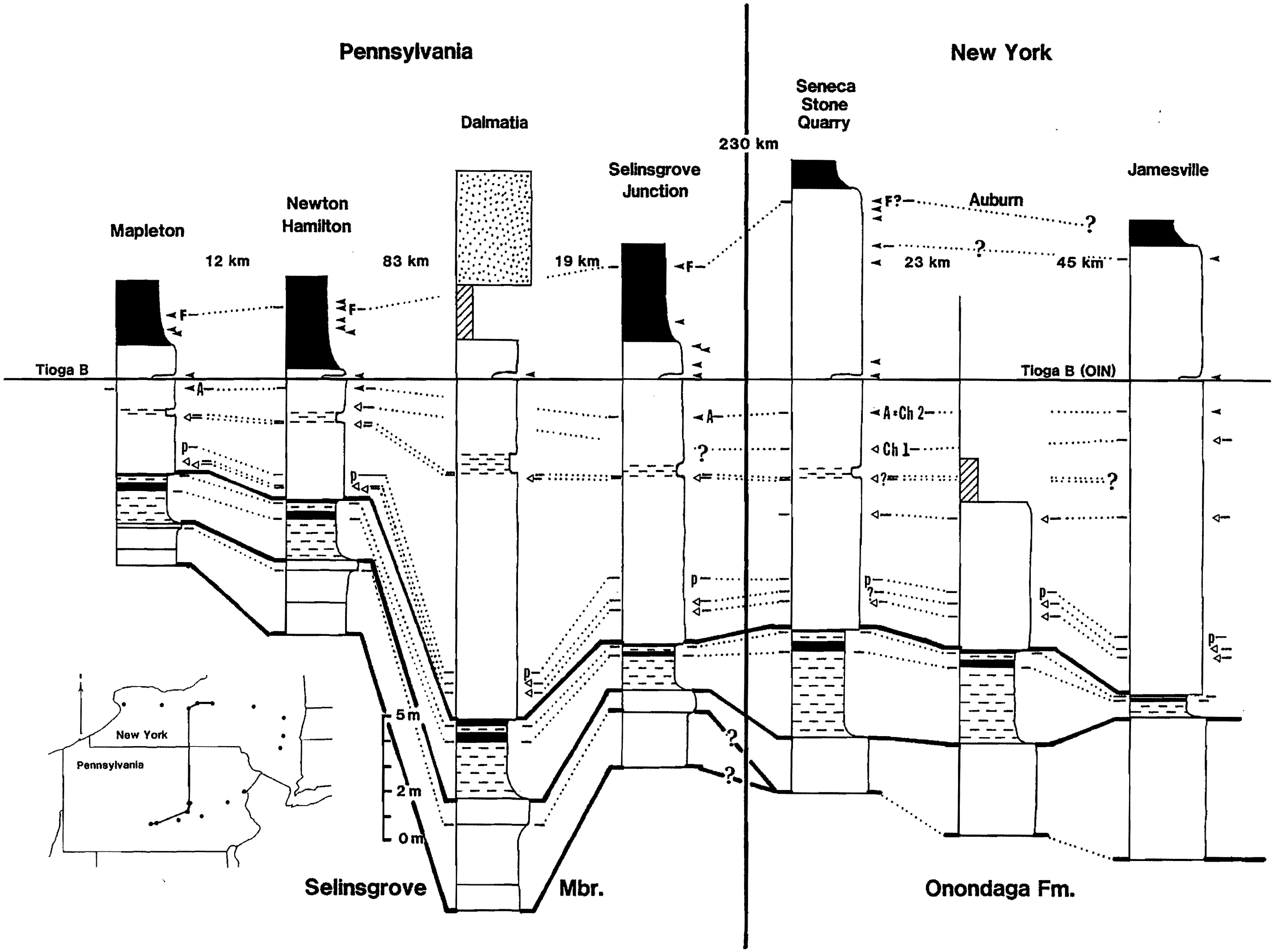
Figure 8. Microstratigraphic correlation of the Selinsgrove Limestone (central Pennsylvania) and the Onondaga Limestone (central New York).



N
CL
HF
Rochester
SS
AU
Syracuse
CV
Buffalo
JM
Albany
CL
New York
CA
Binghamton
KI
Kingston
Pennsylvania
Scranton
Williamsport
ST
SJ
WB
MP
NH
DL
SW
Pittsburgh
MI
Harrisburg
0
50
100 mi
0
100
150
km



Pennsylvania
New York
Mapleton
Newton Hamilton
Dalmatia
Selinsgrove Junction
Seneca Stone Quarry
Auburn
Jamesville
12 km
83 km
19 km
230 km
23 km
45 km
Tioga B
Tioga B (OIN)
F
F?
A
A=Ch 2
Ch 1
p
5 m
2 m
0 m
Selinsgrove Mbr.
Onondaga Fm.
New York
Pennsylvania

interbedded argillaceous limestones and calcareous shales. The Selinsgrove Limestone is characterized by low diversity faunas that include small brachiopods (chiefly *Ambocoelia*, *Pacificocoelia*, and other diminutive forms), trilobites (dominantly two species of *Odontocephalus*), ostracods, and dacryoconariids (Inners, 1975).

Subdivision and microstratigraphic correlation within the Selinsgrove Member in central Pennsylvania is possible through recognition of a number of distinctive lithostratigraphic marker beds. These marker beds include: 1) a thick-bedded to massive limestone, locally with large crinoid ossicles, in the lower part of the member; 2) shale dominated intervals; 3) several thin black shales; 4) the Tioga Ash Beds; 5) thin (ca. <1-5 cm-thick) clay beds that appear to represent additional K-bentonite beds; and 6) pyrite nodule-rich intervals. These beds occur in similar stratigraphic succession along at least 110 km of the central Pennsylvania outcrop belt (see Figure 8).

This microstratigraphic succession is directly correlatable from the type section of the Selinsgrove Member along the Susquehanna River in central Pennsylvania northward 230 km into the Onondaga Limestone at Seneca Stone quarry (Stop 6). At Seneca Stone and nearby central New York localities the same succession of lithostratigraphic marker beds is found (Figure 8). Among the marker beds recognized are the lower and upper (*Schizophoria*) black shale beds at and near the top of the Nedrow Member in New York and an overlying pair of yellow to tan soapy clay beds that appear to represent K-bentonites. The previously noted thick-bedded to massive limestone in the lower part of the member is equivalent to all or part of the Edgecliff Member in New York. The shaly Nedrow Member and the "false Nedrow" shale in the middle of the Moorehouse are also both recognizable in the Selinsgrove Member. The Tioga bentonite cluster is recognized in both regions (see Figure 6 and discussion above), although correlation of some beds from Pennsylvania to New York is problematic at this time.

The term "Onondaga" has at times been used erroneously as a synonym for the Needmore Formation in Pennsylvania. However, as shown here, only the Selinsgrove Member of the Needmore is the direct equivalent of the Onondaga Formation in New York. The two underlying members of the Needmore Formation in Pennsylvania are correlative with strata of the Lower Devonian Tristates Group (Esopus and Schoharie Formations) of eastern New York.

Wider correlation of this microstratigraphic framework into coarser, more shallow-water facies east and west of the central New York trough appears to be more difficult; preliminary work indicates, however, that at least some of the distinctive marker beds are widely recognizable. For example, at Kingston, in the southern part of the Hudson Valley outcrop belt (eastern New York), a distinct dark interval of argillaceous limestone occurs near the top of the Nedrow Member. A short distance above the dark beds occur two thin, clay-rich crevices that may represent K-bentonite layers. This succession corresponds to the two black shale beds and overlying twin clay beds found in the upper part of the Nedrow and lower part of the Moorehouse Members in central New York and equivalent strata in central Pennsylvania.

Preliminary work in eastern Pennsylvania shows a number of the same marker beds occur in the greatly thickened (ca. 83 m-thick), Onondaga-equivalent Buttermilk Falls Formation near Stroudsburg, Pennsylvania. Four members have been reported for the Buttermilk Falls Limestone (Epstein, 1984; Inners, 1975); a lower, relatively coarse, chert-rich limestone (Foxtown Member), an overlying shale-dominated interval (McMichael Member), another chert-rich limestone unit (Stroudsburg Member), and an uppermost Echo Lake Member. A prominent, 30 cm-thick K-bentonite in the upper part of the Stroudsburg Member corresponds to the Tioga B-OIN Ash bed in New York and central Pennsylvania. Recognition and microstratigraphic correlation of the distinctive marker beds shows that the four members of the Buttermilk Falls Formation in eastern

Pennsylvania are equivalent to the Edgecliff, Nedrow, Moorehouse, and Seneca Members in New York.

With all of this in mind, we can see a broader synthesis of lower Middle Devonian carbonates in the northern and central parts of the Appalachian Basin. Figure 9 (a&b) represents dominantly east-west cross sections of Onondaga equivalent strata along approximately 900 km of the outcrop belt in New York and Pennsylvania. Strata equivalent to the Edgecliff-Clarence, Nedrow, Moorehouse, and Seneca Members of the Onondaga Formation in New York are recognized all along the outcrop, except along the Auburn Promontory northeast of Harrisburg, Pennsylvania (Epstein et al., 1974), which represents the southeastern shoreline/margin of the Appalachian Basin during the early Eifelian.

The next stage of this study will extend the research into southern Pennsylvania, Maryland, and the Virginias, to test whether this microstratigraphic framework is correlatable into the southern portion of the Appalachian Basin, and if so, to compare and contrast the sedimentary record across the entire basin during the earliest Middle Devonian.

SUMMARY

The Middle Devonian Onondaga Formation of New York State represents shallow marine carbonates deposited in the northern part of the Appalachian foreland basin. Relatively shallow shoal to shelfal environments in eastern and western New York gently sloped toward a northern tongue of the central Appalachian basinal trough in the central Finger Lakes region of the State. Recent study shows that this basinal tongue shifted through Onondaga time and, in fact represented a center of subsidence in the position of a late Early Devonian uplifted peripheral bulge.

Four members of the Onondaga Limestone are widely recognized across New York State (in ascending order, the Edgecliff, Nedrow, Moorehouse, and Seneca Members; Oliver, 1954). The lower of these, the Edgecliff Member, is characterized by relatively coarse crinoidal to coral-rich packstone to wackestone facies that are locally chert-rich. Coral biostromes and bioherms, which include pinnacle reefs in the subsurface of southern New York, are rooted in a widespread coral-rich zone in the base of the Edgecliff Member. We herein informally propose two dominant, widespread facies of the Edgecliff Member, a cherty, micritic "Clarence facies" (formerly Clarence Member of western New York) and a coarse crinoidal, non- to sparsely cherty "Jamesville Quarry facies."

Overlying calcareous shale-dominated strata and laterally equivalent argillaceous limestones above the Edgecliff are assigned to the Nedrow Member. Chert is generally uncommon to rare in the dark shale facies but is relatively more abundant in laterally equivalent, more carbonate-rich facies of the member. The Nedrow is commonly associated with a diverse brachiopod fauna and numerous platyceratid gastropods. Recent fieldwork indicates that the Nedrow Member is a distinctive interval, correlatable into western New

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Figure 9a. Cross-section of the Onondaga Limestone across New York State. Upper part of Onondaga Formation at Kingston is generally covered and is unknown to the authors.

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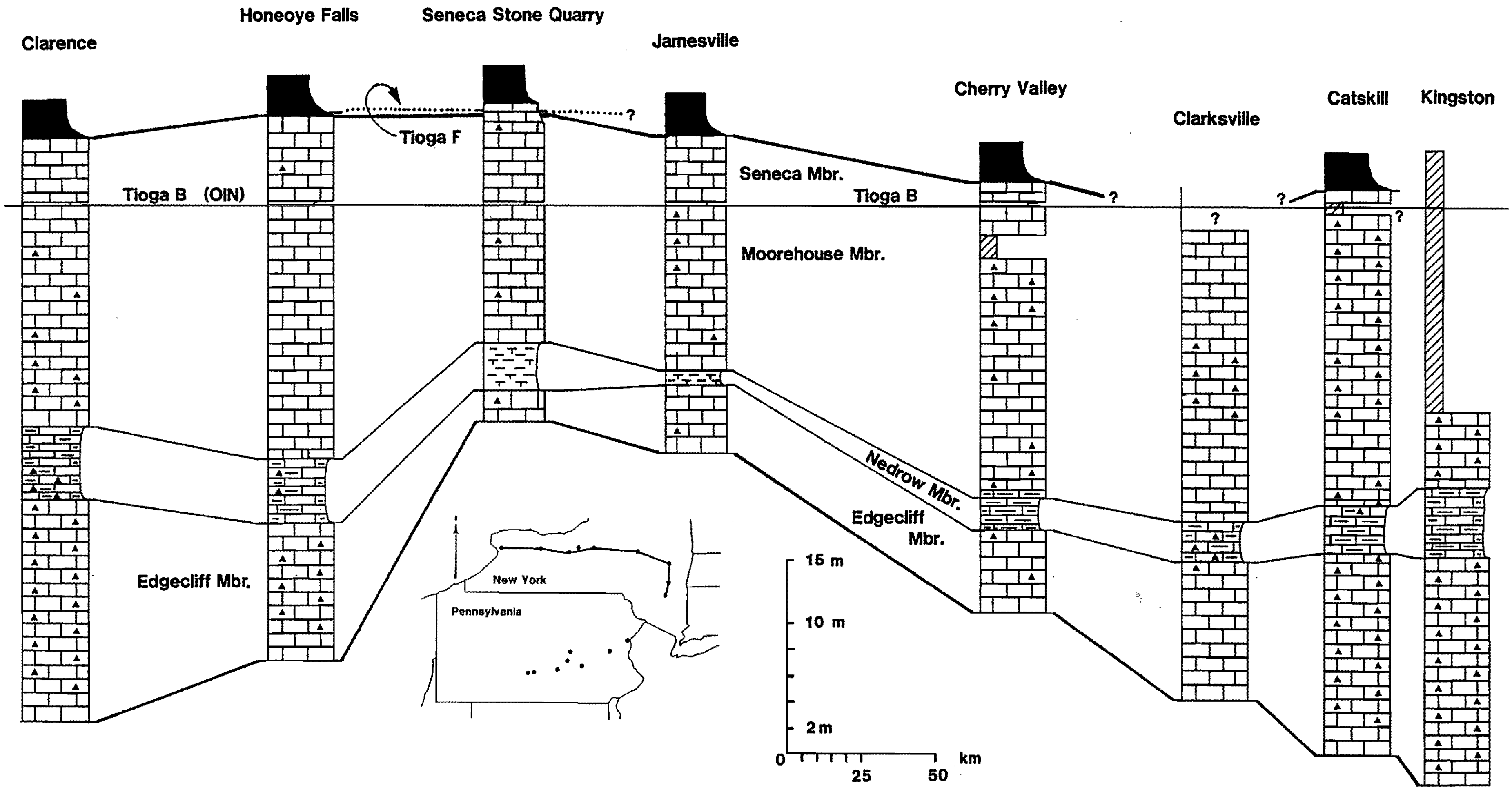
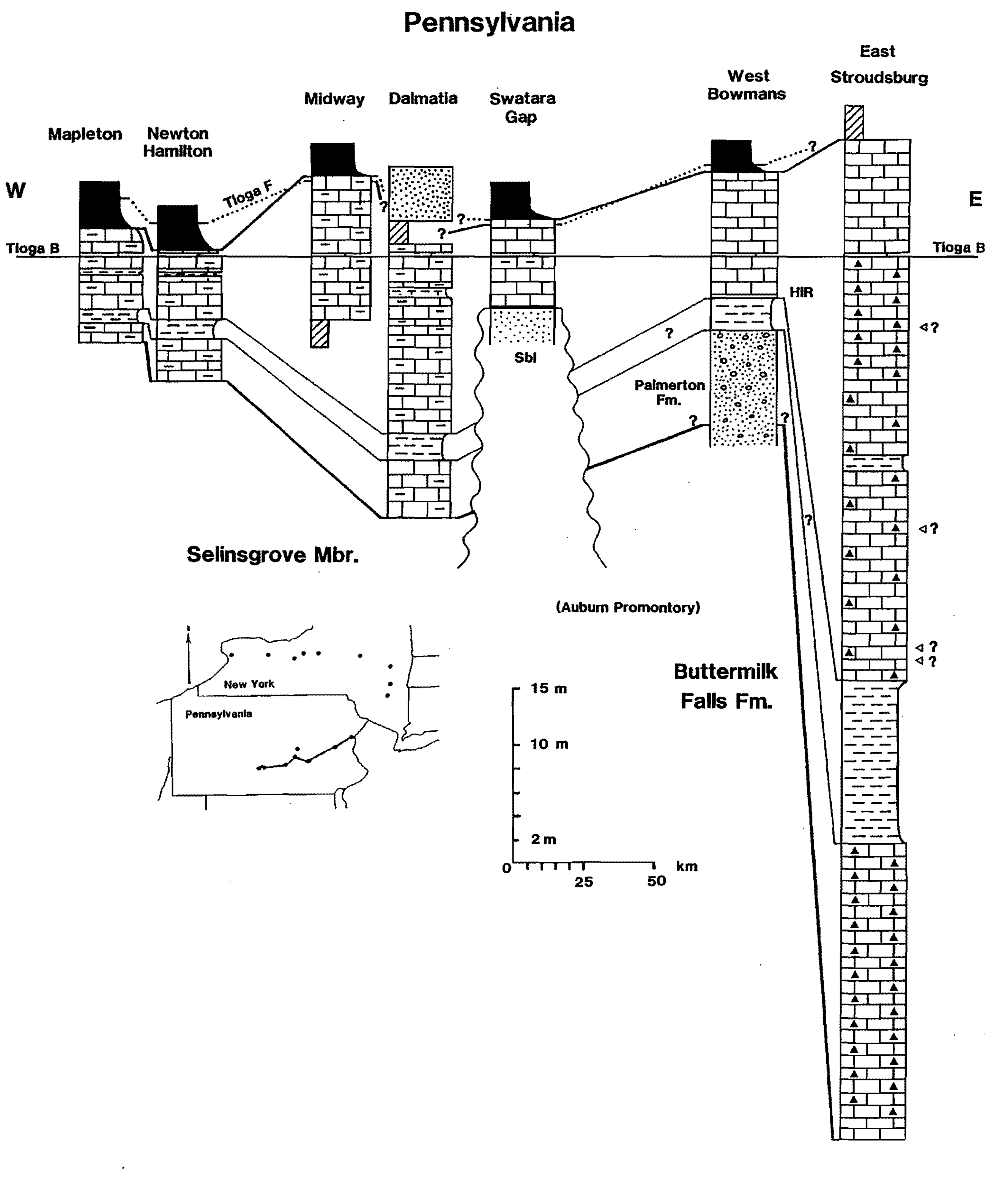
Figure 9b. Cross-section of the Onondaga-equivalent Selingsgrove and Buttermilk Falls Limestones across Pennsylvania. The Selinsgrove is dominantly comprised of interbedded argillaceous limestone and calcareous shale; the Buttermilk Falls chiefly consists of cherty to non-cherty limestone. Thicknesses from Swatara Gap and West Bowmans after Inners (1975) and Epstein et al. (1974). Note thinning and facies change in center of diagram, related to proximity of southeastern shore of Appalachian basin near the Auburn Promontory (see Epstein et al., 1974).

New York



Clarence
Honeoye Falls
Seneca Stone Quarry
Jamesville
Cherry Valley
Clarksville
Catskill
Kingston
Tioga F
Tioga B (OIN)
Seneca Mbr.
Tioga B
Moorehouse Mbr.
Nedrow Mbr.
Edgecliff Mbr.
Edgecliff Mbr.
New York
Pennsylvania
15 m
10 m
2 m
0
25
50
km


Pennsylvania
W
E
Mapleton
Newton Hamilton
Midway
Dalmatia
Swatara Gap
West Bowmans
East Stroudsburg
Tioga F
Tioga B
Tioga B
HIR
Sbl
Palmerton Fm.
Selinsgrove Mbr.
(Auburn Promontory)
Buttermilk Falls Fm.
New York
Pennsylvania
15 m
10 m
2 m
0
25
50
km

York where it overlies Clarence cherty facies of the Edgecliff Member. A pair of black to dark gray shale-dominated beds within the Nedrow Member are widely correlatable across a large area of the northern and central parts of the Appalachian Basin. These beds occur below Oliver's (1954) original position of the Nedrow-Moorehouse Members contact; however, due to the ease of recognition and the widespread mappability of the two black shales, we functionally use them as the upper boundary of the Nedrow Member.

Overlying strata of the lower part of the Moorehouse Member are characterized by poorly fossiliferous limestones, commonly interbedded with thin shales. A widespread calcareous shale near the middle of the member is succeeded by coarser facies in the overlying upper part of the Moorehouse. These rocks generally consist of pack- to wackestone facies similar to parts of the Edgecliff Member. Nodular to bedded chert occurs throughout the member but is more concentrated toward the top of the unit.

Uppermost strata of the Onondaga Formation are represented by medium- to fine-grained limestones of the Seneca Member. The Seneca is marked by a generally lower diversity fauna, and a lesser percent of chert than the underlying Moorehouse Member. Numerous altered volcanic ash beds of the Tioga Bentonites cluster are chiefly concentrated in the Seneca Member.

The contact of the Onondaga Limestone and the overlying Marcellus Shale in the type area of the Seneca Member is treated differently by different authors. The authors herein tentatively follow the usage of Conkin and Conkin (1979, 1984; Conkin, 1987) and place the contact at the position of an apparent discontinuity marked by fish bone material, pieces of hematitic limestone, and a small-scale irregular topography. Overlying interbedded styliolinid limestones, dark shales, and K-bentonite beds (Oliver's 1954 Zone L) are recognized as part of the overlying Marcellus Shale. The Onondaga-Marcellus contact is diachronous across New York State, as seen by the eastward truncation of the Seneca Member below an apparent submarine disconformity at the base of the Marcellus Shale.

Detailed microstratigraphic study across New York and Pennsylvania indicates that the four members of the Onondaga Formation in New York State are widely recognizable across the northern and central portions of the Appalachian Basin. For the first time it is shown that the Selinsgrove Limestone (upper member of the Needmore Formation) in central Pennsylvania is the direct equivalent of the Onondaga Limestone in New York. Underlying strata of the Needmore Formation in Pennsylvania are therefore equivalent to the older Esopus to Schoharie Formations of eastern New York. Furthermore, it is shown that four members of the Buttermilk Falls Limestone of eastern Pennsylvania are directly equivalent to the Edgecliff, Nedrow, Moorehouse, and Seneca Members of New York.

The Edgecliff to lower Moorehouse Members in New York State appear to represent a relatively large-scale sedimentary cycle, probably comparable to those seen in the lower and upper halves of the Helderberg Group, perhaps spanning up to two to three million years. This deepening/shallowing cycle is interpreted as a third order depositional sequence, comparable to those mapped by seismic stratigraphers. It commences with an erosional sequence boundary and combined transgressive surface at the base of the Edgecliff member. The Edgecliff represents a transgressive systems tract while the overlying Nedrow, separated by a surface of maximum sediment starvation early highstand interval; finally the higher Nedrow and its transition into the lower Moorehouse as previously defined by Oliver (1956a) represents a later highstand or regressive interval. However, the upper portions of the Moorehouse and lower parts of the Seneca Member in many areas display a return to coarser skeletal wacke- or even grainstone facies somewhat resembling those of the Edgecliff Member. This abrupt shallowing succession probably reflects a second third-order sequence in the Onondaga Formation as a whole. As in other parts of the Onondaga, the upper Moorehouse and Seneca interval is comprised of a number of smaller-scale cycles. Overall, these display a deepening upward trend, at least above the lower part of the Seneca Member. Bone beds associated with the top of the Seneca Member (see Ver

Straeten et al., this volume) record sediment starvation associated with a major deepening event. This event marks the onset of the second great tectophase of the Acadian Orogeny.

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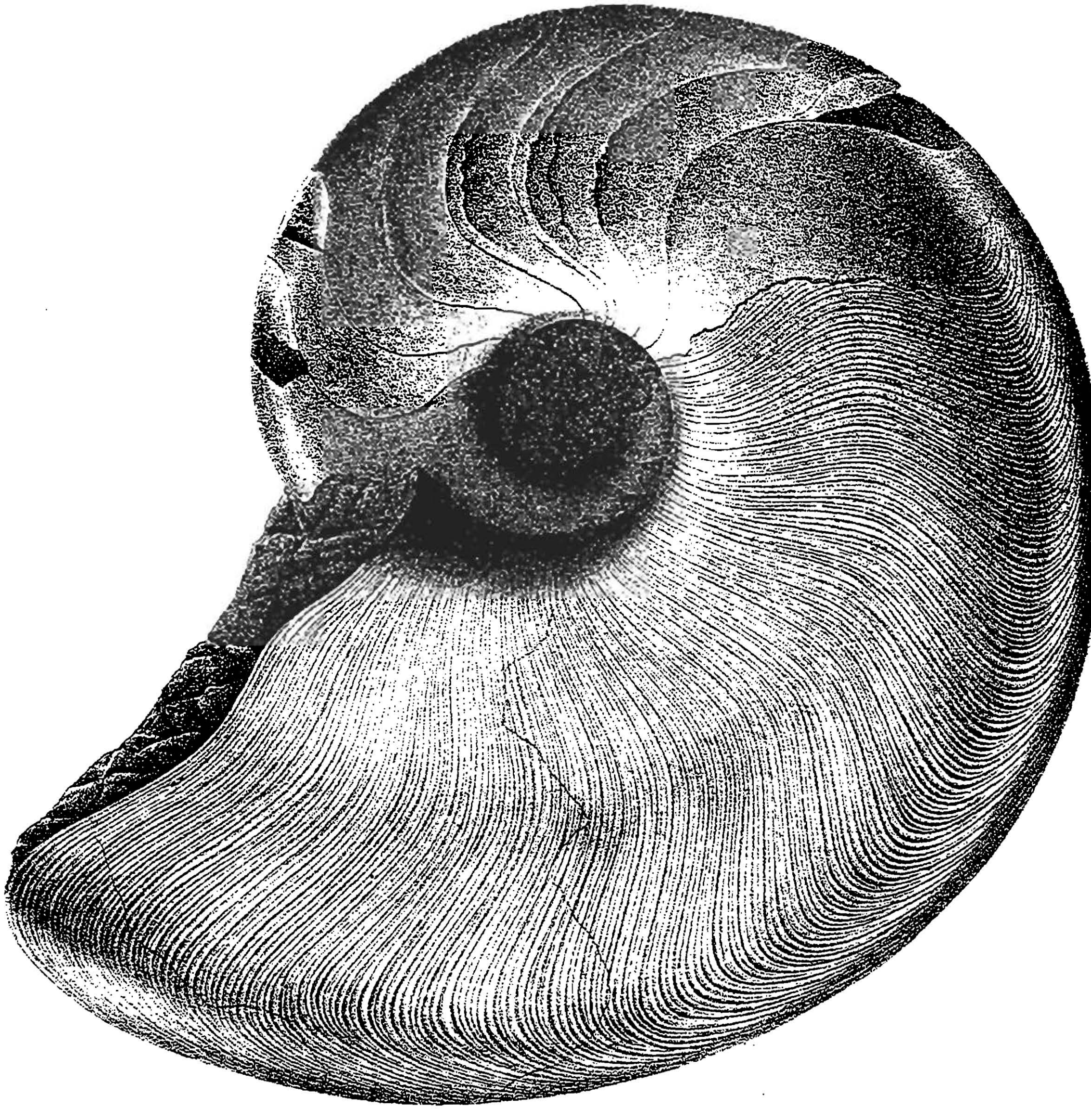
NOTE: Roadlog and stop descriptions for this fieldtrip follow Ver Straeten et al. (this volume).



Agoniatites vanuxemi

[From Hall, 1879, Natural History of New York: Palaeontology, Vol. V, Part II, Plate LXVI, Figure 1]

THE LOWER PART OF THE MIDDLE DEVONIAN MARCELLUS "SHALE," CENTRAL TO WESTERN NEW YORK STATE: STRATIGRAPHY AND DEPOSITIONAL HISTORY

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INTRODUCTION

A major succession of Middle Devonian siliciclastic rocks (Hamilton Group), which ranges from 90 m to approximately 1000 m in thickness, overlies carbonates of the Onondaga Limestone all across New York. The lowest part of these clastics is composed of black shale-dominated strata with minor limestones, and eastwardly coarsening, more sand-dominated, progradational strata, presently referred to as the Marcellus Formation. The Marcellus Formation is the lowest of four formations in the traditional subdivision of the Hamilton Group in New York State (Cooper, 1930, 1933, 1934).

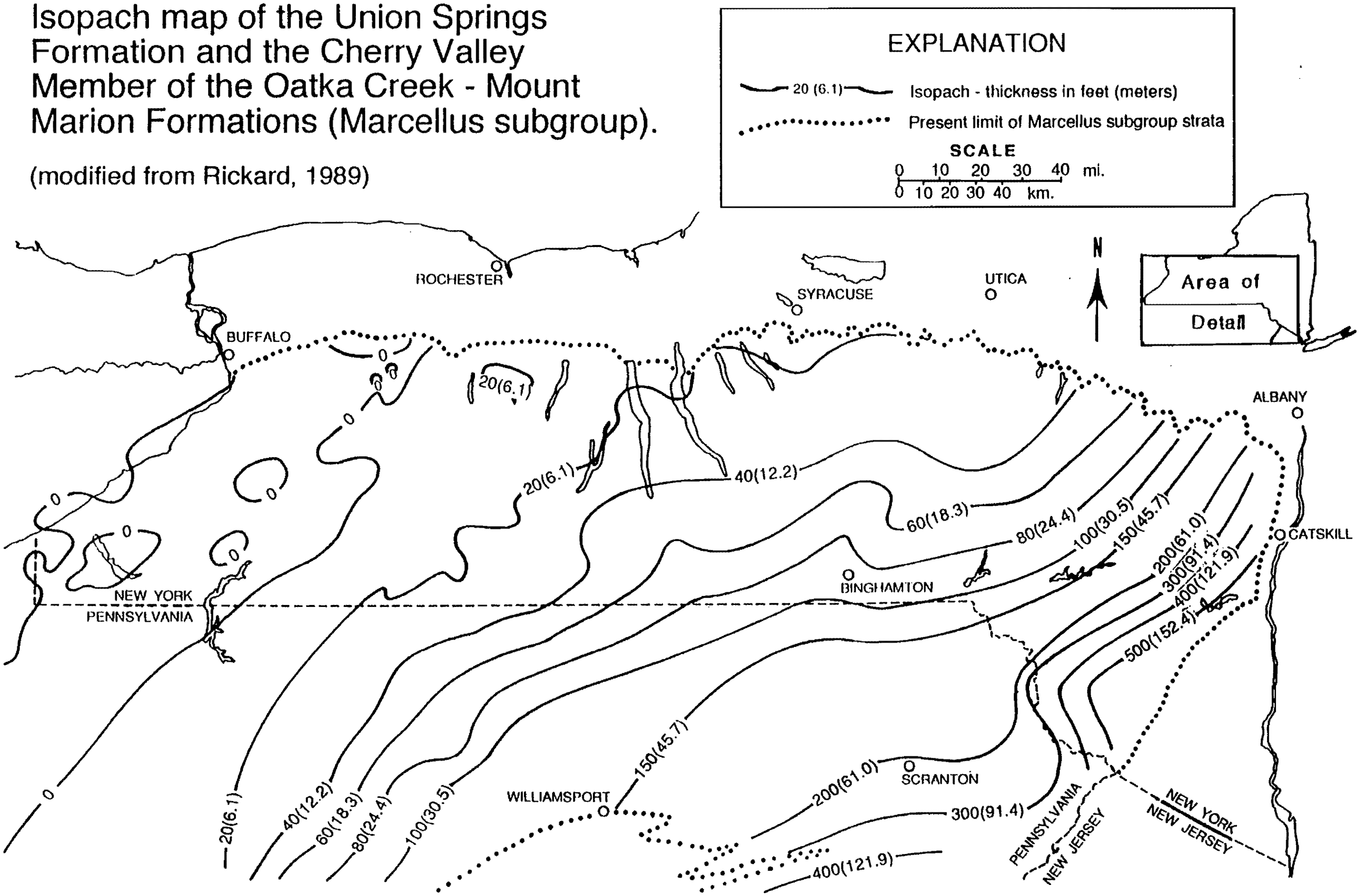
Part of this fieldtrip will concentrate on lower Marcellus strata, which across New York consist of black shales, some thin limestones, and in eastern outcrops, calcareous shale to sandstone facies. These strata range in thickness from zero to less than one meter southwest of Rochester to over 180 m in southeastern New York near Kingston (Figure 1). Recent detailed work across the state indicates that the lower part of the Marcellus Formation comprises a fifth major cycle within the Hamilton Group, which is best developed in eastern New York (Brett and Baird, in press). In addition, the fauna of this interval is distinctly different from that of underlying and overlying rocks. With this in mind, we would like to present an informal preview of a new revision of the lowest part of the Hamilton Group that recognizes the significance of these strata. The revised stratigraphic nomenclature will be formally proposed elsewhere. In this revised scheme, the "Marcellus Formation" will be raised to subgroup status and subdivided into three formations: 1) the Union Springs Formation (=lower part of the "Marcellus subgroup"); 2) the Oatka Creek Formation (=upper Marcellus subgroup basinal black shale-dominated facies of western to central New York); and 3) the laterally equivalent Mount Marion Formation (=upper Marcellus subgroup basinal black shale to shoreface sandstone facies of eastern New York). Uppermost Marcellus strata in eastern New York are represented by non-marine, fluvial-dominated sandstones of the Ashokan Formation (Rickard, 1975, 1989).

The transition from the relatively shallow marine carbonates of the Onondaga Formation to clastics of the Marcellus subgroup marks a substantial reorganization of the Devonian Appalachian foreland basin. Major changes in sedimentation, basin geometry, faunas, and paleoecology are associated with overdeepening of the foreland basin due to thrust load-induced subsidence (Ettensohn, 1985a) and eustatic sea level rise (Johnson et al., 1985) during a second active tectophase of the Acadian Orogeny (Ettensohn, 1985a).

Apparent extinction of some of the endemic Onondaga faunas and immigration of two successive biotas into the Eastern Americas Realm from Arctic Canada and Europe, respectively, are recorded in rocks of the Marcellus subgroup (Koch, 1988). The first migration is marked by a unique brachiopod and coral fauna found in the newly

Isopach map of the Union Springs Formation and the Cherry Valley Member of the Oatka Creek - Mount Marion Formations (Marcellus subgroup).

(modified from Rickard, 1989)



redefined Union Springs Formation in eastern New York. The first appearance of the classic Hamilton Group fauna occurs in the lower part of the overlying coeval Oatka Creek and Mount Marion Formations across New York. In addition, an important world-wide "extinction-radiation-extinction" bioevent involving pelagic goniatite and dacryoconariid faunas, the Kacak-*otomari* Event, occurs through this same time interval (Chlupac and Kukal, 1986; Walliser, 1986; Truyols-Massoni et al., 1990).

Analysis of laterally persistent discontinuities and macrofossil "hash" beds within skeletal limestones of the Marcellus subgroup (specifically those in the Bakoven, Hurley (new), and Cherry Valley Members) indicates condensation by short-term events which reworked long-term, time-averaged accumulations. Although the limestones of the Cherry Valley Member comprise distinctly different facies and represent relatively deeper-water deposition, they share a similar origin with younger Hamilton Group limestone members (Griffing, 1994).

One of the most significant changes between Onondaga and Union Springs/Marcellus time is the geometry of the foreland basin across New York. Deposition of the Onondaga Limestone and equivalent shallow marine carbonates in early Eifelian time was widespread and relatively uniform across much of eastern North America, which resulted in a broad, relatively tabular geometry to these carbonates. This is in contrast to the very distinctive eastward-oriented, wedge-shaped geometry that marks the overlying Union Springs Formation and the Marcellus subgroup as a whole (Figure 1). These key basinal changes were the result of thrust load-induced subsidence of the foreland basin toward eastern New York and uplift of a peripheral bulge in western New York and Ontario during early stages of Acadian tectophase II of Ettensohn (1985a).

Basal black shales and dark gray argillaceous limestones of the Union Springs Formation (Bakoven Member) overlie the Onondaga Limestone across all but westernmost New York. In eastern New York they interfinger with a thick package of calcareous shales to sandstones (Stony Hollow Member). A thin, widespread, fossiliferous package (the newly proposed Hurley Member) occurs at top of the Union Springs Formation along all of its New York outcrop.

The base of the overlying Oatka Creek and Mount Marion Formations is marked by cephalopod-rich limestones and equivalent sand-dominated calcareous strata in eastern New York (Cherry Valley Member). Black shales overlie the Cherry Valley all across New York; in eastern New York, however, they are succeeded by a thick package of progressively coarser, increasingly shallower, progradational clastics that are equivalent to black shale facies in the western part of the state.

Griffing and Ver Straeten (1991) presented the first detailed discussion of lower Marcellus strata in eastern New York. In this paper the authors will examine the equivalent rocks in west-central to western New York in detail, and discuss the larger scale implications of this major transition in the Devonian of the Appalachian Basin.

GEOLOGIC OVERVIEW

PRESENT GEOLOGIC SETTING

Outcrops of the Marcellus Shale/subgroup are exposed along an east-west trending outcrop belt spanning upstate New York from Buffalo to the Albany area. Near Albany the outcrop belt bends southward along the Catskill Front and farther southwest into

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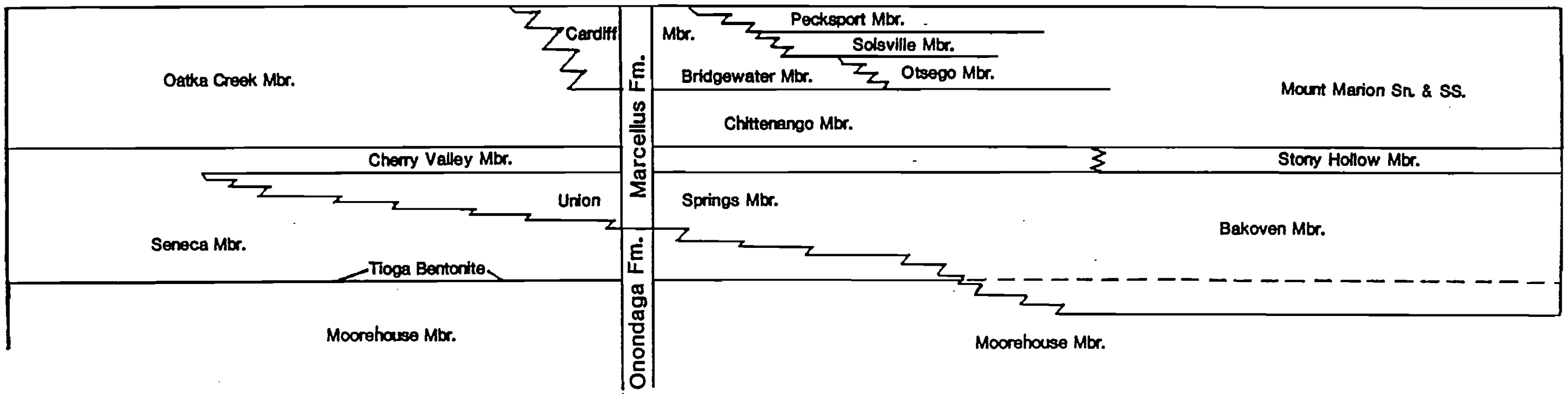
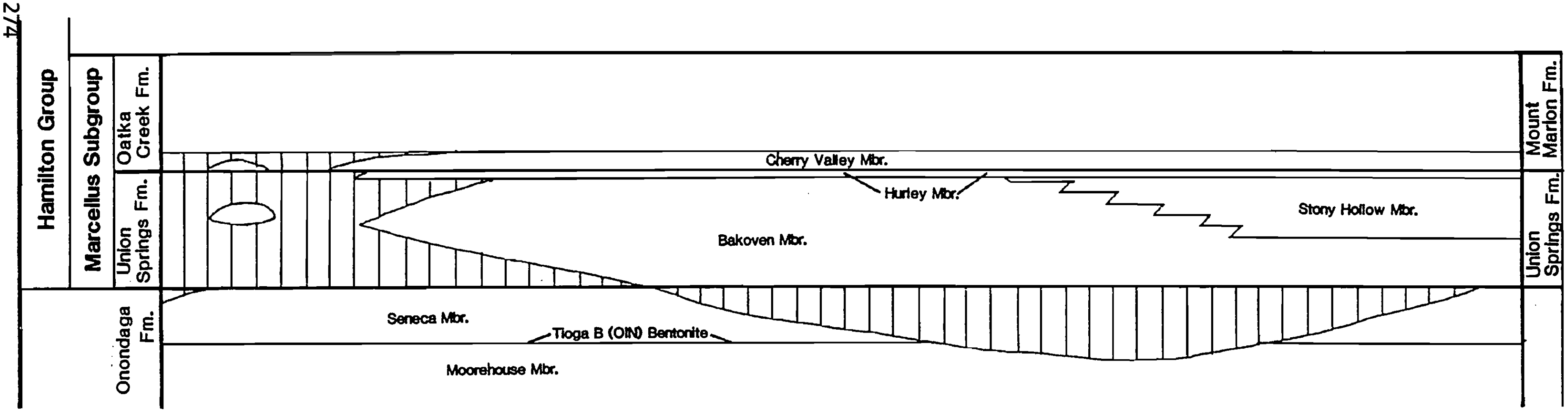
◄
Figure 1. Isopach and outcrop map of lower strata of the Marcellus subgroup (Union Springs Formation and the Cherry Valley Member of the Oatka Creek and Mount Marion Formations) in the Northern Appalachian Basin (modified after Rickard, 1989).

after Rickard, 1975



Proposed Stratigraphic Revision



Pennsylvania, Maryland, and the Virginias through the Valley and Ridge Province.

These lower Middle Devonian strata are moderately to intensely folded and faulted in eastern New York and the central Appalachians (e.g., Bosworth, 1984a,b). Equivalent strata display little deformation in central to western New York; however, broad, low-amplitude folds and minor shear zones and thrust faults occur.

Most natural exposures of Marcellus strata lie in gullies or ravines, where streams have cut through intervals of poorly resistant shale, and where limestones and sandstones form cataracts and caps of waterfalls. Common usage of the underlying Onondaga Limestone for crushed stone in central to western New York also provides valuable quarry exposures of the lower part of the Marcellus Shale.

STRATIGRAPHIC REVISION

In the course of study of the Marcellus "Formation" in New York it has become apparent that this package of strata, which ranges in thickness from approximately 15 m on the Lake Erie shore (western New York) to over 580 m in the Catkill Front (eastern New York; Rickard, 1989), represents a more complex unit than has been previously recognized. A formal revision of these strata is presently in process (Ver Straeten et al., in prep), but the authors would like to take this opportunity to present it as an informal preview. The following is a brief outline and justification of the basis of the new stratigraphic scheme, which is summarized in Figure 2.

a) Raise the Marcellus "Formation" to subgroup status and split it into two formation-level subdivisions: The lower part of the Marcellus "Formation" represents a fifth major cycle in the Hamilton Group (the lowest of the five), equivalent in nature to the four previously recognized formations. Most notably in eastern New York its forms a distinctive succession of black shale to buff-weathering calcareous shale and sandstone easily distinguishable from overlying upper Marcellus strata. In addition, lower Marcellus strata feature a unique fauna related to other late Eifelian assemblages in eastern North America. This fauna is very distinct from those of the underlying Onondaga Formation and the overlying remainder of the Hamilton Group. Therefore, the term Marcellus, which represents rocks from the top of the Onondaga Formation to the base of the Skaneateles Formation, is raised to subgroup status. The Marcellus subgroup will consist of two formation-level packages of strata:

   i) In the lower part of the Marcellus subgroup, the Union Springs Member is raised to the Union Springs Formation.

   ii) The upper part of the Marcellus subgroup consists of the Mount Marion Formation in eastern New York State; in central to western New York the Oatka Creek Member is raised to formational status.

b) Raise the rank of the Union Springs Member to formational status: The Union Springs Member as previously defined incorporates all strata included in the proposed lower formation of the "Marcellus subgroup" in western to central New York. Raising the unit to formational rank maintains the usage of a familiar term which should lead to less confusion and easier acceptance among various workers. At present the Union Springs in western to central New York is the lateral equivalent of several members in eastern New York. The proposed Union Springs "Formation" would incorporate three members across New York State; black shales of the Bakoven Member (revised), calcareous shales to sandstones of the Stony Hollow Member in eastern New York (revised), and the distinctive Hurley

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◄ Figure 2. Previous and revised stratigraphic nomenclature of the Marcellus subgroup in New York State.

this succession of goniatite faunas is known world wide, and recognized as part of the Kacak-*otomari* global bio-event.

Biostratigraphy of other forms (e.g., brachiopods and nowakiids) has received less attention up to the present. In addition, the previous studies have focused chiefly on the condensed and more black-shale dominated succession in central New York. We hope the new detailed stratigraphic work in the thicker and more complete eastern sections (Griffing and Ver Straeten, 1991; Ver Straeten, 1994; Ver Straeten et al., in prep.) will facilitate more highly resolved biostratigraphic study of these rocks.

Due to the lack of a detailed biostratigraphy of this interval, the Eifelian-Givetian boundary in New York is not well defined. Various authors place the boundary anywhere from the Cherry Valley Member (House, 1978) to the Skaneateles Formation above the Marcellus subgroup (Kirchgasser and Oliver, 1993). Rickard (1984) states that the boundary probably lies close to the Cherry Valley Member.

THE LOWER PART OF THE MARCELLUS SUBGROUP IN CENTRAL TO WESTERN NEW YORK STATE

BASAL CONTACT PROBLEM

The basal contact of the Marcellus subgroup with the underlying Onondaga Formation in New York has been widely discussed, and has been the source of long-standing debate. Previous workers have variously interpreted the contact to be: 1) diachronous due to lateral gradation of the limestones of the Seneca Member (upper part, Onondaga Formation) eastward into the black shales of the Union Springs-Bakoven Members (Clarke, 1901; Oliver, 1954; Rickard, 1975); 2) diachronous due to corrasion of upper part of the Onondaga Formation (Brett and Baird, 1990); 3) isochronous (Conkin and Conkin, 1984); or 4) diachronous and disconformable. Rickard (1984, p. 824-826) recently reviewed this problem, and on the basis of subsurface correlations, stated the Onondaga-Marcellus contact is a "major, regional (widespread) unconformity"; Chadwick (1944) and Lindemann and Feldmann (1987) also suggest a disconformable contact in eastern New York.

Detailed study of the Seneca Member of the Onondaga Formation and the overlying lower part of the Bakoven Member (Union Springs Formation) along the New York outcrop provides new insights into this long-standing problem. Brett and Ver Straeten (this volume) discuss the stratigraphy, geometry, and distribution of the Seneca Member in New York, and thus, these relationships will be summarized only briefly herein.

The Seneca Member is thickest and best developed in the vicinity of its type section in the central Finger Lakes region (e.g., Seneca Stone Quarry, Stop 5), where the member is 7.15 m-thick and is characterized by wackestones to packstones interbedded with thin dark shales and bentonites of the "Tioga Ash Zone." The member thins slightly to the west; south of Rochester (Honeoye Falls Quarry, Stop 1) the Seneca is 6.65 m-thick and appears to contain all strata found in the type area, except possibly a thin interval at the top. East of the central Finger Lakes, however, the Seneca Member progressively thins from 5.4 m near Syracuse to 2.0 m at Cherry Valley. No Seneca is reported from the Albany area of eastern New York (however, see Brett and Ver Straeten, this volume). Subsurface studies by Rickard (1989, Plate 31) indicate the Seneca Member reappears south of Albany and thickens into northeastern Pennsylvania.

One of the key features of the Onondaga-Union Springs transition is the occurrence of multiple, thin paleovolcanic deposits of the Tioga Ash Beds. In New York the Tioga beds occur chiefly within the Seneca Member. The Tioga bentonites cluster, as outlined by Way et al. (1986) for central Pennsylvania, appears to be complete in west-central New York (Honeoye Falls to Seneca Stone Quarries), where up to eight beds are associated with the Seneca Member; these include the prominent Onondaga Indian Nation Ash (OIN, Conkin and Conkin, 1984; Conkin, 1987; =Tioga B of Way et al., 1986) and the Tioga F bed near the

Onondaga-Union Springs contact (see Brett and Ver Straeten, this volume). East of the central Finger Lakes region toward Syracuse and beyond, however, the upper bentonites of the Tioga cluster appear to be progressively removed, associated with the top-downwards truncation of the Seneca Member (see Brett and Ver Straeten, this volume). Active search for the bentonite layers in the overlying black shales in eastern New York, analgous to their occurrence in central Pennsylvania (where the Tioga beds occur interbedded with Seneca-equivalent black shales; Way et al., 1986; see Figure 6 in Brett and Ver Straeten, this volume), has been unsuccessful; no Tioga bentonite beds have been found in the lower part of the Bakoven Member in east-central to eastern New York.

The interval of the Onondaga-Union Springs formational contact commonly features fish bone-phosphatic lag deposits. As discussed below, the bone beds appears to represent long-term submarine condensation surfaces associated with transgression and sediment starvation-corrasion. These lag deposits appear to be diachronous across the central to eastern New York outcrop belt, due to their occurrence on top of progressively older strata of the Seneca Member eastward.

BONE BEDS AND SKELETAL LIMESTONES OF THE MARCELLUS SUBGROUP

Bone Beds in central to western New York

Introduction

Several thin intervals of concentrated skeletal phosphate lie within limestones of the uppermost Onondaga Formation and parts of the overlying Marcellus subgroup in west-central New York. Three notable "bone bed" intervals occur: 1) in the Seneca Member at or near the Onondaga-Marcellus contact, 2) in the lower part of the Union Springs Formation, and 3) within the Oatka Creek Formation, at the contact of the Cherry Valley Member with overlying shales. Some these bone beds have been recognized previously, but preliminary analysis of their sedimentological and stratigraphic significance is presented in the following section.

Bone Bed At The Onondaga-Union Springs Contact

An interval of concentrated fish armor, fin spines, and teeth occurs at the contact of the Onondaga Limestone with the overlying Union Springs Formation in west-central New York. Conkin and Conkin (1975) contend that this bed correlates to bone beds that separate Onondaga and Marcellus equivalents in central Ohio, a distance of more than 670 km (Bone Bed 7 of Conkin and Conkin, 1975, 1979). This bone bed is perhaps best developed at the General Crushed Stone quarry, east of Jamesville, where it measures up to six cm-thick. A similar bone bed is recognized at Seneca Stone quarry, where it forms a thinner bone interval in erosional swales above the Seneca Member limestones, and at the Honeoye Falls quarry, where a thin bone interval directly underlies the Tioga F metabentonite.

The Onondaga-Union Springs contact (OUSC) bone bed at Jamesville quarry comprises crinoidal packstone-wackestone on top of otherwise skeletal-poor, bioturbated, dark gray, lime mudstones. Although bioturbated, several 3 to 10 mm-thick sedimentation units are faintly preserved and the basal contact of the bone bed locally forms a sharp, erosional discontinuity. The coarsest skeletal accumulations lie near the base of the bone bed.

The OUSC bone bed features completely disarticulated placoderm armor, large acanthodian fin spines (some *Machaeracanthus sulcatus* spines measure at least 26 cm-long), and individual denticles from the parasymphisial teeth of the marine rhipidistian *Onychodus* (both *O. sigmoides* and *O. hopkinsi*). The coarse, basal fish bone concentration is associated with abundant, large (commonly 1.0 to 1.5 cm-wide and 3 to 6 cm-long) micritic intraclasts which closely resemble rounded segments of the compressed, haloed, horizontal burrows observed within underlying strata. These intraclasts are interpreted as reworked portions of partially cemented burrows. Long segments of exhumed burrows show parallel alignment to each other and were probably exhumed by storm waves or currents. Small, rounded quartz pebbles are also a common component of the coarsest fraction of the bone bed.

In addition to phosphatized fish debris, small pelmatozoan ossicles and the valves of small chonetid brachiopods are phosphatized. Although black in color like unweathered phosphatic debris, the larger pelmatozoan ossicles (particularly large crinoid columnals) display various degrees of pyrite replacement instead. Most large pelmatozoan ossicles are disarticulated, slightly abraded and bored by endoliths, but some short, unworn, articulated stem segments also exist.

Bakoven Bone Bed

Another bone bed occurs within styliolinid grainstones and packstones in the lower black shales of the Bakoven Member (Union Springs Formation). Although several "black" styliolinid grainstones and packstones have been identified within Union Springs black shales at other west-central New York localities, styliolinid-rich bone beds are, at present, only recognized at the Honeoye Falls quarry (Griffing, 1994) and at the Seneca Stone quarry (Baird and Brett, 1986a). Although Union Springs strata are absent west of the Genesee Valley, Conkin and Conkin (1979, 1984) also correlate this bone bed with one in central Ohio (bone bed 8). This Bakoven bone bed occurs in the uppermost bed of a 15 to 19 cm-thick bundle containing 2 to 20 mm-thick, ripple cross-stratified, styliolinid grainstone/packstone beds which are separated by thin beds, drapes or partings of black shale. The interiors of many styliolinids within the grainstones are also filled with black shale. The small, low-relief ripples commonly contain wave discordant cross-laminae, but net migration direction appears to be to the east and southeast.

Like the OUSC bone bed, the Bakoven bone bed contains common disarticulated placoderm armor and *Onychodus* teeth. Also, abundant sand-sized, angular fragments of fish bone intermix with the styliolinids and highlight ripple foresets. Rare, rounded black shale clasts occur with the coarsest fish debris at the base of the bed.

Cherry Valley Bone Bed

A bone bed also occurs at the top of the Cherry Valley Member at several localities in west-central New York. Disarticulated fish armor is concentrated at and near the upper surface of a 1 to 3 cm-thick styliolinid packstone bed which also contains abundant, truncated goniatite and orthocone cephalopod shells. The bone-cephalopod bed overlies a firmground discontinuity that truncates another cephalopod-rich bed; it is directly overlain by fossil-poor black shales. This bone bed has been identified in Cherry Valley sections from Nedrow to Seneca Stone quarry, where it is best exposed. The bone bed occurrence coincides with the most condensed but complete sections of the Cherry Valley Member in New York State.

Most fish detritus consists of individual fish scales and disarticulated placoderm armor (particularly of the arthrodire *Clarkeosteus*). Head shield armor of an even larger arthrodire has been observed at Seneca Stone quarry exposures, where one fragment measured 56 cm in length (Griffing, 1994). Many of the placoderm plates found in this bed are disarticulated, but some remain partially articulated. The bone vesicles and the skeletal matrix of the uppermost Cherry Valley bed are extensively pyritized, and the upper surface of the bone bed forms a firmground or incipient hardground discontinuity. Rare, sub-rounded quartz and quartz sandstone pebbles are also present along the uppermost portion of the bone bed. Although reworked burrow clasts are not common in this bone bed, orthocones within this bed show a roughly unidirectional, offshore-directed current alignment and upper portions of *Rhizocorallium* traces have been eroded.

Discussion and Interpretation

A review of middle Paleozoic through Holocene bone beds by Antia (1979) demonstrates that there are many processes and pathways leading to bone bed formation. Disarticulated bone beds are commonly thought of as either: 1) hiatal deposits representing long-term accumulations concentrated by corrasion of less resistant skeletal, or 2) physically condensed deposits concentrated by hydrodynamic exhumation and density sorting.

Carbonate-hosted bone beds are common in the rock record. For example, a thin, marine vertebrate bone concentration capping the Santee Limestone (middle Eocene) of the South Carolina represents a diastem of several hundred thousand years duration (widespread) to millions of years in duration (locally where the Cooper Marl is absent and the Santee Limestone directly underlies the Pliocene-age Duplin Formation; Sanders, 1974). Personal examination of this long-term hiatal accumulation (by DHG) revealed bone fragments and shark teeth in various states of preservation; from extremely rounded and polished bone with abundant borings, to pristine, unabraded skeletal elements. Vertebrate skeletal material is directly associated with coated-grain phosphate and overlies a pitted, bored, abraded, and partially phosphatized hardground surface at the upper contact of the Santee Limestone. Partially articulated archaeocete whale skeletons from the upper Santee Limestone (Sanders, 1974) suggest that short-term event burial may also have contributed to bone bed concentration. The diverse skeletal assemblages and the rock types surrounding this bone bed indicate formation in very shallow, well-oxygenated waters.

Unlike upwelling models for phosphatization, phosphate enrichment of bone and phosphatization of calcitic skeletal debris can take place in the very shallow subsurface, during extremely slow background sedimentation (Baird and Brett, 1986). Phosphatization of bone can occur in alkalic pore waters of organic-rich sediments (Burnett, 1977; Martill, 1991). Decaying organics within sediments act as one source for the phosphate. Clasts already enriched with apatite (bone) make preferable sites for nucleation, but calcitic grains also provide good nuclei (Antia, 1979). Such "pre-fossilization" by apatite or pyrite may allow normally low density skeletal remnants to remain as part of the coarse lag developed from hydrodynamic reworking (Reif, 1982).

Like the Santee bone bed, Devonian bone beds often also represent diastems of long duration. The 4 cm-thick Upper Devonian North Evans bone-conodont bed of westernmost New York represents a reworked phosphate-rich aggregate of three conodont zones from the Middle and Upper Devonian (Huddle, 1981). The Middle Devonian bone beds of central and western New York may represent shorter diastems, but they comprise similar bone concentrations. Taphonomic and sedimentologic evidence suggests that short-term, event-driven hydrodynamic modification of the sea-floor played a critical role in condensation of these, and many other bone beds. Episodes of storm-generated reworking, mass mortality(?), and rapid deposition (triggering pyrite formation?) alternated with long periods of slow or no background sedimentation. Between episodic events, phosphatization of buried skeletal debris occurred accompanied by corrasion of carbonate skeletons that were exposed on the sea-floor. Later storm events probably disarticulated previously buried fish remains and mixed skeletal debris modified by rapid burial and non-depositional processes. Large quartz pebbles found in some of these bone beds more likely represent rare dropstones rafted in by plant/tree roots, rather than bedload transported sediment. The bone beds described here are all overlain by black shale facies and may represent lags formed during initial transgression or, more likely, during maximum transgression.

Limestones of the Chestnut Street submember (Hurley Member)

Introduction

A thin bundle of skeletal limestone beds form the base of the newly proposed Hurley Member of the Union Springs Formation from eastern New York to Honeoye Falls quarry, south of Rochester. Many of the distinctive elements of the Union Springs fauna which are recognized as part of the Kacak-*otomari* event (discussed above) are concentrated in the micritic packstones, wackestones, and minor grainstones of the Chestnut Street submember.

Lateral Extent and Thickness

The Chestnut Street submember acts as a widespread stratigraphic marker which not only extends across central New York but also persists from eastern New York into the central Appalachians (discussed below). At Kingston, New York, the Chestnut Street submember forms a 7 m-thick series of thin sandstones and interbedded silty shales. However, the entire bundle measures 20 to 43 cm-thick in eastern and east-central New York, where it contains from 2 to 7 individual limestone beds amalgamated into either one composite bed or into two shale-separated beds. The number of amalgamated beds and the overall thickness of the bundle decreases slightly across west-central New York, where it is overstepped by the Cherry Valley Member. The Chestnut Street submember reaches its minimum thickness at Flint Creek, near Phelps, where it forms a single 3 to 12 cm-thick bed which was scalloped and nearly removed by sub-Cherry Valley erosion. An anomalously thick (42 to 48 cm) Chestnut Street bundle of 3 to 4 beds is present along part of the

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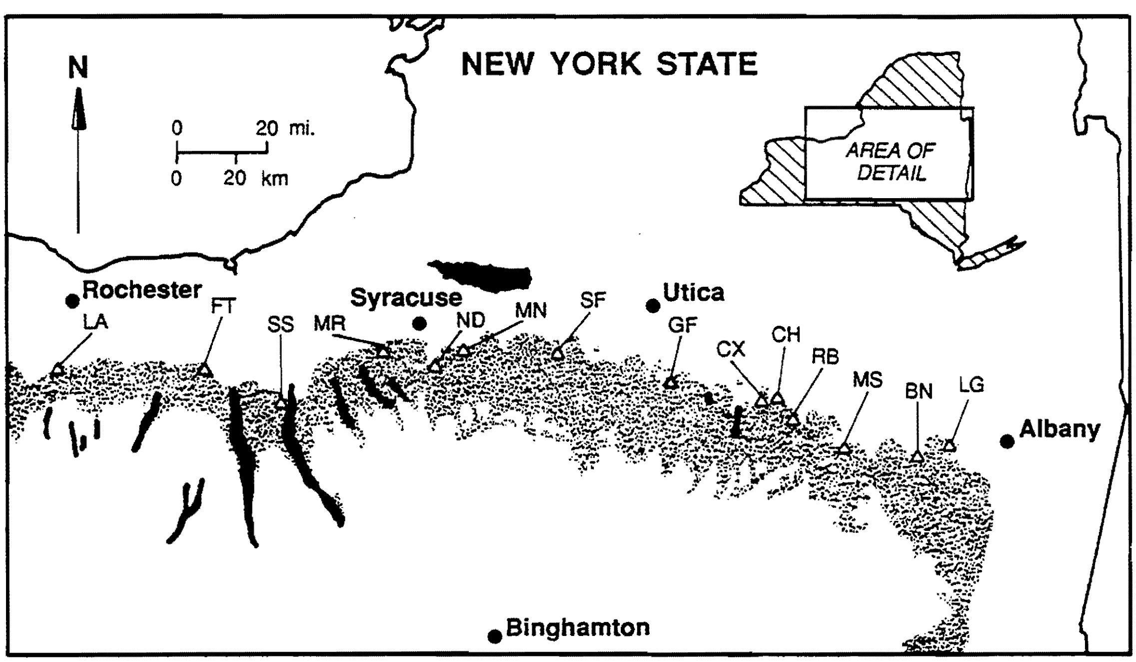
Figure 3. Map of selected outcrops of the Hurley Member (Union Springs Formation) and the Cherry Valley Member (Oatka Creek and Mount Marion Formations)between Albany and Rochester regions of New York State. Localities as follows: LA=Honeoye Falls quarry, near Lima, FT=Flint Creek, SS=Seneca Stone quarry, MR=Marcellus, ND=Nedrow, MN=Manlius, SF=Stockbridge Falls, GF=Gulf Road near East Winfield, CX=Cox Ravine, CH=Chestnut Street, RB=Rosenberg Road near Seward, MS=Mineral Springs, BN=Irish Hill near Berne, LG=Long Road near Thompson Lake.



exposures at the Honeoye Falls quarry. However, the unit is locally truncated to completely removed in other parts of the exposure (see below).

Contacts

The basal contact of the Chestnut Street submember manifests itself as an erosional discontinuity in western New York, where it truncates black limestone concretion beds of the Bakoven Member in places. The basal contact in east-central New York is obscured by "underbed" diagenetic modification, but minor downcutting is implied by reworked *Cabrieroceras plebeiforme* in the lower Chestnut Street beds. The upper contact is distinct but comformable in east-central New York, where overlying Hurley Member siltstones and black shales (Lincoln Park submember) separate the limestones of the Chestnut Street submember and the Cherry Valley Member. The upper contact with the Cherry Valley Member is sharp and forms a widespread disconformity in west-central and western New York.

Internal Stratigraphy

Scalloped hardground and firmground discontinuities separate two to three discrete beds of the amalgamated Chestnut Street submember in central and western New York State. These discontinuities, together with systematic vertical variation of rock types and faunal content, allow the correlation of individual Chestnut Street beds across this portion of outcrop (Figures 3 and 4). The lowermost one or two beds comprise fine-grained, burrow-mottled skeletal packstones/wackestones that weather a very light gray color. Calices of the minute crinoid *Haplocrinites clio* are most commonly found in these micrite-rich beds. These lower beds are separated by a locally manifested firmground contact and are capped by a highly irregular hardground surface. Relict sedimentation units within these lower beds appear to be normally graded. The uppermost bed or beds comprise coarser-grained, crinoid-rich skeletal packstones (and locally grainstones) with relatively more abundant fish bone and conodont remains. This crinoid packstone bed attains a maximum thickness of 11 cm at Pleasant Valley Road, in Marcellus, but most surrounding localities feature a 1 to 5 cm-thick bed that may be locally removed along individual outcrops. A scalloped, pyritized hardground surface forms the sharp contact between the welded Chestnut Street and Cherry Valley limestones in west-central New York (Figure 5). Clasts of the pyritic Chestnut Street crust that occur within overlying Cherry Valley Member limestones indicate synsedimentary pyrite formation.

Interpretation

The fauna of the Chestnut Street submember represents the most diverse fossil assemblage in the lower part of the Marcellus subgroup. Although the diminutive size of most benthic faunal elements in central to western New York suggests that some oxygen deficiency persisted throughout Union Springs deposition, the sea-floor was probably mildly dysaerobic to marginally aerobic in that part of the basin during Chestnut Street deposition. The Chestnut Street submember limestones occur directly above the Stony Hollow Member in eastern New York, which forms a coarsening-upward, calcareous shale-to-sandstone succession. The faunal assemblage, stratigraphic position and bedload transport features of this limestone bundle suggest a relatively shallow-water origin. The large micrite component (of algal origin?) of these limestones and the eastward

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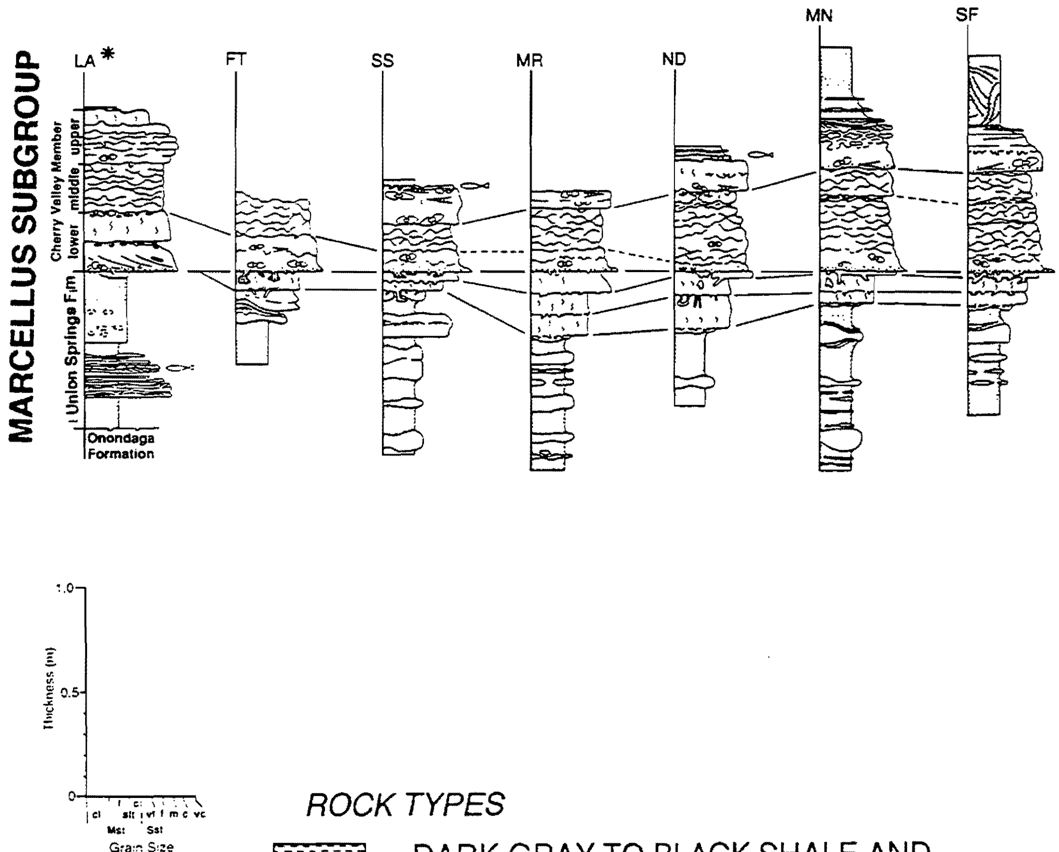
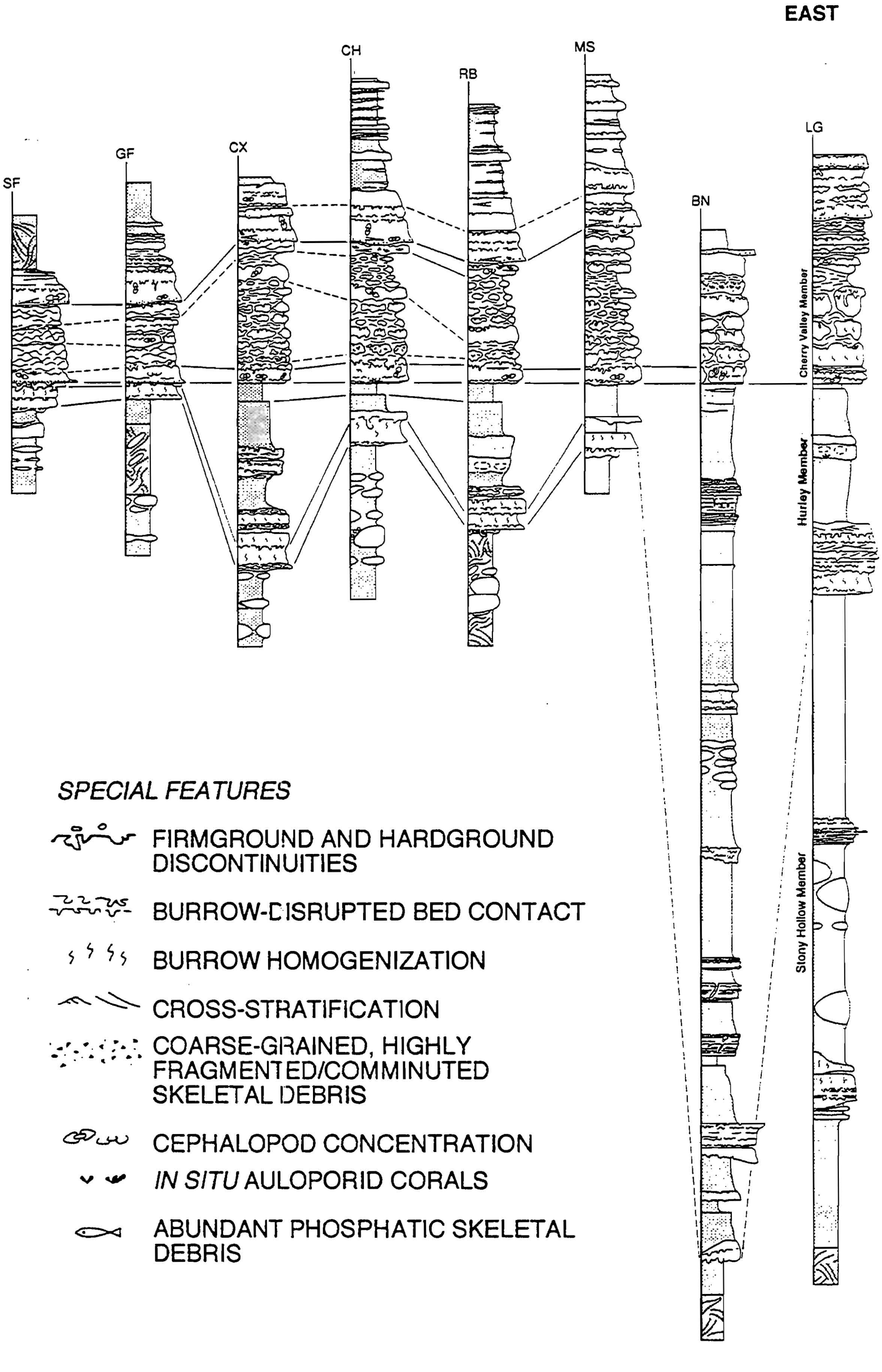
Figure 4. Stratigraphic correlation of the Hurley and Cherry Valley Members between selected sections in New York State. Datum is the base of the Cherry Valley Member (=base of Oatka Creek and Mount Marion Formations). Lower tie lines correlate limestones of the Chestnut Street submember. Localities are same as for Figure 3.

WEST



ROCK TYPES

- DARK GRAY TO BLACK SHALE AND MUDSTONE WITH CARBONATE CONCRETIONS
- TECTONIZED DARK GRAY TO BLACK SHALE
- SKELETAL LIMESTONE
- NODULAR SKELETAL LIMESTONE
- ISOLATED SKELETAL LIMESTONE NODULES IN MARLSTONE
- SILICICLASTIC COARSE-GRAINED SILTSTONE AND SANDSTONE


EAST
CH
RB
MS
LG
GF
CX
SF
BN
Cherry Valley Member
Hurley Member
Stony Hollow Member
SPECIAL FEATURES
FIRMGROUND AND HARDGROUND DISCONTINUITIES
BURROW-DISRUPTED BED CONTACT
BURROW HOMOGENIZATION
CROSS-STRATIFICATION
COARSE-GRAINED, HIGHLY FRAGMENTED/COMMINUTED SKELETAL DEBRIS
CEPHALOPOD CONCENTRATION
IN SITU AULOPORID CORALS
ABUNDANT PHOSPHATIC SKELETAL DEBRIS

overstepping of limestones over Stony Hollow siliciclastics suggest deposition during a hiatus. Relict graded beds even in bioturbated facies of east-central New York indicate event deposition (probably by storms) toward the basin center. The limited basinward expansion and broad facies distribution of both the Chestnut Street and Cherry Valley limestones suggests a broad, poorly-defined trough centered east of the Cherry Valley region with extremely gradual, basin-marginal slopes. Any major bathymetric asymmetry in the initial phases of the Marcellus basin was temporarily reduced before Chestnut Street deposition, probably by infilling of the more rapidly subsiding eastern portion of the basin by Bakoven and Stony Hollow sediments.

Despite aggressive erosional downcutting, thin limestone beds were preserved over hundreds of kilometers, due in large part to syndepositional cementation. Hardgrounds, incipient hardgrounds and firmgrounds formed resistant barriers to erosion during and after Chestnut Street deposition.

Cherry Valley Member in central to western New York

Introduction

A complex bundle of dark, organic-rich, skeletal limestones, marlstones, and shales known as the Cherry Valley Member forms the base of the coeval Mount Marion and Oatka

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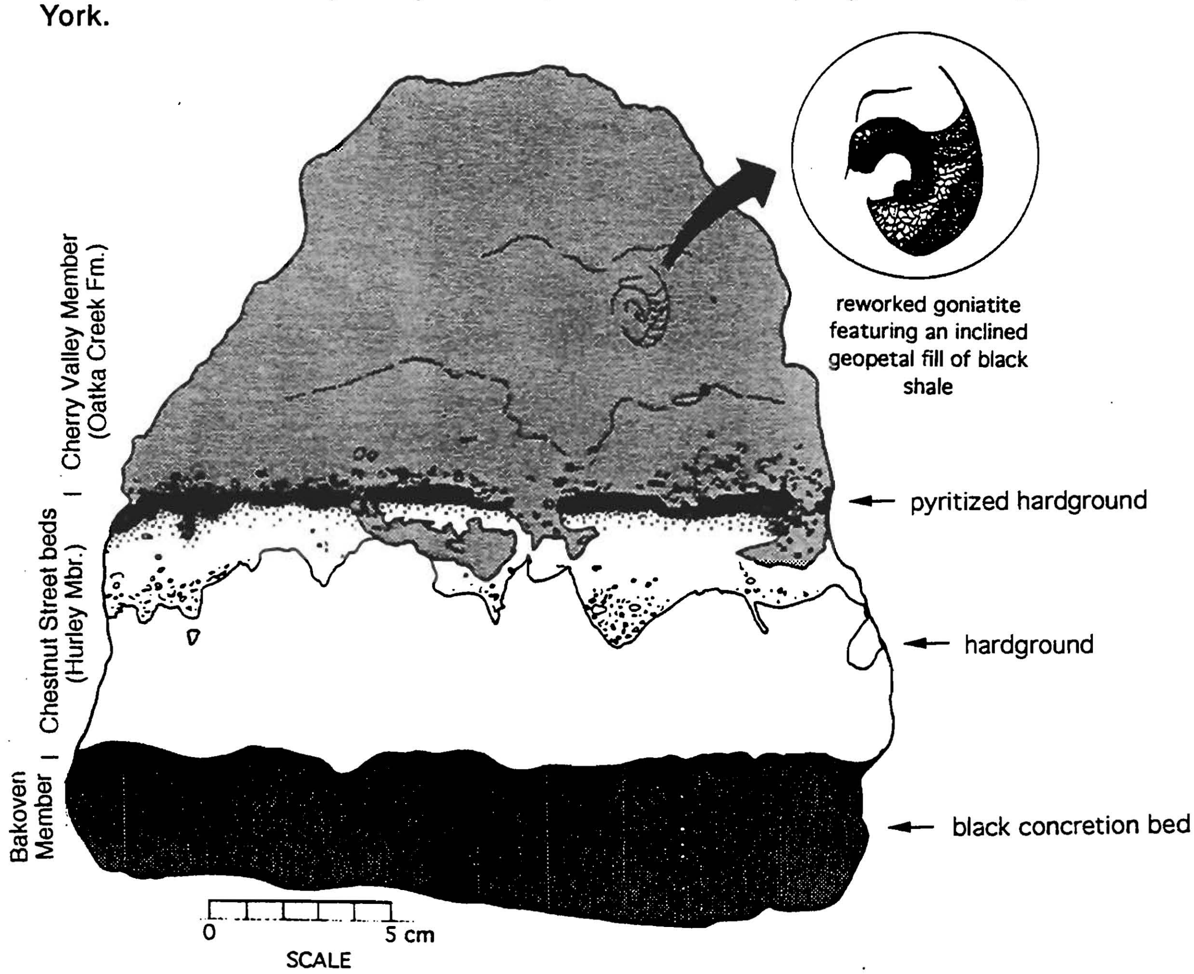
Figure 5. Sketch of polished slab from part of the welded limestone bundle (Chestnut Street submember and Cherry Valley Member) at Seneca Stone quarry, near Canoga, New York.



Creek Formations from Onesquethaw Creek, near Albany, to Honeoye Falls quarry, south of Rochester. Like the Chestnut Street submember limestones, the "Cherry Valley Limestone" (*sensu* Rickard, 1952) forms part of a carbonate-siliciclastic lithosome that extends into the central Appalachians (Griffing and Ver Straeten, 1991; see below). Both the carbonate and siliciclastic portions of the Cherry Valley Member contain a distinctive and widely known cephalopod fauna (Flower, 1936; Miller, 1938) highlighted by the ammonoids *Agoniatites vanuxemi, A. floweri*, and the orthocone nautiloid *Striacoceras typum.* Cherry Valley limestones consist of organic- and pyrite-rich, slightly neomorphosed packstones and grainstones which are dominated by the minute pelagic fossil *Styliolina fissurella* (17 to 39%, avg. 25% of total) and by disarticulated crinoid ossicles (1 to 33%, avg. 7% of total). The limestones also commonly contain remains of auloporid corals, the worm tube *Coleolus aciculatum,* minute rhynchonellid brachiopods, gastropods, nowakiids, ostracodes, bivalves and fish bone. The skeletal component and the common nodular fabric of these limestones is comparable to Devonian pelagic limestones of Europe and North Africa, such as the Cephalopodenkalk and the Griotte (Griffing and Ver Straeten, 1991; Griffing, 1994).

Lateral Extent and Thickness

The Cherry Valley Member thins rapidly along the eastern New York outcrop belt from a 10 m-thick siliciclastic-dominated interval at Kingston to a 1.76 m-thick sandy limestone at the south branch of Onesquethaw Creek. Westward thinning of the limestone bundle across eastern and central New York is very gradual, from 1.4 m-thick at Long Road ravine to about 41 cm-thick at Seneca Stone quarry (Figure 4; Stop 6). Even though the Cherry Valley Member is disconformity bound in west-central New York outcrops, the unit varies less than 10 cm in thickness along individual outcrops in central New York. The Cherry Valley at Seneca Stone quarry represents the thinnest section which contains all three submembers. Farther west the completeness and thickness of the unit vary considerably before disappearing altogether. The 37 cm-thick section at Flint Creek contains only lower and middle submembers below a hummocky erosional surface. The 0.40 m- to 3.15 m-thick section of the Cherry Valley Member at Honeoye Falls quarry (discussed below) demonstrates more thickness variation across a single outcrop than the unit does across the entire rest of central New York.

Contacts

The basal contact is sharp and paraconformable in east-central New York sections but represents a distinct erosional disconformity both east and west of the area. The sub-Cherry Valley disconformity in west-central and western New York is manifested by an irregular, scalloped hardground surface developed on the uppermost bed of the Chestnut Street submember (Hurley Member). Similarly, the upper contact is gradational and conformable with overlying Chittenango shales in east-central New York, as evidenced by the progressive vertical decrease in fossiliferous, calcareous black shale intervals and thin, styliolinid limestone beds. The upper contact becomes disconformable in west-central and western New York and is marked by a partially pyritized, scalloped, firmground or incipient hardground surface associated with cephalopod-bone bed pavements (see bone bed description).

Internal Stratigraphy

Rickard (1952) recognized a three-part subdivision of the Cherry Valley Member in east-central and eastern New York. It consists of: 1) a lower massive limestone, 2) a middle nodular limestone, and 3) an upper massive limestone. Submembers similar to Rickard's subdivisions are recognized by Griffing (1994; this article), but the placement of submember boundaries is based on bedding geometry and internal sedimentary fabrics, whereas Rickard's boundaries were based primarily on the weathering profile.

The lower submember consists of a single, laterally continuous, 9 to 15 cm-thick, bioturbated bed in east-central New York. This submember grades upward into large irregular, isolated to nearly continuous packstone nodules in marlstone (of the middle submember) in this area. Relict, burrow-mottled discontinuities separate several 1 to 3 cm-thick units within the bed in east-central New York. The lower submember thins to a minimum of 5 cm-thick between the Gulf Road and Marcellus localities, where it is commonly represented by a coarse skeletal hash directly above the basal hardground contact. The unit progressively thickens from the Marcellus locality westward, where it separates into at least two lumpy to tabular beds. The lower submember contains several distinctive skeletal concentrations which persist across large parts of central and western New York. The base forms the lowest cephalopod-rich horizon in the Cherry Valley Member. Rare, black shale-filled cephalopod steinkerns in the submember at Seneca Stone quarry suggest that black shales of the Lincoln Park submember (Hurley Member) were deposited and then removed by downcutting (Figure 5). The upper portion of the submember contains a persistent concentration of small, articulated rhynchonellid brachiopods. Patches of *in situ* auloporid corals locally occupy the upper contacts of individual beds within the lower submember at many outcrops across the state.

The middle submember is well-developed at all Cherry Valley localities but the facies vary systematically across the outcrop belt. The basal contact is gradational with the lower submember in central New York but forms a sharp, erosional discontinuity at the westernmost localities. In east-central New York the middle submember consists of several beds of isolated skeletal packstone nodules and a few laterally coalesced, discontinuous nodular to lumpy beds within skeletal marlstone. This packstone-marlstone facies is replaced to the west by a packstone/grainstone facies with small, tightly interlocked nodules and few argillaceous/marly partings. The middle "nodular" submember contains few widespread skeletal marker horizons and cephalopods are extremely uncommon in all but the largest nodules. Patches of *in situ* and fragmented *Aulocystis* are particularly common in the lowest and highest nodular beds in many central New York localities. The contact with the upper submember is sharp and forms an irregular, erosional discontinuity at some west-central New York localities, which is interpreted to be an incipient hardground.

The upper submember is widely identifiable and contains the greatest cephalopod concentrations in the Cherry Valley Member, especially in west-central and western New York. The basal contact is sharp and erosional everywhere and may locally display small scour channels into the underlying nodular limestones. The fining-upward, 35 to 38 cm-thick submember in the Cherry Valley region contains a series of amalgamated, 1 to 4 cm-thick, tabular beds separated by erosion or omission discontinuities. A macrofossil hash at the base of the unit directly overlies the scour surface and contains abundant fragmented and complete cephalopod conchs, gastropods, and fragments of auloporid colonies and *Coleolus* tubes. This cephalopod-rich interval can be recognized at all localities farther to the west. In addition, most of the individual beds within the upper submember bedset contain cephalopods shells, but concentrations in east-central New York are low. The submember thins to the west and these cephalopod beds coalesce into two closely spaced, highly concentrated cephalopod- and bone-rich pavements observable at the upper contact at the Seneca Stone quarry (see previous bone bed description). Truncated cephalopods or "half-cephalopods" occur on both subtle internal bed boundaries and more obvious firmground/incipient hardground discontinuities. Although uncommon, evidence of encrusters and endolithic borings within the interiors of half-cephalopods indicates that shell destruction was synsedimentary and only modified by later pressure solution along bed contacts. Endolithic borings in the upper portion of many complete cephalopod shells also indicate biogenic destruction of the exterior of exposed shell surfaces. Alignment of orthocone cephalopods in the upper submember cephalopod pavements indicate

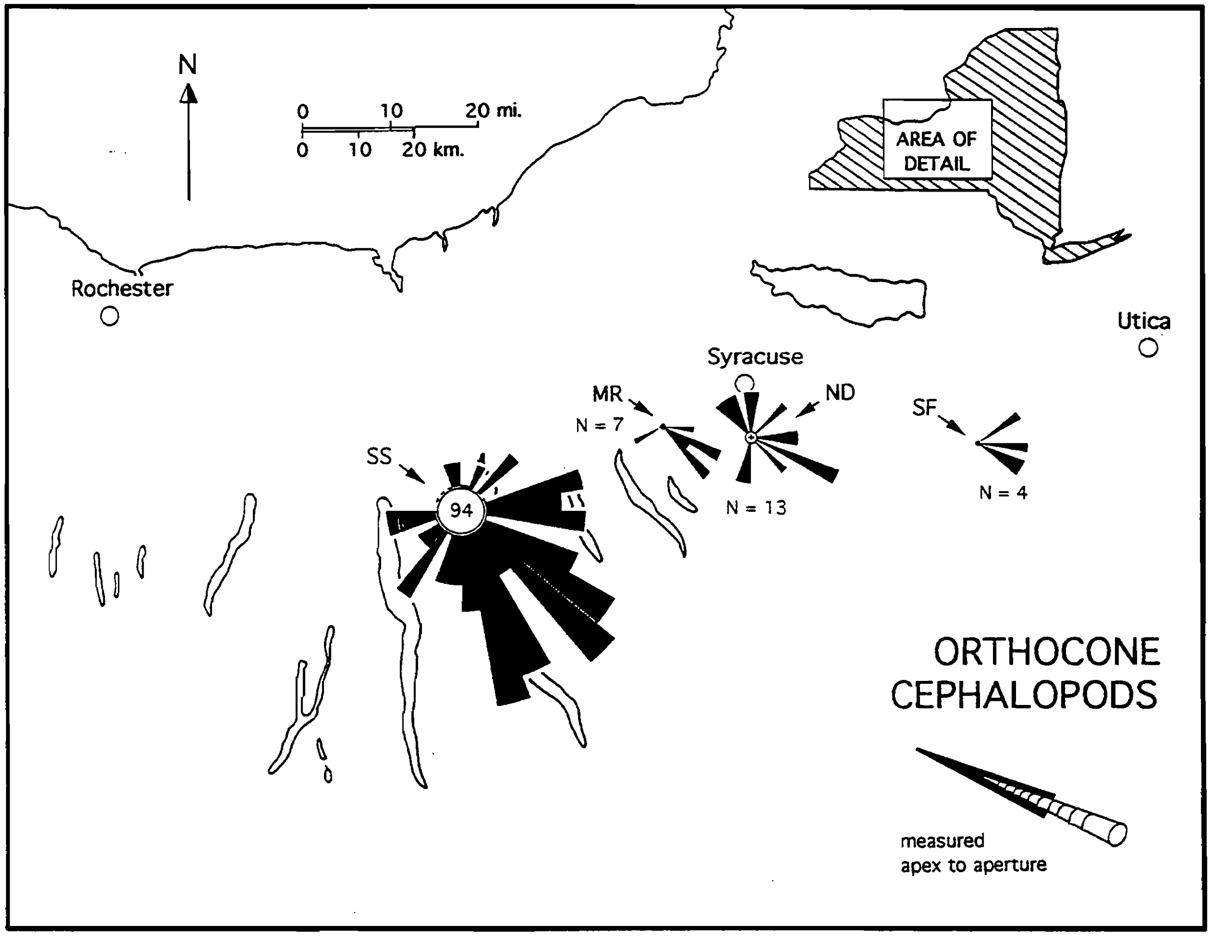


Figure 6. Paleocurrent map of aligned orthocone nautiloids in the cephalopod pavements at the top of the upper submember, Cherry Valley Member. Localities are as follows: SS=Seneca Stone Quarry, MR=Marcellus, ND=Nedrow roadcut, SF= Stockbridge Falls.

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unidirectional flows to the south or southeast, roughly perpendicular to paleodepositional contours (compare Figure 6 with Figure 1).

All three submembers have been identified in the westernmost known exposure of the Cherry Valley Member at the Honeoye Falls quarry (Stop 2), where the lowest beds appear to drape and thicken into erosional swales cut into the underlying Union Springs Formation (discussed below). The lower submember consists of an extremely coarse-grained, crinoid-styliolinid-fenestrate bryozoan grainstone facies which is overlain by the styliolinid-cephalopod packstones at this locality and at Flint Creek. Large-scale cross-stratification and the abundant crinoid content (33% of the total rock) in the coarse grainstones make the Honeoye Falls exposures very similar to the typical shallow-water "encrinites" in the Ludlowville and Moscow Formations. This is the only locality where articulated portions of crinoid skeletons and fenestrate bryozoan debris are commonly found in this limestone.

Systematic Lateral and Vertical Variation

The limestones and marlstones of the Cherry Valley Member represent two stacked, asymmetric, fining-upward intervals that are genetically linked. Although skeletal grain sizes do not vary much in east-central New York sections, a fining-upward trend is expressed by the progressive increase in clay content in the marlstones and by a decrease in nodule size. Farther west this fining-upward trend is expressed by decreasing skeletal grain sizes, whereas easternmost sections show a similar trend in both skeletal and quartz detritus. The lower fining-upward interval is represented by the lower and middle submembers combined. The upper fining-upward interval is represented by the upper submember and overlying dark shale-dominated strata. A sub-symmetrical appearance is given the entire bundle by the repetition: 1) of bedded cephalopod packstone facies in the lower and upper submembers, and 2) of fossiliferous nearly coalesced nodular beds at the base and top of the middle submember.

The nodular packstone-marlstone facies in east-central New York contains the least benthic skeletal component and the fewest hydrodynamic sedimentary structures of all Cherry Valley Member facies, and is interpreted to represent the more basinal facies within the Cherry Valley limestones. Much of the crinoid component in both nodular packstone-marlstones and the associated bedded styliolinid-cephalopod packstone facies in east-central New York is extremely fine-grained and may have been transported downslope. The nodular packstone-marlstone facies oversteps the bedded styliolinid-cephalopod packstone facies and the nodular packstone/grainstone facies in parts of central and eastern New York. All these facies ultimately overstep the crinoid-styliolinid-fenestrate grainstone facies (western New York) and the interbedded sandstone-packstone facies (eastern New York), both of which represent relatively shallow water deposition.

Discussion and Interpretation

The limited benthic fauna and the high organic content (1% organic carbon according to Brower and Nye, 1991) suggest deposition on a dysaerobic sea-floor, especially in more basinal areas. The westward increases in the abundance and size of crinoid ossicles in all facies may indicate increased sea-floor oxygenation upslope and/or closer proximity to skeletal production. In any case, the abundant crinoid ossicles (some partially articulated), articulated brachiopods, auloporids, fenestrates, and solitary cystiform rugosan corals present at Honeoye Falls quarry indicate local benthic production and aerobic sea-floor conditions during Cherry Valley deposition in western New York.

Facies of the Cherry Valley Member closely resemble many of the ancient, mixed pelagic- and benthic-derived skeletal limestones summarized by Tucker (1974), Franke and Walliser (1983), and Wendt and Aigner (1985). Although most commonly attributed to deep-water or open ocean settings, modern pelagic-rich sediments do occur in platform settings (Scholle and Kling, 1972). Sedimentary evidence indicates that many ancient "pelagic limestones" probably formed in shallow platforms in addition to deeper basinal settings. Such carbonates are interpreted to form as hiatal deposits which accumulated at extremely slow rates during periods of siliciclastic sediment starvation. The rate of taphonomic loss of skeletal hard parts in sea-floor environments often exceeds the rate of sedimentation (Cummins et al., 1986). It is unlikely that aragonitic cephalopod and gastropod shells would survive corrasion during long periods of sea-floor exposure associated with slow pelagic sedimentation, and yet both groups are well represented in the Cherry Valley Member. Sea-floor corrasion of aragonitic goniatites and nautiloids is evident in the Cherry Valley Member, especially at erosional discontinuities. The current alignment of orthocones and the coarse macrofossil hash/ reworked lag above scoured surfaces point to condensation by short-term modification and reworking of long-term accumulations, similar to bone bed formation in a lesser degree. Storm-generated mass mortality and rapid burial alternated with long periods of slow, gradual accumulation, during which partially exposed shells were corraded. Subsequent reworking led to further concentration and exposure of additional cephalopods to the destructive agents of the seafloor.

Although limestone facies of the Cherry Valley Member differ markedly from other younger Hamilton Group skeletal limestone bundles, the overstepping of basinal facies over basin-marginal facies within the two stacked asymmetrical, fining-upward intervals in the Cherry Valley Member bundle resembles the facies pattern within the Tichenor Member-Deep Run Member interval or the basin-marginal portion of the Centerfield Member. The main phase of deposition for all these limestones appears to have followed submarine erosion/downcutting associated with peak regression. The limestones themselves represent the initial phase of transgression and relative starvation, as suggested by Brett and Baird (1990). A consequence of the depositional circumstances is that the first transgressive facies have a "shallow-water" appearance and represent more hydrodynamic condensation than later transgressive facies. Bedded styliolinid-cephalopod packstones and grainstones at the base of fining-upward intervals represent erosional lags and initial sediment starvation. Nodular facies appear to represent deeper water deposition in the continued absence of significant siliciclastic input. Sediment starvation persisted longer in west-central and western New York than farther east, as evidenced by the replacement of the gradational limestone-black shale transition with a bone bed/hardground surface. Sediment transfer by downslope currents (compensation currents) on the western slope is supported by the aligned orthocone cephalopods and by large channels preserved at the Honeoye Falls quarry.

LOWER PART OF THE MARCELLUS SUBGROUP AT HONEOYE FALLS QUARRY

Strata of the Union Springs and the lower part of the Oatka Creek Formations are exposed in the southern face of the Honeoye Falls quarry south of Rochester (Stop 1). Quarrying in the late summer and fall of 1993 exposed a discontinuous cross-section approximately 300 m wide along the uppermost bench of the south quarry wall. The section, which is not tectonically deformed, shows a remarkable amount of variation along the exposure in: 1) paleorelief (at two to three levels); 2) thickness of units; and 3) presence-absence of the Chestnut Street submember (Hurley Member) and underlying black shales of the Bakoven Member.

Six separate subunits within the Union Springs and Oatka Creek Formations overlie the Onondaga Formation along the exposure (from the base up): 1) the Tioga F bed (of Way et al, 1986 terminology ; ="Tioga restricted" of Conkin and Conkin, 1979, 1984; Conkin, 1987; see Brett and Ver Straeten, this volume), with a thin, mm-scale black shale at its base; 2) a package of thin, dominantly styliolinid limestones that also include the Union Springs bone beds discussed elsewhere in this paper and several thin, mm-scale, tan clay beds that may represent K-bentonites (Tioga G beds? of Way et al., 1986); 3) black shales (Units 1-3 are included in the Bakoven Member); 4) Chestnut Street Submember (Hurley Member); 5) the Cherry Valley Member (at the base of the Oatka Creek Formation); and 6) overlying black shales of the revised Oatka Creek Formation.

A significant amount of paleorelief occurs below and above the Cherry Valley Member at the Honeoye Falls quarry as shown in Figure 7. A pre-Cherry Valley erosional surface, including channel-like features that cut down through the Hurley and Bakoven Members, is exposed along the uppermost south face of the quarry. Geometry of the erosion surface in places resembles shale-floored submarine channels identified by Brett and Baird (1990) in other basin margin shale strata of the Hamilton Group. At one position along the exposure, almost the entire Union Springs Formation is cut out; only 7 cm of the 15 cm-thick Tioga F (subunit 1) at the base of the Union Springs remain below the Cherry Valley Member at that position. Overlying units 2 and 3 of the Bakoven and the entire Hurley Member (unit 4) is missing at that position; small, rounded limestone clasts are found at the base of the Cherry

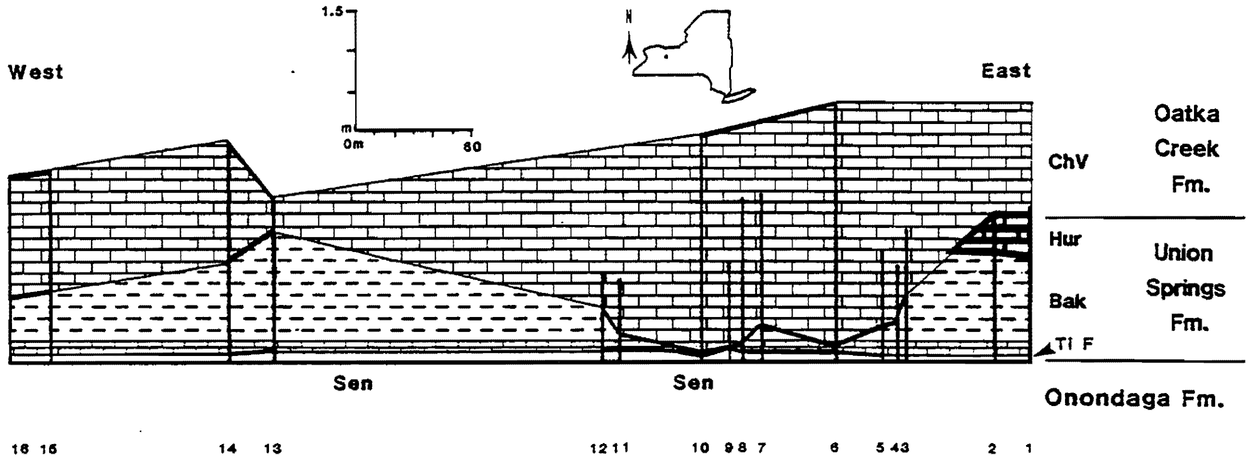

Figure 7. Cross-section of the Marcellus subgroup at Honeoye Falls quarry, south of Rochester (Stop 2 of fieldtrip). Note erosional cutout of strata at base and top of Cherry Valley Member. Bold lines mark known thickness; thinner lines represent projected thickness.

Valley. Elsewhere along the exposure the Bakoven Member ranges up to at least 1.7 m in thickness.

The Chestnut Street beds (Hurley Member) were reported by Griffing and Ver Straeten (1991) to be absent at the Honeoye Falls quarry. Recent excavations, however, show 42 cm of light, cream-colored limestone of the Chestnut Street submember in the eastern part of the exposure, approximately 1.4 m above the planar surface of the Onondaga Formation. The unit has not been found elsewhere along the exposure, except possibly in one area; the strata were apparently removed by pre-Cherry Valley erosion. Another possible surface of paleorelief may occur below the Hurley Member, but the small amount of exposure of the unit makes it very difficult to confirm this.

The Cherry Valley Member varies widely in thickness along the outcrop from as little as 40 cm to as thick as 3.15 m. Erosional downcutting and incorporation of winnowed lags into the basal Cherry Valley Member is evidenced by: 1) abundant black shale-filled styliolinids, 2) angular to rounded clasts of concretions and Chestnut Street packstones, and 3) abraded fish bone plates and fragmented cephalopod shells (including *Cabrieroceras*). As shown in Figure 7, thicker and thinner parts of the member do not necessarily correspond to areas of greater and lesser relief below the Cherry Valley because the top of the member, below overlying shales at the base of the Oatka Creek Formation, also shows variation in paleotopography along the outcrop. This contact across west-central to western New York has been previously discussed by Baird and Brett (1986), who associated relief on the top of the Cherry Valley Member with corrasional removal of carbonate strata during a period of transgression-induced sediment starvation in the region.

DISCUSSION

TECTONIC HISTORY, LATE EARLY TO EARLY MIDDLE DEVONIAN

During the late Early to early Middle Devonian the Northern Appalachian Basin was a dynamic system, marked by major changes in basinal geometry associated with several episodes of subsidence and "rebound." These episodes are associated with two early tectonically-active phases of the Acadian Orogeny in eastern North America (Tectophases 1 and 2 of Ettensohn, 1985a) and an intervening period of tectonic quiescence.

Different mathematical and computer-generated models have been proposed in recent years to describe foreland basin dynamics and stratigraphy associated with orogenic episodes (e.g., Quinlan and Beaumont, 1984; Beaumont et al., 1988; Jordan and Flemings, 1991; Sinclair et al., 1991). The basic premise of these models states that loading of the lithosphere during episodes of tectonic thrusting leads to stress-induced subsidence of a proximal foreland basin and gentle uplift due to relaxation on a cratonward peripheral bulge. Subsequent periods of tectonic quiescence are marked by relaxation and uplift of the foreland combined with subsidence of the peripheral bulge. The timing of subsidence and uplift differs in the models dependent upon an elastic or visco-elastic flexural response of the lithosphere.

Other recent work on foreland basin dynamics has focused on the sedimentary record of the basin. For example, Plint et al. (1993), in studies of Upper Cretaceous strata in the Alberta foreland basin, note depositional patterns that include surfaces of erosive beveling at least 300 km cratonward of the present day Sevier deformation front. They interpret these regional truncations of strata to reflect forebulge uplift and erosion associated either with episodic loading/tectonic rejuvenation in an adjacent fold-thrust belt or continuous loading of lithosphere of laterally varying flexural rigidity.

A model for the evolution of the Devonian Acadian Orogeny in eastern North America was presented by Ettensohn (1985a). Based on the stratigraphic record of the Appalachian Basin, he notes three to four phases of active tectonism in the late Early Devonian to Mississippian associated with oblique convergence of the eastern margin of North America and a landmass termed Avalon. Each "tectophase" is composed of a progression from stages of active tectonism to quiescence, recorded in the basin fill by a succession of clastic- to carbonate-dominated sedimentation.

Upper Lower Devonian, Emsian-age shales and minor sandstones to shaly carbonates of the Esopus, Carlisle Center, and Schoharie Formations (Tristates Group) show a distinctive, eastward-thickening wedge-like geometry that ranges from 0-300 m in thickness across west central to southeastern New York, respectively. These patterns are associated with an actively subsiding foreland basin adjacent to a probable active fold and thrust belt in the New England region during Ettensohn's (1985a) Tectophase I of the Acadian Orogeny. The absence of these rocks across west-central to western New York (Seneca Stone quarry to Buffalo) is the result of active uplift of a peripheral bulge in that region, possibly during later Emsian time (Schoharie Formation). Interestingly, the initial clastic-dominated part of the succession (shales to fine sandstones of the Esopus Formation in New York), which overlies widespread quartz arenites of the Oriskany Sandstone (see Boucot and Johnson, 1967), is restricted to the eastern margin of North America at that time (eastern most parts of the Appalachian foreland basin and deeper water facies in western New England; Rehmer, 1976). The Oriskany-Esopus transition at the base of the clastic wedge is marked by the occurrence of bentonite-rich strata (Sprout Brook Bentonites of Ver Straeten, 1992, ms. submitted; Ver Straeten et al., 1993)

Geometry of the basin was dramatically altered at the beginning of the Eifelian Stage (Onondaga Limestone). The Emsian-age trough of the basin in eastern New York was replaced by a less dramatically subsiding basin center in central New York, near the previous position of the uplifted peripheral bulge of Emsian time (Brett and Ver Straeten, this volume). Relatively deeper water environments in central New York, flanked on either side by relatively shallow carbonate ramps in eastern New York and western New York-Ontario, characterize the basin during deposition of the Onondaga Limestone. Rocks of the Onondaga Formation across New York range from approximately 20 to 60 m in thickness (Rickard, 1989; see Figure 1 in Brett and Ver Straeten, this volume), but nevertheless display a distinctly more tabular geometry than more clastic-dominated rocks of the underlying Esopus-Schoharie interval. Also note that carbonate-dominated, Onondaga-equivalent strata are very widespread across eastern North America at that time (Figure 2 of Brett and Ver Straeten, this volume). An apparent minor amount of volcanism during early Onondaga time (Brett and Ver Straeten, this volume) greatly increased through deposition of the higher part of the Onondaga, as indicated by the presence of as many as ten ash beds associated with the Tioga Bentonites cluster (Brett and Ver Straeten, this volume).

This increase in volcanism was accompanied by apparent subsidence of the eastern margin of the Northern Appalachian Basin that began in eastern New York during late Onondaga (Seneca Member) time. Initial sediment-starved conditions on the subsiding basin floor are indicated by a relatively minor, westward-younging, sediment-starved submarine unconformity at the Onondaga-Marcellus contact across the state.

Clastics of the overlying Union Springs Formation comprise another eastwardly, forelandward-thickening wedge of sedimentary basin-fill that characterizes the Middle Devonian Hamilton Group as a whole. More basinal environments occur in the subsiding trough of the foreland basin in eastern New York. Multiple erosion surfaces through the lower part of the Marcellus subgroup in west-central New York (i.e., Honeoye Falls quarry) give rise to a complete removal of Union Springs and lower Oatka Creek (Cherry Valley Member) strata west of the Genesee River. Strata equivalent to these rocks reappear to the west in the Delaware Limestone of central Ohio and other units as far cratonward as the Spillville Formation of Iowa (Day and Koch, 1994; Koch, 1978); their absence across the intervening region may indicate uplift on a peripheral bulge in western New York to southwestern Ontario and eastern Ohio at that time. The more westward position of this Middle Devonian peripheral bulge, in contrast to that of the late Early Devonian (Tristates Group) is probably associated with the westward migration of the foreland basin through time. Superimposed on this tectonic subsidence trend are two transgressive-regressive cycles that comprise the Union Springs and Oatka Creek-Mount Marion succession of the Marcellus subgroup (T-R cycles Id and Ie of Johnson et al., 1985).

To summarize, two major transgressions mark the upper Lower and lower Middle Devonian of the Appalachian foreland basin. Note, however, that these major transitions from relatively shallow marine orthoquartzites and carbonates to basinal black shales (Oriskany Sandstone into Esopus Shale and Onondaga Limestone into Marcellus Shale) are not due simply to a rise in eustatic sea level. In each case, the regressive, shallow marine orthoquartzite-carbonate suite comprises a relatively tabular body of rock that occurs widely across the Appalachian Basin and onto the eastern North American craton. In contrast, the deep basinal, major transgressive packages of siliciclastics consist of eastwardly-thickening wedge-shaped bodies that are concentrated along the eastern margin of the Devonian eastern interior seaway. These patterns point to a combined tectonic and eustatic control on the Appalachian Basin and the adjacent craton during two early tectophases of the Acadian Orogeny.

THE LOWER PART OF THE MARCELLUS SUBGROUP IN PENNSYLVANIA

Recent work in Pennsylvania shows that much of the same stratigraphic framework of the lower part of the Marcellus subgroup in New York can be recognized across Pennsylvania. In Pennsylvania, subdivisions of the Hamilton Group of New York are not generally recognized, and the usage of Marcellus Formation extends to all lower, black shale-dominated Hamilton-equivalent strata above limestones of the Onondaga-equivalent Selinsgrove Member of the Needmore Formation and the coeval Buttermilk Falls Formation (Ver Straeten and Brett, 1994; Brett and Ver Straeten, this volume; see Figure 8).

Black shales equivalent to the Bakoven Member of New York generally overlie the Selinsgrove Limestone in central and southern Pennsylvania and the Buttermilk Falls Limestone in eastern Pennsylvania. In central Pennsylvania, black shale facies may initially occur lower in the section within the Selinsgrove Member at more basinward

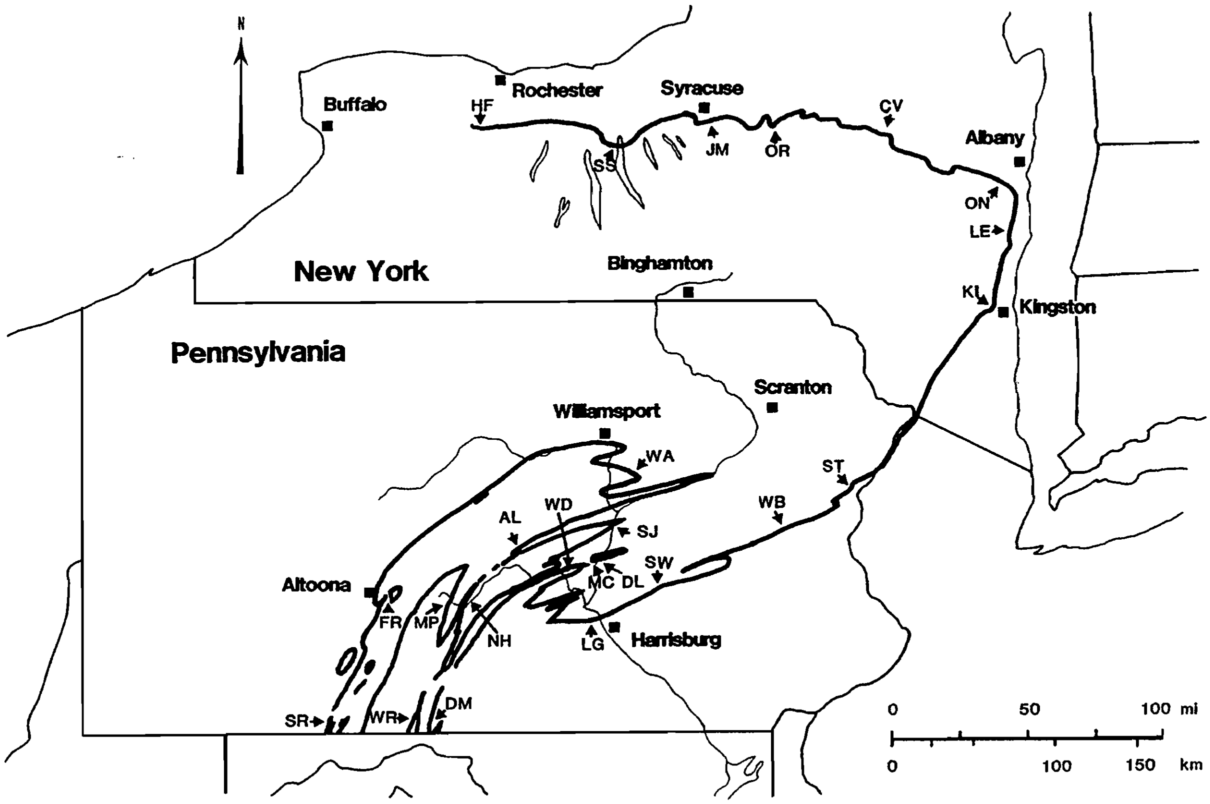

Figure 8. Outcrop map of the lower part of the Marcellus subgroup in New York and Pennsylvania (after Inners, 1975, and Berg, et al., 1980). Key localities are as follows: NEW YORK -- HF=Honeoye Falls quarry, SS=Seneca Stone quarry, JM=Jamesville quarry, OR=Oriskany Falls quarry, CV=Cherry Valley, ON= Onesquethaw Creek, LE=Leeds, KI=Kingston; PENNSYLVANIA -- ST=Stroudsburg, WB=West Bowmans, SW=Swatara Gap, LG=Lambs Gap, DL=Dalmatia, MC=Mahantango Creek, WD=Wardsville, SJ=Selinsgrove Junction, WA=Washingtonville, AL=Alfarata, NH=Newton Hamilton, MP=Mapleton, FR=Frankstown, DM=Dickeys Mountain, WR=Warfordsburg, SR=Stringtown.

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localities (e.g., Selinsgrove Junction and Frankstown, PA; see Figure 8 of Brett and Ver Straeten, this volume).

Calcareous shale and limestone with subordinate amounts of sandstone toward the middle of the Marcellus Formation in Pennsylvania, Maryland, and the Virginias are placed within the Purcell Member (Cate, 1963). The Purcell is generally reported to be equivalent to the Cherry Valley Member of New York (Dennison et al., 1972; Nuelle and Shelton, 1986; Way, 1993; however, see deWitt et al., 1993); detailed comparison between these strata and those of the New York section shows a more complex relationship, as discussed below.

Buff-weathering calcareous shales to sandstones similar to and equivalent, at least in part, to the Stony Hollow Member (Union Springs Formation) are recognizable in eastern Pennsylvania, notably in the Stroudsburg region, and locally through central Pennsylvania, as at Selinsgrove Junction along the Susquehanna River. The lower 11.5 m of strata formerly termed the "upper Selinsgrove Limestone" that are exposed at Selinsgrove Junction are directly equivalent to the Stony Hollow of eastern New York, and display the same set of lithofacies. Overlying strata at Selinsgrove Junction represent the Hurley and

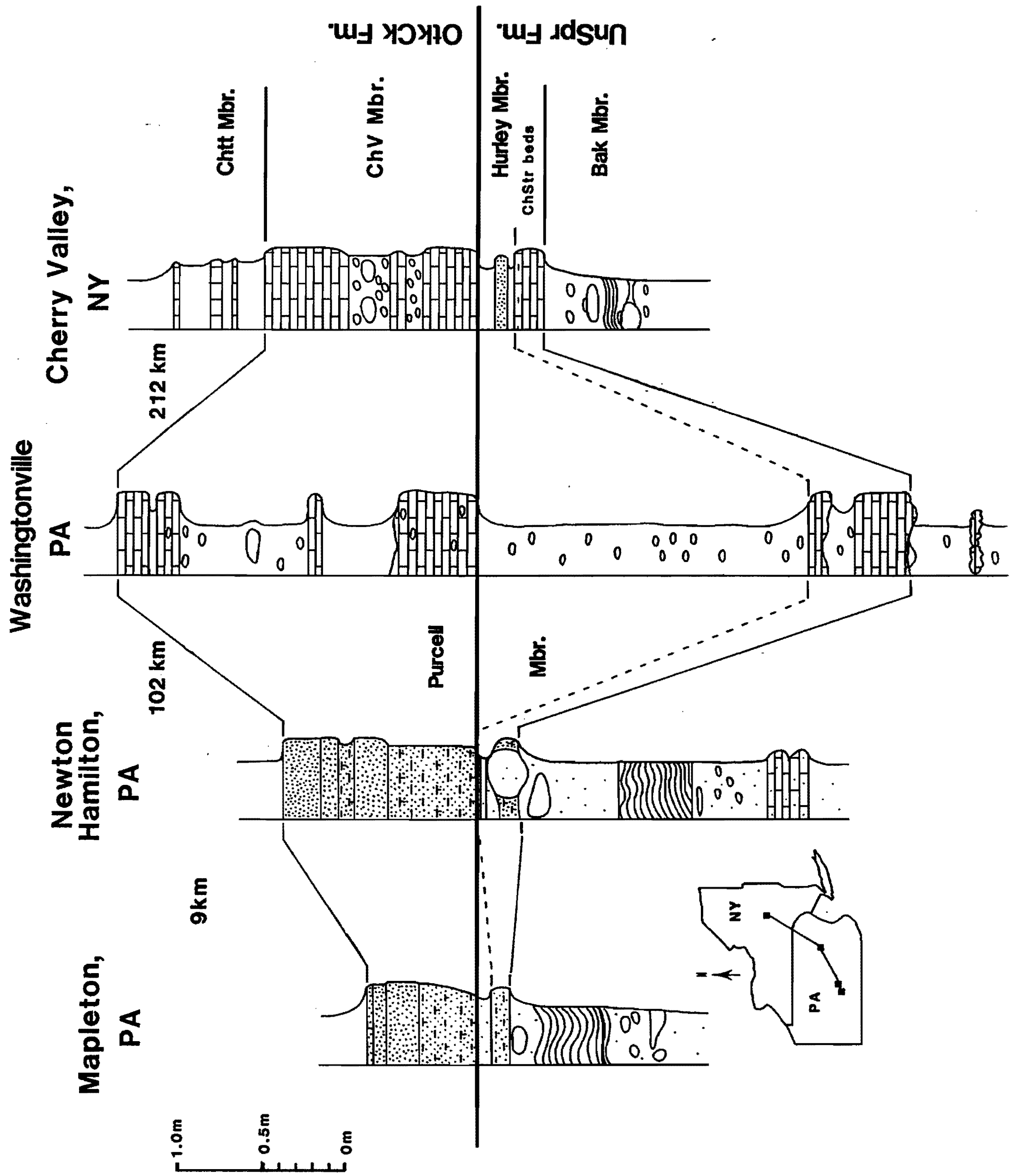

Figure 9. Cross-secton of the Bakoven, Hurley and Cherry Valley Members at Cherry Valley, New York and equivalent strata of the Purcell Member in Pennsylvania (thicknesses of strata at Washingtonville, PA from Way, 1993).

Cherry Valley Members of New York, and include both the proetid trilobite-rich carbonates of the Chestnut Street submember (below) and the classic cephalopod fauna of the Cherry Valley Member (above). In other areas, these Stony Hollow-equivalent strata may be represented by black shales or by dark, pyritic, silty shales to sandstones as can be seen at Newton Hamilton.

Strata correlative with the Hurley Member (Union Springs Formation) are recognizable across most of the Pennsylvania outcrop, most notably the proetid trilobite-bearing Chestnut Street submember (Figure 9). These beds can presently be correlated all across eastern to south-central Pennsylvania to the region along the Maryland border, where they have not as yet been identified. Recognition of the Hurley Member also permits positive identification of strata coeval with the Cherry Valley Member of New York (Oatka Creek and Mount Marion Formations of New York; see Figure 9). These Cherry Valley equivalents may, as in New York, be represented by limestone- or sandstone-dominated facies. The classic cephalopod fauna, dominated by *Agoniatites vanuxemi* and *Striacoceras typum*, is commonly found in more carbonate-dominated exposures.

Nodular to bedded barite, and a lesser component of nodular pyrite, is widely reported from the Purcell Member across the southern and central parts of the Appalachian Basin (Way and Smith, 1983; Nuelle and Shelton, 1986; Way, 1993). These deposits generally occur in the same position as similar, if more commonly pyritic, nodules that are found in the upper part of the Union Springs Formation in New York. Barite nodules in Pennsylvania outcrops are commonly golf ball-sized, although barite may also fill syneresis cracks in large limestone concretions, or even ammonoid cephalopods.

Across central Pennsylvania a previously unreported, widely recognizable K-bentonite bed, generally 3-6 cm-thick, occurs in the middle of the lower part of the Marcellus Formation. This bed often marks a transition from underlying black shales to slightly coarser silty mudstones to sandstones above. The bentonite typically appears as a honey-tan to light gray, soapy-feeling clay bed and forms a prominent, continuous recession along the outcrop. Bleached biotite crystals are visible in less weathered samples; locally the bed contains pyritic concretions up to 5 cm in diameter. This bed has not as yet been recognized in New York or eastern Pennsylvania, but has been found at a minimum of five key outcrops across central Pennsylvania.

This thin bentonite bed may be correlatable into the Harrisburg region of central Pennsylvania, where it commonly overlies typical black shale-dominated strata above the Selinsgrove Limestone and the Tioga Bentonites. Above the bentonite, however, thick-bedded to massive quartz-rich sandstones occur that are generally assigned to the Turkey Ridge Member of the Mahantango Formation (Faill et al., 1978). The Turkey Ridge commonly appears massive and undifferentiated , although in some localities the sandstones appear predominantly cross-bedded. The Turkey Ridge Member generally yields no fossils, which make it difficult to correlate with other strata. A complete section (28 m-thick) of the sandstones is exposed along Mahantango Creek, 40 km north of Harrisburg, west of the Susquehanna River. Roughly 5-10 m below the top of the Turkey Ridge at Mahantango Creek is an interval of sandstone nodules with a barium-rich cement matrix (F. Teichmann, Univ. of Rochester, pers. commun.). These seem to correlate with the barite and pyrite nodule-rich interval previously noted in the Purcell Member from Pennsylvania to the Virginias and in the upper part of the Union Springs Formation in New York. The Turkey Ridge at Mahantango Creek and other nearby localities is overlain by black shales. At least some sandstones assigned to the Turkey Ridge Member in central Pennsylvania, therefore, are equivalent to strata of the Stony Hollow and Hurley Members of the Union Springs Formation and the Cherry Valley Member of the Oatka Creek and Mount Marion Formations of New York.

At Mahantango Creek additional similar sandstones also occur below the thin K-bentonite layer, interbedded with black shales above the Selinsgrove Member. These lower strata, again, are for the most part equivalent to the black shales of the Bakoven Member in eastern

New York, where they occur overlain by the Stony Hollow Member. Across the Susquehanna River from the Mahantango Creek locality, near Dalmatia, the sandstones occur as low as the upper part of the Selinsgrove Member, in strata equivalent to the Seneca Member of the Onondaga Limestone of New York.

Along Interstate 81 at Swatara Gap, 35 km northeast of Harrisburg, 23 m of black shale above the Selinsgrove Limestone equivalent to the Bakoven Member of New York are overlain by a very thick section of sandstone. Plant root traces appear in the middle to upper part of the section. This thick sandstone body may be continuous from lower Marcellus strata (Turkey Ridge) upward into the post-Marcellus Montebello Sandstone within the section exposed; potentially, however, Union Springs-equivalent strata here may even be represented by non-marine facies.

Therefore, we believe we may now be able to widely correlate the New York members and identify a full facies spectrum throughout the lower part of the Marcellus subgroup across the northern and central Appalachian Basin. Much work remains to be done, including refinement of our work in New York and Pennsylvania and an attempt to correlate this new stratigraphic scheme into the southern part of the basin in Maryland and the Virginias.

SUMMARY

Lower strata of the Marcellus Shale comprise a fifth major cycle within the Middle Devonian Hamilton Group of New York State. In this light we informally raise the Marcellus to subgroup status and recognize three formations. They are: a) the Union Springs Formation, comprising strata from the top of the Onondaga Limestone to the base of the Cherry Valley Member all across New York; b) the Oatka Creek Formation, which consists of black shale-dominated upper Marcellus strata in central to western New York (from the base of the Cherry Valley Member to the base of the Stafford Member of the Skaneateles Formation); and c) The Mount Marion Formation, the lateral, progradational shale to sandstone equivalent of the Oatka Creek Formation in eastern New York.

In contrast with the southeasterly thickening wedge of siliciclastics in the lower part of the Marcellus subgroup in eastern New York, this interval is represented in central to western New York by a more tabular, westward thinning body of black shales and skeletal Limestones. Detailed study of the fabric of the limestones of the Bakoven and Hurley Members (Union Springs Formation) and the Cherry Valley Member (base of Oatka Creek and Mount Marion Formations) indicate conditions of long term accumulation of skeletal carbonate modified by short term events. Condensation and submarine erosion/corrasion increased westward across New York associated with a relative decrease in deposition of fine-grained siliciclastics on the distal, cratonic margin of the Middle Devonian Appalachian foreland basin.

Lower strata of the Marcellus subgroup are widely recognizable across the Northern and Central Appalachian Basin. Carbonate-rich strata of the Purcell Member of the Marcellus Formation and underlying black shales in Pennsylvania are shown to be equivalent to the Bakoven, Stony Hollow, and Hurley Members of the Union Springs Formation and to the Cherry Valley Member of the Oatka Creek and Mount Marion Formations of New York State. Furthermore, proximal marine sandstones (Turkey Ridge Member) near to the southeastern margin of the basin appear to be equivalent to more basinward rocks of the lower part of the Marcellus subgroup.

The base of the Marcellus subgroup marks a major reorganization of the Northern Appalachian Basin from a broad, gently sloping carbonate ramp to a rapidly-subsiding foreland basin during a second active tectophase of the Devonian Acadian Orogeny. Subsidence of an eastern trough adjacent to tectonic highlands was accompanied by gentle uplift of a cratonward peripheral bulge in western New York and adjacent parts of southern Ontario. Tectonic flexure of the basin was accompanied by two separate transgressive-regressive cycles that divide the Marcellus subgroup into two distinctive successions, the Union Springs Formation and the coeval Oatka Creek and Mount Marion Formations. Each of these cycles was accompanied by a major faunal immigration into the Appalachian Basin from the Old World Realm. The second fauna, which first appears in the lower part of the Oatka Creek-Mount Marion succession, became well established and thrived throughout the remainder of the Middle Devonian Hamilton Group.

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FIELD TRIP LOG

C.E. Brett, C.A. Ver Straeten, and D.H. Griffing

NOTE: This roadlog accompanies papers by Brett and Ver Straeten and Ver Straeten et al. (this volume) on the revised stratigraphy and depositional history of the Onondaga and Marcellus Formations. In these accompanying papers, we document the details of sequence and event stratigraphy within the Eifelian Onondaga and Marcellus Formations in western and west-central New York State, and provide an interpretation of the sequence of physical and biotic events that these rocks document.

The early Middle Devonian Eifelian age was a time of significant reorganization within the Appalachian Basin both in terms of its tectonic evolution and the evolutionary ecology of its fossil assemblages. This field trip surveys the the revised stratigraphic and facies relationships of early Middle Devonian Onondaga Limestone and the highly disparate Marcellus black shale facies in western and west central New York State, roughly from LeRoy to Seneca Falls, New York.

Detailed stratigraphic correlation of the Onondaga and Marcellus Formations has recently been undertaken on a regional scale. The rather precise event and cyclic and stratigraphic correlations provide a far more refined framework than does biostratigraphy, and in turn permit rather precise resolution of the timing and dynamics of tectonic and sea level events within the Appalachian foreland basin. This detailed stratigraphic framework is also a key to understanding the dynamics of biotic change during the early Middle Devonian interval.

For a mapview of fieldtrip localities in west-central to western New York State, refer to Figure 4 of Brett and Ver Straeten, this volume. Stratigraphic sections of Stops 1, 2, 5, and 6 are shown in Figures 5a and 5b in Brett and Ver Straeten, this volume.

0.0	0.0	Leave parking lot south of Hutchison Hall; <u>turn left (west)</u> out of lot.
0.05	0.05	Junction with Wilson Boulevard; <u>turn left (south)</u> and get into right lane.
0.1	0.05	Junction with Elmwood Avenue; <u>turn right (west)</u>.
0.15	0.05	Cross over Genesee River.
0.5	0.35	<u>Bear slightly to left</u> onto Scottsville Road (end of Elmwood Avenue).
0.9	0.4	Cross over Barge Canal; <u>Immediately over bridge bear right</u> onto I-390 North entrance ramp.
1.35	0.45	Merge onto I-390 North.
3.3	1.95	Cuts in upper part of highly fossiliferous Penfield Dolostone (Middle Silurian Lockport Group); Gates quarry in Penfield and Eramosa (Oak Orchard) Formations to left.
4.0	0.7	<u>Bear left</u> onto I-490 West entrance ramp.
4.4	0.4	Merge onto I-490 West.
4.7	0.3	Cuts on right and in median in lower sandy dolostone of Penfield Formation.
5.5	0.8	Exit for NY Rte. 531; continue ahead on I-490. Outcrops on both sides of highway in upper Penfield and basal Eramosa Formations.
17.2	11.7	Cross over Black Creek at Churchville exit; continue ahead on I-490.
20.1	2.9	Leave Monroe County, enter Genesee County.
23.15	3.05	Enter Town of LeRoy.
23.6	0.45	<u>Exit</u> from I-490 West to NY Rte. 19 and LeRoy.
23.9	0.3	Junction with Rte. 19; <u>turn right</u> onto Rte. 19 southbound.
24.4	0.5	Cross over NY State Thruway (I-90).
24.6	0.2	North Road, Frost Ridge Campground to left.
25.4	0.8	Begin ascent of Onondaga Escarpment.

25.6	0.2	Camillus Shale (Upper Silurian Salina Group) cuts on Rte. 19 and adjacent road to right.
26.2	0.6	Railroad underpass; walk down track to left (east) to Buttermilk Falls of Oatka Creek, a spectacular falls capped by Clarence facies of the Edgecliff Member of the Middle Devonian Onondaga Limestone.
26.3	0.1	Randall Road, route to the LeRoy Bioherm Quarry, site of one of Sunday's trips (see Wolosz, this volume).
27.1	0.8	Enter village of LeRoy.
28.0	0.9	Mill Street on left; route to Oatka Creek exposure of Onondaga-Marcellus contact. Continue ahead on Rte. 19.
28.3	0.3	Bacon Street on left; route to Oatka Creek exposure of Oatka Creek Formation (type section) and contact with overlying Stafford Limestone of Skaneateles Formation. Continue ahead on Rte. 19.
28.4	0.1	Junction with NY Rte. 5; <u>turn right (west)</u>.
28.9	0.5	Railroad underpass.
29.3	0.4	Bethany-LeRoy Road on left; road to classic Middle Devonian Hamilton Group localities.
29.4	0.1	Leave village of LeRoy.
32.1	2.7	Enter village of Stafford.
32.65	0.55	Junction with NY Rte. 237; continue west on Rte. 5.
32.8	0.15	Cross Black Creek.
33.1	0.3	Leave village of Stafford.
33.4	0.3	Berms of quarry visible on left.
33.5	0.1	Stone house on left.
33.75	0.25	Genesee-LeRoy Stone Corp. quarry entrance; <u>turn left (south)</u> into quarry.
34.2	0.45	Check in at quarry office; from office <u>proceed left (east)</u> on gravel road.
34.25	0.05	<u>Turn left (north)</u> and descend quarry ramp road.
34.3	0.05	Continue straight ahead on quarry road.
34.6	0.3	Park cars in northwest corner of quarry and walk down ramp to lower part of quarry.

<u>STOP 1. GENESEE-LEROY STONE COMPANY QUARRY, STAFFORD.</u>

This relatively large quarry provides an excellent overview of the middle and upper parts of the Onondaga Limestone (Clarence facies of the Edgecliff Member, Nedrow, Moorehouse and Seneca Members; see Figure 5 of Brett and Ver Straeten, this volume). This stop serves as a reference section for western facies of the Onondaga Formation.

The quarry operates in five main lifts. The lowest two are commonly below water level but the second is exposed during drier seasons. Walls of the sump pit below the second lift exhibit heavily cherty Clarence facies of the Edgecliff Member. The overlying platform displays numerous large, spheroidal, chert-replaced heads of *Favosites* which are overlain by a thin shaly parting. This contact is thought to represent the base of the Nedrow Member at Stafford. It is overlain by an approximately four m-thick interval with distinctly cyclic alternations of dark gray chert-rich micritic limestone and pale pinkish-gray crinoidal limestone with light gray chert and abundant rugose corals.

The next platform is floored by one of the most distinctive marker beds in the western outcrops of the Onondaga; a greenish-gray argillaceous limestone with skeletal hash of crinoids, bryozoans, and brachiopods in addition to abundant and diverse tabulate and rugose corals. Because of its shaly nature, well-preserved fossils are easily obtained from this bed. It is interpreted as a condensed, time-rich fossil bedwithin the Nedrow Member.

The base of the third lift (5 m high) consists of 1.6 m of dark gray, very cherty, micritic limestone that weathers shaly toward its top and is capped by a 5 cm-thick shale bed. This shale and overlying less cherty, coral-rich, light weathering limestone are also readily correlatable in the Stafford-LeRoy area. The interval is interpreted here as the upper portion of the Nedrow Member. The upper part of the cherty limestone and the 5 cm-thick shale are thought to correlate with the lower, thick black shale bed of the upper Nedrow of central New York (Stops 5 & 6). The overlying light-weathering, slightly cherty limestone and succeeding thin shale appear to represent the white limestone and *Schizophoria* shale bed of the uppermost Nedrow Member as defined herein.

The third lift is capped by another persistent crinoidal hash and coral-rich shaly bed that forms a major platform in the quarry. An old road in the northwest corner of the quarry provides access up to some of the higher beds of the Moorehouse Member. Much of the lower parts of the wall consists of sparsely fossiliferous cherty micritic limestone with dark brownish gray chert. A particularly distinctive bed occurs approximately 1 m below the fourth platform of the quarry. This is a biostrome of thicket-type (fasciculate) rugose corals (*Synaptophyllum* and/or *Eridophyllum*). The light-colored corals contrast with surrounding large, dark gray, ellipsoid chert nodules. Slightly higher on the fourth platform, a variety of solitary and colonial rugosans and large favositids are exposed in the quarry floor. This coral-rich interval and immediately overlying crinoidal pack- and grainstones appear to represent the first of two major cycles in the Moorehouse. Slightly above the fourth platform is a distinctly non-cherty, slightly argillaceous and medium gray lime mudstone unit approximately 1.6 m-thick. This unit displays abundant spreiten of *Zoophycos* (which appear on weathered joint surfaces as gently curved laminae) and thin hash beds of brachiopods. This non-cherty interval weathers with shaly fissility and correlates to a middle Moorehouse shaly, "false Nedrow" interval that occurs widely throughout the Appalachian Basin (see Brett and Ver Straeten, above). It is sharply overlain by coarse-weathering, slightly cherty (pale gray chert) crinoidal pack- and grainstones of the upper Moorehouse, including strata that feature abundant large *Paraspirifer* brachiopods. This upper 3 m-thick interval comprises the *Paraspirifer acuminatus* zone recognized by previous workers. It represents the second major shoaling cycle of the Moorehouse. This crinoidal-rich facies yields excellent specimens of the crinoids *Arachnocrinus* and *Schultzicrinus*. Two prominent, clay-rich partings in the upper part of the Moorehouse (above the Nedrow-like shaly unit) represent the First and Second Cheektowaga Bentonites (Conkin and Conkin, 1979, 1984; Conkin, 1987; 2.7 and 0.5 m below the base of the Seneca Member, respectively).

Highest beds in the Stafford Quarry are in the Seneca Member of the Onondaga Formation. A distinctive, thick (ca. 20 cm-thick) clay bed that is traceable around the quarry represents the Tioga B or Onondaga Indian Nation Ash Bed that marks the base of the Seneca. The member is comprised of crinoidal wacke- to packstones with a relatively minor number of horizons of large, brownish-weathering chert nodules. The upper beds are exceptionally rich in brachiopods which tend to weather pale-cream to pinkish in color; *Leptaena*, *Megastrophia*, *Atrypa*, and *Megakozlowskiella* are all abundant here.

No upper contact of the Seneca Member has been observed here. However, dredge piles of black shale south of the quarry suggest that the contact with the overlying Marcellus subgroup lies a short distance above the upper platform of the quarry at Stafford.

To leave quarry, turn around and retrace route;

34.95	0.35	<u>turn right (west)</u> onto small quarry road.
35.0	0.05	Check out at quarry office on left; then <u>turn right (north)</u> onto main entrance road.
35.4	0.4	<u>Turn right (east)</u> onto NY Rte. 5 (return to LeRoy).
36.05	0.65	Reenter village of Stafford.
37.05	1.0	Leave village of Stafford.

40.0	2.95	Reenter village of LeRoy.
40.8	0.8	Intersection with NY Rte. 19; continue ahead (east) on Rte. 5
40.85	0.05	Turn right (south) into "McDonalds"; **10 minute rest stop.**
40.9	0.05	Leave McDonalds; turn right (east) onto Rte. 5 and continue eastbound.
41.1	0.2	Cross over Oatka Creek; falls over Stafford Limestone immediately north of bridge.
41.9	0.8	Leave village of LeRoy.
42.85	0.95	Perry Road on left; route to 5 or 6 large quarries in the Onondaga Limestone along Perry Road and adjacent Gulf Road.
44.1	1.25	Enter village of Limerock; site of famous stone fences where paleontologists since James Hall have collected crinoids and other fauna of the Onondaga Formation.
44.5	0.4	Church Road; eastward access to Onondaga quarries along Gulf Road to north.
45.2	0.7	Leave Genesee County, enter Livingston County.
47.2	2.0	Enter village of Caledonia.
47.85	0.65	Junction with NY Rte. 36 South; Rte. 36 joins Rte. 5.
48.3	0.45	Center of Caledonia; bear right on Rte. 5 (Rte. 36 turns left).
49.0	0.7	Leave village of Caledonia.
50.3	1.3	Quarry Road on right; old quarries in Onondaga down the road.
53.7	3.4	Descend onto floodplain of the Genesee River.
54.4	0.7	Junction with US Rte. 20; Rtes. 5 and 20 merge; continue straight (east) on combined Rtes. 5 and 20.
54.8	0.4	Bridge over Genesee River.
54.9	0.1	Enter village of Avon.
55.3	0.4	Junction with Wadsworth Avenue; south to classic Middle Devonian Hamilton Group outcrops. Continue straight on Rtes. 5 and 20.
55.5	0.2	Proceed around circle in center of Avon; bear right halfway around circle and continue east on Rtes. 5 and 20.
56.4	0.9	Leave village of Avon.
57.1	0.7	Enter village of East Avon.
57.7	0.6	Junction with NY Rte. 15; turn left (north) onto Rte. 15.
59.95	2.25	Leave Livingston County, enter Monroe County.
60.45	0.5	Intersection with Honeoye Falls #6 Road; turn right (east) onto Honeoye Falls #6 Road.
60.6	0.15	Cross over I-390; view of three drumlins to the left (north).
62.3	1.7	Five Points Road on left; continue straight ahead.
63.2	0.9	Cross Works Road; small outcrop of Onondaga Limestone on right along Works Road.
63.9	0.7	Entrance to General Crushed Stone Corporation, Honeoye Falls quarry, at town of Mendon line; turn right into entrance and bear right.

STOP 2. GENERAL CRUSHED STONE QUARRY, HONEOYE FALLS PLANT. (Five Points quarry).

64.1	0.2	Quarry office; stop and check in.
64.35	0.25	Turn left at high tanks; proceed past white brick building on right.
64.4	0.05	Turn right past building.
64.5	0.1	Descend into quarry on ramp road.
64.6	0.1	Turn left at base of ramp road and proceed under conveyor; note diagonal joints beautifully exposed in lower Moorehouse to left.

64.7 0.1 Fork left around crushed stone piles and proceed down ramp.
64.8 0.1 Turn to right at bottom of small ramp and park.

Stop 2a. Lower part of Honeoye Falls quarry.

This relatively large and active quarry in the southeastern corner of Monroe County exposes an essentially complete section of the Onondaga Limestone and lower strata of the Marcellus Shale subgroup, including the Cherry Valley Member (see Figure 5 of Brett and Ver Straeten, this volume). Strata in this quarry are slightly deformed (Alleghenian?), and rocks display a gentle dip roughly to the south at a steeper angle than the regional 0.5^{o} dip. Two prominent sets of nearly vertical joints occur in the quarry; the middle portion of the Onondaga also shows a distinctive pattern of diagonal jointing oriented at approximately 080-31S and 080-50S. The Honeoye Falls quarry is the only site in the western portion of the state that displays the entire Onondaga Formation (although the Edgecliff Member is only poorly exposed). The Onondaga Limestone at Honeoye Falls is typical of western exposures. The quarry is particularly notable for the unusually coarse facies of the Cherry Valley Member (Oatka Creek Formation) and for the distinctive channeling beneath it that in places chops out nearly the entire Union Springs Formation.

Mining operations work from six main lifts in the Honeoye Falls quarry. The lowest rocks exposed occur in a small, approximately 6.5 m deep sump pit in the northeast portion of the quarry. These rocks are rarely well exposed, commonly covered by water or by a muddy coating. The Upper Silurian (Akron Formation) - Middle Devonian (Onondaga Formation) unconformity is reportedly present in the lower part of the pit (General Crushed Stone Corp. records) although, based on regional trends, we would estimate the contact to lie some meters below. The sides of the small pit are composed of the cherty Clarence facies of the Edgecliff Member. The walls of the overlying larger pit expose a complete section of the Nedrow Member. The base of the Nedrow essentially forms the floor of the large pit in the northeast area of the quarry. Overlying micritic limestones interbedded with more fossiliferous crinoidal wacke- and packstones are characterized by alternating light gray and dark cherts. The middle part of the member features two distinctive subunits; lower, relatively coarse crinoidal pack- to grainstones with scattered corals, somewhat similar to the Jamesville Quarry facies of the Edgecliff Member, and a distinctive interval of greenish-gray argillaceous limestones and green shales 2.8-3.4 m below the Nedrow-Moorehouse Member contact. The green shales are rich in coral and crinoid material, and are correlative with similar strata at the Stafford quarry (Stop 1). The top of the green argillaceous interval occurs at the top of the second wall and forms the base of the third main platform within the quarry, similar to the Stafford quarry. Ellipsoidal and elongate nodular to bedded cherts, which may feature dolomitic rinds, occur scattered throughout generally light gray micrites of the upper Nedrow. A prominent hackly dark chert bed 1.6 m above the green argillaceous interval is probably associated with a prominent, dark to black shale in the upper Nedrow noted widely throughout parts of New York and Pennsylvania.

The Moorehouse Member, as defined herein, begins 2.8 m above the base of the third lift and is relatively thick at this locality (ca. 17 m) Limestones in the lower portion of the member (ca. 5.5 m-thick) appear increasingly fine-grained up to the level of the fourth lift. Nodular chert bands and several layers of bedded, hackly-breaking chert are common. Thin shaly to clay-rich interbeds are also found. Faunal diversity is rather low through the lower part of the Moorehouse.

The middle to upper parts of the Moorehouse at the Honeoye Falls quarry show a general coarsening-up trend. The upper part of the member is generally inaccessible in the high walls of the south face of the quarry, below the Onondaga Indian Nation bentonite (OIN;=Tioga B of Way et al., 1986). Comparison with nearby sections (e.g., Stafford quarry, Stop 1) indicate a rapid coarsening up above a thick, notably more argillaceous unit that occurs above the middle of the Moorehouse (this argillaceous unit is widely recognized across the Appalachian Basin throughout New York and Pennsylvania into at least northern West Virginia; see discussion in Brett and Ver Straeten, above). Where accessible the upper beds of the member are highly fossiliferous and yield well preserved brachiopods and some crinoid crowns (particularly *Arachnocrinus)* on weathered surfaces. These strata represent the coarsest facies of the upper part of the Onondaga Formation.

Return to cars and <u>proceed back up small ramp</u>; retrace route to the high tanks .

65.0	0.2	Drive beneath conveyor again, and immediately <u>turn right</u> and ascend ramp road and continue straight to white brick building.
65.2	0.2	<u>Turn left</u> at white brick building toward high tanks.
65.3	0.1	At main quarry road, <u>turn left</u> and proceed toward old equipment area onto narrow road along north rim of quarry.
65.65	0.35	Fork of lower and higher quarry roads.
65.7	0.05	<u>Enter higher road</u> onto top of Onondaga Limestone (do not drive onto berm pile at back).
65.9	0.2	Park cars.

<u>Stop 2b</u>. <u>Upper part of Honeoye Falls quarry</u>.

The Tioga B (Onondaga Indian Nations bentonite) forms a prominent break within the quarry at the base of the overlying Seneca Member. The Seneca is characterized by 6.6 m of dominantly tabular wacke- to packstones with minor pale gray cherts. Thin clay-rich partings, most notably at 1.75, 3.05, 3.95, 4.7, and 5.4 m above the base of the Seneca, represent altered volcanic ashes (K-bentonites). The upper bed of the Seneca Member, which consists of relatively coarse crinoidal and brachiopod rich limestone is capped by a sharp lithologic break with the overlying Union Springs Formation.

A thin, mm-scale black shale at the Onondaga-Union Springs contact overlies a relatively minor but widespread submarine unconformity across New York. Overlying strata comprise basal deposits of the Middle Devonian, siliciclastic-dominated Hamilton Group. Recent quarry operations in 1993-1994 have exposed a fascinating cross-section of lower strata of the Marcellus Shale subgroup (see Figure 7 and text of Ver Straeten et al., this volume). The rock exposed comprise the Bakoven and Hurley Members of the newly redefined Union Springs Formation and the Cherry Valley and overlying black shale facies of the also newly-redefined Oatka Creek Formation. The Honeoye Falls quarry represents the westernmost known exposure of the Cherry Valley Member and underlying Union Springs strata in New York State (NOTE: terminology used for Marcellus strata above the Onondaga Formation in this field trip guide follows the as yet informal stratigraphic revision presented in Ver Straeten et al., above).

A 15 cm-thick coarse, biotite-rich tuff of the Tioga F bentonite bed immediately overlies the thin skim of black shale at the Seneca-Bakoven Member contact. Overlying thin styliolinid limestones and black, platy, laminated shales of the Bakoven represent deposition under relatively deep, oxygen-starved conditions. A prominent phosphatic fish bone bed that yields abundant conodonts and fish remains, including onychodid teeth and dermal armor of arthrodires (discussed above) occurs within the package of thin limestones above the Tioga F. The Bakoven Member varies in thickness from 7 cm to at least 1.7 m along the outcrop as a result of extensive channeling along the discontinuity surface at the base of the overlying Cherry Valley Limestone.

A relatively thin, light-weathering, richly fossiliferous limestone found toward the eastern part of the Union Springs-Oatka Creek exposures represents the widespread Chestnut Street submember of the new Hurley Member (Union Springs Formation). Pygidia and cephala of the proetid trilobite *Dechenella haldemanni* are the most diagnostic fossils from this unit. Also present are small to medium-sized brachiopods, auloporid corals, sponge spicules and calyces of the diminutive crinoid *Haplocrinites*. The upper and lower contacts of this bed are sharply defined.

Along most of the exposure the Chestnut Street submember appears to have been removed, along with varying amounts of the underlying Bakoven Member by an extensive, channeled erosion surface that underlies limestones of the Cherry Valley Member. The Cherry Valley is extremely variable in thickness in this quarry, ranging from just over 40 centimeters upwards to over three meters along 300 meters of outcrop exposed in fall 1993-spring 1994. It is represented a very atypical facies for the unit, which consists dominantly of crinoidal pack- to grainstones with lesser amounts of fenestrate bryozoan and styliolinid material. As previously stated, the basal surface of this unit is extremely irregular and the Cherry Valley appears to fill low spots cut into the underlying Union Springs Formation. In contrast to most other localities, cephalopods are generally not particularly common at this location, although poorly preserved conchs of *Agoniatites* as well as some orthoconic nautiloids (e.g., *Striacoceras*) have been obtained. The upper surface of the Cherry Valley Member is also quite sharp and distinct; it commonly appears as a pyrite-coated corrosion surface. It displays an abrupt contact with overlying black shales assignable to the Berne Member.

To date, the highest beds observed in the quarry consist of barren, black, laminated shales with minor styliolinid hash beds. This interval at nearby localities (e.g., bed of Oatka Creek at LeRoy) features a richly fossiliferous bed of brachiopods and small corals that represents the first occurrence of the classic Hamilton Group fauna informally termed the LeRoy bed (= gray bed of Baird and Brett, 1986; see Ver Straeten et al., above); the LeRoy bed has not as yet been found at the Honeoye Falls quarry and likely occurs in strata above the erosional contact of the Oatka Creek Formation with overlying glacial till deposits.

As previously noted, the basal contact of the Cherry Valley Member shows significant paleorelief across the quarry exposure, including distinctive channel-like structures. The cause of the channeling is unknown; in all other localities the Cherry Valley displays a nearly planar, although sharp, basal contact. We presume that the erosional scour was developed as a result of submarine channeling by gradient currents during a relative low stand in sea level. Similar erosional furrowing associated with maximum regression and subsequent earliest transgression has been observed at other localities in younger Middle to Upper Devonian strata (for example, see Brett and Baird, 1990).

Return to cars and turn around; retrace route to quarry entrance.

66.1	0.2	<u>Turn right (east) onto quarry road</u> and proceed along north rim of quarry.
66.4	0.3	Drive through old equipment and <u>bear left</u> toward main quarry road.
66.5	0.1	<u>Bear left</u> onto main quarry road.
66.7	0.2	Check out at quarry office.
66.9	0.2	At quarry entrance, <u>turn left (west)</u> onto Honeoye Falls #6 Road.
67.7	0.8	<u>Turn right (north)</u> onto Works Road.
68.7	1.0	Cross Five Points Road.
69.4	0.7	<u>Fork to left</u>; Works Road becomes Phelps Road.
70.9	1.5	Intersect with NY Rte. 15A (Rush-Lima Road).
71.4	0.5	Enter village of Rush.
71.7	0.3	Rte. 15A bends to the right.
71.8	0.1	Cross over Honeoye Creek.
72.05	0.25	Junction with NY Rte. 251; <u>turn left (west)</u> onto Rte. 251.
72.6	0.55	Leave village of Rush.

73.4	0.8	Junction with NY Rte. 15; turn right onto Rte. 15.
73.5	0.1	Bear right onto I-390 North entrance ramp.
75.9	2.4	Rochester skyline view.
76.7	0.8	Exit from I-390 at Exit 12 and proceed to NY State Thruway.
77.0	0.3	Bear left at fork toward Thruway.
77.5	0.5	Toll booth of NY State Thruway; collect ticket.
77.65	0.15	Bear left at fork onto Thruway (I-90) east, toward Albany.
78.1	0.45	Merge onto Thruway eastbound.
89.9	11.9	Exit 45, I-490/Victor; continue east on I-90.
93.7	3.8	Exit 44, Canandaigua.
100.9	3.2	Exit 43, Manchester; exit off of I-90.
101.4	0.5	Toll Booth of Thruway.
101.4	0.1	Junction with NY Rte. 21; turn right (south) onto Rte. 21.
101.6	0.1	Junction with NY Rte. 96; turn left (east) onto Rte. 96.
101.8	0.2	Cross Canandaigua Outlet Creek.
104.1	2.3	Intersect Ontario Co. Rte. 7 at Manchester town hall; continue on Rte. 96. Manchester quarry in Onondaga Limestone is 1.6 mi. south along Rte. 7.
106.2	2.1	Intersect Ontario County Rte. 25; continue ahead.
108.1	1.9	Enter village of Clifton Springs.
109.4	1.3	Enter village of Phelps. Mobil gas station on left is opposite small road to right that leads across Flint Creek. Pull off to right.

Stop 3 (Optional). FLINT CREEK, WEST OF PHELPS.

The bed and banks of Flint Creek just west of Phelps provide an excellent exposure of the lower and middle parts of the Nedrow Member in its typical west-central New York facies. The banks expose medium gray, soft, bioturbated mudstones alternating with thin (up to 30 centimeters thick) tabular, somewhat argillaceous, fossiliferous wackestones. Bedding here appears rhythmic and some bundling of limestone beds is evident toward the top of the exposure. The shales near the stream level are extremely rich in well-preserved brachiopods, particularly robust specimens of *Pseudoatrypa*, *Leptaena*, and *Pentagonia*. Scattered crinoid debris and small rugose corals are also present.

The rhythmic bedding within the Nedrow Member may represent minor climatic oscillations. Neither the carbonate nor the mudstones appear to represent major event deposits. The limestone and mudstone bands both consist of multiple beds, some of which may represent thin event deposits, but with bedding largely obscured by bioturbation. Contacts of the interbedded limestones and shales appear to be relatively gradational and generally burrowed by *Chondrites* traces. Hence the alternations are thought to represent variations in the input of siliciclastic mud. The sharp contrast between of varying siliciclastic content has probably been diagenetically enhanced by early cementation of the more carbonate-rich intervals (discussed in Brett and Ver Straeten, above).

Return to cars and continue east on NY Rte. 96.

109.9	0.5	Railroad crossing opposite old Silver Floss sauerkraut plant.
110.0	0.1	Junction with NY Rte. 88 – turn left (north) onto Rte. 88 and continue to roadcuts just south of NY State Thruway overpass.
110.5	0.5	Pull off along road side just south of Thruway overpass.

Stop 4 (OPTIONAL). ROADCUT ON NEW YORK ROUTE 88, PHELPS.

Rock cuts on both sides of this north–south road display a Silurian–Devonian carbonate section. Because of the relatively strong component of southward dip, a relatively long section is exposed here. The lower part of the outcrops just south of the New York State Thruway bridge are in the upper part of the Upper Silurian Bertie Group that display an unconformable contact with the overlying Devonian strata. The unconformity overlies the Scajaquada Formation of the Bertie Group. Medium to slightly reddish gray shales and shaly dolostones display distinct small casts of salt crystal cubes, but few, if any fossils are obtainable from these beds.

The erosional Silurian-Devonian contact at the Wallbridge Unconformity is marked by thin stringers of quartz arenite, apparently of the Lower Devonian Oriskany Sandstone. Thin fissure fillings or neptunian dykes also of Oriskany quartz sand extend 30 to 40 centimeters downward below the upper surface of the Scajaquada Shale. The contact is overlain by a conglomeratic bed containing phosphatic or cherty clasts within a sandy limestone matrix. This bed is only a few centimeters thick, and is overlain conformably by brownish gray-weathering, very cherty micritic limestone (wackestone). Dark bluish-gray chert nodules up to fist size (10 to 20 cm in diameter) occur within these limestone beds. The rock is sparsely fossiliferous, but contains fragments of brachiopods, corals, and echinoderm debris. This unit is clearly not typical of the lower Edgecliff Member of the Onondaga Limestone; it is tentatively assigned to the upper Lower Devonian Bois Blanc Formation. Oliver (1963) reports minor exposures of apparently similar cherty, micritic limestone along the New York State Thruway in the vicinity of Phelps. The contact of the Bois Blanc and the underlying lenses of the Oriskany Sandstone represents a second, post-Wallbridge and pre-Bois Blanc unconformity.

The highest beds exposed in this roadcut consist of pale, pinkish gray-weathering crinoidal grainstone. These strata appear to be in sharp contact on the underlying Bois Blanc and are assignable to the basal Edgecliff Member of the Onondaga Limestone; this contact represent a third sub-Onondaga unconformity. In addition to the abundant crinoid skeletal material, the lower Edgecliff strata contain scattered solitary rugose and tabulate corals. The basal Onondaga contact contrasts markedly with the basal Devonian contact observed in the Oaks Corners (Stop 5) approximately 5 km to the southeast (discussed below).

110.7	0.2	Return to cars and continue north and turn around in drive north of Thruway overpass; retrace Rte. 88 back to Rte. 96.
111.4	0.7	Junction with NY Rte. 96; turn left (east) and continue through village of Phelps.
112.2	0.8	Cross over Flint Creek; outcrops of Upper Silurian Bertie Group.
114.1	1.9	Leave village of Phelps.
114.3	0.2	Cobblestone house; Scotch Highland cattle on left.
114.7	0.4	Junction with Ontario Co. Rte. 6 (Pre-Emption Road); turn right (south) onto Pre-Emption Road.
115.6	0.9	Entrance to Oaks Corners quarry; turn right (west) into quarry and check in at office; then return to Ontario Co. Rte. 6 (Pre-Emption Road) and turn right (south) on Pre-Emption Road.
115.7	0.1	Junction with Cross Road; turn right (west) onto Cross Road.
115.85	0.15	Pull off along roadside and park; walk down farm lane to quarry to the north (right).

Stop 5. OAKS CORNERS QUARRY, SOUTH WALL.

The old south wall of the active Oaks Corners quarry displays an excellent section ranging from the basal contact of the Onondaga Formation up through the lower two-thirds of the Moorehouse Member (see Figure 5 of Brett and Ver Straeten, this volume). The quarry can be approached by walking through a narrow strip of field just north of Cross Road at about 0.2 miles west of the junction of Ontario Co. Rte. 6 (Pre-Emption Road). Beds exposed at the level of the field and road are in the upper part of the Edgecliff Member (Clarence cherty facies) near its contact with the Nedrow Member. These flats are some 12-15 m above the base of the quarry. By walking to the left (west) and carefully proceeding down an abandoned quarry road from the upper platform it is possible to examine the lower part of the Onondaga down to the contact of the Edgecliff Member with the underlying Upper Silurian strata.

At the Oaks Corners quarry, in contrast to the Route 88 roadcuts (Stop 3), the Onondaga Limestone generally rests directly on beds of the Upper Silurian Akron Formation, a massive, dark brownish buff-weathering saccharoidal dolostone. J.W. Scatterday (personal commun.) reports a thin, locally-occurring sandstone in part of the quarry similar to that seen at Stop 4; it is presently unknown whether these strata are related to Oriskany or Tristates Group sandstones ("Springvale"). Where the sandstones are absent, the Akron-Onondaga contact is typically marked by rust staining, as a result of weathering of pyritic crusts that occur along the erosional surface. The unconformity is approximately four to five meters higher in the Silurian section than at the Rte. 88 exposures (Stop 3), where the unconformity surface is cut down to the level of the older Scajaquada Shale. In the Oaks Corners quarry, the unconformity displays relief of up to one meter. Large channel-like depressions, up to several meters across, occur along this boundary. Again, in contrast to the Route 88 cuts, where possibly four to five meters of Bois Blanc Formation occur above the unconformity, the Bois Blanc is missing at this location and the basal beds of the Edgecliff Member occur within hollows on the combined Wallbridge, sub-Bois Blanc (?), and sub-Onondaga unconformities.

The lowest unit of the Edgecliff Member is a fine- to medium-grained crinoidal grainstone, typically pinkish-weathering, which ranges from 40 to 120 centimeters in thickness as a result of the irregular topography at its base. The transition to chert-rich strata (Clarence facies) is abrupt but gradational and consists of cherty, crinoidal pack- to wackestones. The remainder of the Edgecliff Member is dominated by the Clarence cherty facies, and consists predominantly of pale gray-weathering micritic limestone with 20 to 30% dark gray chert. Skeletal wacke- and packstones, which feature large crinoid columns and tabulate and rugose corals, occur approximately three and six meters above the base. These form two light pinkish gray-weathering bands in the wall of the quarry and thus constitute important markers as well as represent the apparent tops of minor shallowing-up cycles. A 46 cm-thick, medium dark gray, sparsely fossiliferous and very calcareous shale occurs high in the Edgecliff Member. This interval closely resembles the higher Nedrow Member, but it is separated from the latter unit by 1.2 meters of micritic cherty limestone typical of Clarence facies. The Edgecliff is approximately 8.5 m-thick at the Oaks Corners quarry.

A sharp limestone-shale contact at the top of the Edgecliff marks the base of the Nedrow Member. Lowest shales of the Nedrow contain a relatively abundant fauna of *Pseudoatrypa*, *Leptaena*, small rugose corals, and other fossils. The Nedrow Member itself is approximately 5.3 m-thick at this location and consists of medium to dark gray, very calcareous shale or mudstone that alternate with argillaceous lime mudstones or wackestones. Some of the rhythmic bedding and bundling of limestones observed along Flint Creek in the corresponding interval (Stop 2) is also apparent here, although somewhat more subtle in this outcrop. The upper portion of the Nedrow Member contains a distinctive dark gray to black shale interval with small black chert nodules. This black shale marker bed has proved to be very widespread across the Appalachian Basin and is recognized at present across parts of New York to southern Pennsylvania (see above). A 40 cm-thick, light-weathering limestone bed with minor pyrite separates this bed from an overlying brownish-gray shale interval that contains a thin nodular limestone bed rich in the brachiopod *Schizophoria* sp. This bed forms the floor of the second highest lift near the abandoned quarry access road along the south wall of the quarry, where the brachiopod fauna and spectacularly weathered *Chondrites* burrows can be observed.

The Moorehouse Member in the Oaks Corners quarry consists dominantly of medium- to thick-bedded micritic limestones (slightly argillaceous fossiliferous wackestones). Several intervals of medium to dark gray chert occur within this section. Particularly heavy cherty beds occur approximately four to five meters above the base of the Moorehouse. Thin intervals of medium to dark gray, very calcareous shale that closely resemble the Nedrow Member (particularly the *Schizophoria* bed) occur approximately 2.2 and 3.0 m above the base of the Moorehouse. A relatively thick calcareous shale (80 cm-thick) that features well-preserved specimens of *Pacificocoelia* as well as *Schizophoria* occurs 5.4 m above the base of the Moorehouse and is well displayed in a cut along the side of a second, more westerly abandoned roadway that leads into the quarry along the south side. A cherty micritic limestone bed approximately 1.5 m below the *Pacificocoelia* shale contains abundant coiled nautiloid cephalopods (*Gyroceras*) which are well displayed on a glacially-polished surface just east of the second access road.

Highest beds of the Moorehouse are quite fossiliferous, and are particularly rich in brachiopods. Argillaceous beds about one meter below the top of the quarry contain a fauna dominated by atrypids and other brachiopods such as *Megakozlowskiella, Leptaena, Schizophoria*, and others. Trilobite material is rather common in the upper shaly beds and includes specimens of *Phacops* and *Odontocephalus.* Echinoderm material is relatively sparse, although rare specimens of an undescribed stalked rhenopyrgid edrioasteroid have been obtained from the shaly beds. Overall, the Moorehouse appears to have at least four to five minor shallowing-up cycles, each of which commences with calcareous shales and culminates in more massive, cherty, micritic limestones that may contain corals such as *Acinophyllum*. The upper part of the Moorehouse Member and the entire Seneca Member are absent at this location due to post-Devonian erosion.

The lower part of the Onondaga Formation at Oaks Corners is similar to more western sections in that the Edgecliff Member is dominated by Clarence cherty facies. Only a relatively small proportion (approximately 10% of the member) consists of the more classic chert-poor crinoidal packstones characteristic of the Jamesville Quarry facies (see body of paper) of the Edgecliff. The remainder of the member is relatively cherty, although not as chert-rich as the Clarence in Erie and Genesee counties. The Edgecliff here (ca. 8.5 m-thick) contrasts markedly with that seen at the Seneca Stone quarry (Stop 5; 21 km southeast) where the member is quite thin (< 3 m-thick) and consists mainly of the crinoidal wackestone facies with only a single band of cherty micritic limestone (see stop description below). The Nedrow Member at Oaks Corners is considerably shalier and less fossiliferous than the equivalent beds seen at the Honeoye Falls quarry (Stop 1; 50 km west) where the member is composed of micritic limestones with light gray to dark chert and greenish-gray argillaceous limestones rich in a diverse assemblage of rugose and tabulate corals, gastropods, brachiopods and other fossils. At the Oaks Corners quarry the Nedrow is only sparsely fossiliferous and carries a fauna dominated by a few species of brachiopods. Moreover, the upper portion of the member contains the distinctive black marker bed, which in the Stafford and Honeoye Falls quarries (Stops 1&2) appears to be represented by fossiliferous greenish shales. The Nedrow interval at Oaks Corners is, on the other hand, quite similar to that seen at Seneca Stone quarry (Stop 5).

The Moorehouse Member at Oaks Corners displays features suggestive of an intermediate setting between those seen in the Honeoye Falls and Seneca Stone quarries (Stops 1&5). It is thicker, more chert-rich, and more fossiliferous than the equivalent interval at Seneca Stone quarry. However, it contains considerably less crinoid material and is dominated by brachiopod-rich wackestone lithologies as opposed to the crinoidal wacke- and packstones of the Moorehouse at the Honeoye Falls quarry. It carries a fauna rich in brachiopods assignable to the *Megakozlowskiella, Atrypa*, and *Pacificocoelia* communities recognized by Feldman (1980).

Overall, the Onondaga appears to be uniformly of more fine-grained, more argillaceous, and probably deeper water facies than those seen in the corresponding intervals at and to the west of the Honeoye Falls quarry (Stop 1). However, the relationship of facies to those of the Seneca Stone quarry is not so clear cut. The lower Edgecliff Member appears to be distinctly finer-grained, more cherty, and less fossiliferous than the equivalent strata at Seneca Stone quarry. We suggest that only the upper beds of the Edgecliff at Oaks Corners quarry are continuous into the Seneca Stone quarry, where the lower units have pinched out over the irregular topography below the combined Wallbridge and sub-Onondaga unconformities. Those upper Edgecliff beds show distinct facies changes between Oaks Corners and the Seneca Stone quarry 21 km to the southeast. In particular, the chert-rich intervals appear to grade southeastward into crinoidal wacke- and packstones, whereas the dark gray Nedrow-like shale interval approximately 7.3 m above the base of the Edgecliff appears to grade into a single chert-rich micritic bed at the Seneca Stone quarry. Hence, it would appear that the Edgecliff Member is uniformly of deeper water character at Oaks Corners than at Seneca Stone quarry, which suggests a more basinal setting to the west at that time. The Nedrow Member appears very similar at both localities. The Moorehouse Member, in total contrast, appears to represent distinctly shallower water facies at the Oaks Corners quarry than it does at the Seneca Stone quarry, a complete reversal of the trend seen in the Edgecliff Member. This suggests that the basin axis or center of subsidence shifted through time from a more westerly position (probably west of Oaks Corners) eastward to the vicinity of the Seneca Stone quarry during deposition of the Onondaga Limestone.

Return to vehicles and turn around on Cross Road, then retrace route to Pre-Emption Road.

116.0	0.15	Intersection with Pre-Emption Road (Ontario Co. Rte. 6); Continue straight across (east) on Cross Road.
116.1	0.1	Railroad crossing.
117.1	1.0	Large gravel pit on left.
117.8	0.7	Intersection with NY Rte. 14; continue straight on Cross Road.
118.6	0.8	Junction with Town Line Road; leave Ontario County, enter Seneca County.
119.0	0.4	Intersection with NY Rte. 96. Turn right (southeast) onto Rte. 96.
121.9	2.9	Cross Whiskey Hill Road; continue on Rte. 96.
124.1	2.2	Junction with North Road, Rte. 96 turns right (south). Proceed straight east onto North Road.
124.8	0.7	Seneca County Fairgrounds.
126.0	0.8	Junction with NY Rte. 414. Turn right (south) onto Rte. 414.
125.95	0.35	Turn right (west) into "McDonalds;" **10 minute rest stop.** Return to Rte. 414 southbound.
126.0	0.05	Junction with Rtes. NY 5-U.S. 20; continue straight (south).
126.1	0.1	Bridge over Seneca River.
126.15	0.05	Junction with River Road; turn left (east) onto River Road.

127.8	1.65	Junction with Kingdom Road (opposite lumber yard); turn right (south) onto Kingdom Road.
128.8	1.0	Junction with County House Road, jog left; name changes to Disinger Road. Continue south.
129.8	1.0	Intersect Tom Allen Road; Continue straight on Disinger Road.
130.8	1.0	Junction Yellow Tavern Road; turn left (east) onto Yellow Tavern Road.
131.6	0.8	Junction NY Rte. 414, continue east on Yellow Tavern Road (Seneca Co. Rte. 21).
132.0	0.4	Bend in road; name changes to Canoga Springs Road.
132.7	0.7	Entrance to Seneca Stone Quarry; turn left (north) into quarry and check in at office, then proceed straight (north) into quarry.
133.1	0.4	Fork left onto ramp road, beyond sheds on left. Proceed to lower level.

STOP 6. SENECA STONE QUARRY.

The Seneca Stone quarry is another relatively large quarry that exposes a complete section from the upper part of the Lower Devonian Manlius Limestone to the top of the Middle Devonian Cherry Valley Member of the Oatka Creek Formation (Marcellus subgroup; see section in Brett and Ver Straeten, this volume, Figure 5). Strata within the quarry generally appear undeformed, with the exception of a prominent north-directed thrust fault that cuts up through the upper part of the Onondaga Formation (visible in the southeast and southwest parts of the quarry).

Limestones of the Lower Devonian Manlius Formation (Helderberg Group) are found in a small sump pit in the bottom of the quarry, overlain unconformably by quartz arenites of the Lower Devonian Oriskany Formation (lower part of the Tristates Group). These two units are separated by the true Wallbridge Unconformity that marks the base of the Kaskaskia Megasequence. Clasts of the older Manlius Limestone are visible in the base of the Oriskany, where they may be overgrown by tabulate coral colonies. The white quartz arenites of the Oriskany Sandstone also feature numerous large, robust brachiopods (including *Costispirifer arenosus*, *Rensselaeria*, and *Hipparionyx*), platyceratid gastropods, and rare rugose corals.

Twenty-five km northeast of the Seneca Stone quarry, at Auburn, sand-dominated strata of the upper part of the Tristates Group (Carlisle Center and Schoharie Formations) are present, similar to strata reported for the Syracuse area (see Brett and Ver Straeten, above). At Seneca Stone quarry, however, these rocks are absent except as reworked phosphatic cobbles in a basal sandstone bed of the Edgecliff Member (Onondaga Formation). Reworked, phosphatized clasts of the Manlius and Oriskany Formations are also found in the basal Edgecliff sandstone bed.

A thin interval of coral-rich limestone is succeeded by two cyclic packages of Edgecliff strata that appear to be equivalent to the two upper Edgecliff cycles seen at Oaks Corners (Stop 5). Well developed fine-grained, cherty Clarence facies are only developed within the lower part of the second cycle at Seneca Stone; the Edgecliff member in general here appears coarser than equivalent strata at Oaks Corners (Stop 5).

A sharp change into overlying calcareous dark shale-dominated strata marks the base of the Nedrow Member. The Nedrow and lower to middle parts of the overlying Moorehouse Member are best exposed in the east wall of the quarrybelow the third platform. The lower part of the Nedrow consists of alternating dark shales and argillaceous wackestones. Several distinctive crevices within the middle to upper Nedrow here may represent altered volcanic ash beds. The distinctive "black bed" in the upper Nedrow, which at Seneca Stone quarry consists of black shale with several bands of black chert nodules, is highly visible in the eastern wall. The bed, as observed at the previous locality (Stop 5), is overlain by a light weathering limestone and the dark "*Schizophoria* bed" at the top of the Nedrow.

Overlying strata assigned to the lower part of the Moorehouse Member consist of alternating calcisiltite limestone beds (< 1 m-thick) interbedded with thin (ca. 5-15 cm-thick) calcareous shales. This interval is relatively similar to facies that characterize the Onondaga-equivalent Selinsgrove Limestone of central Pennsylvania. Two paired crevices approximately 0.9 and 1.0 m above the *Schizophoria* bed represent apparent K-bentonite beds that are widely traceable across the central part of the Appalachian Basin (see Brett and Ver Straeten, above). Quartz pebbles and granules have been noted in a 14 cm-thick calcareous shale bed 2.5 m above the base of the Moorehouse. A 45 cm-thick argillaceous unit 6 m up in the Moorehouse, just below a platform on the east wall, correlates with the prominent "false Nedrow" shale seen at Stafford (Stop 1) and other localities across much of the Appalachian Basin (see Brett and Ver Straeten, above).

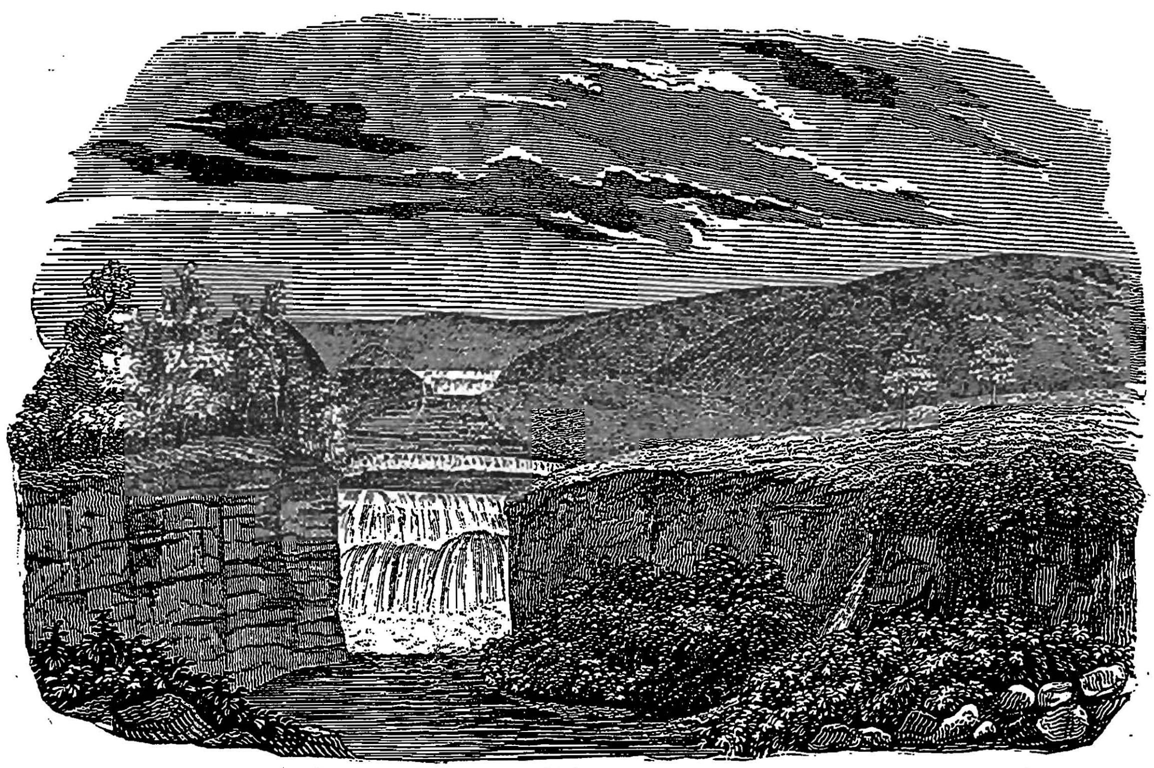
More resistant, coarsening-up strata of the upper Moorehouse Member Form the wall of the lift above the Nedrow-like shale unit up to the thick Tioga B-OIN Bentonite bed at the base of the overlying Seneca Member. Two conspicuous crevices, which feature biotite-rich claystones, represent the First and Second Cheektowaga Bentonites of Conkin and Conkin, (1979, 1984; Conkin, 1987). A 10 cm-thick, dark chert bed forms the cap of the 9.8 m-thick Moorehouse Member at Seneca Stone, immediately below the Tioga B-OIN bentonite bed.

The Seneca Member represents a general fining-up trend from the relatively coarse upper part of the Moorehouse Member. The Tioga B-OIN bentonite marks the base of the member. Five and possibly seven additional bentonites occur within the Seneca Member and the base of the overlying Union Springs Formation at Seneca Stone. Two fining- to coarsening-up cycles within the Seneca represent relatively smaller scale cycles superimposed on the general deepening-up trend (see Brett and Ver Straeten, above). Lithology of the member ranges from poorly fossiliferous wackestones to low diversity brachiopod-rich packstones (e.g., "Pink *Hallinetes* (formerly *Chonetes*) Zone," 3.2-4.0 m above the base). Nodular cherts and impure dolomitic cherts occur scattered throughout the member. The Seneca Member, as defined herein, totals 7.15 m in thickness. The upper contact of the Seneca Member at Seneca Stone quarry as defined herein is marked by a distinctive bedding plane with large horizontal burrows, scattered fish debris, and some pinkish weathering, apparently hematitic, limestone "clasts." The bed is immediately overlain by eight cm of black shale and a thin (3 cm-thick) k-bentonite bed. Overlying dark, fine-grained, styliolinid limestone beds are assigned herein to the Bakoven Member of the Union Springs Formation. The prominent, rusty-weathering Tioga F bentonite (ca. 12 cm-thick) occurs 65 cm above the base of the Bakoven at Seneca Stone.

The lower 1.3 m of the Bakoven is dominated by the previously mentioned thin styliolinid limestone beds. Black shales become more prevalent above, but occur interbedded with numerous thin (ca. 2-20 cm-thick) micritic limestone beds. The top of the Union Springs Formation at Seneca Stone quarry is represented by the richly fossiliferous Chestnut Street submember of the Hurley Member. The Chestnut Street beds, with their characteristic lighter-weathering color and fauna (e.g., proetid trilobites and the microcrinoid *Haplocrinites clio*), form a thin ledge of limestone approximately 10 cm in thickness that is firmly welded to the base of the overlying Cherry Valley Member (Oatka Creek Formation).

The Cherry Valley Member at Seneca Stone is more typical of its occurrence along much of its outcrop west of the Albany area. Three subdivisions, a lower massive, middle nodular, and upper massive limestones comprise an approximately 50 cm-thick section of organic-rich, pyritic, bedded to nodular wacke- to packstones. The classic goniatite and nautiloid cephalopod fauna of the Cherry Valley, which includes *Agoniatites vanuxemi* and *Striacoceras typum*, is well displayed on the excellently exposed upper bedding plane seen above the south wall of the quarry. Truncated cephalopod shells, fish bone material, and scattered pebbles, along with the development of a firm- to incipient hard-ground and pyritic crusts on the upper surface of the Cherry Valley Member, indicate a period of submarine sediment starvation prior to deposition of overlying shales. Overlying strata are not at present exposed, but Baird and Brett (1986) report dark shales that include a distinctive bed (LeRoy bed=gray bed of Baird and Brett, 1986; see Ver Straeten et al., above) in which the classic fauna of the Middle Devonian Hamilton Group makes its first appearance in the Appalachian Basin.

END OF TRIP. To return to the University of Rochester, retrace route to NY Rte. 414 at "McDonalds"; continue north on Rte. 414 past North Road and past intersection with NY Rte. 318 to entrance onto NY State Thruway (I-90). Take Thruway west (toward Buffalo) to exit for I-390 at Rochester and proceed north on I-390 to the University of Rochester.



Upper and Middle Falls of Portage. Sketch by Mrs. Hall
[From Hall, 1843, Geology of the Fourth District, Figure 96, p. 224]

FRASNIAN (UPPER DEVONIAN) STRATA OF THE GENESEE RIVER VALLEY, WESTERN NEW YORK STATE

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INTRODUCTION

History of Investigation

The Frasnian section in the Genesee Valley is located in the western part of the central Devonian outcrop belt of New York State that stretches from the Catskill Mountains to Lake Erie. The Genesee Valley was close to the center of the Fourth Geologic District of the first state survey, the district between Cayuga Lake and Lake Erie, covered by James Hall. Hall's final report, the monumental Part IV of the Geology of New York (Hall, 1843), is the starting point for Devonian studies in the region (Figure 1). The rocks of the classical Genesee, Portage and Chemung divisions consist of a succession of marine siliciclastics beginning with dark basinal shales (Genesee) that contain a pelagic molluscan fauna, which are overlain by lighter colored shales and siltstones that contain a pelagic and benthic fauna (Portage and Naples to the east) that were deposited in basin-and-slope environments. These strata are overlain by siltstones and sandstones bearing a benthic brachiopod fauna (Chemung) that were deposited on the slope and outer-shelf. Most of the fossils in these units were described in monographs by Hall (1879) and Clarke (1899, 1904). The excellent exposures in the Genesee Valley made this section a center for geological studies of the distal rocks of the Catskill Delta, which record stages in the filling of the Appalachian Foreland Basin. Beginning with Hall's (1843) observation of the lateral changes in lithology within each division ("constant increase of arenaceous matter" to the east and "increase in mud or shale" to the west) the Genesee Valley section came to play a central role in the application of the facies concept to the Appalachian Devonian.

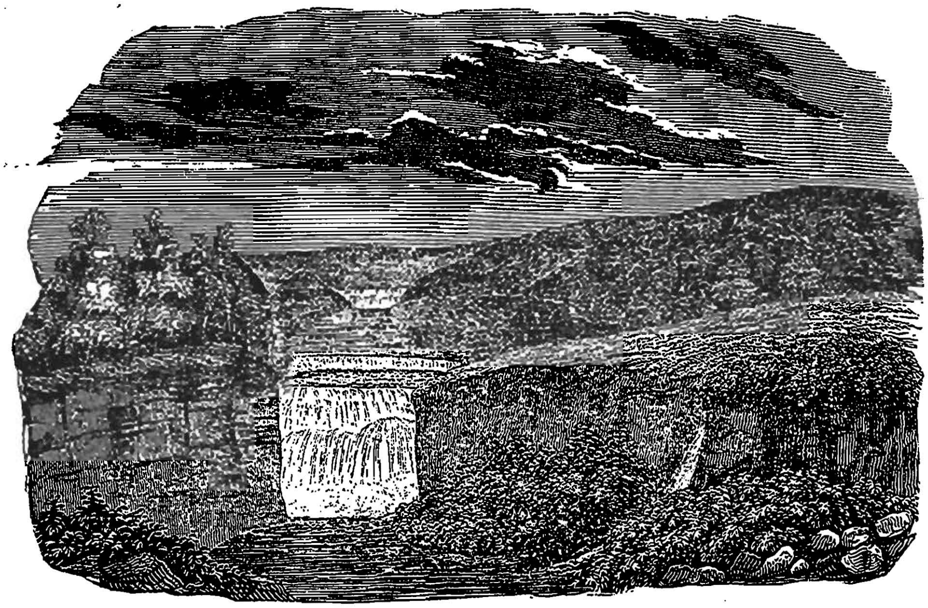

Figure 1. "Upper and Middle Falls of Portage". From a sketch by Mrs. James Hall. From Hall (1843).

Application of facies concept

When Chadwick (1935) declared that "Chemung is Portage" he could have added "Portage is also Genesee" to the title of his paper. It was recognized that the original divisions were facies-equivalents of each other, each one thickening shoreward (eastward) and grading into and interfingering with the adjacent facies. The dark shales of the offshore basin facies of the Genesee were seen to pass shoreward into the lighter colored shales and sands of basin-margin and slope facies (clinoform/ramp) of the Portage (and Naples). The Portage, in turn, intertongued with the sandy Chemung facies of the outershelf, which grades into nearshore shelf and shore deposits. The broadly upward-coarsening and vertical stacking of the facies seen in the Genesee Valley region (in the order Genesee-Portage-Chemung) is a manifestation of the upward shallowing and the seaward (westward) shift of each the facies as the Catskill Delta built into the basin. Rickard (1981) provided a concise review of the general relations and Kirchgasser (1985) and Kirchgasser et al. (1986) reviewed aspects of the history of correlations and stratigraphic classification.

Black-shale correlations

While the names Genesee, Portage and Chemung (and Catskill for the terrestrial deposits) survive as useful general descriptors of the major facies, only the name "Genesee" survives as a unit (group) in the modern lithostratigraphic classification (Figure 2). The boundaries of the major stratigraphic divisions and subdivisions employed today are defined by the sequence of key black shales. Each black shale records a deepening event or transgression, and can be correlated from the basin shoreward toward the shelf. Robert Sutton and his students at the University of Rochester and James Pepper, Wallace deWitt, Jr. and George Colton of the U.S.G.S. were instrumental in the recognition of the black shale correlations. These correlations led to the development of modern classification which were synthesized in the correlation charts of Rickard (1964, 1975) and employed in the revised Geologic Map of New York State (Rickard and Fisher, 1970).

Major subdivisions

The major subdivisions of the New York Frasnian (Genesee, Sonyea, and West Falls groups) consist of similar cyclic sequences of basal black shale (respectively Geneseo, Middlesex, and Rhinestreet) overlain by gray and green shales [e.g,. Penn Yan-West River shales in the Genesee Group; Cashaqua Shale in the Sonyea Group; Angola and Hanover shales in the West Falls Group (Figure 2)]. Each group (cycle) thickens and coarsens shoreward (eastward) as the shales of the basin grade into and interfinger with the leading edges of clastic wedges of turbiditic silt and sand. The major units of the coarse clastics from the Genesee Valley and eastward are the Ithaca (Genesee Group), Rock Stream (Sonyea) and Nunda-Wiscoy (West Falls). The Genesee and Sonyea cycles, which together are about 90 meters thick in the Genesee Valley, thin westward to less than 18 meters at Lake Erie, and thicken eastward to almost 500 meters at Cayuga Lake. The cyclicity of black shale-gray shale couplets, so apparent in the major divisions, is also seen at smaller scales down to centimeters and millimeters. Within the succession, numerous thin and widely traceable "event horizons" have been identified (including carbonates, erosional pyrite and distal tubiditic siltstones; Figures 3, 4). All of these subdivisions provide the stratigraphic framework for defining the biostratigraphic sequences of ammonoids (goniatites) and conodonts. They are also the starting point for discussion of the role of eustatic sea-level change, basin subsidence, and tectonics in explaining the cyclicity and facies migrations.

Lithostratigraphy

The Genesee Valley Frasnian lithostratigraphy is based on data from exposures in the deep canyon of the Genesee River at Letchworth Park, from scenic glens and gorges tributary to the Genesee River, wells drilled for

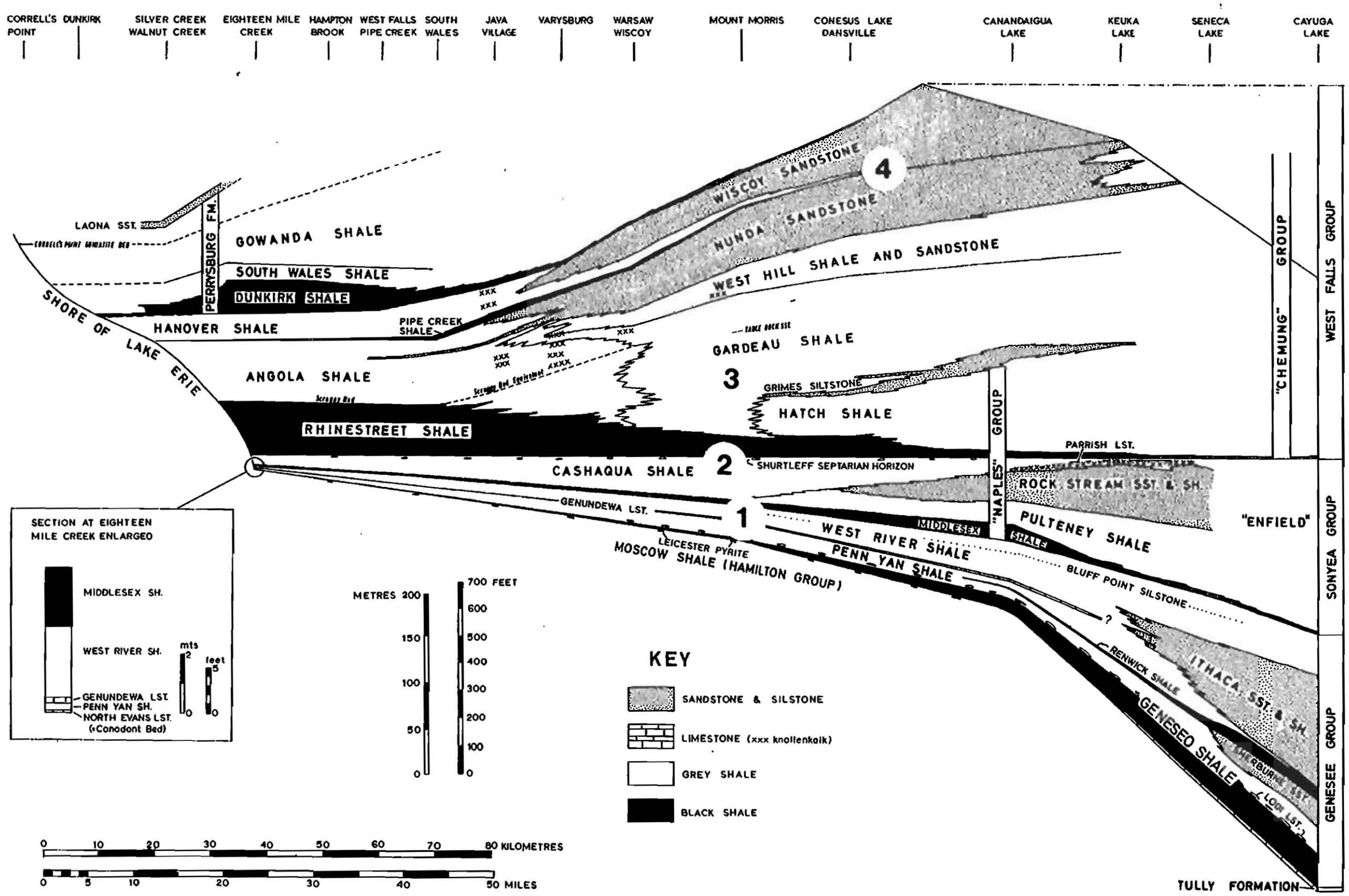

Figure 2. Frasnian facies in New York State from the Finger Lakes to Lake Erie showing stratigraphic position of field-trip stops. Modified from House and Kirchgasser (1993).

hydrocarbons, salt, water and engineering studies, and along road-cuts. Taken together, data on rock type, sedimentary structures, fossils, rock unit sequence, rock unit thickness, and other parameters facilitate a reasonable and broadly applicable lithostratigraphy, one which forms the basis both for stratigraphic correlations within the Appalachians and beyond, and for geologic mapping.

Upward-coarsening of the section

Upper Devonian strata of the Genesee Valley demonstrate the upward coarsening characteristic of the Late Devonian throughout the central and southern Appalachians (Figure 2). When considered with the Onondaga/Hamilton sequence below, the strata exposed in the Genesee Valley south of Rochester represent the classic sedimentary response to tectonics envisaged by many pre-plate tectonics workers and summarized by Pettijohn (1975). Friedman et al. (1992) provide a more current summary based on plate tectonics and note that the Appalachian Devonian sequence exemplifies a foreland basin where sediments were derived from a thrust faulted orogen that resulted from plate interactions. Deposition was in a basin floored by subsiding continental crust. Correlation of the Genesee Valley Frasnian strata across New York and Pennsylvania demonstrates the displacement of deeper water facies by the lateral shifting of shelf, shore, and coastal plain facies to the west and northwest as basin-filling proceeded.

Stratigraphic marker-beds

Though clearly developed, upward coarsening and lateral displacement of facies is complicated by recurring cyclic sequences that make a simple regional pattern locally more complex. Correlation within and between complex stratigraphic sections is made possible through use of the black and dark gray shales (Geneseo, Middlesex, Rhinestreet, Pipe Creek, and Dunkirk) as marker beds. Additional secondary stratigraphic control is based on thin carbonates (e.g., Genundewa, Parrish) and distinctive thin siltstones (e.g., Bluff Point Siltstone), sandstones (e.g., Crosby Sandstone), and less obviously on ash beds.

Black and dark-gray shales serve as the primary basin-wide stratigraphic markers and are utilized as Frasnian formation and group boundaries across New York and throughout the Appalachian Basin (Rickard, 1975; Woodrow, 1985; Woodrow et al., 1989). The shale-bounded rock units, thus defined, enclose sequences of coarser-grained strata, which are also recognized in the stratigraphic terminology. Localized compaction and tectonic effects notwithstanding, the black and dark gray shales are thought to represent a deepening of basin water with resultant transgression of the shore.

Unlike the black shales, thin carbonates have restricted distributions within the basin, but whenever developed they make excellent stratigraphic markers and yield fossils of primary importance to biostratigraphy. Of lesser importance, or of indeterminate value as stratigraphic markers in this sequence, are distinctive

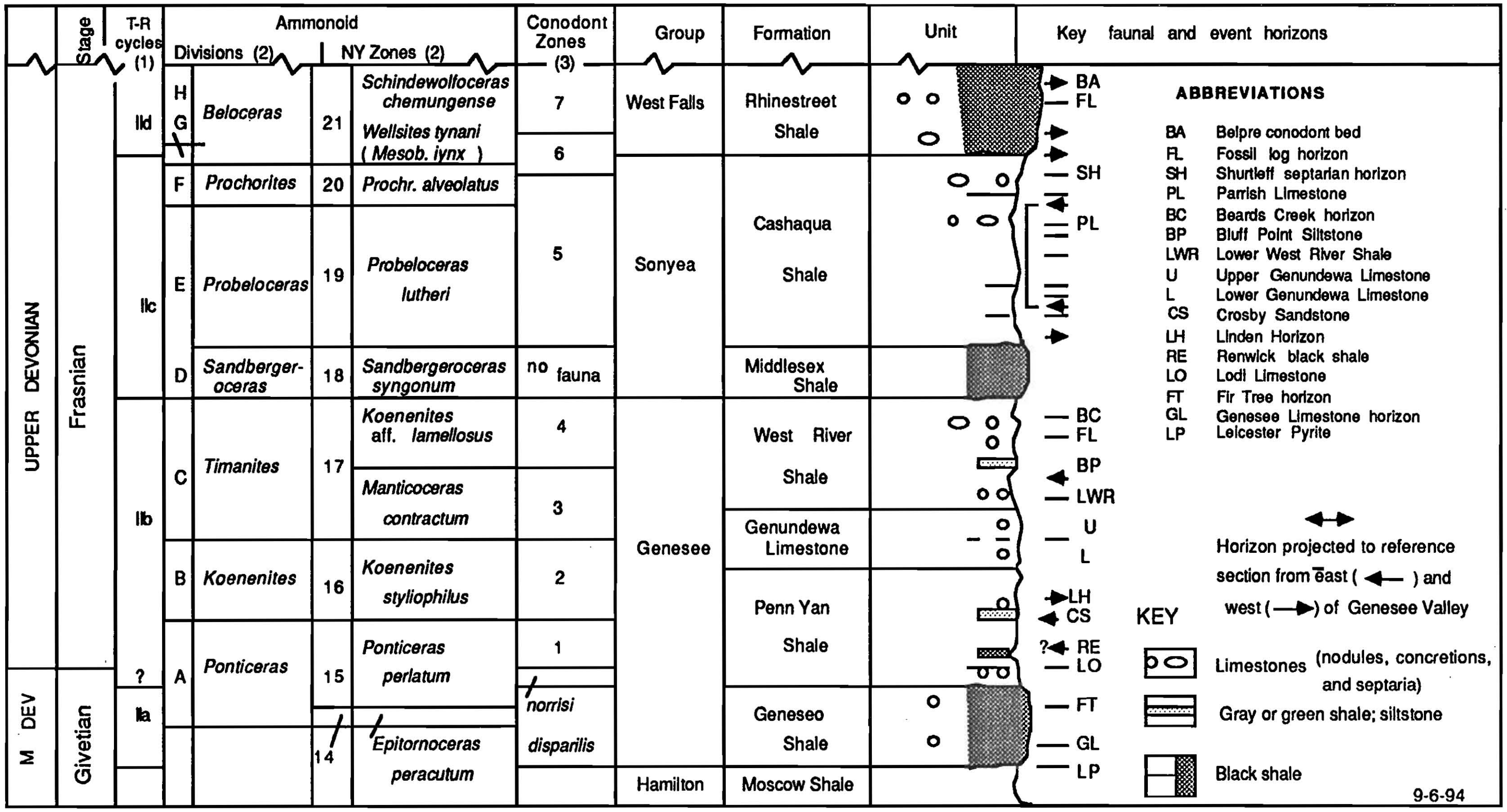

Figure 3. Idealized lower Frasnian section in Genesee Valley based on a composite of sections between the Finger Lakes and Lake Erie. 1. Transgressive-regressive cycles (T-R cycles) from Johnson et al. (1985); 2. Ammonoid (goniatite) sequence follows House and Kirchgasser (1993); 3. MN (Montagne Noire) conodont zones of Klapper (1989; et seq).

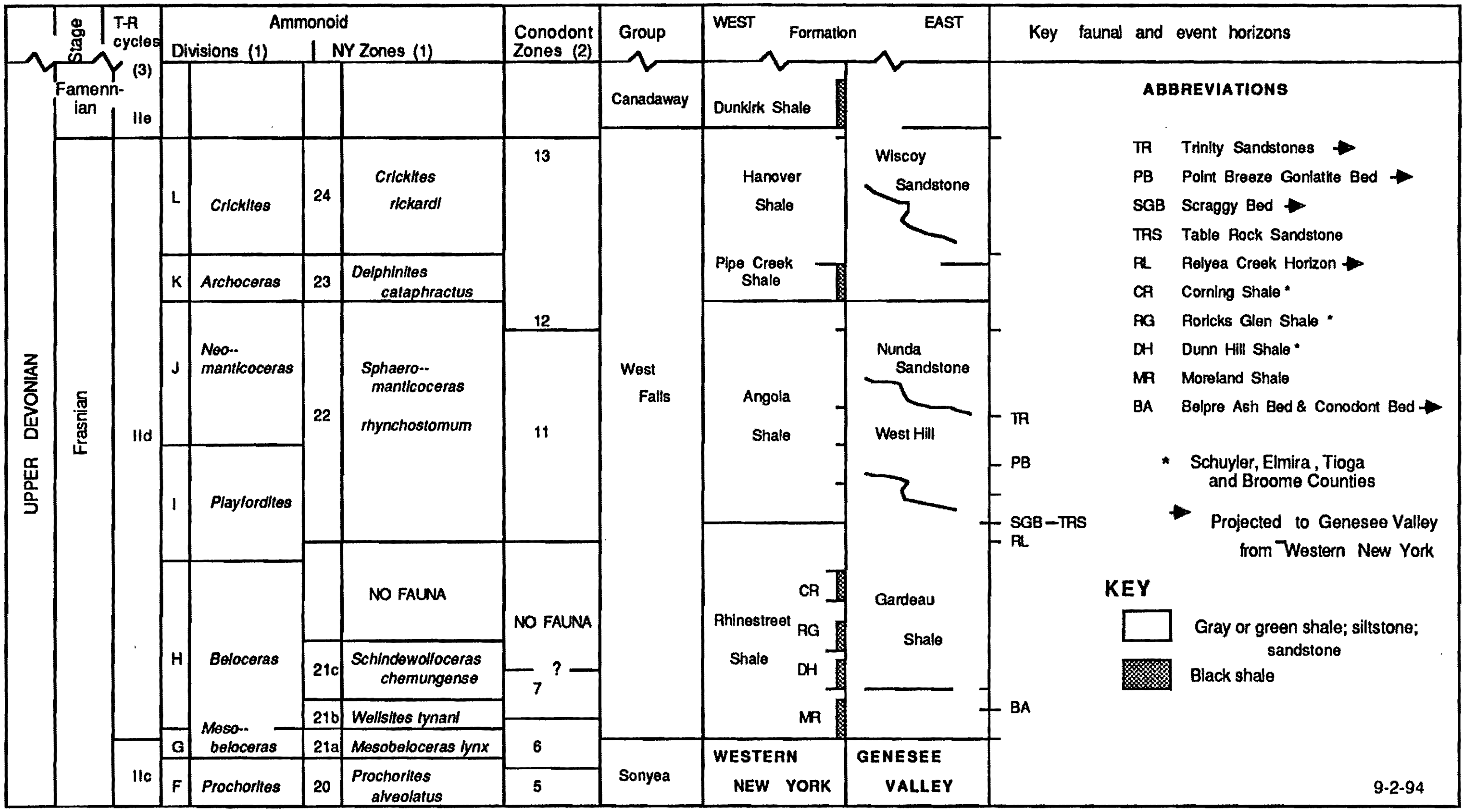

Figure 4. Idealized upper Frasnian section in western New York based on a composite of sections between the Finger Lakes and Lake Erie. Sources as in Fig. 3.

locally traceable siltstones or sandstones and volcanic ash beds. The Bluff Point Siltstone, a single stratum characterized by clear cross-lamination, much of it convolute, was originally recognized by Sutton and Lewis (1966) and correlated within the valley of Keuka Lake. Later, Colton and deWitt (1978) correlated the Bluff Point into the Genesee Valley. Frasnian ash beds, although well known from the surface and subsurface in other states (e.g., Belpre and Chimney Hill in the Chattanooga Shale; Roen, 1980; Roen and Hostermann, 1982) have not been identified with certainty in New York State. The Belpre Ash Bed is apparently represented in the Rhinestreet Shale at Stop 2, the Mount Morris Dam (Wahler, 1984; Levin and Kirchgasser, 1994).

Unique rock types and erosion surfaces

The entire late Devonian sequence thins markedly across western New York (Figure 3) and erosion surfaces developed within the sequence are more pronounced in that direction. Some of these surfaces are marked by phosphate nodules, quartz pebbles, pyritized shells and burrow fills, fish bones, inarticulate brachiopods (orbiculoids and lingulids), conodonts, and various features of erosion and dissolution typical of dysoxic and anoxic environments (Baird and Brett, 1986a, 1986b; Baird et al., 1989; Brett and Baird, 1982). The most prominent example is the Leicester Pyrite which is developed on an erosion surface west of the erosional edge of the Tully Limestone (Figure 5).

SEA-LEVEL CHANGES IN THE LATE DEVONIAN

The cyclic alternation of facies so characteristic of the Genesee Valley Upper Devonian is taken to represent variations of water depth. Johnson et al. (1985) proposed the most comprehensive hypothesis of Devonian sea-level fluctuations. House and Kirchgasser (1993) provide a more recent consideration of the evidence for sea-level fluctuations within the New York Frasnian (Figures 3, 4, 8).

Johnson et al. (1985) recognize within the Upper Devonian of New York cyclic deposition at three scales: 1) long-term "depophases" which span more than a stage; 2) shorter-lived, shallowing-upward, transgression-regression cycles ("T-R cycles") which span 1 to 4, or as many as 14, conodonts zones; and 3) PAC-like, small-scale, 1-5 m thick T-R cycles. In the view of Johnson et al. (1985) the cyclicity in Frasnian strata cannot be a response to glacially-induced eustatic sea-level changes because there is no evidence for glaciation at that time. Instead, the large-scale depophases are thought to reflect changes in the volume of the ocean basins caused by the "...growth and decay of oceanic ridge systems." The smaller scale T-R cycles are thought to reflect " ...mid-plate thermal uplift and submarine volcanism..." The apparent synchroneity of transgressions over parts of North America and western Europe is considered in the analysis to rule out the local effects of tectonics, basin subsidence, compaction and climate.

In New York State the Geneseo, Rhinestreet, and Pipe Creek black shales, as well as the Genundewa Limestone, represent the deepening phase of major T-R

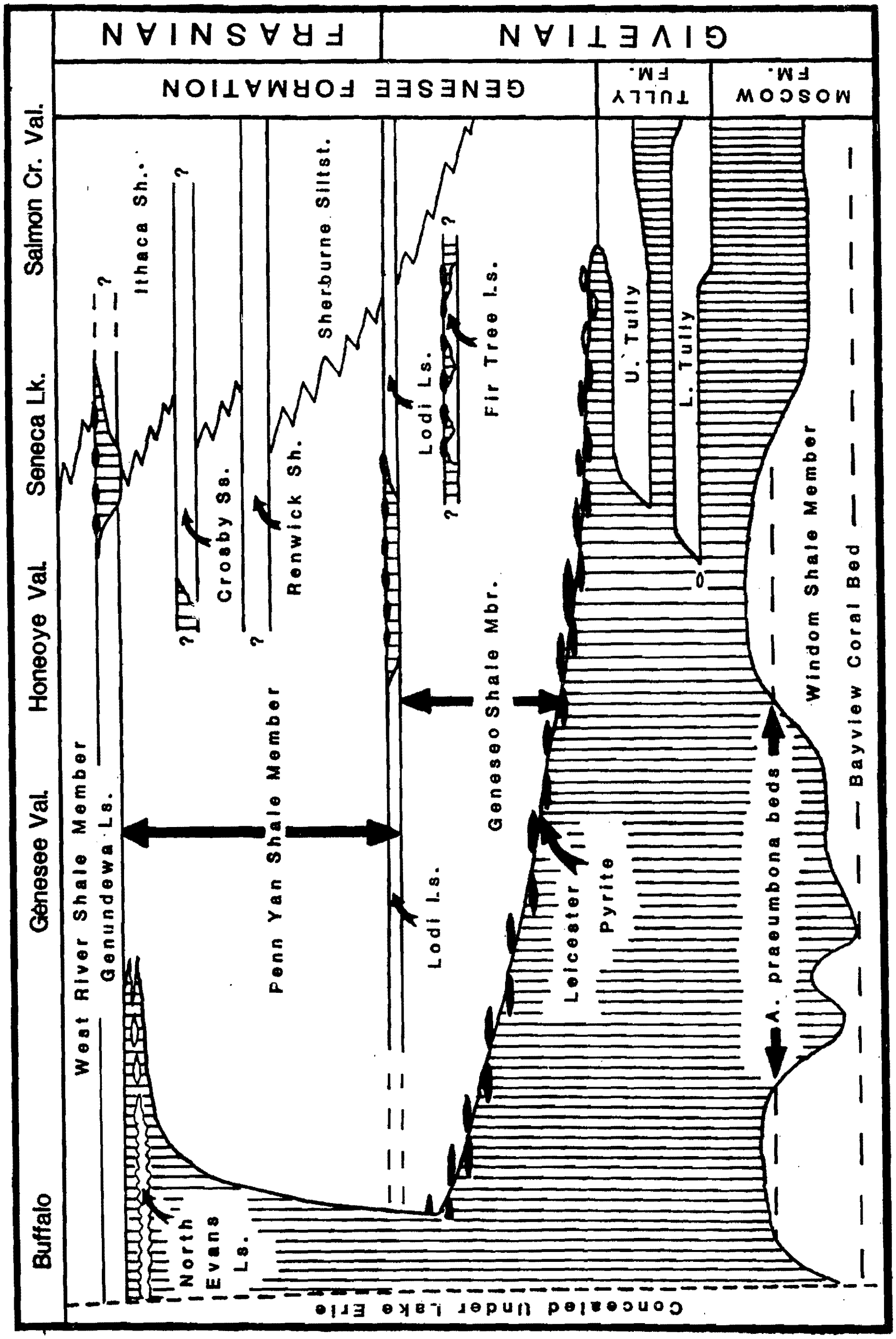

Figure 5. Cross-section of lower Genesee Group showing sub-Genesee Taghanic Unconformity. Lenses of detrital Leicester Pyrite record this long period of submaine erosion prior to burial during the Taghanic onlap of the black muds of the Geneseo Shale. A similar black-shale roofed discontinuity above the Lodi Limestone is at the Givetian-Frasnian (Middle-Upper Devonian) boundary to the east of the Genesee Valley (Kirchgasser et al., 1989). From Baird et al. (1989) and based on Brett and Baird (1982) and Baird and Brett (1986a).

cycles. The coarser clastics within each cycle are the shallowing or fill-in phase (House and Kirchgasser, 1993). The relatively sharp basal contacts of the shales may indicate that deepening (with its attendant shoreline transgression) occurred relatively rapidly while the gradational tops of the black shales indicate relatively slower shallowing as clastics were introduced to the deeper water. Minor T-R cylces are based on the Middlesex and Renwick black shales.

In the cycles based on black shales, it is inferred that deepening brought the pycnocline up the clinoform/ramp and on to the shelf, thus spreading poorly oxygenated or anoxic water across sediment interfaces which were previously relatively well-oxygenated. Benthic faunas were greatly reduced in variety and number, or were displaced entirely. Along the shore, transgression trapped clastics in newly established estuaries or along streams in which gradients had been reduced, effectively cutting off the introduction of clastics to the clinoform/ramp and deeper basin.

Not every episode of deepening resulted in broadly extensive black shales as demonstrated by the development of the Genundewa Limestone, other styliolinid limestones in the Genesee Group, and a few scattered limestones that contain a pelagic fauna higher in the Frasnian [e.g., pre-Middlesex Beard's Creek septarian horizon (BC) and the pre-Rhinestreet Shurtleff Septarian horizon (SH)]. In situations when clastics were cut off from the shelf and clinoform/ramp, and a pelagic fauna flourished, a limestone developed. The formation of carbonates during an episode of deepening (instead of black shale) suggests cut-off of clastics to the basin and the enhanced production of pelagic fauna, perhaps due to upwelling. The limestones are condensed strata that contain evidence of erosion, reworking of shelly materials, and scattered phosphate nodules and lag accumulation of conodonts.

In the terminology of sequence stratigraphy, the Taghanic Onlap of the Genesee Group onto the pre- and post Tully unconformity (Figure 5) is clearly a first order sequence boundary. However, it is not clear how the multi-scale cycles higher in the succession fit into the hierarchy of sequence cycles (the numbered orders). The regional episodic influx of turbiditic silt and sand and the erosive effects of bottom currents and submarine dissolution all would serve to disrupt the accumulation of uniform and symmetrical sequences that might be related to a particular scale of eustatic cyclicity.

BIOSTRATIGRAPHY

Conodonts and goniatite (ammonoid) cephalopods are the principal groups for age determination of marine rocks of the late Devonian age, and both groups are well represented in the New York section. Most of the key goniatites were described by Hall (1843, 1879) and Clarke (1899) and re-illustrated by Miller (1938; see also Linsley, 1994). Modern work dates from House (1962), and the sequence is outlined in Kirchgasser and House (1981) and House and Kirchgasser (1993). Modern work on conodonts in late Devonian strata of New York begins with Huddle (1968, 1981). The sequence is reviewed in Klapper

(1981), and an outline of recent collaborative work with Gilbert Klapper (University of Iowa) is presented here. Work on other important biostratigraphic groups [e.g., miospores, dacryoconarids (styliolines, nowakiids)] is still in a preliminary stage (see Woodrow et al., 1989).

Faunal horizons and "event beds"

While fossils are scattered through the various marine facies of the New York Frasnian, the sources of most identifiable goniatites and conodonts are the "event beds" identified in Figures 2-4. Key fossil beds include the thin styliolinid limestones, argillaceous limestones, concretionary (septarian) or nodular bands, or calcareous siltstones and sandstones. The offshore styliolinid bands (e.g., Genundewa Limestone), which yield datable faunas from the pelagic conodont biofacies, are condensed levels condensed (as well as current-reworked) believed to have accumulated during times of low clastic influx. By comparison, the concretion, septarian, and nodular carbonate horizons may have accumulated in somewhat shallower water during times of "normal" sediment influx. The calcareous siltstones and sandstones are distal siliciclastics carried down the slope and into the basin. The black shales that represent the deepest facies are mostly devoid of macrofossils, but the transgressive lower boundaries of some black shales overlie discontinuity horizons of carbonate dissolution and lag-concentrations of reworked pyrite (including goniatite interiors and burrows), phosphate and quartz pebbles, fish scales and bone, inarticulate brachiopods (orbiculoids, lingulids) and conodonts. The major basal Genesee disconformity is the Leicester Pyrite, similar but thinner "cryptic" horizons occur through the succession.

New York and Frasnian Composites

The idealized sections for the Frasnian of the Genesee Valley illustrated in Figures 3 and 4 are composites of the sequence of units and "event beds". The sections are aligned to the Montagne Noire (MN) conodont zonation of Klapper (1989, et seq.) and are correlated to the global ammonoid divisions following House and Kirchgasser (1993).

For the lower Frasnian of New York, the lithostratigraphic (black shale boundaries and other "event beds") and biostratigraphic data (ranges of conodonts and goniatites) from 20 sections have been compiled into a single New York Composite by the method of graphic correlation based primarily on the positioning of key beds (Klapper and Kirchgasser, 1992). The section in the Genesee Valley at Beard's Creek, Leicester, N.Y., between the base of the Geneseo black shale and the base of the Middlesex, was selected as the reference section for the lower Frasnian. The resulting New York biostratigraphic composite is a compilation of the ranges of 28 conodont species and 24 ammonoid genera and species. The New York Composite in turn has been correlated by graphic correlation into a developing Frasnian Composite that consists of data from some 27 sections from around the world (Klapper et al., 1993; in press). The ranges of key zonal conodonts from the lower Frasnian in the New York and the Frasnian Composites are illustrated in Figure 6; note that the ranges in New York (white bars) mostly fall "within" the ranges in the Frasnian Composite, reflecting the obvious "facies control" of occurrences of conodonts and goniatites in New York State. Construction of a New York Composite for the New York upper Frasnian (Rhinestreet to Dunkirk) is in progress. Some of the characteristic goniatites of the Genesee, Sonyea and West Falls groups are illustrated in Figure 7. Many of the key goniatite horizons also yield datable conodonts. The T-R cylces shown in Figures 3 and 4 were aligned by Johnson at al. (1985) by conventional correlation to the conodont zonal scheme proposed by Ziegler (1962) and most recently revised by Ziegler and Sandberg (1990).

Genesee Group

In the Genesee Valley section, late Givetian conodonts occur in the Leicester Pyrite (with the goniatite *Tornoceras uniangulare*) and at levels within the Geneseo Shale associated with poorly preserved *Ponticeras* (Huddle, 1981; Figures. 3, 10). In the Lodi Limestone in the lower Penn Yan Shale, *Ponticeras perlatum* (Figure 7) and rhynchonellid brachiopods occur with *Skeletognathus norrisi,* the indicator of the latest Givetian *norrisi* conodont zone (Klapper and Johnson, 1990). In the Genesee Valley, the Lodi horizon projects by graphic correlation into Montagne Noire (MN) Zone 1 of the Frasnian (Figure 6). East of the Genesee Valley, MN Zone 1 conodonts (*Ancyrodella rotundiloba* early form) occur in black shales immediately above the Lodi (Kirchgasser et al., 1989; Kirchgasser, 1994; Figure 5). In the Genesee Valley, MN Zone 1 conodonts occur in the SB black shale higher in the Penn Yan Shale, and still within the range of *Ponticeras* (Stop 1: Fall Brook and Dewey Hill; Figure 10).

MN Zone 2 conodonts with *Ancyrodella rotundiloba* (late form) enter in the middle of the Penn Yan Shale. The goniatite *Koenenites styliophilus* first appears near the level of the lowest entry of *Acanthoclymenia.* The key horizons are a styliolinid band, locally with white barite, called the Linden Horizon (LH), and the equivalent or approximately equivalent Crosby Sandstone (CS) around Keuka Lake. The characteristic bivalve in the dark Penn Yan Shale and West River Shale is *Pterochaenia fragilis.*

The Genundewa Limestone is an important key bed in the middle of the Genesee Group, and this prominent condensed styliolinid limestone is thought to record a short-lived deepening event of possible eustatic origin (Figure 8). The upper Genundewa marks the entry of the MN Zone 3 conodonts *Ancyrodella rugosa* and *Ad. alata,* and a distinctive group of short ranging species of *Ancyrodella* represented by *Ad.* sp. B (Kralick, 1991, in press). The upper Genundewa also marks the entry of *Manticoceras* and *Acanthoclymenia genundewa* (Figure 7).

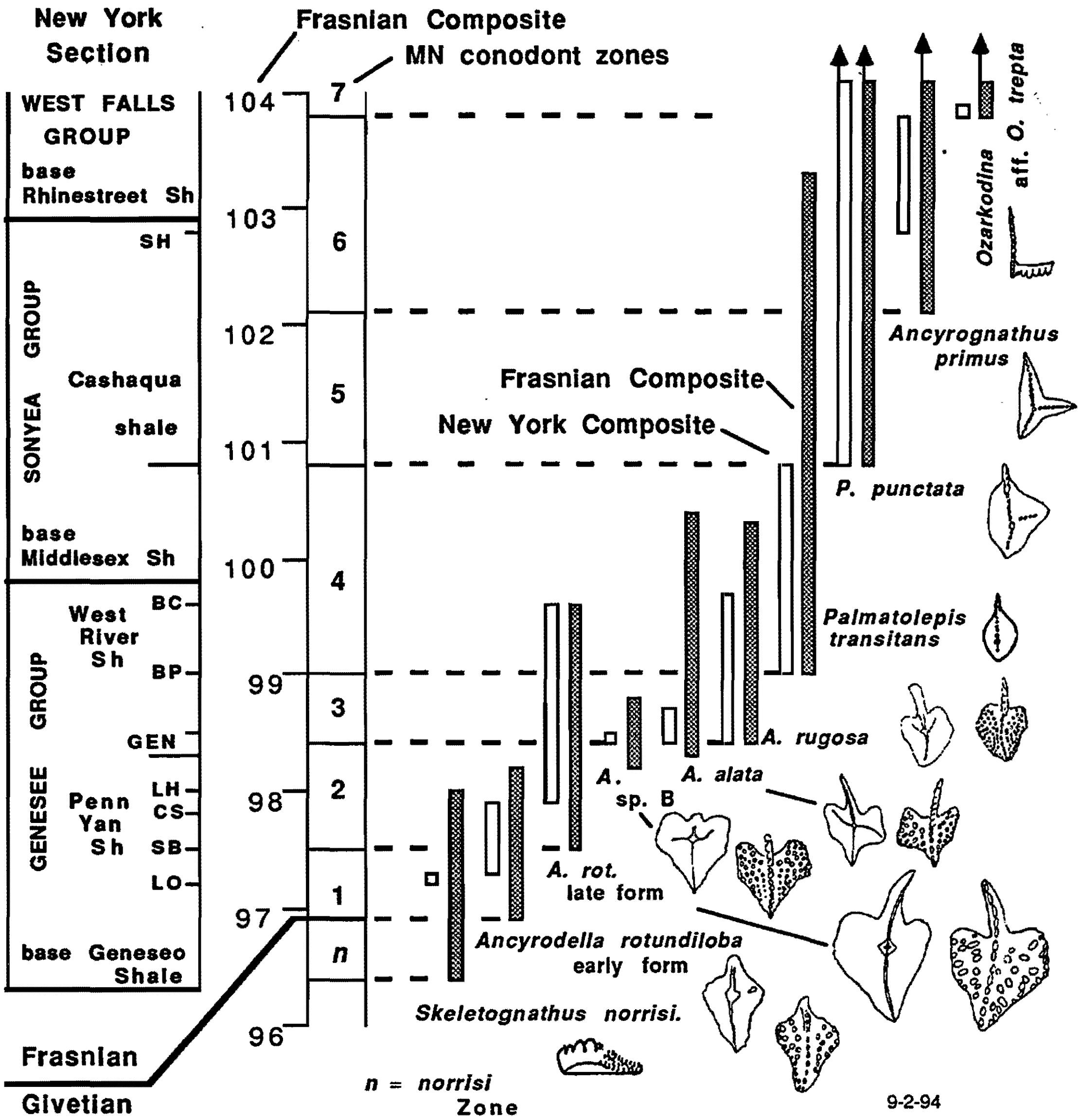

Figure 6. Correlations of lower Frasnian conodont sequence in New York with Montagne Noire (MN) conodont zones (Klapper, 1989; et seq.) and Frasnian Composite (Klapper et al., 1993; in press). The New York seqence is a composite of range-data from 20 sections compiled by graphic correlation (see text). The Frasnian Composite is compiled by graphic correlation of 27 sections from the around the world. Note that in most cases the species ranges in New York fall within the ranges in the Frasnian Composite. The distibution of New York species are strongly controlled by facies, and the ranges are restricted because of the frequent facies shifts.

MN Zone 3 conodonts continue into the lower West River Shale. Faunas are characterized by *Ancyrodella alata* (Figure 10). In the middle West River Shale conodonts of MN Zone 4 (with *Palmatolepis transitans*) enter just below the Bluff Point Siltstone (BP) in a bed with *Koenenites* aff. *Ko. lamellosus*. The MN Zone 4 fauna also occurs in the Beards Creek horizon (BC) near the top of the West River Shale in concretions with species of *Koenenites, Acanthoclymenia* and *Manticoceras* (House and Kirchgasser, 1993). Huddle (1981) illustrated many specimens of conodonts from this level at Keuka Lake.

Sonyea Group

Conodonts are unknown in the black Middlesex Shale and only one ammonoid species has been identified. MN Zone 5 with *Palmatolepis punctata* begins in New York at the base of the Cashaqua Shale. *Probeloceras lutheri* and *Manticoceras sinuosum* are present at several levels within the unit and will likely be seen at Stop 2 (Figure 7).

The facies sequence of the Sonyea to basal Rhinestreet is strikingly symmetrical and follows the pattern ABCBA, with A being black shale (Middlesex and Rhinestreet); B, dark gray shales interbedded with thin black shales (lower and upper Cashaqua); and C, green shale and mudstone (middle Cashaqua). The B facies and distinctive olive green shale and mudstone facies of the middle Cashaqua (facies C) have argillaceous concretions, burrowed horizons, and a moderately diverse molluscan fauna, sometimes with large bivalves (*Lunulicardium*). This fauna and sedimentological changes suggest the middle Cashaqua records a shallowing event (Figure 8). In the upper Cashaqua, fissile dark gray shales (facies B), like those of the lower Cashaqua, recur. This dark shale suggests renewed deepening that precedes the major highstand recorded by the lower black shale of the overlying Rhinestreet Shale (Moreland Shale). The lower Rhinestreet may be the deepest water facies in the entire section (Figure 8).

The styliolinid and baritic Shurtleff Septarian horizon (SH) within the upper dark shale facies of the Cashaqua is the source of *Ancyrognathus primus,* the zone indicator of MN Zone 6, although by graphic correlation with the Frasnian Composite, MN Zone 6 aligns to a position lower in the Cashaqua (Figure 6). In addition to its rich conodont fauna of *Palmatolepis punctata* and *Ancyrodella nodosa*, the Shurtleff Horizon locally contains a diverse molluscan fauna. Cephalopods, preserved in pink and white barite, were described by Clarke (1899) and include *Manticoceras sinuosum* and *Acanthoclymenia neapolitana*, and the distinctive *Prochorites alveolatus* with its concave ventral margin (Figure 7). *Probeloceras lutheri* and *Prochorites alveolatus* occur in western Australia, aligned with the same conodont sequence as in New York (Klapper and Kirchgasser, 1992; Becker et al., 1993; Figure 8.).

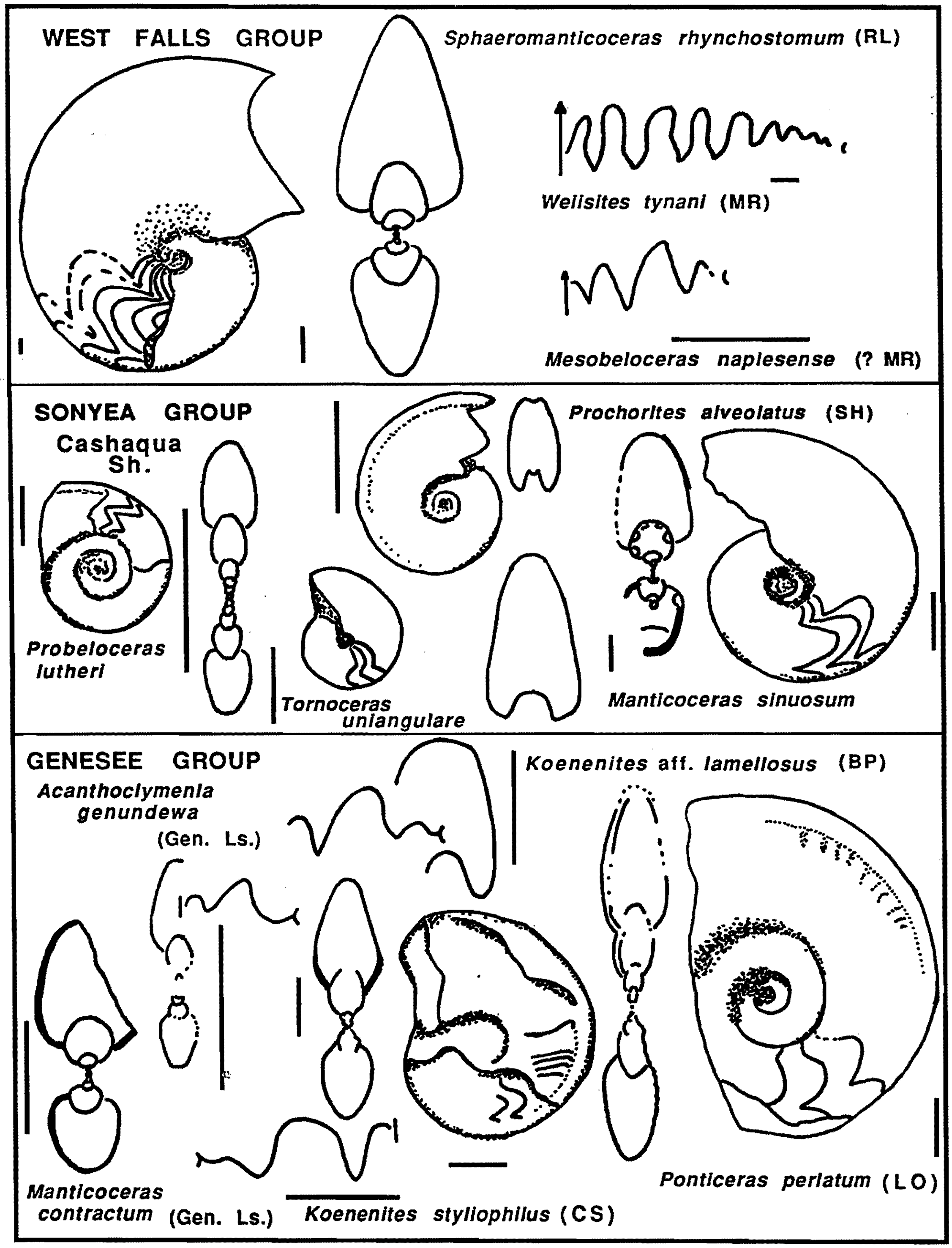

Figure 7. Characteristic goniatites in the Genesee, Sonyea and West Falls groups. Abbreviations of horizons as in Figs. 3 and 4. Bar scale is 1 cm. Modified from Kirchgasser (1973) with revisions based on House and Kirchgasser (1993).

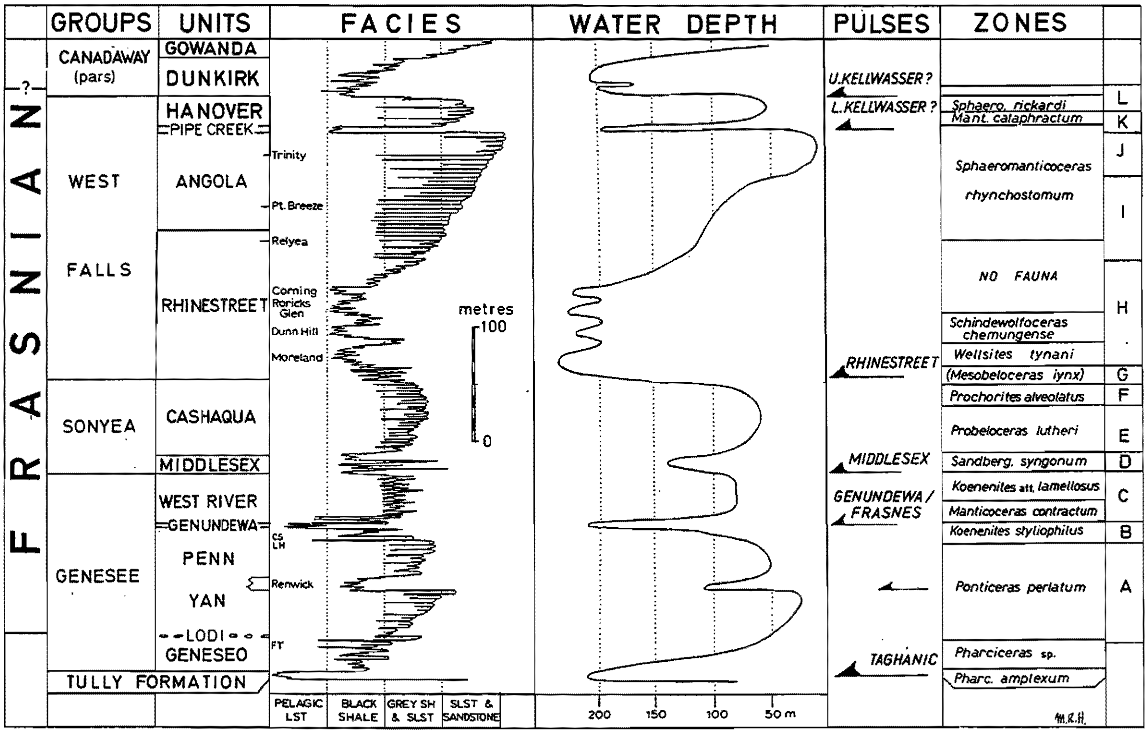

Figure 8. Facies shifts and major sea-level changes in the New York Frasnian. The major transgressive pulses are aligned to the goniatite divisions and zones illustrated in Figs. 3 and 4. From House and Kirchgasser (1993).

West Falls Group

Conodonts recovered from the base and lower part of the Rhinestreet black shale suggest that the Rhinestreet deepening begins in the interval of MN Zone 6. MN Zone 7, marked by the entry of *Ozarkodina* aff. *Oz. trepta*, begins about 6 meters above the base of the Rhinestreet at Cazenovia Creek in western New York. An MN Zone 7 fauna occurs 8 meters above the base at the Lake Erie shore, about a meter below the Belpre Ash Bed (Levin and Kirchgasser, 1994). These Zone 7 occurrences correspond to the Moreland Member of the Rhinestreet Shale.

Identifiable goniatites are rare in the black and gray shales of the Rhinestreet, and only a few crushed specimens are known in the Genesee Valley. However, to the southeast toward the Finger Lakes and the Southern Tier, there are records of distinctive evolute and multi- lobed genera and species from just below, within, and just above the Moreland Shale, as well as from middle levels within the Rhinestreet (House and Kirchgasser, 1993; Figures 4, 7).The occurrence of these cosmopolitan "belocerids," known elsewhere in the Old World and Australia, suggests that the Rhinestreet records a major eustatic deepening.

Near the top of the Rhinestreet in the Oatka Creek Valley (Warsaw, N.Y.), the Relyea Creek concretionary horizon (RL) contains a MN Zone 11 fauna and the large manticocerid *Sphaeromanticoceras rhynchostomum* (Figure 7). Based on conodonts from over- and underlying strata the middle and upper Rhinestreet interval corresponds to MN Zones 8, 9 and 10, but no faunas of these zones have been found. The top of the Rhinestreet in western New York is the pyritic-concretionary Scraggy Bed. A black shale beneath that horizon is believed to project into the Genesee Valley section to about the level of the Table Rock Sandstone, the siltstone bed that caps the Table Rock observation platform in Letchworth Park near the Lower Falls (Kirchgasser, 1973; Kirchgasser and House, 1981; Figure 2). This correlation indicates that the Gardeau Shale below the Table Rock correlates westward with the Rhinestreet, and that the Gardeau above the Lower Falls correlates westward with the Angola Shale (sections in Pepper et al., 1956). The section at Stop 3 is within the Gardeau Shale in strata equivalent to the Angola.

In the Genesee Gorge at Letchworth Park, the base of the 107 foot high Middle Falls is in the upper Gardeau Shale, and the section from the top of the Middle Falls to the top of the 70 foot high Upper Falls is within the West Hill Shale and Sandstone (Figure 2). The massive beds of the Nunda Sandstone (the Portage Sandstone of earlier workers) begin at the top of the Upper Falls. In western New York, the Angola Shale equivalents of the West Hill and Nunda have concretion horizons that contain Zone 11 conodonts and goniatites of the *Sphaeromanticoceras rhynchostomum* Zone. Above the Nunda, in a tongue of the upper Angola at Varysburg, MN Zone 12 conodonts have been recovered in a nodular (knollenkalk) horizon associated with burrowed shales. The Angola Shale and equivalent Nunda Sandstone in western New York is succeeded by the black Pipe Creek Shale, which, in turn, is overlain by the green-gray Hanover Shale (Figures 2, 4). Over (1992, 1994) has recovered conodonts of MN Zone 13 (the highest Frasnian Zone) and Lower and Middle *triangularis* Zone (lowest Famennian) in the upper Hanover Shale, several meters below the base of the black Dunkirk Shale, the horizon where the Frasnian/Famennian boundary has traditionally been drawn.

At the south end of the Genesee Valley towards Dansville, the Pipe Creek Shale separates the Nunda Sandstone from the overlying Wiscoy Sandstone, the coarse-grained (sandstone) equivalent of the Hanover Shale. The section on I-390 near Cohocton (Stop 4) is believed to be close to the Nunda-Wiscoy boundary. Conodonts of the *Polygnathus* biofacies characteristic of Zone 12 or younger have been recovered from a channel-fill, but none of the indicator species of the MN zones have been recovered. The brachiopod fauna is also typical of Mn Zone 12 (Day, pers. comm, 1994). The goniatites recovered so far from the channel fill bed are small and not well enough preserved to be compared to any of the late Frasnian zone fauna (Figure 4).

SUMMARY

The predominantly marine siliciclastic Frasnian strata exposed in the Genesee River Valley record the fine-grained offshore filling of the central Appalachian Basin. Cyclic sequences consist of black shale that grade upward into lighter colored shale and coarser clastics that record large- and smaller scale regional and eustatic sea level fluctuations and westward progradation of the Catskil Delta. Conodonts and cephalopods from key marker horizons allow regional and global correlation of Frasnian strata, as well as timing of cyclic sequence events.

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ROAD LOG

(Begin at I-390 South, Exit 8 (Geneseo), stop sign at US 20A; distances in miles)

00.0	0.0	US 20A West.
03.1	3.1	Village of Geneseo.
04.9	1.8	Junction Rt. 39, SUNY-Geneseo to right, continue around sharp left turn on US 20A.
05.9	1.0	Rt. 63 and US 20A/Rt. 39 split, turn right and continue on US 20A/Rt. 39. The highway was closed in March 1994 due to road and bridge damage west of the Genesee River near Cuylerville that resulted from roof collapse of the Retsof salt mine and subsequent subsidence and formation of sink holes. The salt beds are approximately 300 m (1000 ft) below the valley floor in the Upper Silurian Vernon Formation.
06.6	0.7	Cross Fall Brook and turn left into alfalfa processing plant, turn around, recross Fall Brook.
06.7	0.1	Pull off road and park at uphill (east) end of guard rail. Walk down path, cross fence, and follow paths or stream up to high banks and falls.

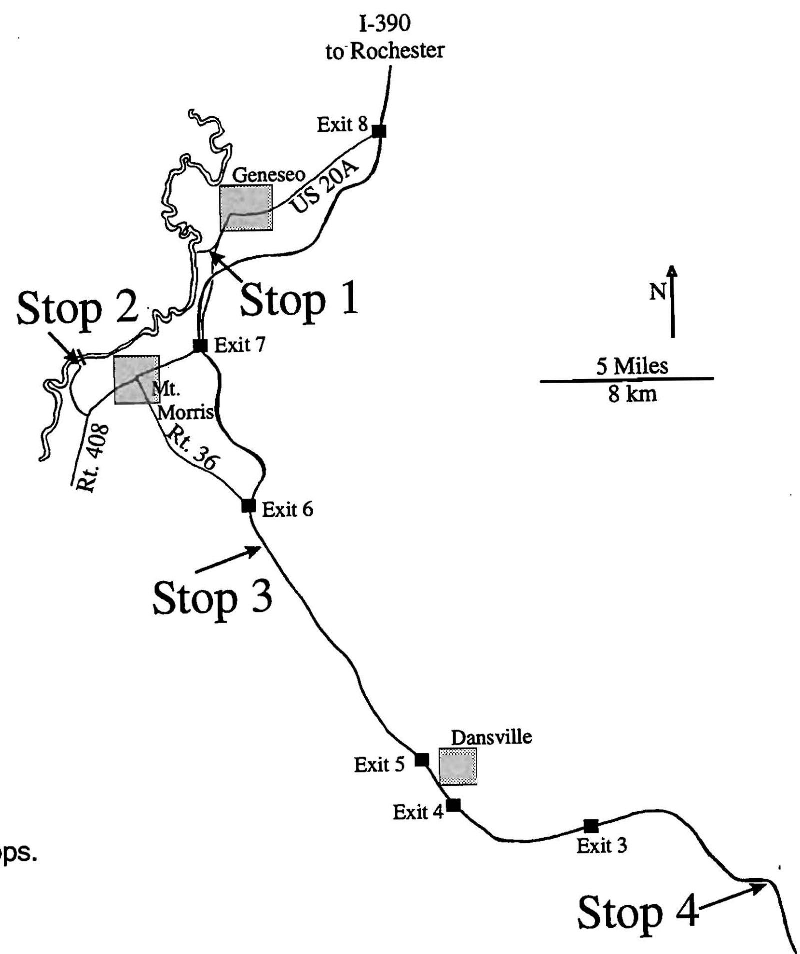


Figure 9. Map of field-trip stops.

Stop 1A - Fall Brook (Property is posted, access by permission only.): Erosional base of Genesee Group, pyritic lag, dark colored petroliferous shales, interbedded carbonates, Givetian-Frasnian boundary, and major deepening of the basin represented by planktic-fauna carbonates. Leicester Pyrite, Geneseo Shale, Penn Yan Shale, Genundewa Limestone (Figures 9, 10).

Strata of the Moscow Formation (Hamilton Group) are exposed in the stream bed and banks below the falls. Fossiliferous bluish-gray shales of the Kashong Member are overlain by bioturbated fossiliferous medium gray calcareous shales and carbonate beds of the Windom Member (Figure 10). The steep banks prevent close investigation of Genesee strata. Stratigraphic relationships and fallen debris can be examined in the gorge, in place collections can be made from a side creek or Stops 1B (Dewey Hill) and 1C (above falls) where strata are more accessible. **Caution: large blocks have been known to fall from overhangs and strike eminent geologists.**

The Genesee Group consists of the Leicester Pyrite, Geneseo Shale, Penn Yan Shale, Genundewa Limestone (resistant bed forming falls), and West River Shale. The base of the Genesee Group is marked by the Leicester Pyrite, a 0-20 cm thick sharp based accumulation composed almost entirely of pyrite nodules, pyritized burrows, and pyrite replaced fossils that represents a condensed lag accumulation above a major disconformity. The bed contains a mixed conodont and cephalopod fauna of Hamilton, Tully, and Genesee nature (Figure 5). The unconformity is the result of submarine erosion, corresponding to the Taghanic Onlap (Baird and Brett, 1986a). Look for pyritized cephalopods (*Tornoceras uniangulare*), crinoids, bivalves, burrow fills and rare brachiopods and trilobites in fallen blocks of the Leicester, which is distinctive due to the orange-brown weathered surface.

The Geneseo Shale gradationally overlies or is locally interbedded with the Leicester Pyrite. Here at Fall Brook the Geneseo is 8.2 m thick (base of lower black shale to top of upper black shale), characterized by a lower 1.7 m thick and an upper 1.4 m thick dense black petroliferous shale units separated by 5 m of medium-dark gray shale and thin carbonate mudstone beds. [In deWitt and Colton (1978) only the lower black shale is considered to be the Geneseo.] The Geneseo black shale thickens eastward, absent at Lake Erie (possibly represented as a parting between the North Evans Limestone and Genundewa Limestone) to about 30 m thick in the Finger Lakes (Figure 2).

The continuous limestone band in about the middle of the gray Geneseo shale interval is the Genesee Limestone (GL), with *Pharciceras* and a *disparalis* Zone conodont fauna that includes *Polygnathus linguliformis* gamma and *Po. dubius*. The Fir Tree Limestone, represented here by a thin carbonate mudstone bed 1.2 m above the Genesee Limestone, is a distinctive horizon in the Finger Lakes region that contains a pyrite-bone bed lag developed on the corroded surface of a carbonate bed (Baird et al., 1989).

The Penn Yan Shale lies above the upper thick black shale of the Geneseo and is characterized by medium-dark shale and interbedded thin carbonate

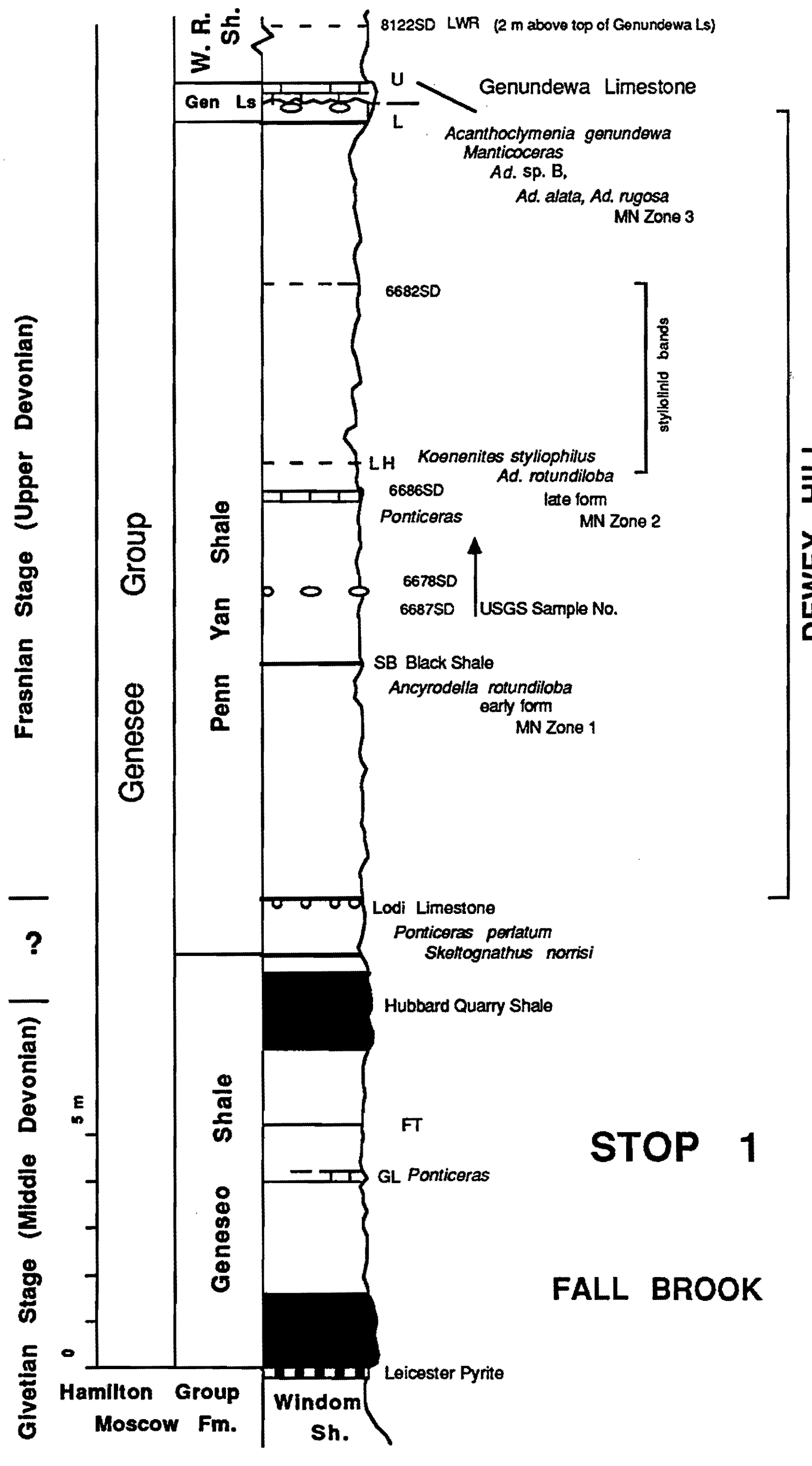


Figure 10. Section of Genesee Group at Stop 1, Fall Brook and Dewey Hill, near Geneseo, New York. Revised from Oliver and Klapper (1981). USGS Samples Numbers (SD) from deWitt and Colton (1978) and Huddle (1981).

mudstones, styliolinid packstone/grainstones, and thin black shale beds. The Lodi Limestone, a nodular carbonate 1.1 m above the base of the Penn Yan, marks the local first occurrence of *Ponticeras perlatum* and *Skeletognathus norrisi*, indicative of the *norrisi* Zone, the latest zone of the Givetian. *Ancyrodella rotundiloba* early form, which defines the base of the Frasnian and MN Zone 1 of Klapper (1989), and the Givetian-Frasnian boundary, occurs just above the Lodi Limestone east of the Genesee Valley (Figure 3, 5). Based on graphic correlation with the Frasnian Composite Standard of Klapper et al. (1993), however, the Givetian-Frasnian boundary would lie in the Geneseo-Penn Yan transition (Figure 6). The Lodi Limestone can be collected from the side creek, and will also be seen at Dewey Hill (Stop 1B).

The Genundewa Limestone caps the water fall of Fall Brook. The Genundewa is a 0.3 to 0.5 m thick, wood-bearing, cephalopod styliolinid packstone-grainstone that consists of several amalgamated beds separated by discontinuous shale laminae. The grainstone consists almost entirely of the small conical shells of *Styliolina fissurella*, an enigmatic organism that may be a protist (Yochelson and Lindemann, 1986). The lower surface is irregular, indicating scouring of the sea floor prior to or during deposition. Reworking and erosion are indicated by current alignment of the styliolinids and corrosion/wear features. The Genundewa is interpreted as the result of deepening of the basin; consequent flooding of the terrigenous sediment source and near shore sediment trapping resulted in accumulation of planktic organisms, without dilution by siliciclastics. Several styliolinid laminae and thicker packstone-grainstones occur in the upper Penn Yan indicating cyclic deepening and consequent sediment starvation. The base of the Genundewa is recognized by a regionally persistent, thin (10 cm) black shale bed that underlies the prominant carbonate band. The Genundewa will be viewed in place at Stop 1C; however, samples can be easily collected from large slabs in the stream bed.

06.7 0.0 Proceed up hill on US20A/Rt. 39.
07.0 0.3 Pull off and park before guard rail on right. Exposures are in ditch and cuts on both sides of the road; the section described is on the south side.

Stop 1B - Dewey Hill: Penn Yan Shale including Lodi Limestone Member, SB bed, and Linden Horizon.

Here we will be able to closely examine beds below the Genundewa Limestone in the side walls of Fall Brook at Stop 1A. The Penn Yan Shale is comprised of dark to medium gray shale with interbedded carbonate and black shale. The Lodi Limestone, exposed in the road ditch, is the lowest key bed. The Lodi consists of thin (5 cm) discontinuous beds and nodules that are traceable to the Finger Lakes region and represents the latest Givetian (Middle Devonian) in New York State; *Ponticeras* is the predominant goniatite, but it is rare this far west. The Givetian-Frasnian (Middle-Upper Devonian boundary) is taken as the base of the shales overlying the Lodi, recognized by the first occurrence of *Ancyrodella* *rotundiloba* early form, which is used to define the base of MN Zone 1 (lowest Frasnian). Gastropods and rhynconellid brachiopods may be found in the nodular Lodi at this locality.

The SB bed is a 20 cm thick distinctive black shale interval that is characterized by a conodont and lingulid lag horizon in the lower few centimeters. The horizon interval represents a phase of benthic anoxia-dysoxia with preservation of organic material, as well as a hiatus/erosion surface and concentration of phosphatic bioclasts. Conodonts of MN Zone 1 can be recovered from the shale.

In contrast to the larger scale sequence marked by the disconformity at the base of the Genesee Group, smaller scale sequences in the Geneseo, Penn Yan, Genundewa, and West River formations (basin facies) seem symmetrical, preserving sediments that represent both the deepening and shallowing without development of a pronounced hiatus. The SB phosphate lag is an example of a cryptic disconformity within a minor cycle that marks a deepening maxima phase in the basin.

Nodular and continuous carbonate beds in the middle-upper Penn Yan contain cephalopods (genera include *Koenenites, Acanthoclymenia*, and *Tornoceras*) and conodonts. The key styliolinid bed is the Linden Horizon (LH) that contains *Koenenites styliophilus*. In this horizon the goniatite shells are often filled with crystalline pink and white barite. This level marks the base of MN Zone 2 in New York (*Ancyrodella rotundiloba* late form).

07.0	0.0	Proceed uphill (east) on US 20A/Rt. 39.
07.3	0.3	Turn right (south) at intersection onto Rt. 63.
07.8	0.5	Pull off to right and park at Fall Brook overlook.

Stop 1C - Fall Brook above falls: Genundewa Limestone and West River Shale.

Extreme Caution: the edge of the Genundewa is unmarked and the drop to the base of the falls is over 100 feet.

The Genundewa Limestone consists of approximately 35 cm of several amalgamated beds of styliolinid grainstones separated by discontinuous shale partings that can be traced from Lake Erie to Keuka Lake. The base of the Genundewa is marked by a persistent black shale bed that separates it from carbonate beds in the upper Penn Yan. Note here the irregular contacts in the Genundewa and proximal strata and apparent rippled surfaces. The styliolinids are aligned, indicating current orientation, however, the nature of the current(s) responsible is enigmatic: contourites, waves on pycnocline boundaries, storm waves, ocean gyres, or density currents are possible mechanisms.

The upper and lower Genundewa contain distinctive faunas. The lower beds are characterized by *Ancyrodella rotundiloba* late form, indicative of MN Zone 2. The upper Genundewa contains the first appearance of *Manticoceras* and conodonts of MN Zone 3, including *Ad. rugosa, Ad. alata*, and *Ad.* sp. B (Kralik, in press; Figures 3, 6).

The lower West River Shale is exposed in the creek bed and banks above the falls. The West River Shale consists of dark to medium gray shales and thin carbonate beds, lithologically similar to the Penn Yan Shale, that indicate shallowing and a return to fine clastic sedimentation following Genundewa deposition. A thin (3 cm) styliolinid carbonate bed approximately 2 m above the base of the shale contains numerous conodonts of MN Zone 3 [USGS Silurian-Devonian cat. No. 8122SD (Huddle, 1981)]. The upper West River Shale is exposed at the base of the Mt. Morris Dam at Stop 2.

07.8	0.0	Proceed south on Rt. 63.
10.3	2.5	Junction of Rt. 408, continue south on Rt. 408.
10.6	0.3	Junction I-390, continue south on Rt. 408.
12.0	1.4	Village of Mt. Morris.
12.5	0.5	Turn left at light.
12.6	0.1	Turn right at light, continue south on Rt. 408.
14.3	1.7	Mt. Morris Dam Road, continue to next intersection.
14.5	0.2	Turn right at park entrance to Mt. Morris Dam, continue to parking lot at dam overlook.
16.4	1.9	Park. Exposures are along the dam access road and "Burma Road" that descends gorge wall upstream from dam.

Stop 2 - Mt. Morris Dam (access by permission from US Corps of Army Engineers): Gorge overlook: dark gray shale of upper West River Shale at base above and below dam (river level); black shale of Middlesex Shale and green/gray shale of Cashaqua Shale, Sonyea Group; and gray/black shale of Rhinestreet Shale, lower West Falls Group to the top of the exposure (Figure 11). An apparent correlative of the Belpre Ash Bed may be visible.

The Letchworth Gorge of the Genesee River is a Pleistocene hanging valley cut into shales, siltstones, and sandstones of the Genesee, Sonyea, and West Falls groups. Three main waterfalls in the upper gorge are formed on resistant sandstones of the upper West Falls Group. The section exposed at the dam includes gray shale of upper West River Shale at the base above and below the dam (river level); black Middlesex Shale and green/gray Cashaqua Shale, Sonyea Group; and gray/black Rhinestreet Shale, lower West Falls Group up to the top of exposure. The two black shales represent major transgressions.

The lower Rhinestreet Shale and upper Cashaqua Shale are exposed in the access road to the top of the dam. The Cashaqua consists of green-gray shale and mudstone that are darker at the base, lighter colored and highly bioturbated in the middle, and darker gray at the top. This lithologic change indicates a shallowing cycle, superimposed on which there are numerous shorter-duration cycles indicated by darker shale bands and concretionary horizons. The medium gray shale of the upper Cashaqua exposed at the dam contains numerous bivalves and coalified plant fragments. The prominent concretionary band in the upper Cashaqua is the Shurtleff Septarian Horizon. These concretions contain a diverse conodont fauna of MN Zone 6 (*Palmatolepis punctata, Ancyrodella*

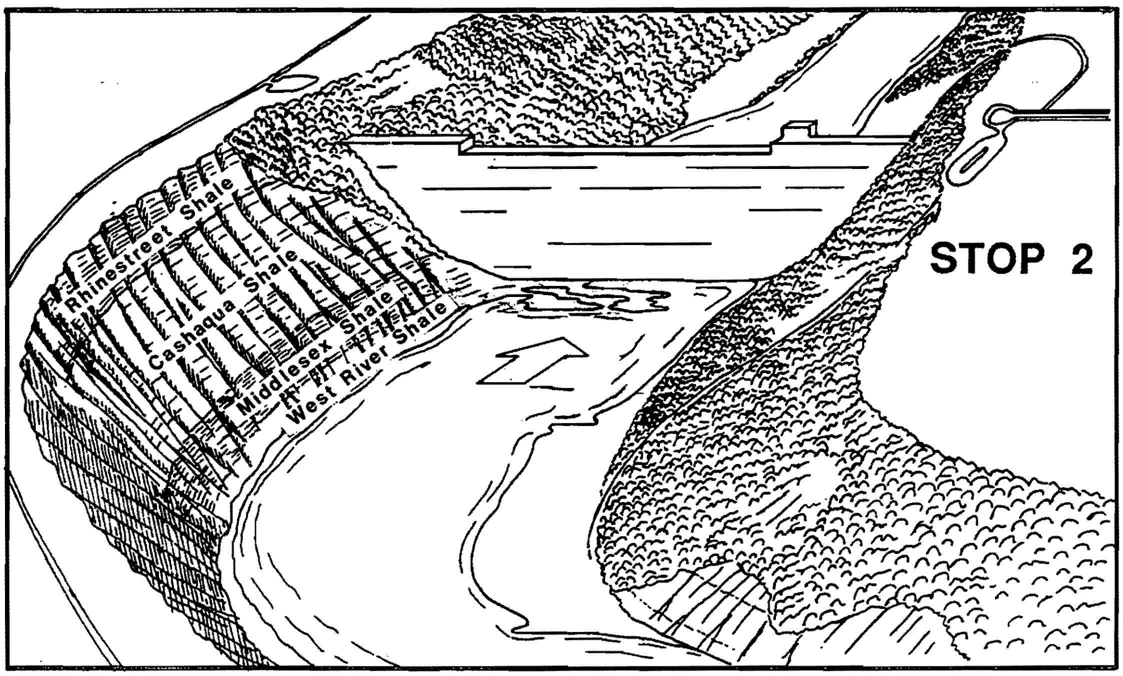


Figure 11. Schematic drawing of section of upper Genesee (West River Shale), Sonyea (Middlesex and Cashaqua Shales) and lower West Falls Group (Rhinestreet Shale) in the Genesee Gorge at Stop 2, Mount Morris Dam at Mount Morris, New York. From Van Diver (1980) with permission of the author.

nodosa), as well as mollusks replaced or filled with pink or white barite, including the goniatites *Manticoceras sinuosum* and *Prochorites alveolatus* (Figure 7; House and Kirchgasser, 1993).

The base of the Rhinestreet Shale and the West Falls Group is marked by a sharp transition from medium-to-dark gray shale below to thick black shale. The Rhinestreet is the thickest and most widely distributed black shale bed in the Upper Devonian of New York and represents a major transgression and deepening (Figures 2, 8). Plant remains, rare cephalopods, conodonts, and other fossils can be recoverd from bedding planes.

If time and weather permits the entire Cashaqua can be accessed on the "Burma Road" which enters the gorge upstream of the dam. If pool level is low the upper West River and Middlesex may also be exposed, however, out-croppings are likely to be partially covered by mud and debris. The base of the Cashaqua denotes the start of MN Zone 5 (*Palmatolepis punctata*) and *Probeloceras lutheri* Zone. A diverse molluscan fauna can be recovered from the greenish shales, including *Probeloceras lutheri* and numerous bivalves; among the most common are *Buchiola, Ontaria,* and *Pterochaenia* (see Clark, 1904).

16.4 0.0 Return to Rt. 408.
18.3 1.9 Turn left (north) on to Rt. 408.
19.3 1.0 Village of Mt. Morris.
20.1 0.8 Signal and junction with Rt. 36, turn right (south) onto Rt. 36.
23.7 3.6 Groveland Correctional Facility.
25.1 1.4 Junction with I-390 (Exit 6 - Sonyea), turn right and proceed south (toward Corning).
26.2 1.1 Pull off shoulder and park at Mile Marker 32, exposures are in road cut on both sides of highway; measured section is on the west side of the southbound lane.

Stop 3 - I-390 Mile 32: slope-basin muds and turbidites of Gardeau Shale, West Falls Group.

In the gorge section, the Gardeau Shale overlies the Rhinestreet and extends upstream to the Middle Falls, consiting of approximately 100 m of interbedded green-gray shale, black shale, and silts that represent an influx of fine and coarse clastic material into the basin. The exposure here of the Gardeau Shale consists of interbedded medium gray and dark shales that may represent small scale cycles that are overlain by interbedded silty shales and thin silt-sandstones that may also represent small scale cycles (Figure 12). The shales have yielded a low diversity and sparse fauna of nowakiids. Inarticulate brachiopods and trace fossils occur in the coarser (silt-sandstone) beds. The brachiopods and trace fossils in the silt-sandstone beds indicate development of dysoxic-oxygenated benthic waters and a suitable substrate for benthos. The silt-sandstones are graded, consisting of a scoured base, massive lower sand, and ripple to hummocky cross-laminated upper surfaces that are sharply overlain by gray shale. Scour features on the bases of silt-sandstone beds are oriented east-west. Many of the scours are laterally asymmetric, possibly the result of helical flow of the scouring current. The silt-sandstones are interpreted as turbidites, however, the upper surfaces have been reworked by subsequent currents.

26.2 0.0 Proceed south on I-390.
33.5 7.3 Dansville, Foster-Wheeler is a specialty steel plant.
41.4 7.9 Wayland (Exit 3), continue south on I-390.
44.0 2.6 High exposures on both sides of road between mile markers 15 and 14, pull well off shoulder at southern end of the exposures near the crest of the hill and park.

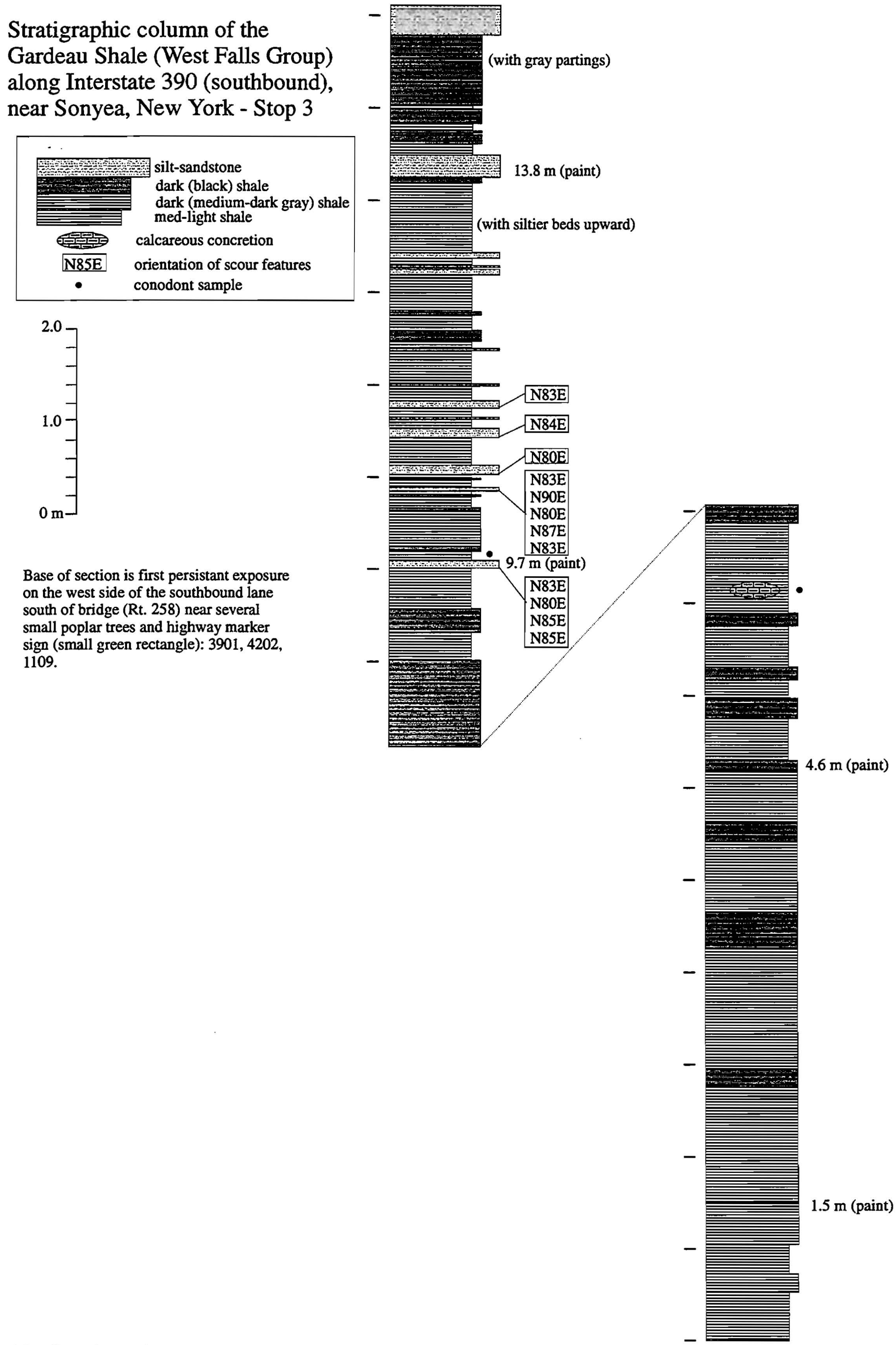


Figure 12. Section of Gradeau Formation at Stop 3, I-390 Mile 32.

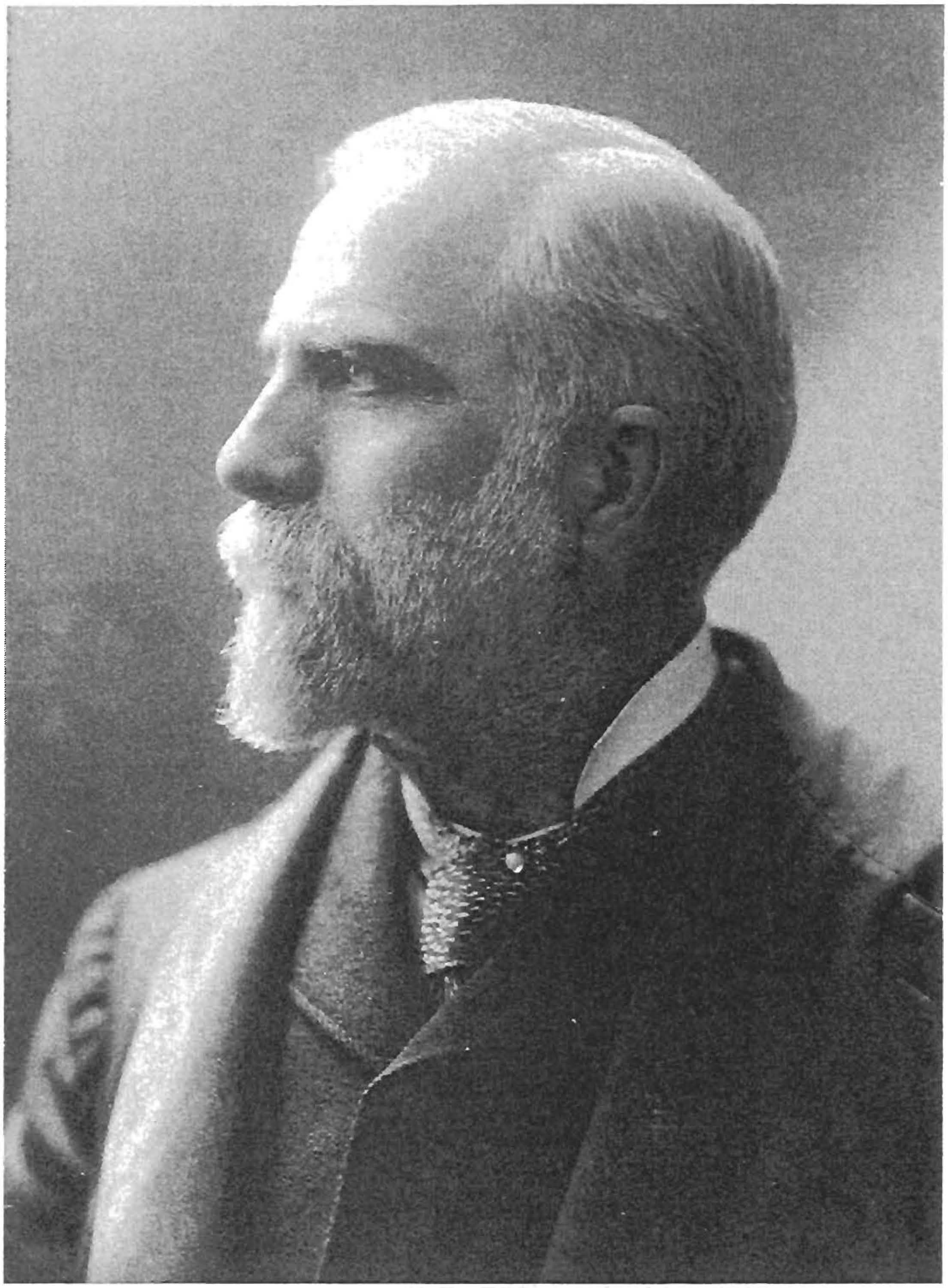
Stop 4 - I-390 Between Mile Markers 15 and 14 : Shelf sands of upper West Falls Group, Nunda or Wiscoy sandstone; ichnofossils, channel sand and shell-rich basal channel fill.

These sandstone exposures are questionably of the Nunda Sandstone which represents the upper West Falls Group and coarse-grained lateral equivalent of the Angola Shale in the Genesee River Valley. The sandstones are composed of tabular to lenticular thin to medium beds. Brown weathering sands are muddier, often low angle cross-laminated, and may contain rip-up clasts of brown mudstone and plant fragments. Medium blue-gray sands are ripple surfaced and cross-laminated, locally bioclastic (bivalves, brachiopods, and crinoids), and moderately to completely bioturbated. Identifiable trace fossils include *Skolithos, Diplocraterion, Scalarituba*?, and small *Chondrites*?, as well as poorly preserved lined and unlined horizontal to subhorizontal burrows. These sands are similar to the bluestones that were a common quarry stone near Portageville at the head of the Genesee gorge in the 1800's and early 1900's. Also notable in the section are resistant spherical weathering zones, suggestive of concretions developed in the sandstone.

In the southern most outcrop of the south bound lane is a channel cut filled with light gray weathering sandstone. The base of the channel consists of a 40 m long, 0.4-0.7 m thick dark-brown weathering fossiliferous sandstone to sandy carbonate. The stratum contains numerous brachiopods of a deepshelf environment *(Cyrtospirifer, Douvillina, Nervostrophia, Spinatrypa)*, as well as cephalopods, crinoids, fish remains, and a diverse polygnathid conodont fauna. The fauna is characteristic of MN Zone 12.

The current reworked and bioturbated sands, brachiopod fauna, *Skolithos-Curziana* ichnofacies, and polygnathid conodont fauna are indicative of a moderate energy shelf environment. These sands correlate to finer grained green-gray and dark gray shale of the Angola to the west (Figure 2). The Pipe Creek Shale, a thin transgressive black shale unit, separates the Nunda from the overlying Wiscoy Sandstone, the lateral equivalent of the Hanover Shale (highest Frasnian). The Pipe Creek is not well developed in this region and has not been locally identified. More nearshore channel and deltaic sandstones are found to the south in Famennian strata. These shallow water shelf and near shore sandstones record the late stages of progradation of the Catskil Delta across New York State and the filling of the Appalachian Foreland Basin.

END OF ROAD LOG.



Herman Leroy Fairchild

Professor of Geology, University of Rochester
[1888-1920]

LAND INSTABILITY IN THE GENESEE RIVER VALLEY - LIVINGSTON COUNTY

Types of instability

Land instability in the Genesee River Valley in Livingston County takes three forms: River-bank failure during meander migration, landsliding, and subsidence over abandoned salt mine workings. River-bank failure is a completely natural process possibly influenced by controlled discharge from the U.S. Army Corps of Engineers Mount Morris Dam in Letchworth State Park. Landsliding is a natural process along the Genesee River. Subsidence over abandoned salt-mine workings is a process induced by mine collapse. Only the subsidence stop (1) and the meander migration stop (2) will actually be visited on this trip.

Subsidence and other events in the Boyd-Parker area[2]

At 05:43 EST on 12 March 1994, an energy release equivalent to that from a magnitude 3.6 earthquake took place from a source beneath the flat valley floor of the Genesee River under Boyd-Parker State Park near Cuylerville in the SW Geneseo 7½' quadrangle (Figure 6).[3] This energy release was recorded on seismographs managed by Lamont-Doherty Observatory in Palisades, New York and by other seismographs in the U.S. and Canada. This energy-release event was first reported as an earthquake, but subsequent analysis seems to indicate that the energy release was simultaneous with the fall of roof rock in the Akzo salt mine beneath the bridge on Route 20A over Little Beard's Creek.

This event at depth was accompanied by dramatic events occurring at the surface. These included sharp shaking felt by residents immediately above the failed mine workings, lateral displacement of the Route 20A bridge over Little Beard's Creek, failure of the west abutment of the bridge, ground-surface subsidence, and open fracturing of the ground surface.

It was soon learned that changes had occurred below ground. These included the collapse of pillars previously supporting the mine roof, failure of the roof, and inflow of water into the mine at an initial rate estimated at 4000 gal/min. The area in which subsidence has taken place is referred to here as the Boyd-Parker area after the park located close to the center of the area. The initial subsidence area is designated as S-1 on the map accompanying the road log (Figure 9).

Episodic as well as gradual subsidence of the ground surface and steady inflow of water into the mine continued from 12 March 1994 to early April. On 06 April 1994 at 05:00 EDT, this slow subsidence was punctuated by the sudden drop of

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[2] This summary is based largely on Lundgren (1994a and 1994b)

[3] See road log for figures cited in this section.

an approximately 4000 square foot area on the south side of Route 20A. This subsidence, estimated to be on the order of 10 ft (3 m) created a lake along this reach of Little Beard's Creek. This subsidence area is designated as S-2 on the map (Figure 9).

The gradual subsidence of area S-1 was punctuated by another event on 23 April 1994 when an open fissure having a circular plan formed around the outer edge of S-1. Then on 25 May 1994, yet another subsidence event occurred in the area designated as S-3 (Figure 9). This event was manifested by the formation of a sinkhole 600 ft in diameter and 70 ft deep in the center.

The inflow of water into the mine instantaneously created severe problems for the Akzo Salt company, problems that have been the main focus of attention on the part of Akzo and the New York State Department of Environmental Conservation. These problems have been extensively covered by the Rochester newspapers and have received daily mention on TV and radio. Closing of the mine and the continued occurrence of events have already had substantial economic and other impacts on the community.

Relationship between subsidence and mine workings

The map accompanying the road log (Figure 9) illustrates that subsidence area S-1 lies directly above a rectangular section of the salt mine set off from the main mine area at its southern end. In this section, the salt pillars left to support the roof were approximately 20 ft on a side. This is in contrast to the extensive mined area immediately to the north where the salt pillars were approximately 80 ft on a side. Subsidence area S-3 also lies above a rectangular section of the mine in which the pillars were 20 ft on a side. The 2 mine sections in which pillars 20 ft x 20 ft were left are bordered on the west and south by unmined salt or by a continuous "wall" of salt. The map (figure 9) shows the dimensions of the partial walls of salt on the eastern side of these two areas. Since the seismic event recorded on 12 March 1994 has been attributed to the fall of a volume of roof rock approximately 12 ft x 600 ft x 600 ft, then the implication is that a number of 20 ft x 20 ft columns failed almost simultaneously in order to make the fall of such a large block of roof rock possible. Subsidence at the surface (S-1) was simultaneous with the seismic event, indicating the rapid propagation of fractures (faults) upwards from the mine 1000 ft below.

Although extensive drilling has been carried out and surface surveys have been made, none of the findings are available from either Akzo Nobel or from the New York State DEC at this time (June 1994). Therefore the actual displacement of rock layers and surficial materials is not known. The subsidence event in area S-3 also is apparently located directly above a second area of the mine in which the 20 ft x 20 ft mining method was used. This subsidence event differed from that at S-1 in a significant respect. Whereas subsidence at S-1 has been relatively gradual, except for the occurrence of event S-2 within the S-1 area, subsidence at S-3 was instantaneous creating a "sinkhole" reported to be 70 ft deep in the center. This geometry has suggested to observers who do not have access to the site that unconsolidated materials may have been transported downwards in the phenomenon known as "piping."

Implications

The subsidence events to date have occurred above a stratigraphic section that consists of approximately 400 ft of unconsolidated valley fill deposits that lie on eroded Onondaga limestone. Therefore failure above the mine involves failure in a 450 ft sedimentary rock section and the overlying unconsolidated materials. Since the Onondaga is in direct contact with the valley fill, there has been much debate about the "source" of the water flowing into the mine. Akzo Nobel representatives have generally maintained that this source is the Onondaga, which is a confined aquifer west of the valley. Many other geologists who have examined the situation have argued, however, that the source more likely is a combination of the Onondaga and the valley fill above. The importance of this is that to the extent that the valley fill serves as "source" inflow of water into the mine can affect hydrologic relationships over a much larger area than if the source is a confined aquifer (Onondaga).

MEANDER MIGRATION - GENESEE RIVER

Meanders

Even cursory examination of the Geneseo 7.5' quadrangle or of aerial photographs reveals that the Genesee River meanders freely across the valley floor in the Geneseo quadrangle. The most recent oxbow lake to have formed is readily evident in the SW ninth of the Geneseo quadrangle. There, the photorevised 1978 edition shows the 1950 course of the Genesee as relatively straight. The 1978 photorevision shows an oxbow lake, documenting that a major meander was formed and cut off within little more than 20 years.

Meander activity at two locations south of Route 20A has been of substantial concern. Meander migration immediately south of Route 20A (see figure 6 and 8 with road log) threatened the bridge over the highway and the highway itself. Therefore the U.S. Army Corps of Engineers controlled this migration by placing rip-rap on the concave bank of this meander. As is evident from the map, this meander is also very close to cutoff .

A prominent set of meanders at the southern edge of the Geneseo quadrangle has been migrating rapidly since 1938, cutting into an unpaved road sometime in the 1950s and progressivley removing farmland from cultivation. The main meander here, informally designated as the Christiano meander (see figure 7 with road log), illustrates the meander migration process very well. The actual physical process very much resembles the failure process seen at Stop 2 at Irondequoit Bay in the Monroe County segment of this trip. This process and the results are described in the road log below.

Implications

Since complete meander loops can form in the valley on a time scale of 2 to 3 decades, migration and cutoff are factors in land-use decisions in the valley. The most visible evidence of this is the engineered approach to meander control at the Route 20A meander. In addition, land owners engaged in agriculture have apparently attempted to create artificial (premature) cutoffs to control meander migration. Cutoff of the Christiano meander will eventually stop its further progression but with unkown implications downstream. Meander migration that impinges on the walls of the valley also has the potential to trigger landslide activity. This may have been the case at the site of the Oxbow Lane landslide (see figure 6 with road log). No structures existed on that landslide at the time of failure, but the event illustrated the potential for damage should structures be placed in similar settings in the future.

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ROAD LOG (MONROE COUNTY SEGMENT)

This road log begins at the red light at the intersection of Empire Boulevard and the exit ramp from Exit 8 of Route 590N (See figure 1). To reach this point from the University of Rochester, we will follow I-390S to 590N and follow the white 590N signs to Exit 8 (Empire Boulevard, Route 404, Webster)

Note: This road log is designed explicitly for free-lance use. The copies of parts of the Rochester East 7.5' quadrangle reproduced below show stop locations and provide coordinates using the 10,000 foot New York Coordinate system grid shown on all USGS topographic maps of New York State.

Mileage

Total	Incremental	
00.0	00.0	Empire Boulevard (Route 404) at Exit 8 of Route 590N. Turn right (east) onto Empire Boulevard. Once you cross Irondequoit Creek, be prepared to make a left turn with caution.
01.1	01.1	Left into parking area at New York State Historic Marker (Irondequoit Bay). Walk through the parking lot of the Bounty Harbor Restaurant (assuming permission) and along the shore to the site shown in figure 1.)

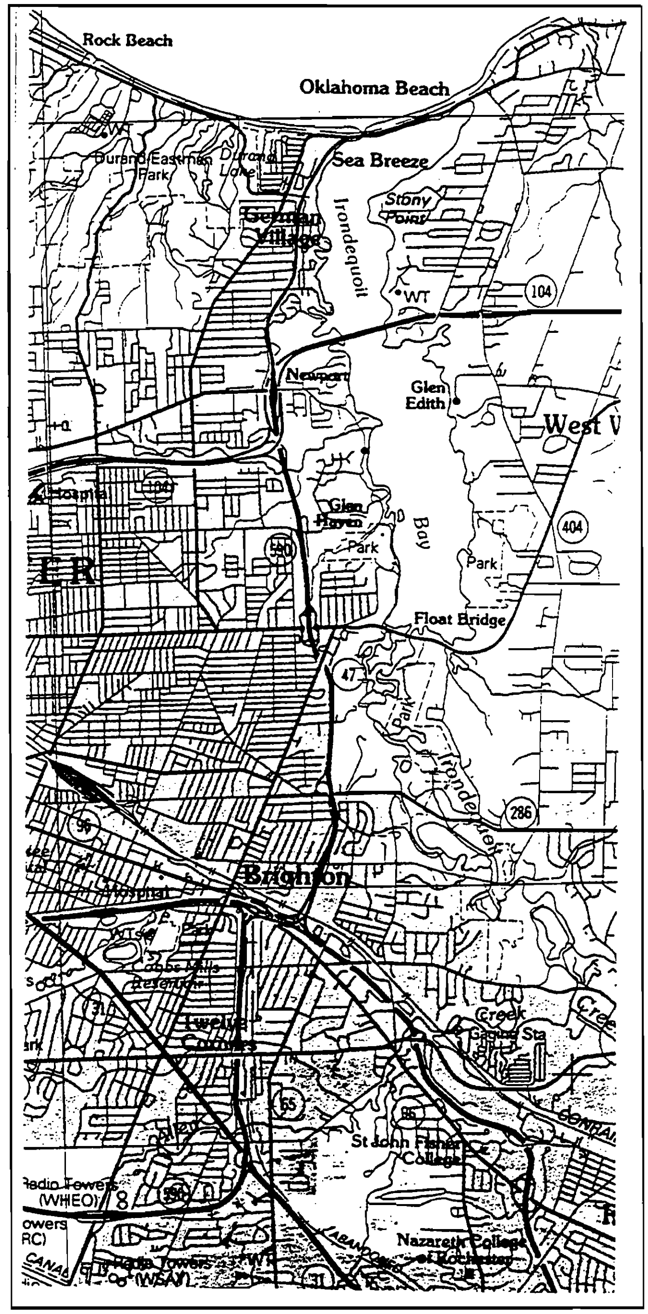


Figure 1 Map of Irondequoit Bay showing approaches on 590N and Route 404. From USGS 1/100 000 Rochester quadrangle.

STOP 1 (Wang sites SE-01-88 and SE-02-88: Note that these sites are only accessible with permission of the land owner. They may or may not be accessible at the time of the NYSGA field trip. If they are not, then they will be viewed from the parking area at the New York State Historic marker site.

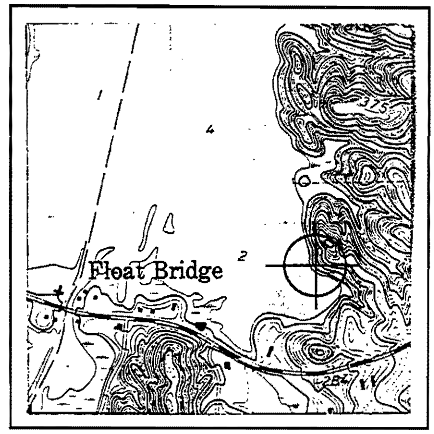


Figure 2 Stop 1 (Wang site SE-01-88 and SE-02-88. NYS 10 000 ft grid coordinates (784100E/159500N)

Discussion topics: (1) The inception of slope instability on previously stable forested slopes. (2) The initial removal of the soil-zone slab. (3) Interaction among different types of slope failure.

These sites are the key to our interpretation of slope instability and evolution. Landslides have been evolving dramatically in the 6 years since observations were begun by Wang.

The pre-landslide character of the slopes at this site is evident from the still vegetated slopes around and above the active landslides on the east side of the bay. Oak, maple, poplar, and other large trees are rooted in the meter-thick soil blanket developed on slopes as steep as 40°. Up to 1951, the entire slope at this site was stable and completely covered by bushes and trees, many of them 50 years old or older. The slope first became unstable sometime between 1953 and 1960.

In 1988, when Wang made his first observations, multiple landslides were already evident (Figure 3). Figure 3 illustrates what we infer to be the initial stages in the destabilization processes affecting Irondequoit Bay slopes. The meter-high scarp surrounding the landslide complex reveals a meter-thick soil zone. Initial slope failure entailed the downslope movement of slabs the base of which is approximately the base of the soil zone. This slab carried with it all of the trees rooted in the soil zone. Even large oak trees have been transported down slope while initially remaining in upright position. Slab A began to move down slope in April 1990 to the position shown in April 1993.

This initial movement of the soil-zone slab effectively removes the protective cover from the slope, but it does not change the value of the slope angle. This process of block sliding takes place on differing scales as may be seen from the way in which the main scarp of this initial and largest landslides is itself cut by the scarps of several smaller landslides. Some of these smaller landslides are similar

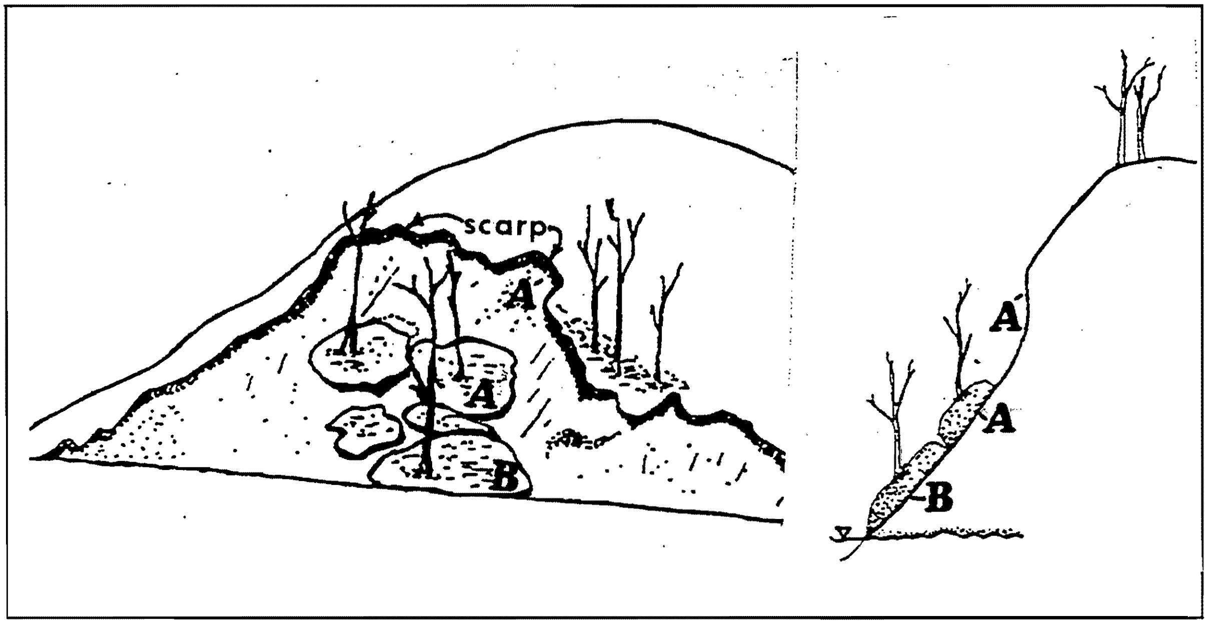


Figure 3 Views of landslides at Stop 1 (Wang site SE-01-88). Based on photograph taken April 1993. Highest point on scarp is 17 m (56 ft) above water level. Slab A was in place at A' until April 1990. Since then it has been moving slowly downslope. Slab B has reached water level and is being undercut.

in character to, but younger than the largest scarp. In other words, some of these smaller landslides have the same character as the largest slide

In addition to these slab-like landslides, the lower part of the slope displays slumps,landslides created by failure along a spoon-shaped slip surface. These slumps typically form in material from which all vegetative cover has been removed. The most recent of these have all formed in the summer and fall of 1993. These slumps apparently are triggered by undercutting at the shoreline, especially if the slope above has been denuded of vegetation. The formation of these slumps creates regions where the slope is very steep, commonly vertical. These vertical faces are inherently unstable, and failure occurs here through topple and fall as illustrated at stop 2. It appears that once this stage has been set in motion, a threshold has been crossed so that slope steepening becomes the rule.

01.1 01.1 Exit left from parking area using extreme caution. Proceed uphill on Empire boulevard. We will wait on the shoulder until everyone has safely exited.

02.8 01.7 Left onto Bay Road (Seaway Trail Sign)

03.1 00.3 Bayview Family YMCA. Park in lot.

STOP 2 (Wang site SE-09-88)
Access from Bayview YMCA Parking lot. From Bayview YMCA walk down road to burned-out building and through it to the ORV trail marked by orange paint blazes. Continue to Stop 2 at the shoreline. (Alternative access from Smith Road at Empire Boulevard. Smith Road terminates at the east shore of Irondequoit Bay adjacent to a quonset hut owned by the Rochester Canoe Club.)

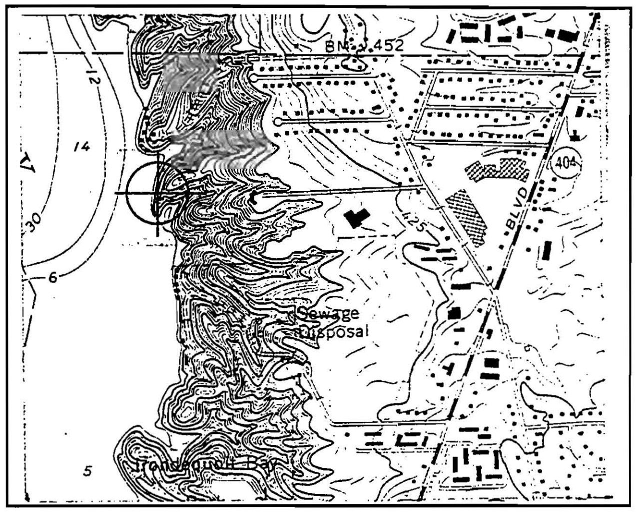


Figure 4 Stop 2 (Wang site SE-09-88) NYS 10 000 ft grid coordinates: 783500E/1165200N

Discussion topics: (1) Lake Iroquois sediments. (2) Present-day processes, especially topple and slump observable in real time. (3) Inception of bluff development in the 1960s.

This site displays 3 bluffs (very steep unvegetated slopes), each of which is vertical in its upper part. The highest bluff rises 26 m (85 ft) above water level. The vertical upper face of these bluffs is created in a cohesive clay unit. Earth slump and earth flow of the non-cohesive materials that underlie this clay unit set the stage for the opening of vertical joints in the clay unit. Blocks of the clay unit as large as 2 m on a side then fall from this free face, sometimes remaining intact and sometimes breaking into numerous small blocks.

Below the vertical face are fans of material that has fallen from above and then accumulated at the base. Some vegetation has taken root in this material. During the summer and fall of 1993 and the late spring of 1994, these aprons were being eroded by wave action, and this erosion and undercutting of the base was triggering earth slumps 2.5 to 3.6 m (8 to 12 ft) high at the base of the bluffs. Earth flow is also common in the fans. Most of the steps in the process can be observed in real time whenever waves and boat wakes break along the base of this bluff.

These 3 bluffs first appear in the 1961 aerial photographs; they cannot be recognized in older photographs. They are interpreted as the end product of landslide processes that first operated on steep but vegetated slopes. Remnants of these slopes are still present. Complete removal of the original soil blanket and the vegetation rooted in that blanket set the stage for the operation of slump processes. Undercutting by wave action during periods of high lake level (as in 1973 and 1993) removed the protective aprons, cut a shelf under the vertical bluff, and set the stage for earth topple noted above.

03.4 0.2 Return to Bay Road. Left (north) on Bay Road.

05.0 1.6 Over bridge above Route 104 and left onto the entrance ramp to 104 west. Keep right.

05.4 0.4 Right into parking area overlooking the Bay Bridge.

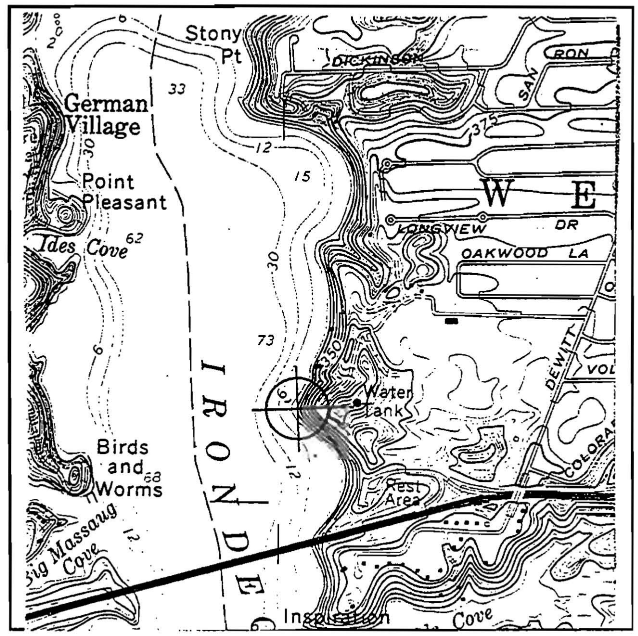


Figure 5 Stop 3 (Wang's site NE-05-88). NYS 10 000 ft grid coordinates: 780200E/1173700N

STOP 3 (Wang site NE-05-88) - Walk north along trail to water tower and then down the west-facing slope west of the water tank at the Webster Water pumping facility.

Discussion topics: (1) Use of multiple scarps in establishing landslide history (oldest and highest scarp dates from 1930 or earlier). (2) Role of high water level and wave activity in setting the stage for slump activity. (3) Probability of future landslide activity. (4) Nature of Lake Iroquois sediments.

This slope displays at least 6 different landslide scarps ranging in age from 1930 (or older) to 1994. The oldest scarp (top at 33 m - 110 ft) above water level outlines a landslide area (probable earth block slide) visible on the 1930 aerial photograph. The entire slope within this oldest scarp displays none of the large trees that are uniformly present in the area outside of the scarp line. There were few large trees on this slope even in 1930. The upper part of this oldest landslide is covered by grass, bushes, and by a few trees less than 15 cm (6 in) in diameter.

All of the other scarps lying within the slope area bounded by the oldest scarp are displayed at lower elevations ranging from 8 m to 33 m (27 ft to 108 ft) above water level. Each scarp bounds a separate landslide, some of which are earth slumps. The youngest of these landslides, re-activated in 1993 when lake level was at a 20-year high, illustrates the conversion of a steeply sloping landslide site to an earth slump and vertical bluff.

The earth slump displayed in the lowermost 15 m (50 ft) of this slope began to form prior to 1978. It exposes an excellent sample of the type of section generally seen north of the Route 104 bridge. A clay layer at the base of the slope is overlain by at least 6 m (20 ft) of sand and silt displaying the climbing ripples characteristic of all northern sections. The upper edge of the scarp of this slump was at 15 m (50 ft) above lake level in May 1994. The upper 2 m of section display clay and silt layers with prominent pillow structures.

The inferred history of this site is as follows: (1-pre 1930) Undercutting at the base; large-scale earth-block sliding of blocks made up of the soil horizons. There were few trees left on this slope after this event. (2) Removal of tree cover, followed by episodes of high lake level led to the inception of the processes active in 1993-1994. (3) Undercutting of the base of the slope by wave action during periods of high water level. Major periods of high water level occurred in 1952, 1973, and 1993. (4) Failure by earth slumping, flow, and topple. These three types of landslide activity occur in conjunction with one another as illustrated by relationships expected to be visible in October 1994.

END OF MONROE COUNTY SEGMENT OF TRIP

ROAD LOG (LIVINGSTON COUNTY SEGMENT)

If continuing to Livingston County segment of trip, cross the Irondequoit Bay Bridge and follow 590S to the I-390S ramp marked for Corning. The road log starts at the entry to this ramp. The potential stops are shown on the map of the SW ninth of the Geneseo 7.5' quadrangle (Figure 6). The road names used here are those used on this map and are spelled as on the map. As of June 1994, access to this area is controlled by security guards, and it is not known what the situation will be on 09 October 1994.

Cumulative	Incremental	
00.0	0.0	I-390S ramp from 590. (Monroe County)
22.6	22.6	Exit 8 (Geneseo exit) onto Route 39/20A . Follow Route 39/20A signs west.
27.8	5.2	Main Street - Geneseo. Continue to follow 39/20A
28.9	1.1	Follow 39/20A signs and descend into the Genesee River Valley.
30.1	1.2	Bridge over Genesee River. Meander on east side of bridge protected by riprap.

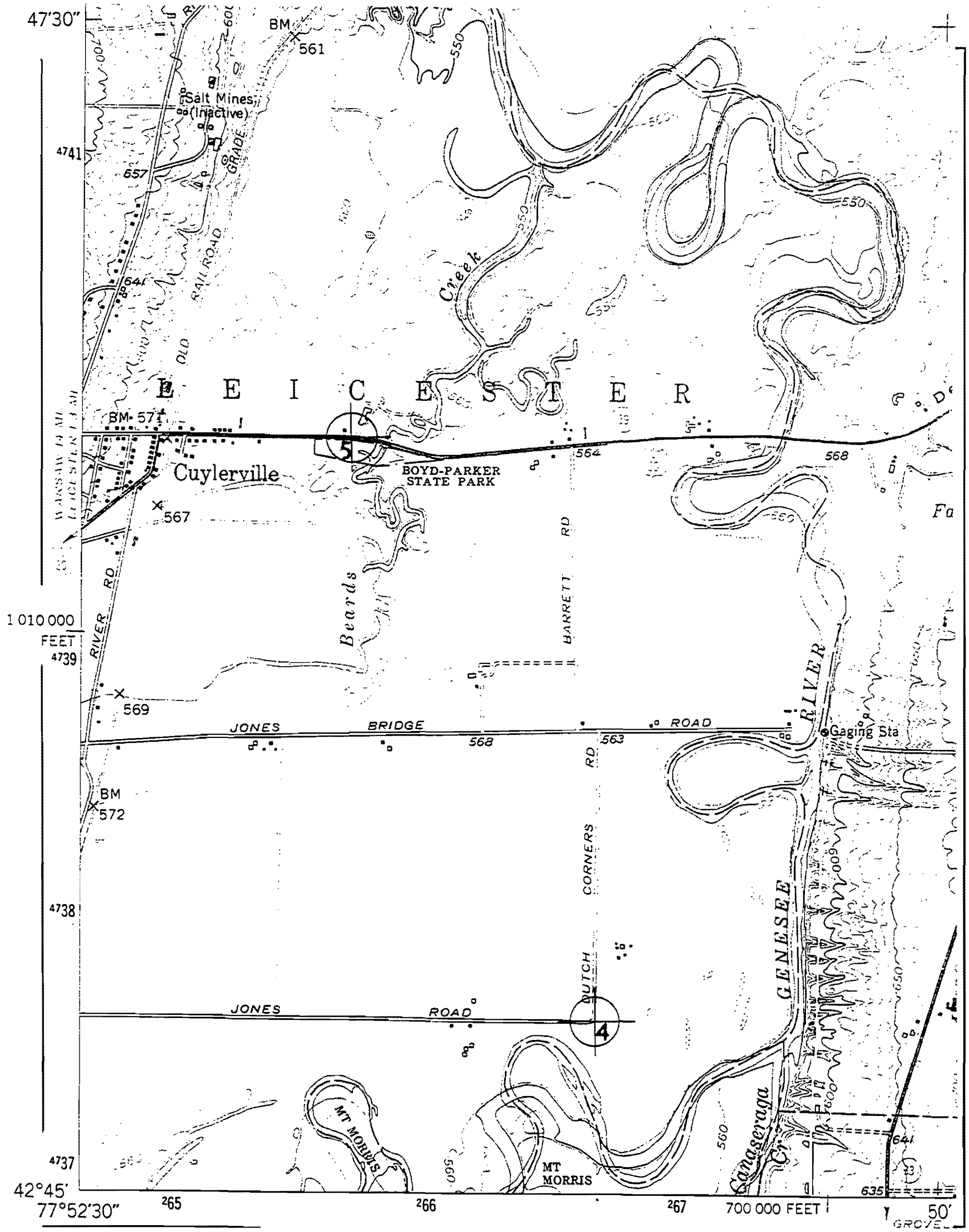


Figure 6 SW ninth of Geneseo 7.5′ quadrangle (1/24000). Photorevised 1978. Meanders are numbered for reference to figures 7 and 8 and text. Possible locations for discussion of subsidence in the Boyd Parker area are the west edge of Boyd Parker Park or a site to be chosen on the trip.

Discussion topics: (1) Rates of migration of Christiano meander (this stop) and Route 20A meander (compare figures 7 and 8), (2) Meander migration process at times of high discharge, (3) Possible role of controlled discharge controlled by the Mount Morris dam.

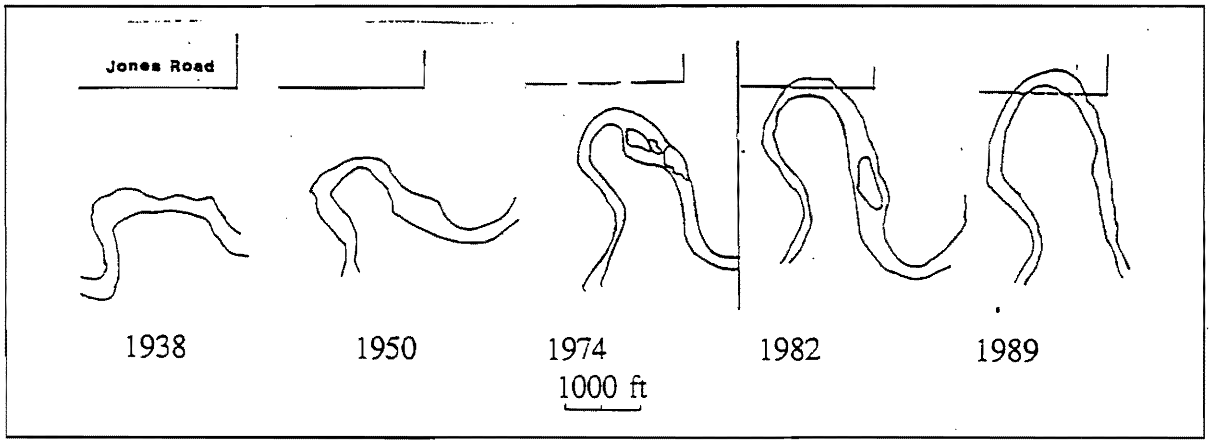


Figure 7 Map of Christiano meander(MS-3) position 1938 - 1989. Plotted from aerial photographs by BOCES summer students 1989.

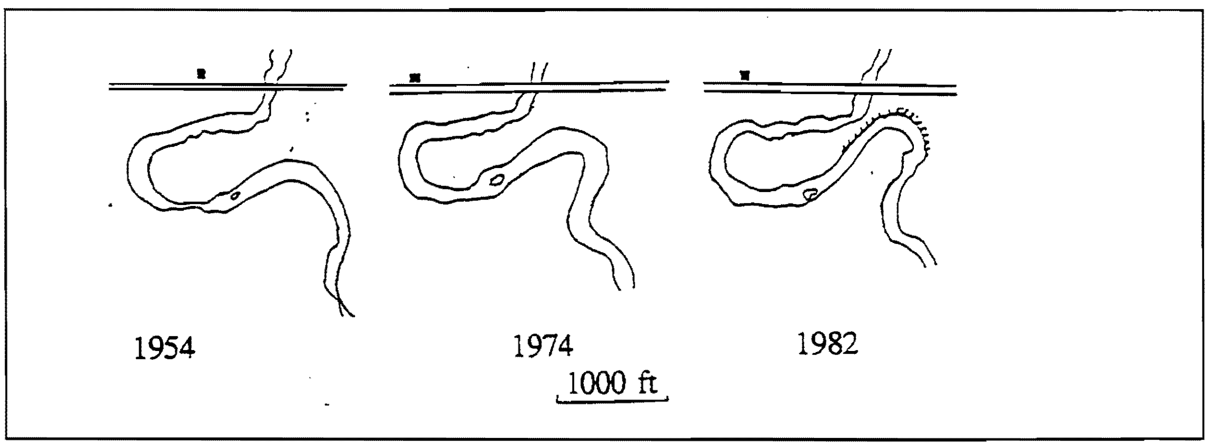


Figure 8 Map of Route 20A (MS-1) meander 1954-1982. Compiled from aerial photographs by BOCES summer students 1989.

Stop 5 Route 20A east or west of the site of the bridge over Little Beards Creek at Boyd Parker Park. This bridge was destroyed by subsidence 12 March 1994. It is not known at this time (June 1994) if this site will be accessible. See figure 9.

30.6 0.5 Intersection of Barrett Road (as spelled on USGS map) and Route 39/20A. South on Barrett Road.

31.3 0.7 Intersection of Jones Bridge Road and Dutch Corners Road. South on Dutch Corners Road to barrier.

31.9 0.6 Barrier.

Stop 4 See figure 6. Genesee River at intersection of Jones Road and Dutch Corners Road (NYS 10 000 ft grid coordinates: 703100W/1002800N).

Discussion topics: (1) The sequence of subsidence events (see figure 8), (2) The relation between subsidence and the mine workings, (3) Hydrologic questions, (4) Future prospects for subsidence and hydrologic impacts.

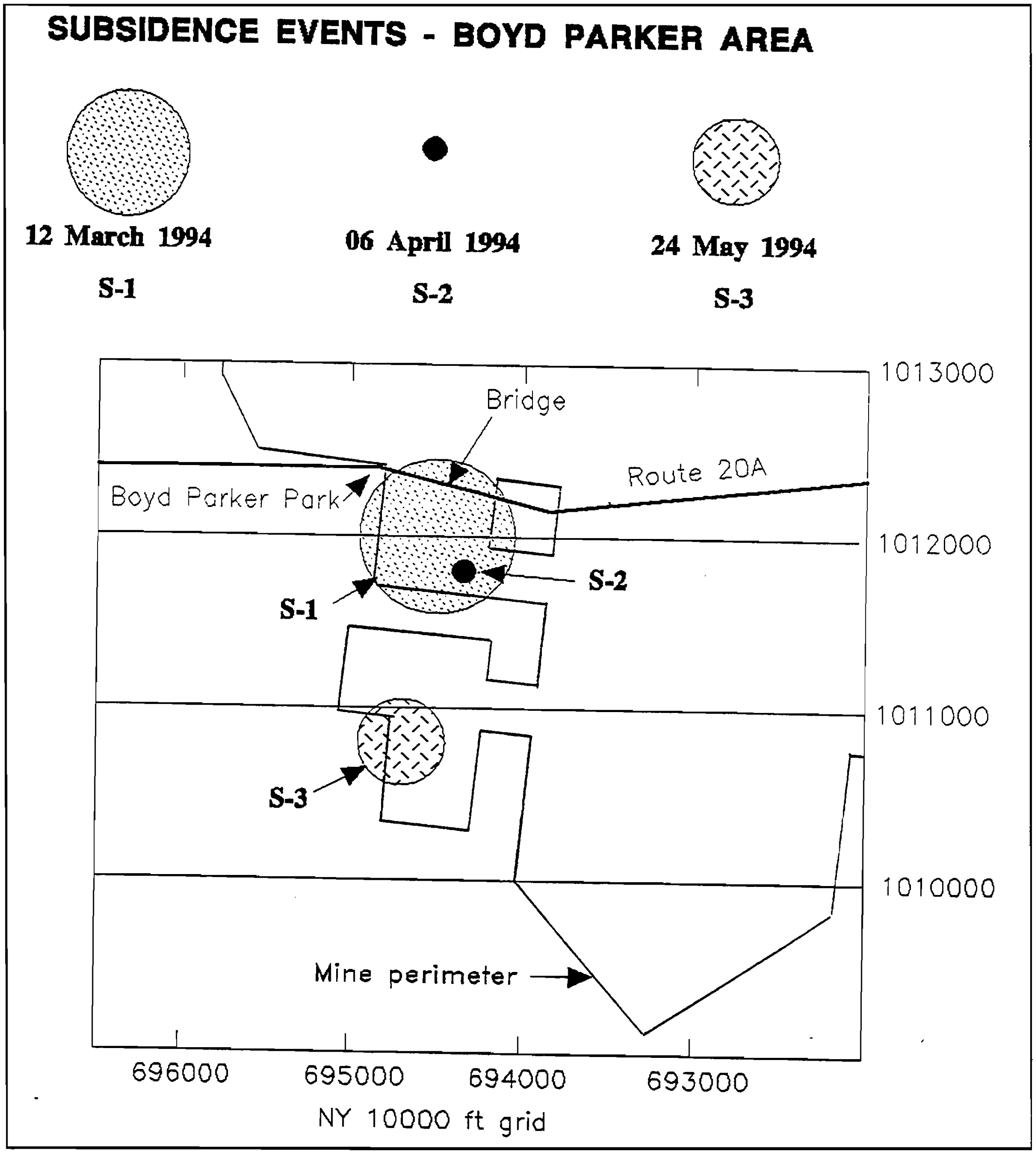
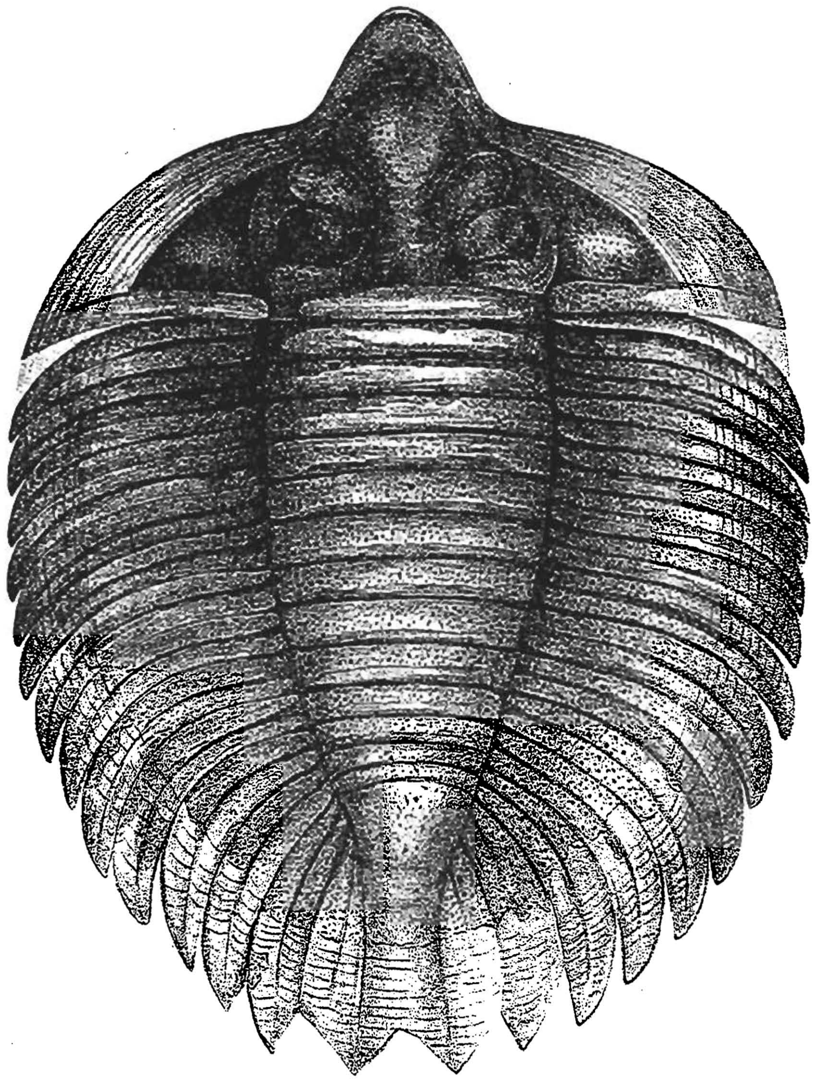


Figure 9 Approximate locations of subsidence features and mine perimeter. Subsidence features based on field mapping. Mine perimeter from map provided by Akzo Salt, Inc. All boundaries schematic.



Actinurus boltoni

[From Hall, 1843, Geology of the Fourth District, Plate 19, Figure 1]

ORDOVICIAN AND SILURIAN STRATA IN THE GENESEE VALLEY AREA SEQUENCES, CYCLES AND FACIES

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INTRODUCTION

Upper Ordovician and Silurian strata in New York State are well-known from their spectacular exposures in the Niagara and Genesee River Gorges. Because of the excellent exposures in western New York, the descriptive stratigraphy of these units is well-refined, and the interval has recently been used as a test case for models of eustatic and tectonic stratigraphic dynamics. Important modern works concerning the litho- and biostratigraphy of these strata include Gillette (1947), Fisher (1953, 1960, 1966), Kilgour (1963), Rexroad and Rickard (1965), Zenger (1965, 1971), Rickard (1969, 1975), Martini (1971), Brett (1983), Duke (1991) and LoDuca and Brett (1991). Allostratigraphic analyses have been performed by Duke and Fawcett (1987), Brett et al. (1990a, b; 1991) and Goodman and Brett (1994).

The Upper Ordovician and Silurian strata of the Genesee Valley and surrounding areas were deposited near the northern reaches of a dynamic and rapidly evolving Appalachian foreland basin. The striking patterns of unconformity and lateral basin axis shift are striking given that the Late Ordovician and Silurian Periods are generally considered to be times of tectonic quiescence between the Taconic (late Middle Ordovician) and Acadian (Middle to Late Devonian) Orogenies. Foreland basin flexure produced asymmetric marginal unconformities, but the differential subsidence did not obscure relative sea-level trends in litho- and biofacies that correlate with patterns in other basins. Thus, the stratigraphic interval exposed in the Genesee Gorge is an excellent field laboratory for the interaction between tectonic foreland basin flexure and eustatic sea-level changes.

During the Late Ordovician and Silurian, the northern Appalachian Basin was situated within a subtropical climatic belt, probably at about 20 to 25 degrees south latitude (Van der Voo, 1988, Witzke, 1990). Climates apparently oscillated from relatively arid during the Late Ordovician to humid during the Early Silurian and back to very arid during the Late Silurian. During the Late Ordovician and Early Silurian, the bulk of the basin fill in western New York represents marginal marine to nonmarine settings. Siliciclastic sediment that infilled the basin appears to have been shed from rejuvenated Taconic source areas primarily located to the east and

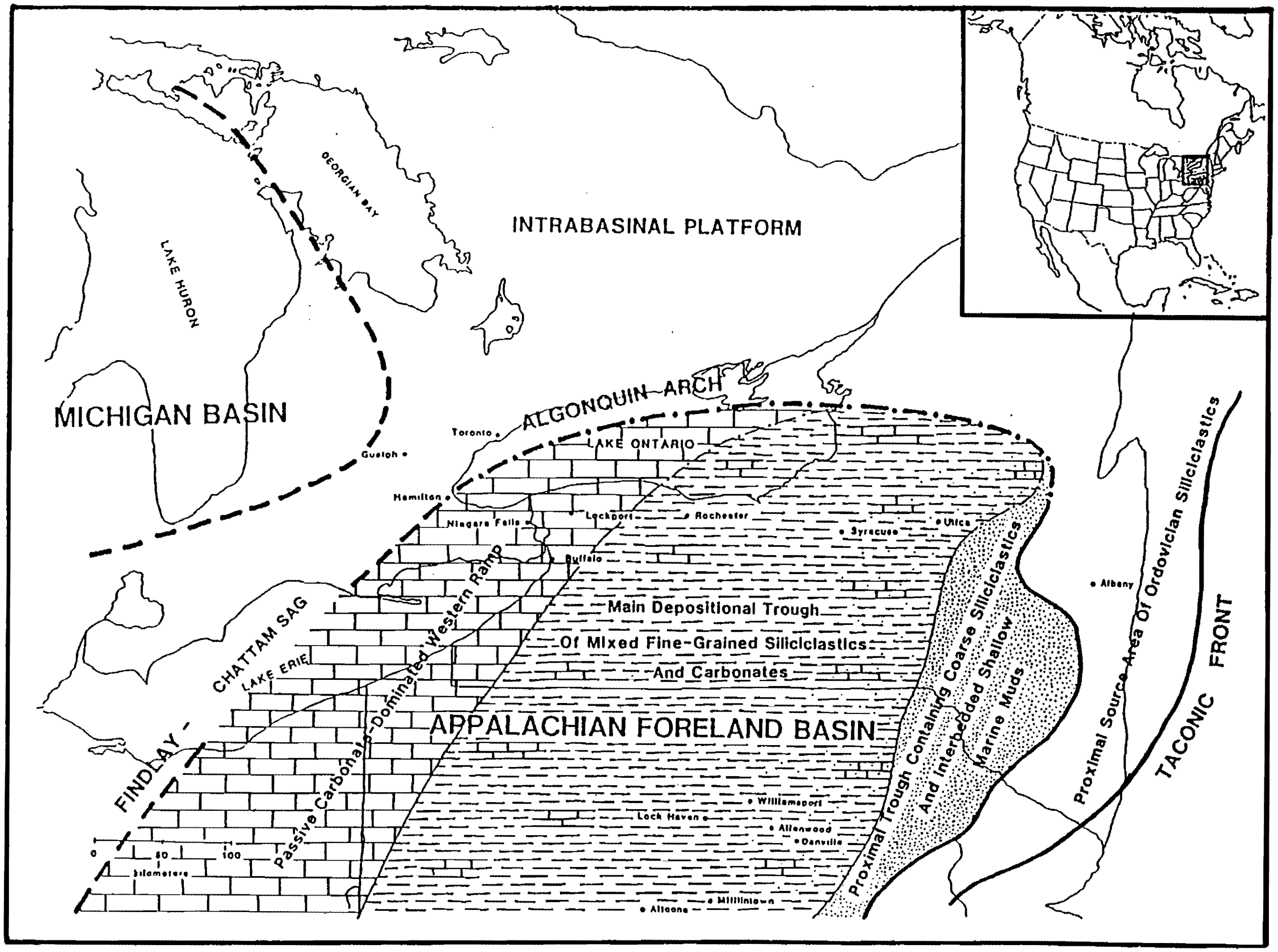

Figure 1. Paleogeographic map of the northern Appalachian foreland basin during medial Silurian time; inset map shows position of the study area within North America. From Brett and others (1990).

southeast. Open marine facies that are time-correlative to the Queenston Shale and Medina Group are situated northwest of the Niagara region.

A low forebulge, Algonquin was intermittently uplifted along to northwestern rim of the basin (Figure 1).

Subtidal ramp carbonates are preserved in the lower Clinton Group between Niagara and Wayne Counties, New York. The onset of clean carbonate deposition stratigraphically above a low angle, regional unconformity at the top of the Medina Group suggests structural uplifting of the Algonquin Arch during lower Clinton time. Unfortunately, a late Llandoverian unconformity that divides the Clinton Group into two units removed a significant portion of the lower Clinton stratigraphic record in western New York. Consequently, paleogeographic reconstructions can only be indirectly inferred from biofacies and textural trends in the truncated carbonates and siliciclastic facies belts situated farther east.

During the Late Silurian (Wenlockian and Ludlovian Epochs), upper Clinton and Lockport carbonates accumulated on the western basin ramp. Except for the Wenlockian eustatic sea-level highstand during which the fossiliferous Rochester Shale accumulated over significant portions of the basin, persistent siliciclastic sedimentation was restricted to the eastern ramp and basin center. Shales and mudstones commonly occupy the central, offshore facies belt. The eastern strandline facies belt is dominated by sandstone and conglomerate.

Late in the Silurian (Pridolian Epoch), circulation within the basin became restricted in response to increased aridity and fringing reefs of the Guelph Formation (Rickard, 1969). Increased siliciclastic sedimentation associated with the formation of the Bloomsburg-Vernon deltaic complex may signal rejuvenation of the eastern basin flank and cannibalization of Upper Ordovician and Lower Silurian strata during the Salinic Disturbance. Following accumulation of deltaic red beds across western New York, a thick sequence of peritidal carbonates, mudstones and evaporites (Salina Group) accumulated in western New York and surrounding areas. Resumption of more normal shallow marine conditions is recorded in the Cobleskill biohermal dolostone which is the youngest pre-Helderberg Silurian unit preserved west of Cayuga County.

In this paper, details of upper Ordovician and Silurian stratigraphy are provided in a regional context so that a framework may be established for the Rochester area facies. Where interpretations of allostratigraphy have been developed, depositional sequences and smaller, component cycles are discussed. This presentation proceeds in ascending stratigraphic order.

QUEENSTON FORMATION

The Upper Ordovician Queenston Formation is the oldest formation that crops out in western New York. In western New York and Ontario, the Queenston Formation consists of predominantly red shales and subordinate fine-grained sandstones, in contrast to the sandstone-dominated facies of the Queenston of central New York and of the correlative Juniata Formation in Pennsylvania.

The Queenston Formation represents the accumulation of molasse during late stages of the Taconic orogeny. In New York and Ontario, the Queenston Formation conformably overlies unfossiliferous peritidal to nonmarine sandstone (Oswego Sandstone) and fossiliferous grey and green marine shales and

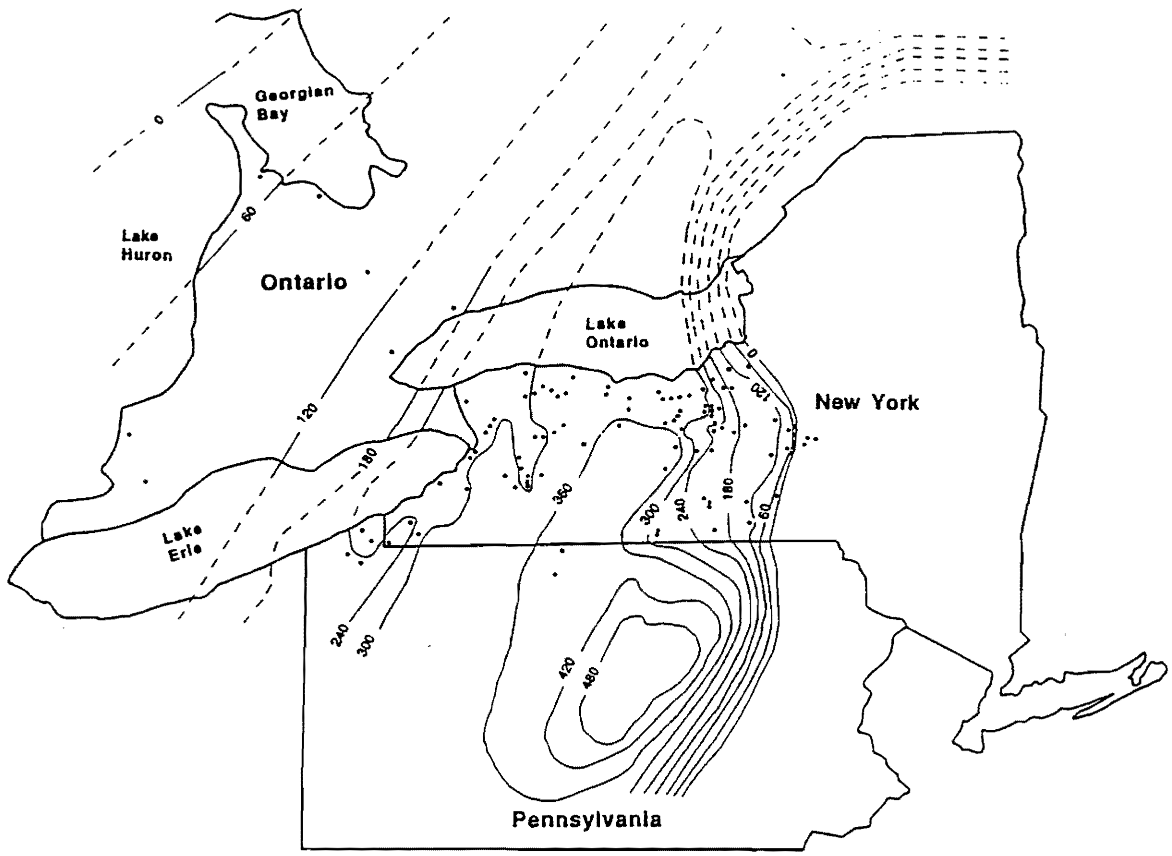

Figure 2. Isopach map of the Queenston Formation. Dashed lines are conjectural. Isopachs in Pennsylvania are based primarily on work by Swartz (1946).

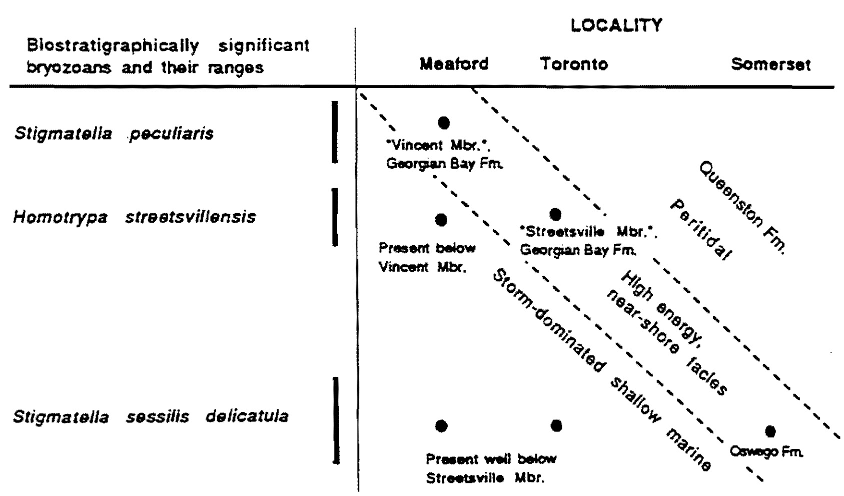

Figure 3. Bryozoan biostratigraphy of Queenston and correlative strata. High energy nearshore facies and Queenston red-bed facies appear to be older to the southeast. Uppermost Oswego strata at Somerset, N.Y. are time-correlative with middle Georgian Bay strata at Meadford, Ontario.

sandstones. The base of the Queenston, defined by the lowest occurrence of red beds, largely corresponds to the onset of peritidal conditions. The Queenston reaches a maximum thickness of approximately 360 m in western New York (Fig. 2).

The Queenston Formation is part of a progradational succession that filled the Appalachian foreland basin during the Late Ordovician glacioeustatic sealevel fall. The regressive maximum is recorded by the Cherokee unconformity that serves as the Queenston-Medina contact in western New York and Ontario. Queenston strata comprise transgressive-regressive cycles of a number of scales. These cycles are most apparent in the lower part of the formation in western New York and throughout the section in Ontario where green and red shale facies are interbedded. For example, in the Bruce Peninsula region, a green calcareous tongue of the Queenston is sandwiched between gypsum-rich, red shales containing desiccation-cracks. These green calcareous strata include biomicrites and biosparites that contain a normal marine fauna. In the Toronto area, this "mid-Queenston" marine incursion is represented by peritidal, gypsum-rich, calcareous strata bounded by non-marine, deltaic deposits (described below). This mid-Queenston transgressive interval can be divided into three, smaller, transgressive- regressive cycles which are manifested as alternating fossiliferous and gypsum-rich strata in the Bruce Peninsula region and as alternating gypsum-rich and non-gypsiferous, unfossiliferous strata in the Toronto area.

Although the Queenston strata in New York are generally unfossiliferous, the age of the Queenston can be constrained by the biostratigraphy of fossiliferous deposits which directly underlie it. Fritz (1926, 1982) noted that different stratigraphic levels in the underlying Georgian Bay Formation of Ontario contained characteristic bryozoa and, to some degree, defined a bryozoan biostratigraphy. Her work indicates that the uppermost marine strata below the Queenston Formation in the Toronto area contain a slightly older bryozoan assemblage than do lithostratigraphically correlative strata from farther northwest (Meaford, Ontario). Fossiliferous, lower Queenston bryozoa beds from Somerset, New York (RG&E core, stored at the University of Buffalo) contain an even older bryozoan fauna, indicating that the base of the Queenston is slightly diachronous, becoming progressively younger to the northwest of Somerset (Figure 3).

The depositional settings of the Queenston strata in western New York and Ontario include storm-dominated normal marine, tidal flat, and delta plain environments. The tongues of normal marine strata are restricted to the lower part of the formation. For example, a *Hebertella*- and bryozoan-rich bed is present 3 m above the base of the Queenston in the Hamilton, Ontario area. This bed, in which some fossils are oriented in life position, is bounded by red mudstones containing desiccation cracks.

A similar lower Queenston marine incursion is represented in drill cores from eastern Niagara County, New York. At Somerset, the basal 9 m of Queenston strata consists of unfossiliferous, cross-stratified sandstone and mudstone containing desiccation cracks. These peritidal strata are overlain by 4.5 m of red and green shale with interbeds of white, fossiliferous, calcareous sandstone, which, in turn, is overlain by unfossiliferous Queenston strata. This tongue of calcareous sandstone contains bryozoans and brachiopods, some oriented in life position. In general, fossiliferous beds are more closely associated with green shale than with red shale, although rip-up clasts of red shale are present in some fossil beds. Based on the presence of large (>7 cm) rip-up clasts and the vertical proximity of these beds to peritidal deposits, we postulate an extremely shallow shelf depositional setting for this marine tongue.

Red and green, gypsum-rich, variably calcareous mudstone and cross-stratified fine- to very fine-grained calcareous sandstone are characteristic of the middle Queenston strata in the area between Meaford and Toronto. This same facies assemblage is characteristic of the lower and upper Queenston strata of the Bruce Peninsula. In these rocks, laminae around gypsum nodules are typically distorted, suggesting that the nodules formed in unlithified sediment and displaced over- and underlying sediment during nodule growth.

Imbricated shale clasts are present at the base of some sandstone beds, and the top of sandstone beds contain current-, wave-, and interference ripple marks. Some mudstone beds contain desiccation-cracks, and birdseye structures are present in calcareous mudstones. Sandstones with interference rippled tops and mud-cracked bases are common, indicating that sands were deposited during flood tides or during storms.

In general, beds are not heavily bioturbated, and body fossils, with the exception of rare leperditiid ostracodes, are absent in these gypsum-rich strata. Along Workman's Creek near Meaford, Ontario, a small, partially dolomitized algal mound is present near the base of the Queenston. A leperditiid ostracod and algae-dominated biota seems to be characteristic of marginally hypersaline settings throughout the middle Paleozoic (Walker and Laporte, 1970).

These middle Queenston strata represent muddy tidal flat and sabkha settings in which periodic or episodic arid climatic conditions led to the precipitation of gypsum nodules. Well-developed, stable channels may have been lacking in this setting; we have not seen any large channel or bar forms from these intervals in outcrop.

Whereas the gypsum-rich lower Queenston strata of Ontario seem to represent hypersaline tidal flat conditions, normal marine tidal flats are represented in the lower Queenston strata of western New York. Unfossiliferous peritidal facies occur in the basal 9 m and the upper 40 m of Queenston formation in the Somerset drill cores. These peritidal successions, separated by the aforementioned 4.5 m of marine Queenston, include intervals of variably bioturbated, cross-stratified sandstone (often containing imbricated shale clasts) up to 15 cm thick interbedded with green and red shale. Intervals of predominantly red mudstone with minor, thin, soft sediment-deformed sandstone beds are also present. Recognizable trace fossils include shallow *Skolithos* (<6 cm deep) and small *Chondrites*. Within individual trough cross-stratified sandstone beds, trough size increases upwards indicating sand was deposited during times of increasing water velocity and/or decreasing flow depths; these sandstones may record unusually strong ebb tides. Load casts are relatively common at the bases of sandstones, especially in mudstone-rich intervals. Desiccation cracks are also present, indicating periodic subaerial exposure of tidal flat muds.

Mudstone-dominated strata are characteristic of the Queenston as a whole in western New York and are characteristic of lower and upper Queenston in the area between Toronto and Corbetton, Ontario. Four facies comprise these strata: 1) intensely bioturbated, arenaceous mudstone; 2) hackly-fracturing mudstone; 3) thin (<8 cm) couplets of fine sandstone and shale; 4) cross-stratified, upward-fining sandstone beds up to 70 cm thick.

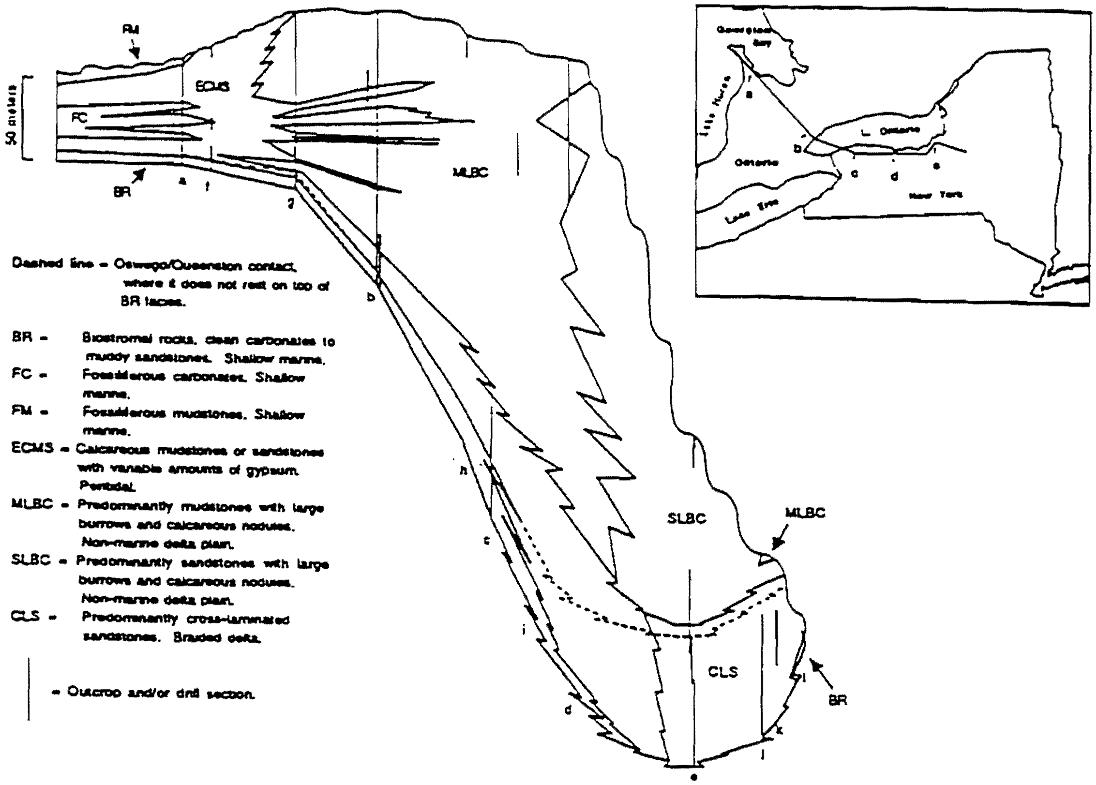

Figure 4a. Paleoenvironmental interpretations of the Streetsville (and Vincent) Mbr., Georgian Bay Fm; Oswego Fm, and Queenston Fm. Outcrop and drill sections are shown as vertical lines.

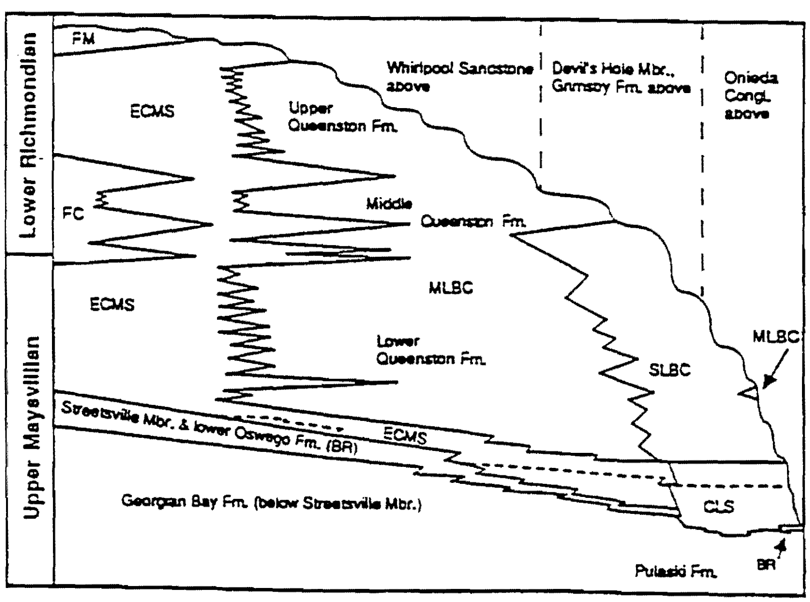

Figure 4b. Chronostratigraphic relationships of Upper Maysvillian and Lower Richmondian Strata of New York and Ontario.

The intensely bioturbated mudstones contain small (<2 cm) nodules of micrite and calcite-replaced detrital quartz sand and silt. These features are characteristic of caliche (Sakar, 1988). The upper surfaces of mudstone intervals typically contain desiccation-cracks, indicating that the mudstones were subjected to sub-aerial conditions. Hackly fracturing mudstones, gradational in character with the intensely bioturbated mudstones, contain sub-horizontal to sub-vertical slickensides. In outcrop, these mudstones weather to prismatic, angular blocks. This weathering pattern, in conjunction with possible caliche nodules, suggests that these bioturbated and hackly-fractured mudstones may be paleosols. They have characteristics of B-horizon soils, as outlined by Retallack (1988). We have not yet recognized distinctive A-horizons within these mudstones, but this, in part, may be due to erosional truncation of the upper horizon.

Sandstone/shale couplets form successions up to 7 m thick. Sandstones typically exhibit some degree of grading and are cross-stratified. Superified cross stratification (climbing ripples) is common in thicker (>2 cm) sandstone beds indicating rapid deposition. We interpret couplets as representing flooding events in a delta plain setting, somewhat proximal to distributary channels.

Upward-fining sandstones, up to 70 cm thick, occur as both individual and amalgamated beds, forming arenaceous intervals up to 2.5 m thick. Within the thicker upward-fining sandstone beds, a consistent internal stratigraphy is present. Graded intervals containing imbricated shale rip-up clasts and/or calcareous nodules are present at the bases of most sandstones. These graded lag deposits are overlain by cross-stratified sands. Sandstones typically form distinctive accretionary bar forms (up to 1.5 m high), shallow (<75 cm) sand-filled channels, and distinctive sheets. Orientations of accretion surfaces and foresets from cross-stratified sheet sands at Rochester, New York, indicate northward directed paleoflow (Zerrahn, 1978).

Sandstone beds contain distinctive sub-vertical to sub-horizontal trace fossils, essentially identical to putative terrestrial arthropod burrows (Retallack, 1985; Feakes and Retallack, 1988) described from the correlative Juniata Sandstone of Pennsylvania. These burrows are up to 25 mm in diameter and may penetrate 30 cm or more of strata.

Low-Mg calcite ensheathes mud-filled burrows and also decreases upward from the bases of sandstone layers, so that sandstone beds typically exhibit downward color gradations from white to pink. The gradational, downward increase in these calcite cements is reminiscent of groundwater dolocretes described from the Triassic strata of the Paris Basin by Spotl and Wright (1992). The preferential cementation around burrows and the thinness of calcite-cemented sandy intervals, however, leaves open the possibility that the Queenston cements had a pedogenic origin.

The upward-fining sandstone beds represent tidally-influenced, non-marine distributary channel deposits. The presence of mud drapes within accretionary bar forms indicates periodic or episodic cessation of channel flow, possibly associated with slack tides, and the lack of thick channel deposits suggests that the distributary channels were migratory.

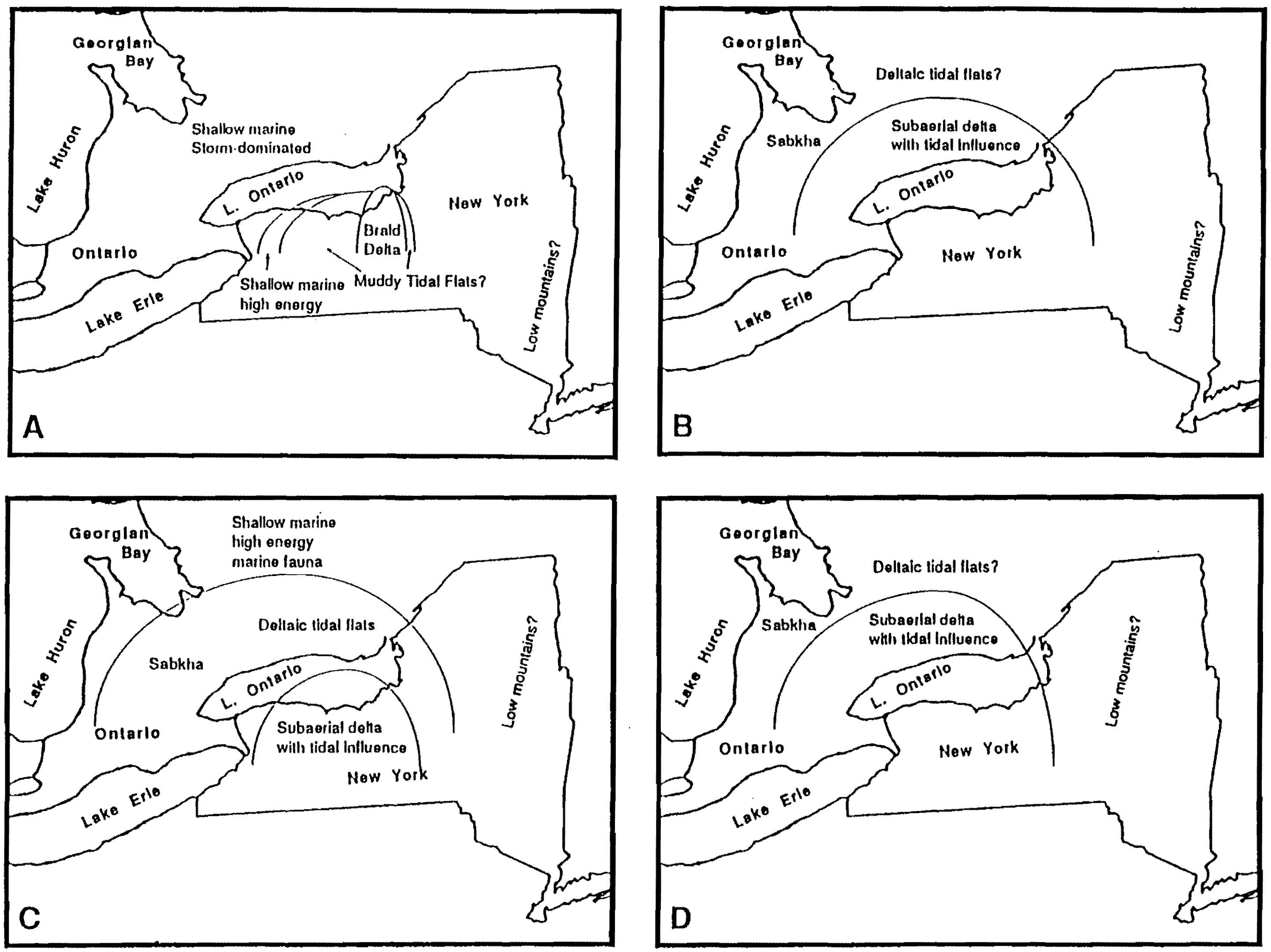

Figure 5. Paleogeographic reconstructions of the Queenston delta. A) During Oswego deposition in north-central New York. B) During lower Queenston deposition in southwestern Ontario. C) During middle Queenston transgression. D) During upper Queenston deposition in southwestern Ontario.

Thus, sedimentologic, stratigraphic, and biostratigraphic data indicate that the Queenston Formation represents a northward prograding complex of shallow marine to non-marine environments (Figures 4, 5). The presence of synsedimentary to very early diagenetic evaporite minerals in tidal flat deposits indicates hypersaline conditions in southwestern Ontario, whereas similar evaporite minerals are generally absent in peritidal paleoenvironmental settings in New York. It seems likely that proximity to freshwater (i.e., the Queenston/Juniata delta) may have influenced evaporitic mineralization.

Regional trends in stratal thickness, maximum grain size, and paleocurrent data suggest that the molasse present in New York and Ontario is part of a clastic wedge which prograded from southeastern Pennsylvania (Figure 1; see Lehmann, 1993, for discussion). The molasse accumulated in a gently north-plunging elongate basin. Based on correlation of gypsum beds and facies relationships, sabkhas were greater than 40 km in width, and tidal effects influenced sedimentation on delta plains more than 400 km inland from fully marine settings. The gentle topographic gradient responsible for these expansive facies belts was presumably created and maintained as sediment supply overcompensated for accommodation space produced by basin subsidence.

The Queenston red beds represent a large deltaic complex that prograded into a shallow marine basin. At Manitoulin Island, approximately 120 km northwest of Wiarton, Ontario, blue-gray to buff colored limestones and dolostones are time correlative with the Queenston red beds which are present to the southeast. Reports of fossiliferous Queenston strata at Russel, Ontario (27 km southwest of Ottawa) suggest that fully marine conditions existed to the northeast of Rochester along the basin axis as well (Wilson, 1946).

The lobe of the Queenston-Juniata siliciclastic wedge that prograded from Pennsylvania into Ontario is presumably one of many coalescing red bed lobes which developed adjacent to Taconic highlands along eastern North America. Although the deltaic complex discussed herein apparently thins and grades into marine-dominated facies to the north, at least 400 to 600 m of Upper Ordovician non-marine red beds (Becancour Formation) have been reported from the subsurface of the St. Lawrence lowlands, approximately 35 km northwest of Montreal (Belyea, 1952). The thickness of the Becancour Formation suggests that a regional depocenter, and perhaps a local sediment source, also existed in that region.

Additional work on the physical stratigraphy of the Queenston Formation remains to be completed before a detailed sequence stratigraphy may be interpreted. The facies analyses presented in this paper serves as a prerequisite for synthesis of the regional-scale physical sequence stratigraphy of the Queenston Formation.

SILURIAN SEQUENCES

Because of good exposure, the litho- and biostratigraphy of the Silurian strata in western New York and Ontario have been progressively refined over the past 150 years (Figure 6). Because of the vast amount of knowledge that has accumulated on the Silurian strata, the interval is an ideal laboratory for the testing of various stratigraphy-based models. In recent years, the present authors have evaluated prevalent models of allostratigraphy and foreland basin dynamics underlying the Silurian strata of the northern Appalachian basin (Brett et al., 1990a, b,; Goodman and Brett, 1994).

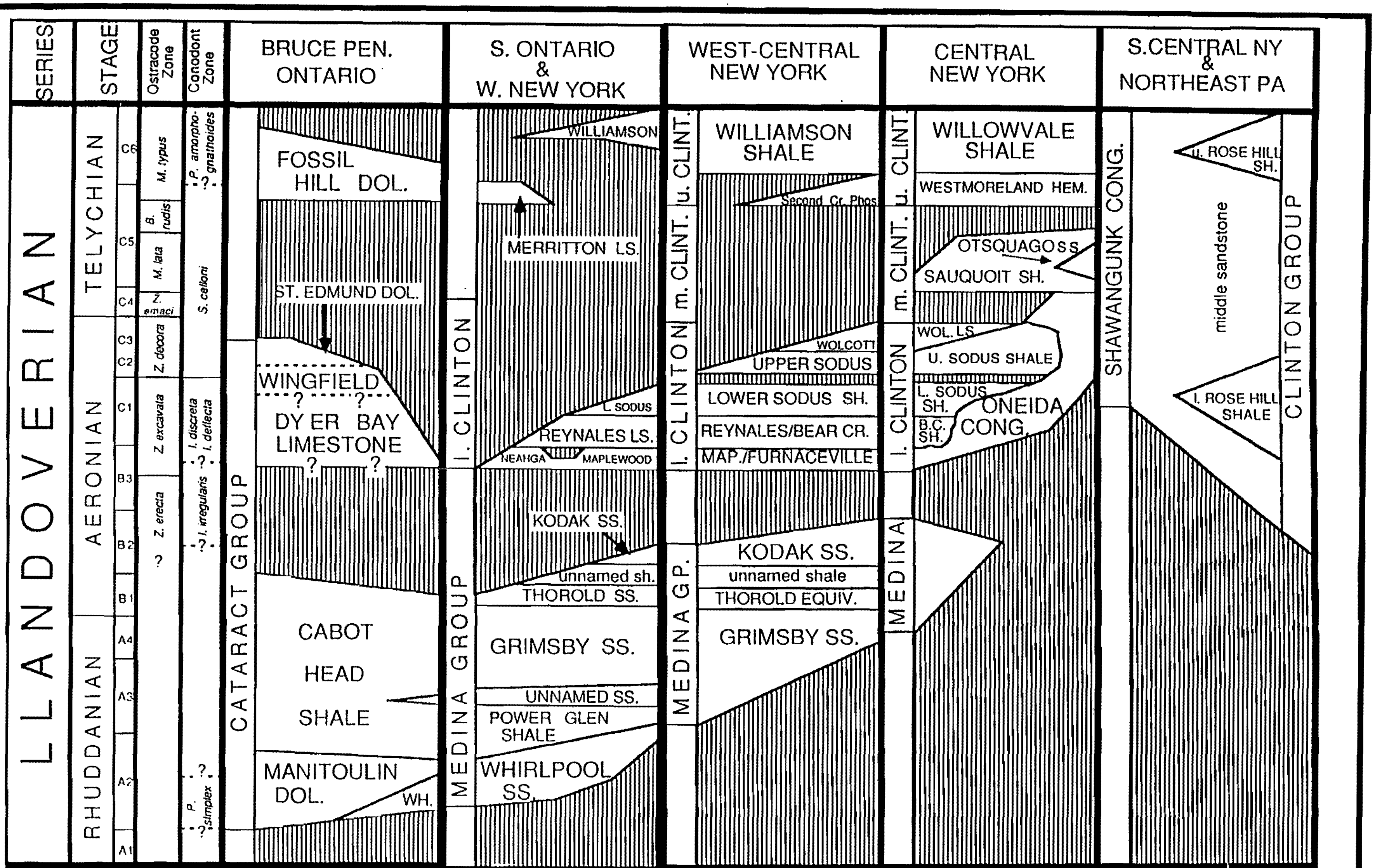

Figure 6a. Chronostratigraphic relationships of Lower Silurian (Llandoverian) strata of sequences I and II in Ontario (Bruce peninsula), western, central and eastern New York State. Vertical ruling indicates unconformities.

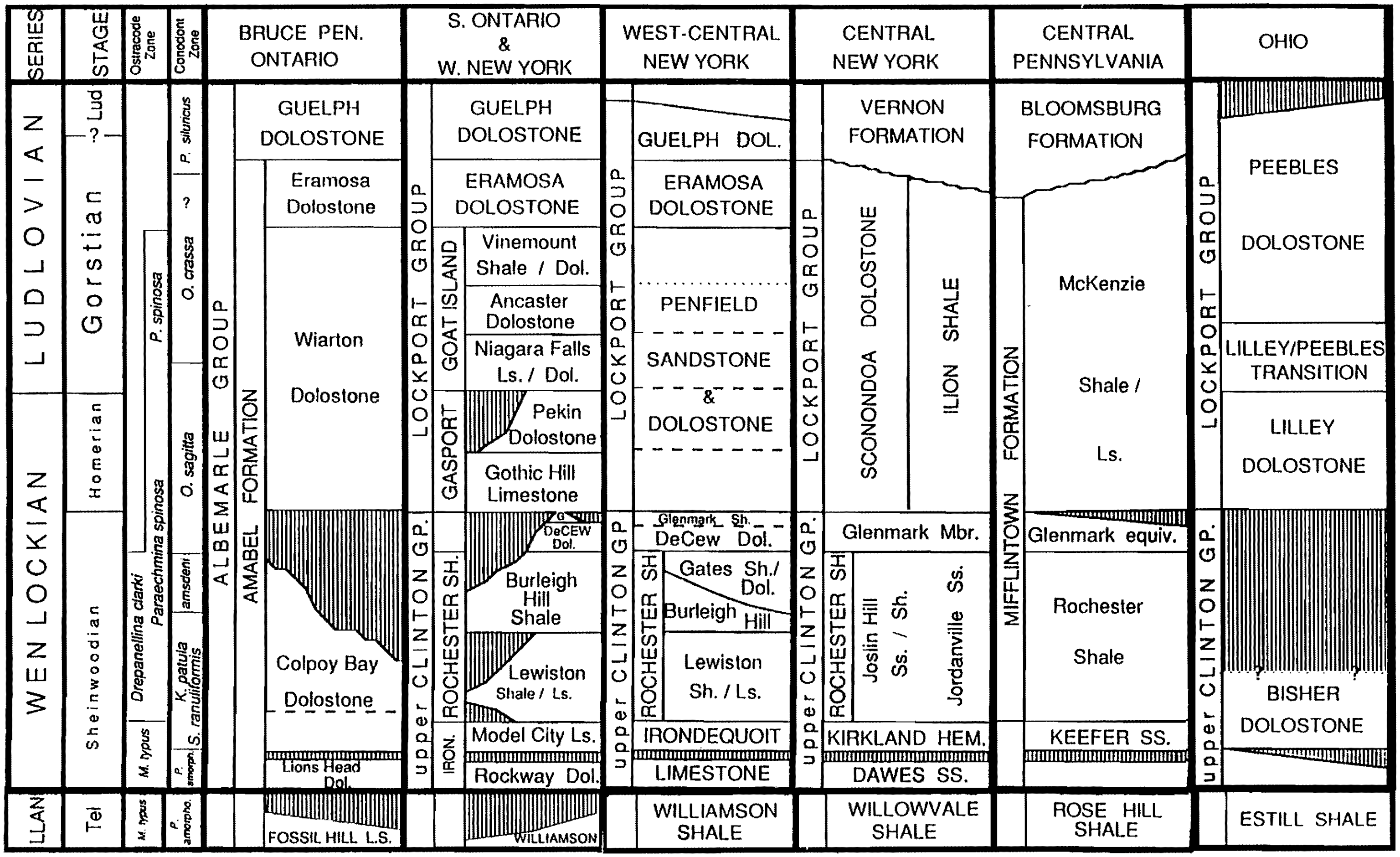

Figure 6b. Chronostratigraphic relationships of units within the upper Clinton (Sequences IV, V) and Lockport (Sequences VI) Groups in Ontario, New York, Pennsylvania and Ohio. Formation names are listed in upper case, members in lower case.

The Llandoverian to Ludlovian (Medina through Lockport Group) succession of the northern Appalachian foreland basin has previously been divided by Brett and others (1990a, b) into at least six large-scale, unconformity-bounded stratal packages comparable in many respects to *depositional sequences* of Vail and others (1977, 1991). For the most part, these sequences correspond to previously recognized group-level stratigraphic units. The first sequence corresponds to the Medina Group, the second and third sequences to the lower and middle portions of the Clinton Group, respectively, the fourth and fifth sequences to portions of the upper Clinton Group, the sixth sequence to the lower Lockport Group. A newly recognized seventh sequence corresponds to the upper Lockport Group and Vernon Shale (Figure 7).

The sequences recognized herein are at least crudely divisible into *systems tracts* that are analogous to those within depositional sequences of seismic stratigraphers (e.g. Vail et al., 1977; Van Wagoner, 1988; Posamentier and others, 1988). In terms of temporal magnitude, Silurian sequences, like those of seismic stratigraphers, encompass approximately 1 to 5 million years and, therefore, can be classified as third-order cycles (Vail et al., 1977).

The Silurian sequences are divisible internally into very prominent and basin-wide *sub-sequences*, which are of lesser temporal magnitude and display sharp, slightly erosive disconformities. Sub-sequences are considered to reflect shorter duration, roughly 1.0 to 1.5 million year cycles in the range referred to by Busch and Rollins (1984) as fourth-order cycles. Sub-sequences are also comparable to Carboniferous mesothems described by Ramsbottom (1979) that apparently record rapid regressions followed by progressive, stepwise transgressions. Other researchers, including Vail and others (1991), have recently broadened the temporal range of third-order cycles to approximately 0.5 to 5.0 million years. This broader definition of third-order cycles would also include sub-sequences of Brett and others (1990b). We choose to retain the Busch and Rollins (1984) classification of allocycles and to consider the smaller-scale sub-sequences as a discrete scale in a hierarchy, because of the following: a) the bounding discontinuities (except those which also coincide with sequence boundaries) involve less erosional truncation than do third-order sequence boundaries; and b) two or more sub-sequences consistently occur within each interpreted Silurian sequence, suggesting a nested hierarchy of allocycles. Sub-sequences are, in turn, divisible into smaller-scale cycles that correspond roughly to members or submembers in lithostratigraphic terminology. Each sub-sequence, thus, is divisible into two to three cycles that are analogous to *parasequence sets* or fifth-order cycles. The smallest scale cycles that can be correlated over significant portions of the basin tend to be arranged within parasequence sets in groupings of two to five. These *parasequences* commonly exhibit an asymmetrical upward-shallowing motif and correspond to sixth-order cycles or PACs (Goodwin and Anderson, 1985).

In the following sections, we describe the seven major sequences and their internal cyclicity. Details of the outcrops in the Genesee Valley are emphasized.

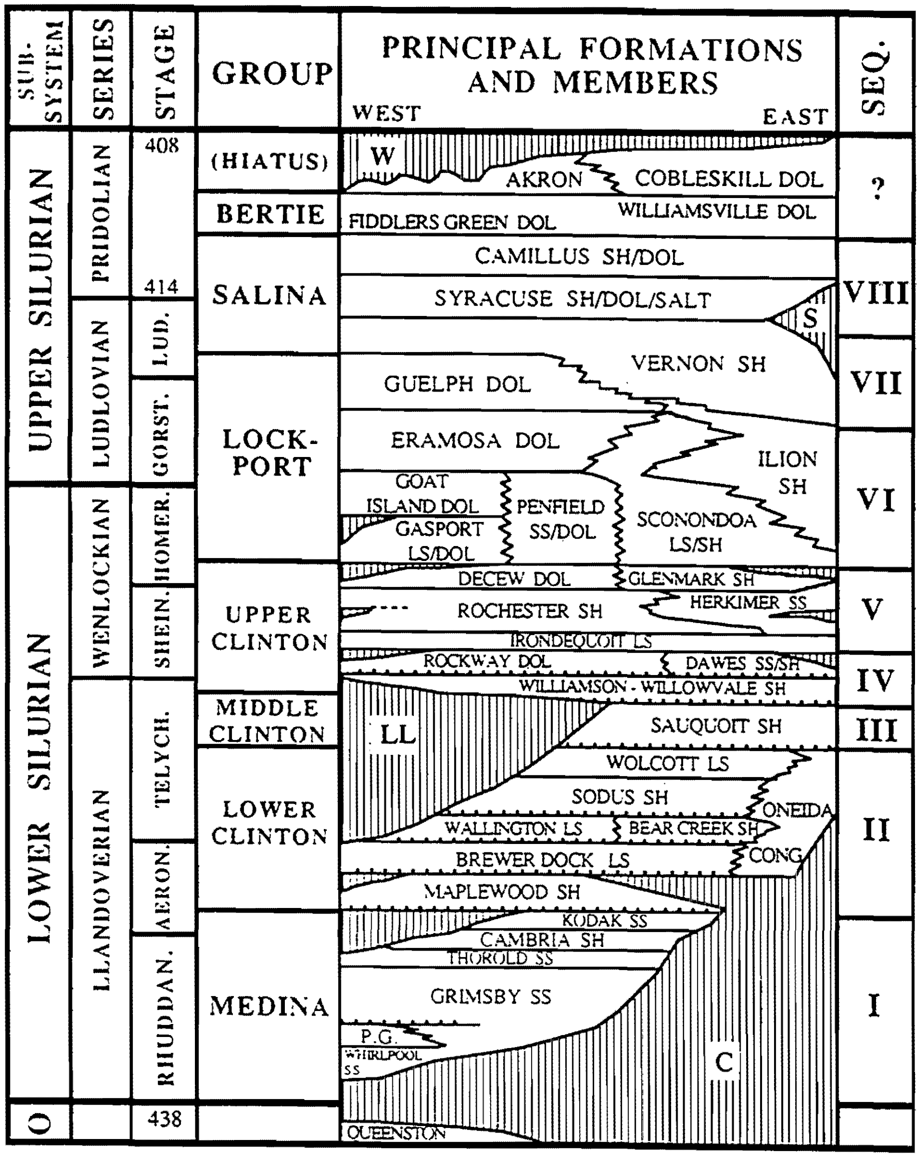

Figure 7. Generalized Silurian stratigraphy of New York State illustrating division into third-order depositional sequences. Vertical axis is scaled to time; horizontal scale is west to east distance along the New York outcrop belt from Niagara Gorge to near Utica, N.Y., approximately 300 km. Abbreviations: O = Ordovician; Rhuddan. = Rhuddanian; Aeron. = Aeronian; Telych. = Telychian; Shein. = Sheinwoodian; Homer. = Homerian; Gorst. = Gorstian; Lud. = Ludfordian. Ages listed are estimates for beginning and end of Silurian Period and end of Ludlovian (in millions of years). For formations and members: P.G. = Power Glen Shale, DOL--dolostone, LS = limestone, SH = shale, SS = sandstone; dotted lines indicate persistent hematitic/phosphatic beds; major unconformities are labeled with upper case letters C = Cherokee (basal Silurian): LL--Late Llandovery; S = Salinic, W = Wallbridge; SEQ. = sequences. Chart modified from Rickard (1975).

Sequence I: Medina Group

The Medina Group at Rochester Gorge consists of 17 m (50 ft) of red- and white-mottled, interbedded sandstone, siltstone and mudstone (Figure 8). Prior to recent allostratigraphic analyses (Duke and Fawcett, 1987; Brett and others, 1990a, b; Duke, 1991; Goodman and Brett, 1994), the Medina Group strata near Rochester were assigned to only two units, the Grimsby and Kodak Formations. The refined correlations that have resulted from recent investigations suggest that the Rochester area section contains intervals equivalent to at least four newly defined formations in the Niagara region (Figure 9).

The basal three meters of red, coarse-grained sandstone appear to correlate with the upper half of a thick basal sandstone interval quarried near Albion, Orleans County and with the Devils Hole Sandstone which overlies the Power Glen Shale in Niagara County. At Rochester, the Devils Hole Sandstone is a massive bed of planar laminated to cross-stratified, subarkosic, medium to coarse-grained sandstone. The bed also contains iron and phosphate-rich ooids and scattered crinoid ossicles. The sharp, slightly undulatory, basal contact with the Ordovician Queenston Formation is a low angle regional unconformity of considerable magnitude. This contact is the Cherokee Unconformity of Dennison and Head (1975).

The unconformity and unusual petrography of the Devils Hole Sandstone are readily understood in the context of the regional sequence stratigraphy (see Figures 8 and 9). The Medina Group has recently been interpreted to be a single depositional sequence sensu Wilgus and others (1988) and to represent stratigraphic accumulation during a single, third-order cycle of sea-level rise and fall (Brett and others, 1990a, b; 1991; Duke, 1991). The Cherokee Unconformity at Rochester is the surface of maximum starvation that separates strata of the transgressive systems tract (Whirlpool Formations of the Niagara region) from overlying strata of the highstand systems tract. In the Niagara region, both transgressive and highstand strata can be observed in the same stratigraphic section. Because marine transgression progressed from west to east through the Rhuddanian and Aeronian Stages, the oldest Medina Group strata pinchout west of Rochester. Only after the maximum sea-level highstand did sufficient accommodation space exist on the local paleotopographic high of the Rochester area for accumulation of condensed, onlapping, marginal marine strata. Ten km to the east at Webster, the Medina section expands to approximately 25 m (Fred Amos, personal communication). The increased thickness is attributable to increased preservation of basal strata. This pattern reflects the regional-scale paleotopographic relief of the Cherokee Unconformity that, for the most part, appears planar at the outcrop scale.

Medina Group strata that overlie the Devils Hole Sandstone, therefore, comprise the highstand systems tract of Silurian Sequence I. These strata are now assigned to the Grimsby (restricted sense), Thorold, Cambria and Kodak Formations (see Figure 8). These formations comprise three (subsequence) coarsening cycles on the scale of 3-5 m that record a pulsed, westward progradation of the eastern strandline during a late Rhuddanian to early Aeronian sea-level highstand. The basal sandstones of each cycle are intensely bioturbated by *Daedalus* and *Arthrophycus* which probably are mining traces (fodinichnia) of large annelid worms. Where primary sedimentary structures are preserved, however, the sandstones contain lateral accretion surfaces (Thorold Formation) or alternating tabular beds of sandstone and shaley siltstone (Kodak Formation).

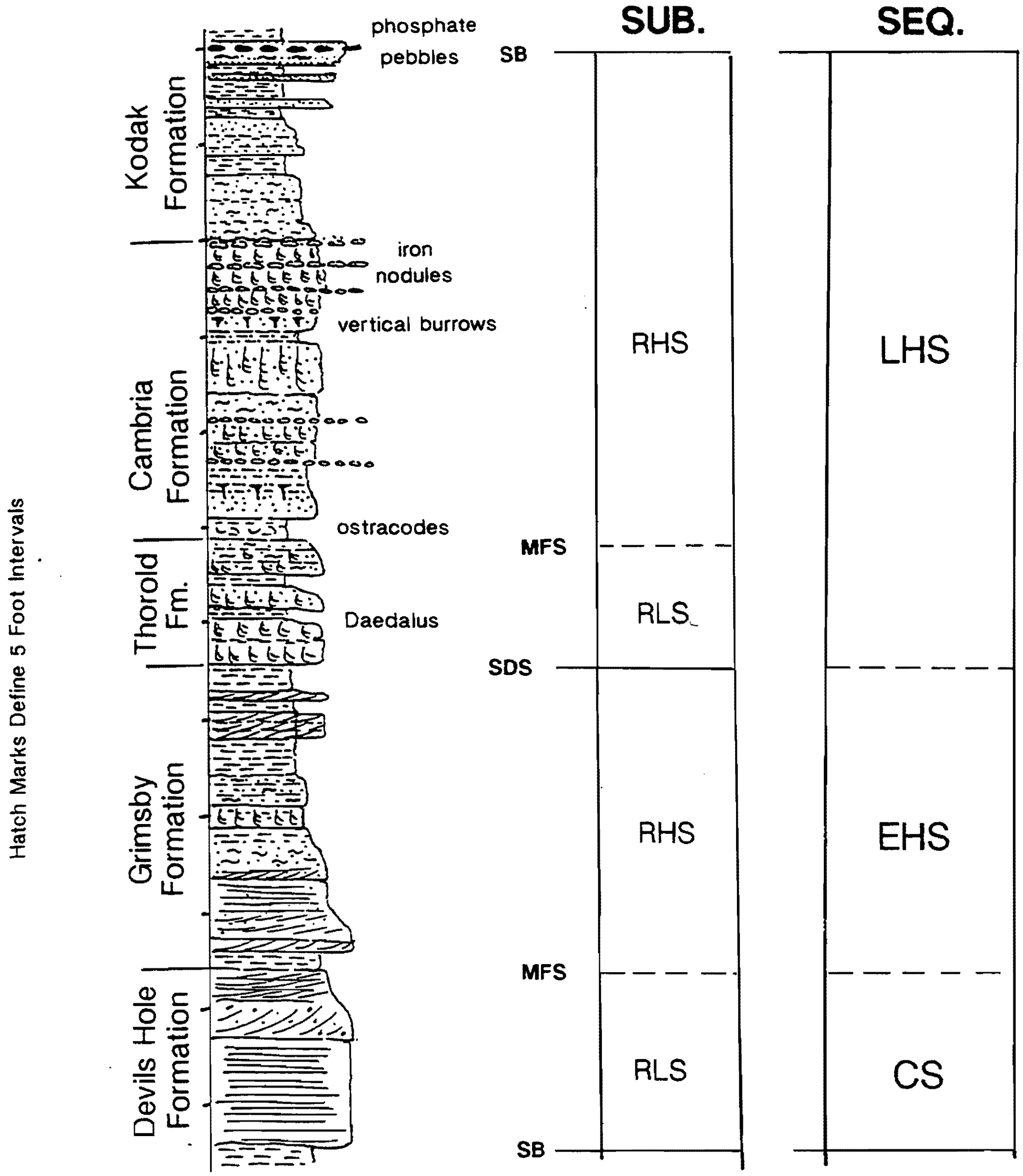

Figure 8. Stratigraphic column for the Medina Group (Sequence I) from a drill core in Rochester, N.Y.

Bars on right side of figure indicate subdivisions of subsequences (SS; left bar) and for systems tracts the sequence as a whole (SEQ; right bar). Abbreviation for subsequences: NM = non-marine lowstand deposit; RLS = relative lowstand (shallow marine deposit); RHS = relative highstand; MFS = marine flooding surface; SDS - sealevel drop surface; CI = condensed interval. For sequences SMT = shelf margin systems tract (lowstand deposit); TST = transgressive systems tract; MFS = marine flooding surface; CS = condensed section; EHS = early highstand; LHS = late highstand; SB = sequence boundary; TS = transgressive surface.

The Grimbsy consists primarily of red and white mottled sandstones, commonly with basal lags of shale pebbles interbedded with maroon silty shales. At Rochester marine shelly fossils are absent. Lingulids are abundant tot he west near Lockport. Minor coarsening-upward cycles, capped by tidal channel sandstones were recognized in the Grimsby by Duke (1991).

Consequently, the regressive phases of the sea-level cycles resulted in the progradation of tidal channel and tidal flat facies over muddy subtidal to lower intertidal facies that contain ostracodes, lingulid brachiopods, pelecypods, trilobite trace fossils (*Rusophycus*), and, possibly, hematized stromatolites.

In the Genesee Gorge, the upper boundary of the Medina Group is placed at the contact between the Kodak Sandstone and the overlying Maplewood Shale of the Clinton Group. A phosphate pebble bed marks a low angle, regional unconformity at this contact. To the west, upper Medina strata (Thorold, Cambria, Kodak) are beveled beneath this unconformity. The area of maximum erosion may be located between Hamilton and Guelph, Ontario. Conversely, this region was close to the basin center (deepest water facies during deposition ofthe lower Medina (Cabot Head) strata. This region appears to have become a cratonic arch that separated a subsiding foreland basin with an eastward migrating depocenter in New York from the more stable-positioned, southeast margin of the intercratonic Michigan Basin near the end of Medina deposition. Consequently, although this unconformity is likely to have been formed during a sea-level lowstand following progradational infilling of the Appalachian Basin by upper Medina strata, strong evidence of early Aeronian foreland basin flexure with concomitant uplift of the cratonic arch is apparent in the stratigraphic architecture of the region.

Sequences II-V: Clinton Group

The Clinton Group at Genesee Gorge consists of a diverse assemblage of shale and carbonate formations with associated hematitic ironstones and phosphatic beds. The total thickness of the Clinton Group is approximately 55 m (170 ft). The Clinton strata span the upper Aeronian through Sheinwoodian stages in the Niagara region. In western New York, the Clinton Group is divided into upper and lower halves by a major unconformity (Gillette, 1947; Rickard, 1975; Lin and Brett, 1988). Farther west in Ontario, Canada, the entire lower Clinton is truncated beneath the major unconformity. Farther east in the type Clinton area of east-central New York, the lacuna of the unconformity decreases; lower and middle Clinton strata only briefly present at Rochester are preserved. The middle Clinton strata are only briefly discussed in this paper but have been studied in detail by Gillette (1947) and Muskatt (1972).

Sequence II: Lower Clinton Group

The lower Clinton Group has been previously interpreted as a discrete depositional sequence (Sequence II of Brett et al., 1990a, b, Figure 10). Lower Clinton strata, however, have proven to be a challenge to interpret in terms of systems tracts because of the following attributes: 1) undulatory onlap surface (Medina-Clinton contact) and resultant local, lateral discontinuity of basal Clinton units; 2) complete absence of lower Clinton strata west of St. Catharines, Ontario, due to truncation beneath the middle Clinton unconformity; 3) lack of a diagnostic benthic assemblage (sensu Boucot, 1975) in the Maplewood Shale; and 4) inverse relationship between benthic assemblages used to define relative sea-level by

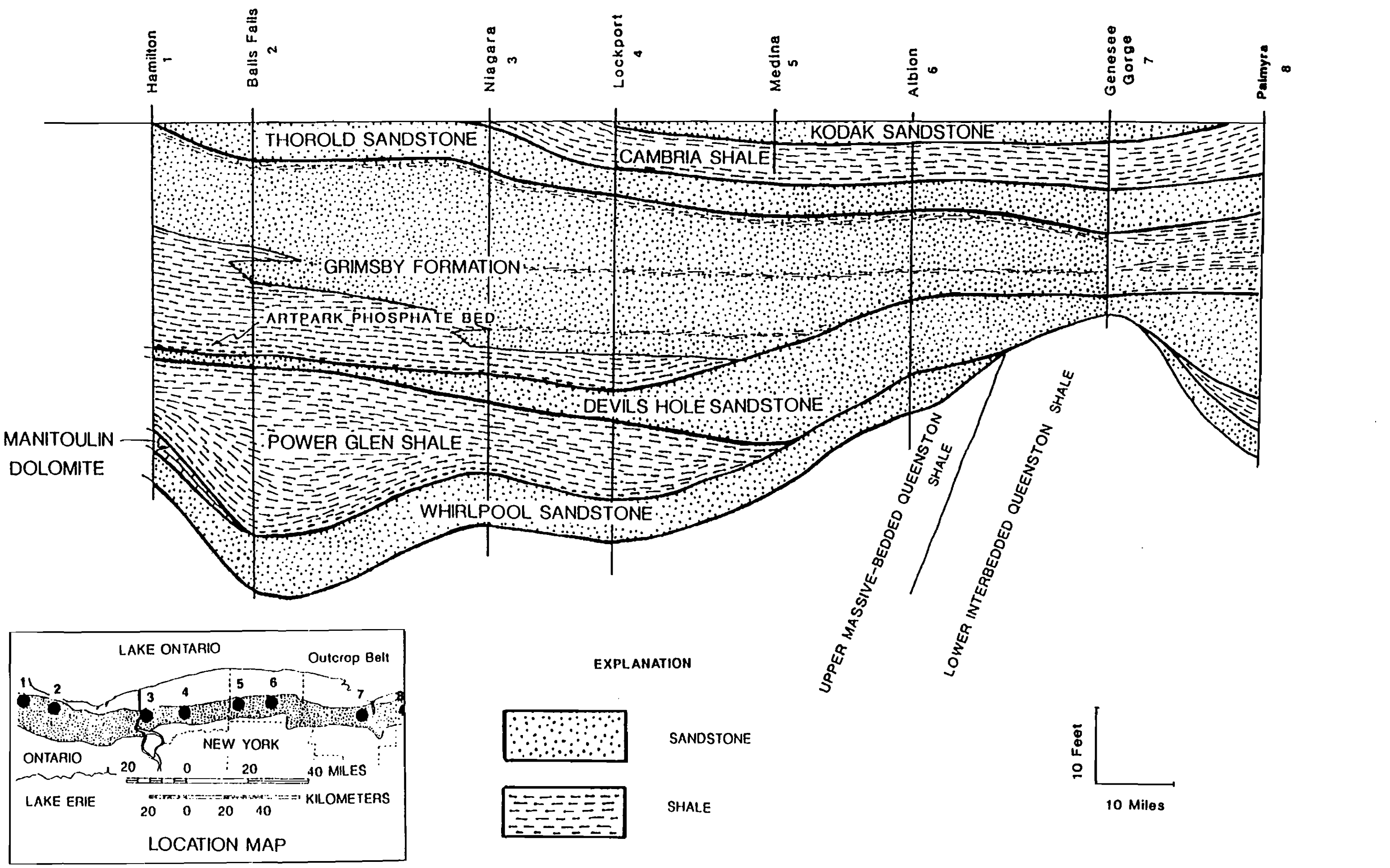

Figure 9 -- Diagrammatic stratigraphic relations in the Medina Group, with regional correlations between Hamilton, Ont. and the Genesee Gorge in Rochester, N.Y. Length of vertical line at each locality indicates units observed in the field or in drill cores. (Modified from Brett and others, 1991, fig. 7.)

paleontologists and carbonate grain-size trends used for the same purpose by sedimentologists. Detailed stratigraphic performed by LoDuca and Brett (1994) has clarified some of the relationships between strata at Rochester and those in Wayne County. Additional work with drill cores between Wayne and Madison Counties will be helpful in completing the correlations across the basin. Drill cores were once available (see Gillette, 1947), but they were apparently lost in a fire at the office of the state geological survey.

The Maplewood Shale, lowest of the Clinton units, consists of 6 m (18 ft) of soft, green-gray, clay shale (Figure 11). Despite its prominence in the Genesee Gorge, the unit pinches out at Webster, only 10 km east of Rochester (Fred Amos, personal communication). The Maplewood thins to about 2 m at Albion, Orleans County and to less than 30 cm at Lockport, Niagara County, where it is mapped as Neahga Shale. The Maplewood-Neahga Shale is about 2 m thick in the Niagara Gorge.

The Maplewood Shale is bounded both at its base and top by phosphatic pebble beds. The basal bed, designated the Densmore Creek Phosphate Bed by LoDuca and Brett (1994), may be up to 10 cm thick and contains large clasts of phosphatized, arenaceous carbonate and black, fossil steinkerns of fluorapatite (Paxson, 1985). Phosphatic clasts and fossils are often piped downward 5-10 cm into burrows that penetrate the underlying uppermost bed of the Medina Group. The phosphate bed is traceable westward from Rochester to St. Catharines, Ontario where it is represnted by phosphatic limestone. The Densmore Creek Phosphate Bed grades eastward from Rochester into the base of the hematitic and phosphatic carbonate intraclast and oolite facies (Webster Bed) of the Furnaceville Formation (LoDuca and Brett, 1994). In a sequence stratigraphic context, this lower phosphate bed records the basal Clinton marine transgression over the beveled upper Medina surface. No nonmarine strata are preserved immediately above the unconformity along the New York-Ontario outcrop belt. Consequently, if a lowstand systems tract exists within the basin, it may be found in the sections of central Pennsylvania and/or in the subsurface of southern New York. The Maplewood Shale has been tentatively interpreted as part of the transgressive systems tract of Silurian Sequence II (Brett and others, 1990a, b).

The upper phosphate bed, designated the Budd Road Phosphate Bed, occurs at the contact of the Maplewood Shale with the overlying Brewer Dock Member of the Reynales Formation. The upper phosphate bed is well developed at the Genesee Gorge but is generally less distinctive than its lower counterpart. In a sequence stratigraphic context, the upper phosphate bed may signify the onset of clear-water, carbonate deposition with progressive sea-level rise during the middle Aeronian (C_1) stage.

Macrofossils are very rare in the Maplewood Shale and most that have been reported likely come from the basal phosphate bed (Fisher, 1953a). Nonetheless, fully marine conditions are suggested by the fauna. Fisher 1960 reported a nektonic trilobite from the shale. More recently, Sam Ciurca (personal communication) has obtained lingulid brachiopods, nautiloids, eurypterid fragments, and complete specimens of a new species of crinoid from the Maplewood Shale. The authors have observed gastropods and brachiopods (*Eocoelia, Hyattidina* and *Leptaena*) preserved as steinkerns in the lower phosphate bed. The shale also contains a diverse and well-preserved microflora of acritarchs and chitinozoans (Fisher, 1960; Miller and Eames, 1982). Based upon the paleontology and lateral stratigraphic relationships, the Maplewood-Neagha Shales were probably deposited

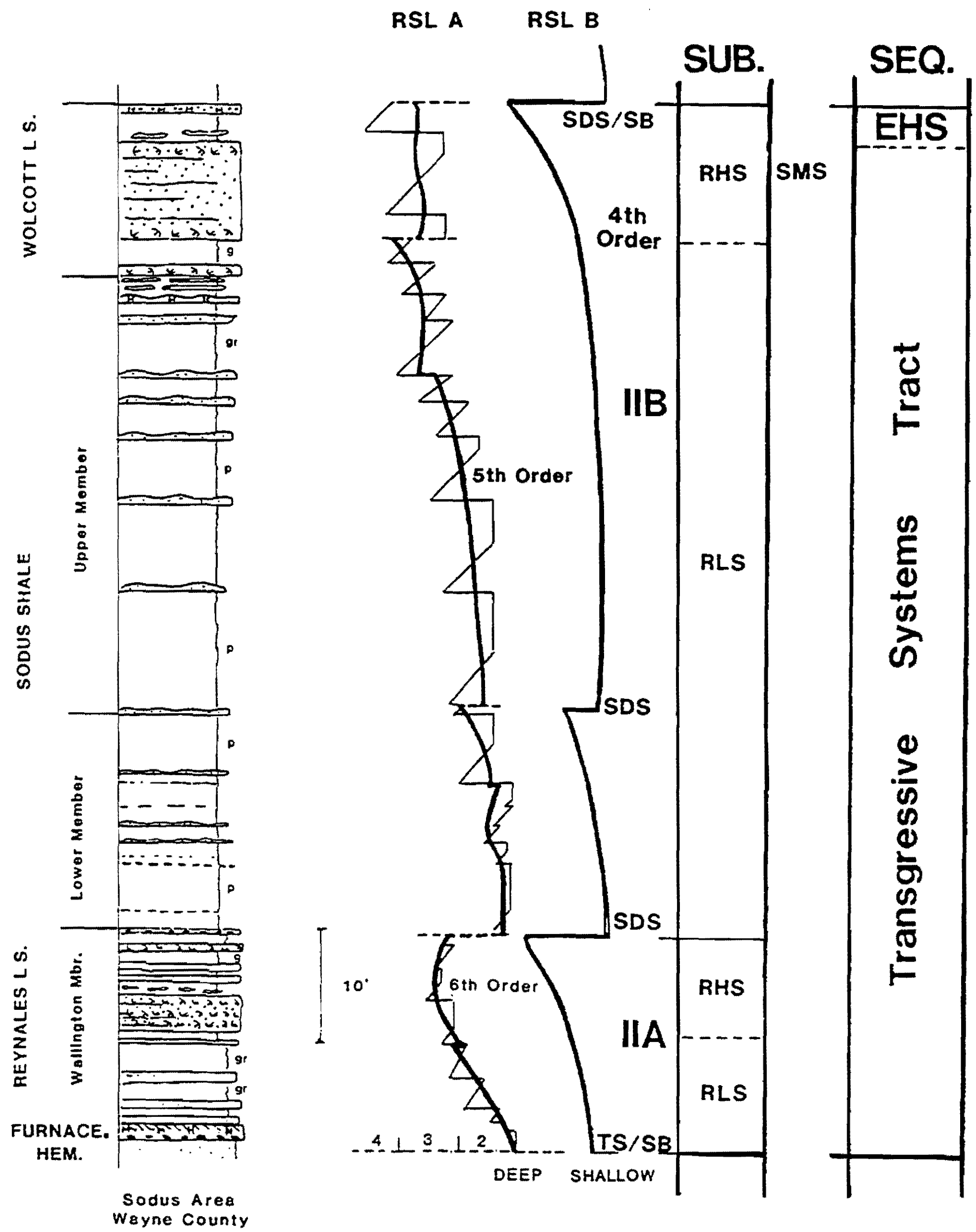

Figure 10. Lithostratigraphy, inferred relative sea-level, and sequence terminology of the lower Clinton Group (Sequence II) near Sodus, Wayne County, N.Y.: RSLA = relative sea-level curves for the 5th- and 6th-order cycles. Calibration based on biofacies: 2 = *Eocoelia* association (BA-2); 3 = pentamerid association (BA-3): 4 = stricklandid association (BA-4).

in a variably deep, stagnant trough immediately offshore of a high energy strandline located near Webster, Monroe County. The thinness of the unit near Lockport, Niagara County may suggest that the Neahga and Maplewood Shales were deposited in weakly differentiated sub-basins separated by a minor topographic high.

The Reynales Formation is a carbonate-rich interval that is well exposed along the RG&E access road locality, (Stop 1A) within the Genesee Gorge. The age of the Reynales has been the subject of some debate (see LoDuca and Brett, 1994). On the basis of ostracodes and extensive conodont sampling in recent years, however, the entire Reynales Formation is presently believed to be Aeronian (C_{1-2}) in age (Maxwell and Over, 1994;Mark Kleffner, personal communication). To the west, the formation (Hickory Corners Member) has been recognized only as far as Queenston, Ontario. To the east of Rochester, the Reynales Formation grades laterally into the type Furnaceville Hematite and overlying Bear Creek Shale.

In the Rochester area, the formation has been divided into three units which are in ascending order: the Brewer Dock Member; the Furnaceville Hematite Bed; and the Wallington Member. The name of the hematite bed at Rochester has been changed to Seneca Park Bed by LoDuca (1988) and LoDuca and Brett (1994), because the distinctive bed within the Reynales correlates with only part of the type Furnaceville Hematite exposed in now flooded quarries near Ontario, Wayne County. The type Furnaceville Hematite represents a thoroughly hematized equivalent of the entire Brewer Dock Member at Rochester.

In the Genesee Gorge, the Brewer Dock Member is approximately 60 cm thick and consists of the basal Budd Road Phosphate Bed, and interbedded pelletal calcisiltites and green-gray shale. Soft sediment deformation features (ball and pillow structures) are locally exposed in the gorge. The member is fossiliferous; the fauna consists primarily of bryozoans, brachiopods and crinoid ossicles. The primary brachiopods are *Eocoelia* and *Hyattidina*.

The Seneca Park Hematite Bed overlies the Brewer Dock Member. In the Genesee Gorge, the hematite stands out as a striking red bed within the light gray carbonate sequence. The bed is approximately 30 cm thick and consists of hematited, bryozoan-rich grainstone. The hematite bed is distinctly cross-bedded. The direction of bedform migration was to the northwest. The Seneca Park Hematite and the Furnaceville Hematite likely represent sandwaves deposited on the interior margin of storm wave-base environments. These bioclastic beds differ petrographically from phosphatic, oolitic hematite beds such as the Webster Bed that appear to have been generated during sediment starvation associated with marine flooding surfaces (LoDuca and Brett, 1994).

The Wallington Member is a 5-6 m (16-18 ft) thick interval of medium- to thick-bedded brachiopod packstones and pelletal grainstones. At the Seth Green Drive section, imbricated beds with sigmoidal geometries occur at one level. Although large bioclasts disrupt internal patterns of lamination, similar beds in the Devonian Becraft Limestone have been argued to be tidal bundles by Ebert (1987). Thin partings of green-gray, silty shale occur near the base of the member. Several beds in the lower portion of the member also contain blue-gray chert nodules. Sedimentary structures and biofacies suggest deposition at and above storm wavebase. Fossils include the brachiopods, *Eocoelia* and *Pentamerus*, stromatoporoids, and complete specimens of an unusual crinoid *Stipatocrinus* (Eckert and Brett, 1987). Stromatoporoids are common in the uppermost beds.

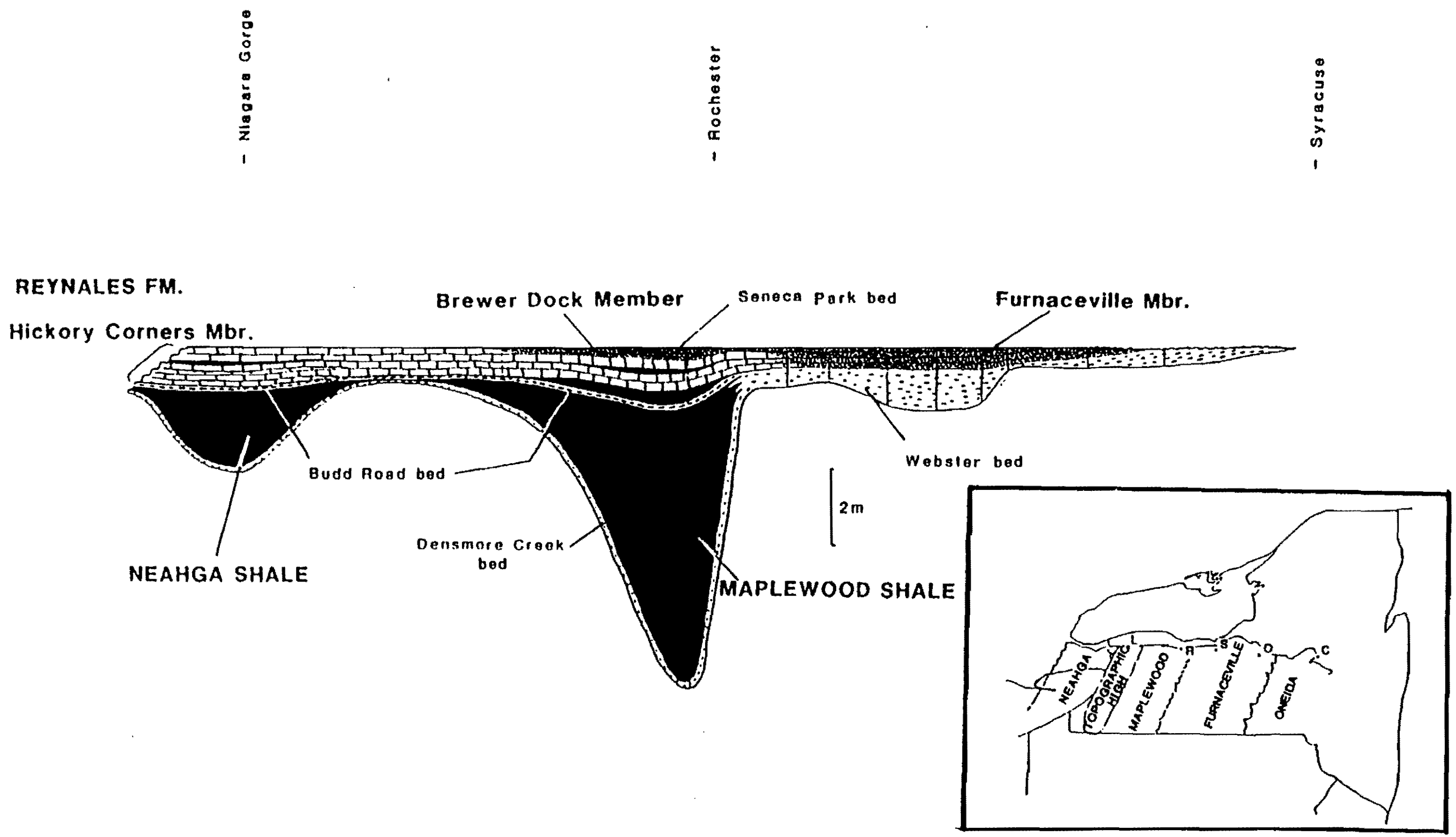

Figure 11. Regional cross-section of the lower part of the Clinton Group between Syracuse, New York, and St. Catharine's, Ontario. Note the presence of two minor depocenters (for Neahga and Maplewood Shales, respectively) separated by a minor arch. Also, note eastward passage of Maplewood Shale into a highly condensed, multigenerational phosphatic conglomerate (Webster Bed). West of Niagara Gorge, the Neahga Shale is truncated by the major mid-Clinton Group angular unconformity.

One distinctive carbonate bed containing stromatoporoids is separated from the main carbonate body of the Wallington Member by a prominent purple shale bed that resembles the overlying Sodus Shale Formation. Thus, although the lithologies of discrete beds are distinct, a stepped transition in bedding sequence occurs at the Wallington-Sodus Shale contact. The formation boundary is typically placed at the upper contact of the distinctive stromatoporoid-bearing bed.

Six pentamerid-rich grainstone beds recognized within the Wallington in the Genesee Gorge define the bases of small-scale, asymmetrical carbonate-to-shale cycles that resemble punctuated aggradational cycles (PACs). The pentamerid beds are positioned at the base of the Wallington Member, and at approximately 1.4 m, 2.0 m, 2.8 m, 3.2 m and 4.0 m above the base. The uppermost cycle of the Wallington in the Genesee Gorge consists of the condensed, dark-stained stromatoporoid bed referred to previously. This distinctive bed appears to grade eastward into the 10-20 cm thick Sterling Station Hematite.

Westward from Rochester, outcrops of the Wallington Member are rare although pentamerid-rich limestones are known from scattered patchy outcrops in the bed of Salmon Creek between Spencerport and Brockport. At least one distinctive pentamerid bed has also been observed in drill cores near Albion, Orleans County. At Middleport and Lockport, however, pentamerids are not present; equivalent strata consist of crinoidal pack- and grainstone. Neither the Brewer Dock nor the Wallington Member have previously been recognized in Niagara County. These strata have been collectively assigned to the Hickory Corners Member.

The Sodus Shale consists of alternating green and purple, fossiliferous shales. At the Genesee Gorge, approximately 4.4 m (13 ft) of Sodus Shale are preserved beneath the middle Clinton (Sequence II-IV) unconformity. The western pinchout of the Sodus Shale has not been precisely located, but occurs east of Albion, Orleans County. To the east in Wayne County, the full thickness of the formation as well as younger, lower Clinton strata (Wolcott Limestone, Wolcott Furnace Hematite) are preserved beneath the unconformity. All of these strata appear to condense farther east and grade into the Oneida Conglomerate.

The predominant fauna of the Sodus Shale consists of *Eocoelia* brachiopods and diminutive ostracodes. The litho- and biofacies are reminiscent of the much thicker and, in part, time correlative strata of the Rose Hill Formation of Pennsylvania (see Cotter, 1983; 1988; 1990).

In the Rochester area, a major unconformity below the Williamson Shale progressively bevels, the upper and lower Sodus Shale. This shale-on-shale unconformity is easily missed in local sections without careful inspection.

Sequence III: Middle Clinton Group

The middle Clinton Group, as defined by Gillette (1947), consists of 35-45 m of green Sauquoit Shale in the Clinton type area of east-central New York. These same strata have been designated as Silurian Sequence III by Brett et al. (1990a, b). A thin tongue of greenish shale, possibly assignable to Sauquoit occurs between the Furnace beds and Williamson Shale at Second Creek (Stop 6). No strata correlative to the Sauquoit Shale are preserved west of central Wayne County along the New York outcrop belt. These strata have been truncated beneath the unconformity at the base of the Williamson Shale.

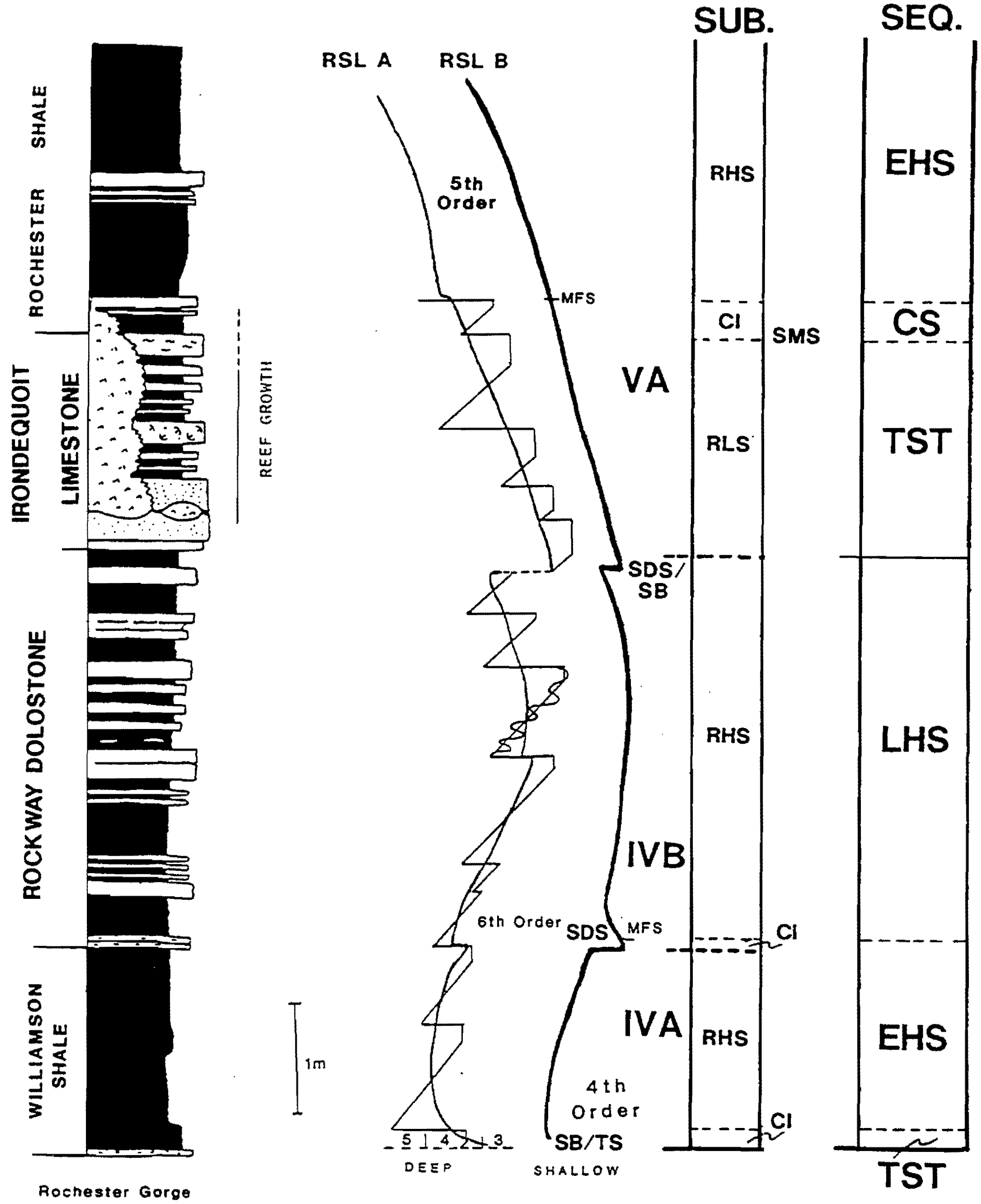

Figure 12. Lithostratigraphy, inferred relative sea-level and sequence interpretation of the late Llandoverian to early Wenlockian upper Clinton Group (Sequence IV and lower V) at Genesee Gorge, Rochester, New York. Sharp but nearly conformable sea-level drop surface (SDS) separates Sequences IV and V at base of the Irondequoit Formation. Calibration of relative sea-level curves (RLS) based on biofacies as follows: 3 = large *Whitfieldella* association. benthic assemblage 3: 4 = diverse brachiopod *(Dicoelosia-Atrypa)* association (BA-4): 5 = deep-water, dysaerobic shales with graptolites and

Sequence IV: Upper Clinton Group (Williamson Shale-Rockway Dolostone)

Sequence IV contains the basal two formations of the upper Clinton Group in western New York. These two formations, the Williamson Shale and Rockway Dolostone, are a mixed carbonate and siliciclastic mudstone succession of latest Llandoverian to early Wenlockian age. The basal boundary of Sequence IV has long been recognized (Gillette, 1947; Kilgour, 1963) as a major, albeit cryptic, unconformity. The shale-on-shale unconformity may be observed at on the west side of the Genesee Gorge at Maplewood Park and at the Tryon Park locality along Browncroft Creek. The surface is commonly overlain by a 1-10 cm thick phosphate and quartz pebble bed designated by Lin and Brett (1988) as the Second Creek Phosphate Bed. The bed is traceable for approximately 240 km between Syracuse and Niagara Falls, New York. East of Syracuse, the Second Creek Bed grades laterally into the Westmoreland Hematite which has been interpreted by Brett et al. (1990a, b) to be a condensed transgressive systems tract. Thus, the Second Creek Bed records a marine flooding surface at the top of the transgressive systems tract of the same sequence. The Williamson Shale is interpreted to be the early highstand systems tract of Sequence IV (Figure 12). West of Niagara Falls, the bed appears to be represented by a glauconitic zone at the top of the *Pentameroides*-bearing Merritton Dolostone. The Second Creek Bed yields conodonts indicative of the *P. amorphognathoides* Zone that establish it as late Telychian C_5 to C_6 age. The biostratigraphy and stratigraphic relationships also suggest correlation of the Merritton Dolostone and the Westmoreland Hematite. Both units represent the condensed transgressive systems tract of Sequence IV.

In the Rochester area, the Williamson Shale is a 2-6 m thick sequence of black and greenish-gray, graptolite- and brachiopod-bearing clay shale. The Williamson Shale contains the deepest marine biofacies in the Silurian System of western New York. The graptoloid-*Eoplectodonta* biofacies of the Williamson Shale represents a transitional BA-5 to BA-6 benthic assemblage (sensu Boucot, 1975). In polished drill core, small *Chondrites* burrows are also visible.

Polished drill cores also reveal decimeter-scale rhythmic alternations of shale beds exhibiting different shades of green and gray that are reflective of varying oxidation states of iron. These color alterations must be primary depositional features, because the *Chondrites* burrows in one bed are infilled by clay that is the same color as the overlying bed. These small-scale rhythms are reminiscent of limestone-marl couplets that have been interpreted elsewhere to record 10 ka. precessional cycles in the Milankovitch band (Hallan, 1986; Ricken, 1986). A similar bedding motif is also apparent in the overlying Rockway Dolostone (Figure 13).

The Rockway Dolostone is marked at its base by a thin (10-30 cm), quartz and phosphate pebble bed, designated the Salmon Creek Phosphate Bed by Lin and Brett (1988). The Rockway Dolostone consists of rhythmically alternating green-gray shales and fine-grained carbonates. Both lithofacies are generally fossiliferous and contain brachiopod assemblages dominated by *Costistriklandia* west of Niagara Falls and *Eoplectodonta* and *Clorinda* at sections in western Monroe County and Wayne County. These brachiopods suggest that the *Chondrites* -rich shales and carbonate mudstones were deposited in below-wave-base environments that were slightly shallower than the Williamson Shale paleoenvironment. Thus, the Rockway is interpreted to represent the late highstand systems tract of Sequence IV.

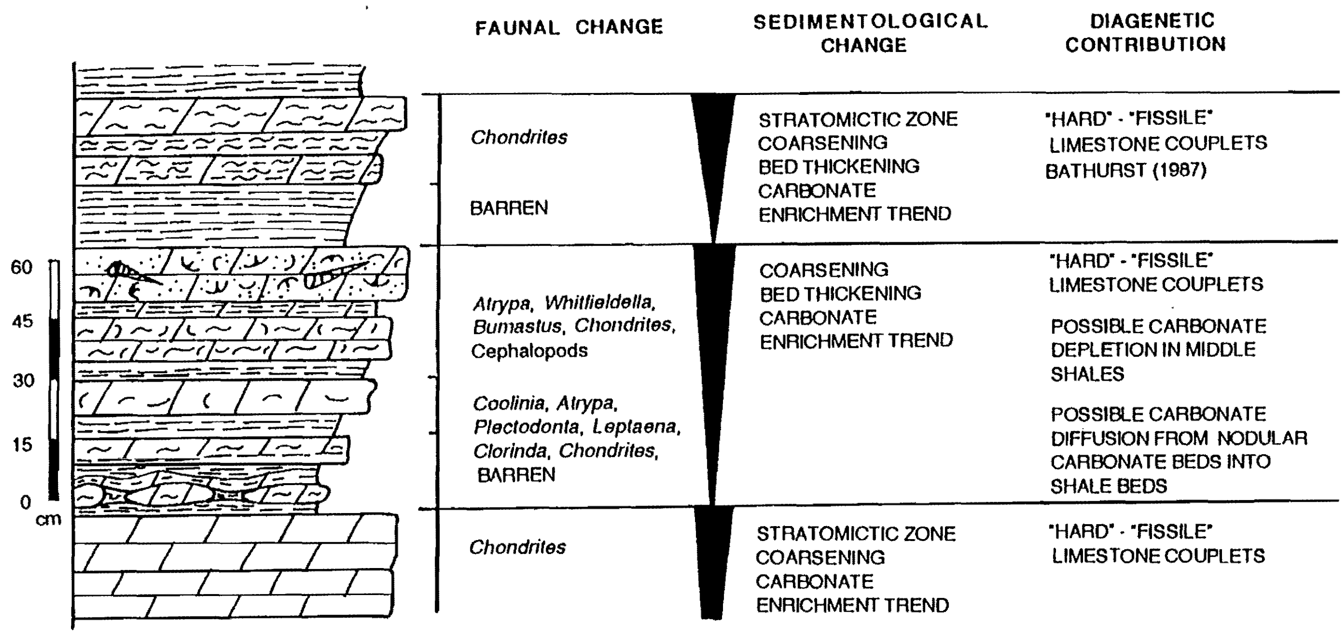

Figure 13. Example of Rockway sixth-order cycle in the Genesee River Gorge, Rochester, NY. Description of faunal, textural, and diagenetic components of same sixth-order cycle are provided to the right of the stratigraphic column.

The decimeter-scale rhythmic couplets seen in the Rochester area sections of the Rockway are stacked into meter-scale asymmetric to subsymmetric cycles reminiscent of Punctuated Aggradational Cycles (PACs) of Goodwin and Anderson (1985). These meter-scale or parasequence cycles are apparently traceable as discrete allocycles between Hamilton, Ontario and Wayne County, New York. Tentative correlation of these cycles with similar bedding patterns in the Dawes Sandstone of central New York and the upper Rose Hill Shale of Pennsylvania suggests that these are circumbasinal allocycles.

Sequence V: Upper Clinton Group (Irondequoit Limestone-Rochester Shale-DeCew Dolostone-Glenmark Shale)

The Rockway Dolostone shares a sharp, distinctive contact with the overlying Irondequoit Limestone. This contact may be traced basin wide as a distinctly sharp surface between equivalent formations and was recognized by Dennison and Head (1975) as representing a basin-wide sea-level drop surface. In east-central New York, the sharp surface occurs between the Dawes Sandstone and the overlying Kirkland Hematite. In Pennsylvania, the surface generally has been used to mark the contact between the Rose Hill Shale and Keefer Sandstone. Given the ubiquity of the distinctive surface across much of the basin, the Rockway-Irondequoit contact has been interpreted to be the Sequence IV-V bounding unconformity.

The Irondequoit Limestone consists of buff-weathering, crinoidal and brachiopod packstones and wackestones with green-gray shale interbeds. Small, bryozoan-rich bioherms are intercalated in the bedding sequence. The bedding sequence generally reflects a stepped, retrogradational facies trend. The shallowest facies appear to be in the basal beds of the formation and consist of crinoidal packstone. Beds of relatively shallow water, packstone facies that are intercalated with calcareous shales and bioherms appear to cap meter-scale allocycles. These cycles are traceable westward to Niagara County where they coalesce into a massive formation of well-sorted, crinoid and brachiopod grainstone reflecting wave-base shoal environments. The better articulation of body fossils and the diverse biofacies dominated by the brachiopod *Whitfieldella* in the Rochester area suggest that the sections observed in the Genesee Gorge reflect a more basinal setting than the classic section at Niagara Gorge.

The retrogradational trend initiated in the Irondequoit Limestone continues into the lower member of the overlying Rochester Shale. The basal bed of the Lewiston Member contains a diminutive brachiopod fauna (Tetreault, 1994) and consists of an intensely bioturbated calcareous shale. Drill cores reveal a profusion of highly compacted, millimeter-scale *Chondrites* burrows. The bed appears to represent a period of slow net accumulation of siliciclastic muds and may be interpreted as a condensed section at the base of the Sequence V highstand systems tract. Given this interpretation, the basin-wide, nearly isochronous facies change from coarse-textured carbonates and siliciclastics to shale records a major Wenlockian marine transgression. A major Silurian sea-level highstand in the early Wenlockian (Sheinwoodian) is also apparent in other marine basins. (Johnson, 1987; Johnson et al. 1985, 1991).

The strong evidence of eustatic control on the major facies change, however, cannot explain regional stratigraphic relationships in lowermost Rochester Shale strata that suggest that differential subsidence of the basin continued to

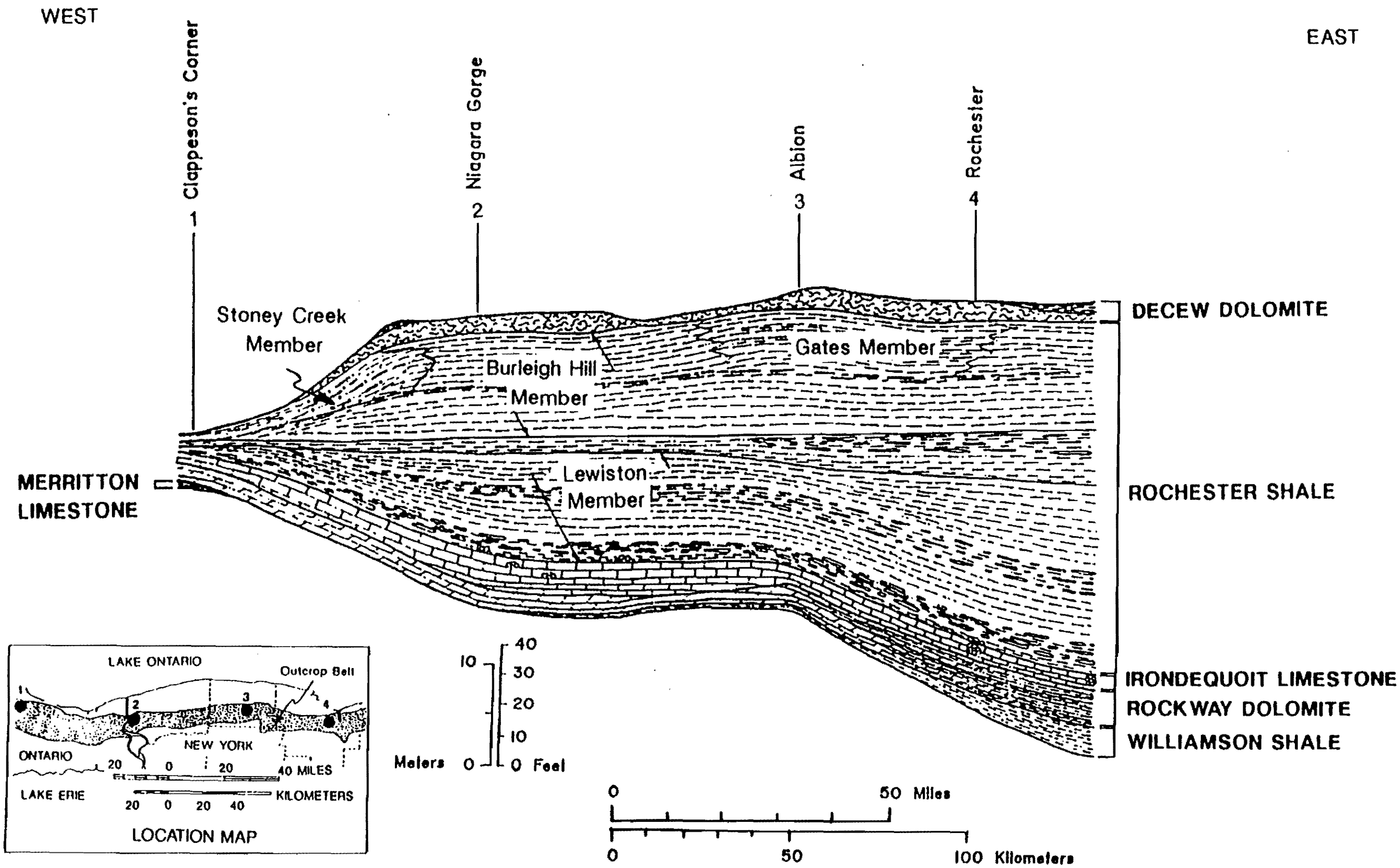

Figure 14. Diagrammatic stratigraphic relations in the upper part of the Clinton Group, with regional correlations between Clappison's Corners, Ont., and Rochester, N.Y. Reference datum is the contact between the Lewiston and Burleigh Hill members of the Rochester Shale.

operate. First, complex pinch and swell architecture of basal beds in the Rochester Shale of Pennsylvania suggest the formation of intraformational saddles and sags. Second, the retrogradational pattern of the Irondequoit continues beyond the condensed section for approximately 10 meters higher into the lower Lewiston Member before maximum relative sea-level is attained. This pattern may be explained by an episode of basin subsidence that outpaced deposition of siliciclastic muds and bioclastic carbonates that comprise the Lewiston Member.

The top of the Lewiston Member of the Rochester Shale is capped by a bryozoan biostromal layer that extends far into the basin off the western ramp. A mirror image fossiliferous and hematitic sandstone tongue extends into the basin from the eastern basin margin. This horizon, designated the Lewiston E submember by Brett (1983a, b), has been interpreted by Brett and others (1990a, b) to mark the boundary between the early and late stages of the highstand systems tract.

The beds overlying the Lewiston E submember at Rochester were assigned to the Gates Member by Brett (1983a, b). These strata consist of silty clay shale and are only sparsely fossiliferous. The paleontology of these beds is less noteworthy than that of the Lewiston Member except for some spectacular crinoid (*Dimerocrinites*) pavements observed at the Penfield Quarry (Stop 3).

The Rochester Shale attains thicknesses of approximately 35 m in the Rochester area (Figure 14). Consequently, the formation does not crop out completely at any one locality. The most complete section, albeit an inaccessible one, occurs at the Upper Falls on the Genesee River. At this section, river level below the falls occurs in the lower beds of the Lewiston Member. At the crest of the falls, the contact between the Gates Member and the DeCew Dolostone may be seen on the east bank.

Numerous exposures of the Rochester Shale-DeCew Dolostone contact are exposed along the area expressways and in the Erie Barge Canal. Although apparently conformable, the contact is always distinctive because of the often spectacular soft-sediment deformation in the basal beds of the DeCew Dolostone. The horizon of soft-sediment deformation is commonly overlain by hummocky cross-stratified calcisiltite and pelletal calcarenite beds. This association of bedding features is suggestive of submarine slumping of storm-generated beds due to widespread liquifaction and remobilization of sediment it may represent a seismite.

The DeCew Dolostone is only sparsely fossiliferous. The fauna includes mainly disarticulated crinoid ossicles and lingulid brachiopods. At Rochester, arenaceous beds in the DeCew contain bulbous *Skolithos*-like vertical burrows. Locally, both the Gates Member and the DeCew Dolostone apparently accumulated on the platform of an intrabasinal paleotopographic high referred to by Don Crowley (unpublished manuscript) as the Penfield Shoal. Brett and others (1990a, b) interpreted these strata as part of the late highstand systems tract of Sequence V. In Monroe County, the DeCew Dolostone is unconformably overlain by the Gasport Limestone-equivalent part of the Penfield Formation of Sequence VI.

Thin remnants of Sequence V strata that overlie the DeCew Dolostone can be observed in Wayne County. These younger strata consist of fossiliferous shales and thin brachiopod-rich limestones. The diagnostic brachiopod of this interval is *Nucleospira pisiformis*. This interval of *Nucleospira*-rich black limestones extends eastward into the type Clinton area of central New York and into Pennsylvania and

Maryland where it has previously been mapped as the basal unit of the McKenzie Shale (Brett et al. 1990).

Sequence VI: Lower Lockport Group (Gasport Limestone, Goat Island Dolostone equivalents of the Penfield Sandstone)

Identification of the Glenmark Shale in Wayne County confirms that the sharp, erosional contact between the DeCew and Gasport Formations in the Niagara region is a low angle unconformity. Thus, the unconformity defines the base of Silurian Sequence VI in western New York and Ontario outcrop belts. This sequence includes the formations of the lower Lockport Group. In the Rochester area, the Lockport carbonates are particularly arenaceous. For this reason, Zenger (1965) defined the Penfield Sandstone Member of the Lockport Formation to include all strata between the underlying DeCew Formation and the overlying Oak Orchard Formation. Recent work by Brett and others (1994) has demonstrated the need for substantial nomenclatural revision for three strata because of the following reasons: 1) the Lockport strata are presently given group strata because of internal, mappable units which are larger in scale than beds and, thus, may be defined as members and formations; these member- and formation-scale units are recognizable within the Penfield Sandstone. For example, the Gasport Limestone and Goat Island Dolostone horizons may be readily identified at the Penfield Quarry; 3) the name "Oak Orchard" must be abandoned and substituted by the names "Eramosa" and "Guelph" because of miscorrelation of strata across the United States-Canada border and the fact that none of the strata previously designated as "Oak Orchard" are actually exposed along Oak Orchard Creek. The nomenclature for Lockport strata in the Rochester area is in the process of being revised (Brett and others, 1994). Brett and others (1990a, b) have promoted use of the Niagara region formation names to identify allostratigraphic units that are relatively time-parallel and cross-cut local facies belts. The benefit of defining allostratigraphic as opposed to purely lithostratigraphic units lies mainly in improved correlations and more precise paleogeographic reconstructions (Figure 15).

Gasport-equivalent strata in the Rochester area consist of brown-gray, medium- to thick-bedded, planar to bimodally trough cross-stratified, dolomitic sandstone and arenaceous dolostone. The thickness of the Gasport-equivalent, part of the Penfield averages approximately 5 m. At localities in western Monroe County, lenses of crinoidal grainstone, i.e. classic Gasport facies, are still common in the lower half of the interval. The upper half of the interval is often darker colored and slightly more argillaceous. The upper half is commonly intensely bioturbated. Small-diameter *Chondrites* burrows weather out in relief a render a "vermicular" texture to the upper half of the Gasport interval.

The Gasport equivalent units are overlain by more argillaceous, slightly deeper water Goat Island equivalents. The Goat Island interval contains a relatively massive bed of arenaceous dolostone at its base. LoDoca (1991) has mapped remnants of the *Medusaegraptus* epibole within this massive basal Goat Island equivalent at localities in western Monroe County. The massive bed is overlain by more argillaceous, thin-bedded, fossiliferous wackestones that probably represent the deepest water facies of the Lockport Group in the Rochester area. These facies are particularly well exposed at the Route 531 exit off of Route I-490 on the west side of the city. On this road cut and many others around the city, massive, vuggy, biostromal beds previously assigned to the Oak Orchard Member and now assigned to the Eramosa Formation abruptly overlie these relatively open marine, thin-bedded carbonates.

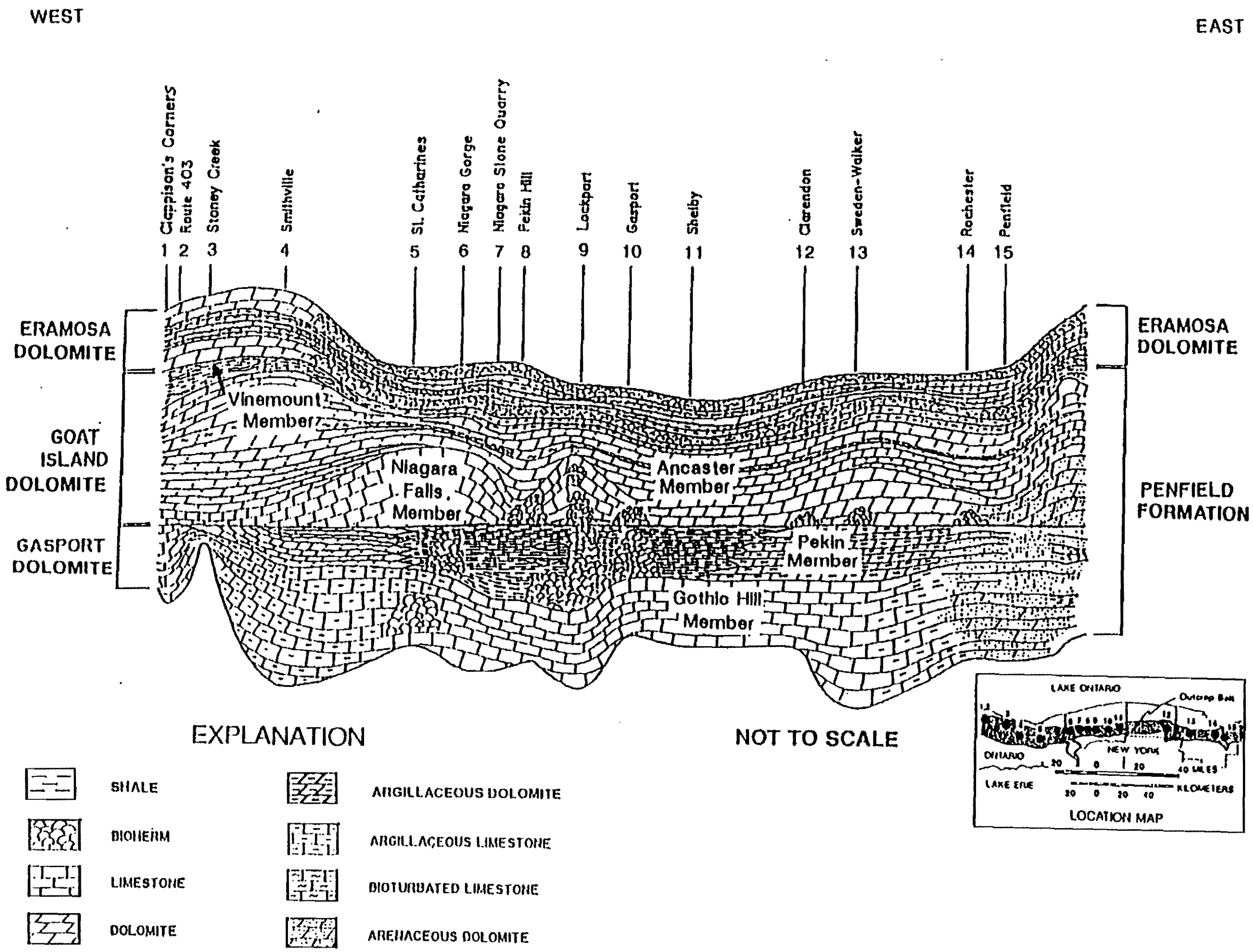

Figure 15 -- Diagrammatic stratigraphic relations in the Lockport Group, with regional correlations between Clappison's Corners, Ont., and Penfield, N. Y. Reference datum is the contact between the Gasport Dolomite and the Goat Island Dolomite.

Sequence VII: Upper Lockport Group (Eramosa and Guelph Formations) and Salina Group

The abrupt facies change from deep water, fine-grained carbonates of the upper Goat Island Formation to the shallow water, biostromal facies of the lower Eramosa Formation can be traced from Guelph, Ontario to Wayne County, New York. Although the interpretations are preliminary, this contact may define the base of a seventh Silurian third order Sequence. The Eramosa is the only unit within this tentatively defined sequence that crops out across western New York. The Eramosa generally contains a tripartite stratigraphy. The upper and lower portions of the formation consist of massive biostromal beds that are separated by a middle unit of medium-bedded, tabular, slightly more argillaceous dolostones. In the Niagara region, the Eramosa Formation is capped by distinctive stromatolite reefs of the Guelph Formation. These stromatolite reefs have not been documented in the Rochester area. In fact, Rickard (1969) suggests that the 20-25 m thick Guelph interval of the Niagara region grades eastward beyond Lockport, Niagara County into Vernon facies. Consequently, the sharp gamma ray inflection used to mark the Lockport-Vernon contact in the subsurface of the Finger Lakes region may correlate with the Eramosa-Guelph contact of the Niagara region.

DISCUSSION AND SUMMARY

The Late Ordovician to medial Silurian strata of the Genesee region represent marginal marine to deeper basinal deposits of mixed siliciclastics, derived from the erosion of the Taconic orogenic terrains, and intrabasinal carbonates. These strata record cyclic variations at several scales. The largest are equivalent to third order sequence (1 to 3 million years; Vail et al., 1991) that are bounded on one or both margins of the forehand basin by prominent erosion surfaces (see Figure 7). These large scale sequences are also subdivisible into two or more fourth order subsequences, (Bush and Rollins, 1984; Brett et al., 1990), which display a similar internal pattern to the large scale sequences of which they are a part, but also possess less significant erosion surfaces at their boundaries. We recognize a total of seven large scale or third order sequences, roughly corresponding to parts of originally defined lithostratigraphic groups in the Silurian of the Rochester area, and at least 15 subsequences. The erosional sequence boundaries correspond to relative low stands of sea level. Transgressive surfaces are generally superimposed upon the lowstand erosion surfaces with no intervening non-marine (lowstand) sediment. Transgressive or rising sea level phases are recorded in thin, but typically widespread intervals of reworked, clean sandstones (e.g., Devils Hole Sandstone) or carbonates (e.g., Irondequoit). Highstand deposition is represented by thicker shales such as the Williamson and Rochester, or by argillaceous limestones (Goat Island equivalent portions of the Lockport). Maximum flooding of shoreline areas and alluviation in estuarine environments are recorded in the offshore facies as highly condensed deposits and/or surfaces of sediment starvation near the tops of transgressive sandstone or limestone intervals. Such surfaces may be sharp discontinuities, even with evidence for erosion and are typically marked by phosphatic debris, or, in the medial, Silurian by hematitic or glauconitic beds.

The Silurian sequences and their bounding discontinuities recognized in western New York appear to correspond to those seen in other parts of the world, at least within the rather loosely constrained limits of biostratigraphy. (Johnson et al., 1985; Brett et al., 1991). Because of this, we argued that at least the larger cycles (third and fourth order) or 0.5 to 3-million year duration are global and reflect eustatic sea level changes.

Silurian sequences can also be subdivided into a number of smaller-scale cycles which range from nearly symmetrical limestone shale cycles to markedly asymmetrical, shallowing upward cycles. These fifth and sixth order cycles (100 to 400 thousand year cycles) also appear to correlate widely within the Appalachian basin. The shallowing upward shale to limestone cycles in western New York, for example, appear to correlate with coarsening upward shale to siltstone or sandstone cycles in the central New York type Clinton area. At present, we are unable to determine whether or not these cycles can also be correlated outside the Appalachian basin.

As a generality, the succession from Queenston Shale to Vernon redbeds represents one broad scale trend towards deepening culminating in the Williamson and Rochester Shales in the mid Silurian (late Llandoverian to Wenlockian) interval, followed by a general trend towards shallowing. This large scale, second order cycle is also observable within the Silurian in other parts of the world, and it may also reflect a global scale, perhaps tectono-eustatic driven cycle of rise and fall in sea level.

Three very broad scale trends are superimposed upon the cyclic patterns of the Silurian stratigraphic record. The first is a rather obvious trend toward a decreasing influence of siliciclastic sedimentation upward in the Silurian, at least to the middle portion of the Clinton group. Thus, Sequence 1, the Medina Group is almost entirely siliclastic sands and muds in western New York, whereas as Sequence 2, the lower portion of the Clinton Group, comprises mixed siliciclastics and limestones. A reversal of this trend occurs within the middle portion of the Clinton Group with deposition in central New York of relatively thick (30 to 40 m siliciclastic wedge of the Saquoit-Otsquago Formations Sequence 3). Siliciclastic-dominated sedimentation continued in the eastern and central New York through the accumulation of Sequences 4 and 5, (Willowvale-Williamson shale, Herkimer Sandstone and its lateral equivalent, the Rochester Shale). This pulse of increasing siliciclastic input, including relatively coarse sands in New York and Pennsylvania (Keefer Sandstone, Colter, 1983, 1988), is timed with a reorganization of the basin and a change in its direction of migration. The trend towards increasing siliciclastics is reversed through much of the later Wenlockian and Lludovian with deposition of the Lockport Group, at least in western New York. However, the appearance of dolomitic sandstone within the Penfield facies of the Lockport Group at Rochester, and their eastward gradation into finer-grained Ilion Shale may indicate that siliciclastics were largely trapped in the eastward portions of the basin due to partitioning associated with the development of a "Penfield high" or shoal area (Zenger, 1965; Crowley, 1973). In any case, the interval of carbonate-dominated sedimentation in the Lockport in the west was terminated in the later part of the Ludovian by the westward spread of the Vernon siliciclastics. These latter appear to have over-filled the shallow eastward portion of the basin, such that the upper parts of the Vernon and its equivalent in Pennsylvania, the Bloomsburg Formation, consist of nonmarine red beds.

A second, related trend, less obvious from the local frame of reference, is a large scale cycle in the migration direction of the Appalachian of the foreland basin axis. During the late Ordovician, (Ashgillian) to the mid Silurian (late Llandoverian) time, some 15 million years, the depocenter or area of thickest sediments, as well as the area of apparent deepest water facies appears to have shifted by over 300

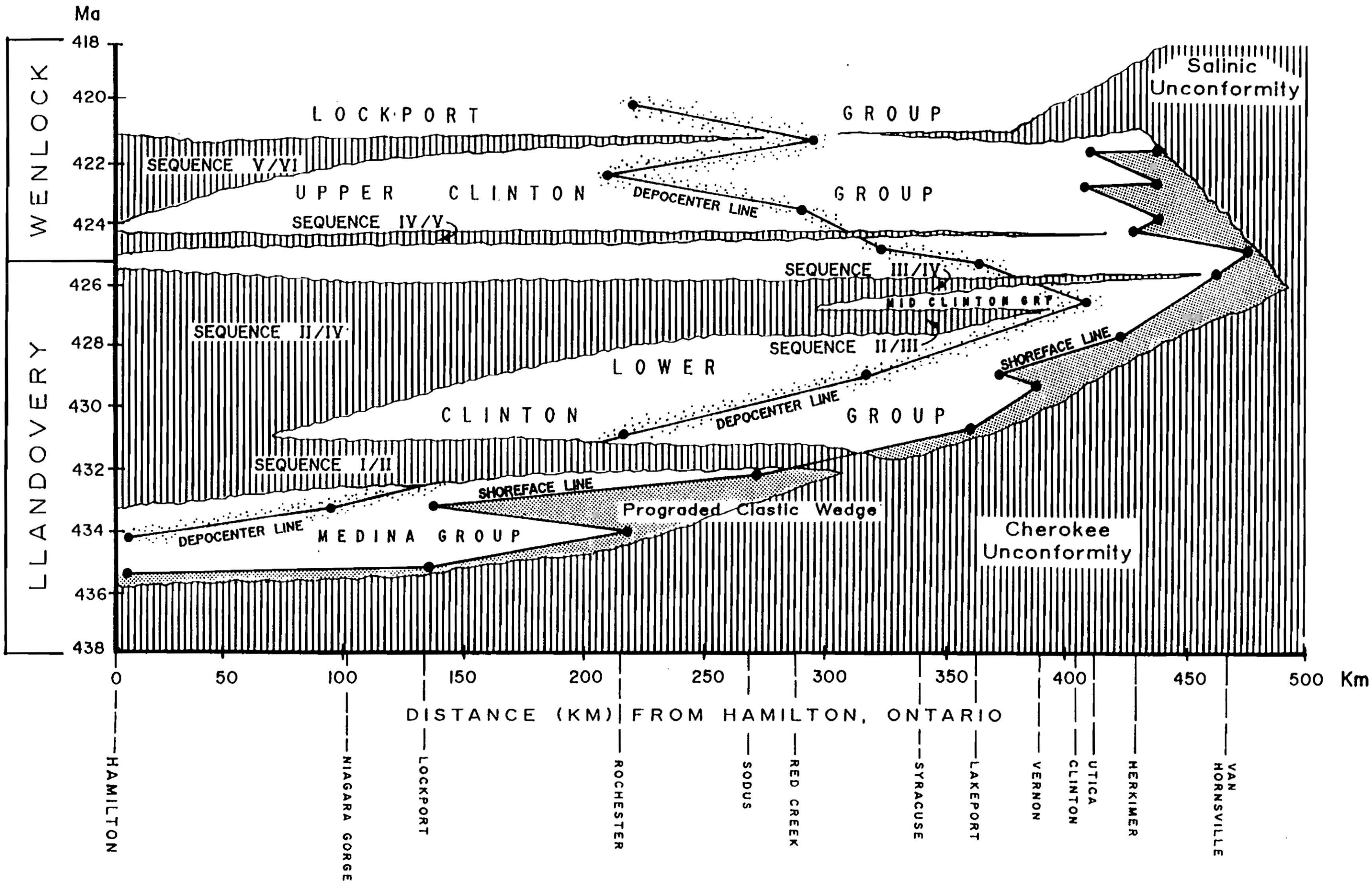

Figure 16. Migration of shoreface and depocenter for the Appalachian Basin during early to medial Silurian time in section along New York State outcrop belt. Major sequence-bounding unconformities are depicted by vertical ruling.

kilometers from western Ontario to central New York State (Figure 16). This was followed by an abrupt change during the latest Llandoverian, at which time the basin axis appears to have migrated back westward some 200 plus km during deposition of Sequences 4 and 5. A return to eastward shifting in the basin center is evident, starting within the upper portion of the Lockport Group, and certainly continuing into the time of deposition of the overlying Salina Group (late Llandovian to Pridolian times).

Finally, a third phenomenon relates to the development of unconformities that bound the Silurian sequences. Some of the sequence bounding unconformities become more extensive in an eastward direction (i.e., proximally or toward the source of the sediments), while others in the mid part of the Silurian open in a westward or Cratonward direction (Figure 17). The basal Cherokee unconformity opens eastward such that the Queenston shale and then the underlying Oswego, Lorraine, and parts of the underlying Utica siliciclastics are progressively beveled in central New York. Likewise, the Salina unconformity which is a minor discontinuity separating the Vernon Shale from the overlying Syracuse Formation in western New York bevels the Vernon and ultimately oversteps the underlying Herkimer, Willowvale and Otsquago siliciclastics in east central New York.

Conversely, unconformities bounding sequences 2, 4, 5 and 6 all appear to become more prominent toward the west, approaching the so-called Algonquin Arch; these discontinuities appear to decrease to conformities in central New York State. For example, the very major unconformity that underlies the late Llandoverian Williamson Shale throughout west central New York (Lin and Brett, 1988) is a regionally angular beveled surface. This unconformity appears to be a minor discontinuity between the West Moreland hematite and underlying Sauquoit-Otsquago shales in central New York. However, to the west the unconformity becomes more prominent and progressively bevels almost the entire Clinton Group between the area of Clinton, New York and Grimsby, Ontario.

This pattern suggests erosion of a differentially uplifted arch, perhaps centered in the region of Hamilton, Ontario during the medial Silurian. We interpret this feature as a transient forebulge, as predicted in most models of basin dynamics (e.g., Quinlan and Beaumont, 1984; Beaumont et al., 1988, Jordan and Flemings, 1991). Unconformities of the basal parts of the Medina Group, upper Clinton and Lockport Groups also display lesser, but still significant erosional truncation toward the west.

Thus, three phenomena appear to be interrelated: (1) the pattern of decreasing and increasing siliciclastic input, (2) the eastward to westward oscillation of basin axis migration over several tens of millions of years, and (3) the alternation of eastward opening and westward opening sequence bounding unconformities. Taken together, these observations suggest the action of a long term tectonic cycle.

Following the viscoelastic thrust loading model of Quinlan and Beaumont (1984) and Beaumont et al. (1988), we have previously argued (Brett et al., 1981; Goodman et al., in press), that phases of tectonic thrusting alternating with intervals of quiescence are recorded in this long term pattern. Specifically, we have observed that intervals of westward basin axis shift, e.g. during the late Ordovician, and again during the mid part of the Silurian, are coincident with increasing coarse siliciclastic sedimentation and the development of minor

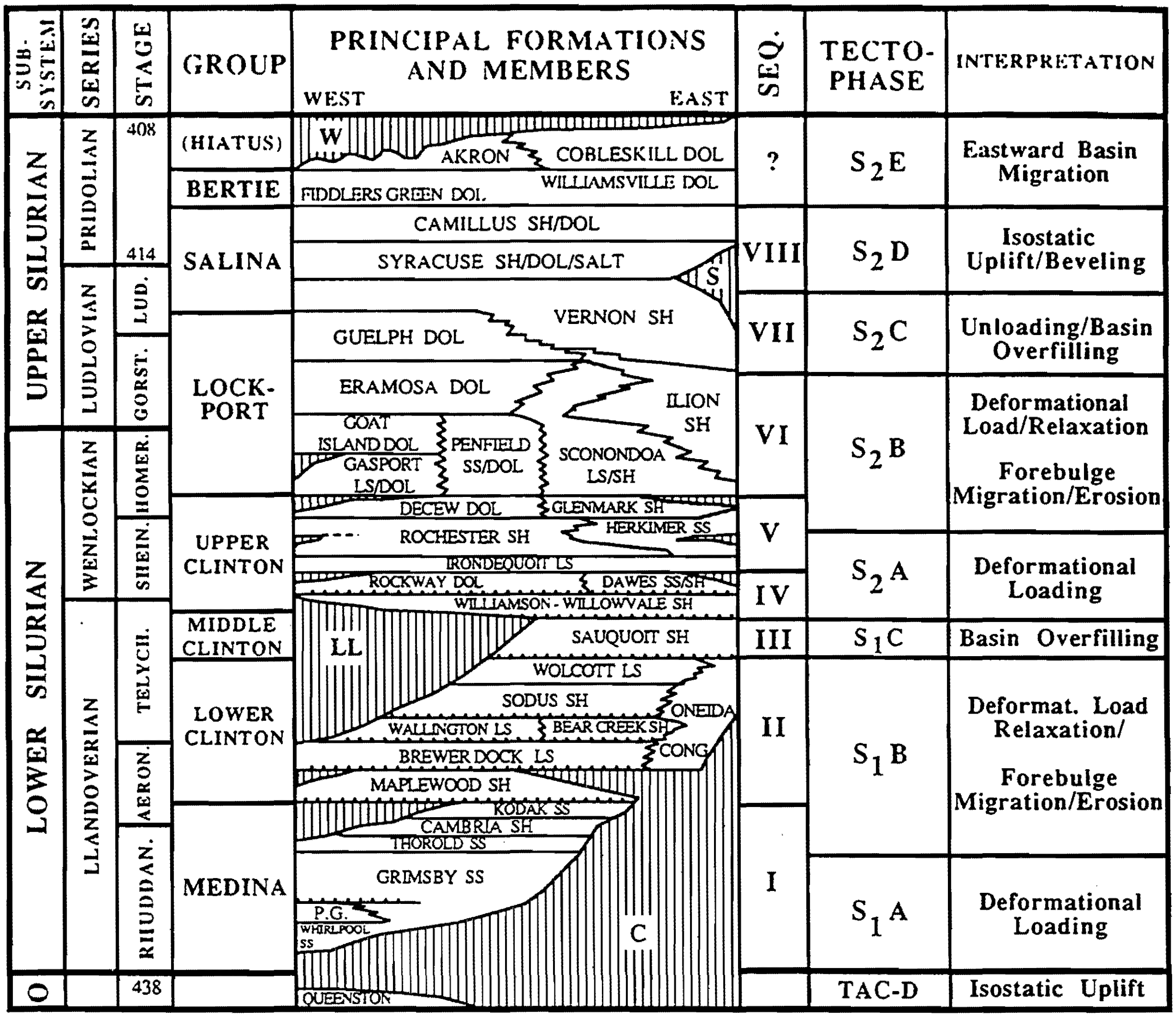

Figure 17. Tectophases of Sequences I-VIII across New York State. Rightside presents interpretation of phases of tectonic cycles. Tectophases designated as TAC-D = Taconic late phase; S_1A-S_1C and S_2A-S_2E are inferred phases of tectonic cycles during the Silurian.

westward opening unconformities. All these phenomenon may be related to one another and appear to have occurred during times of active thrust loading in the Taconic Orogen. Intervals of eastward basin shift, decreasing coarse siliciclastics, and westward opening major unconformities appear to be associated with interludes of tectonic quiescence and thrust load relaxation. A migration of the foreland basin toward the Hinterland and accentuation of the peripheral forebulge are predicted in the viscoelastic model of Beaumont et al., (1988). Accentuation of the eastern basin and the western forebulge is evident at times such as the early Silurian interval from at least the upper portions of the Medina Group deposition into the middle Clinton group.

A phase of basin overfilling by fine grained siliciclastics and some carbonates occurred during the later quiescent interludes. The Ordovician Queenston-Juniata and the late Silurian Bloomsburg-Vernon red mudstones would exemplify this phase. Finally, very late phases of uplift in the proximal clastic wedges were followed by an eastward erosive beveling of the proximal molasse. For example, major eastward opening unconformities cut the top of the Queenston and the Vernon red beds. In accordance with the Beaumont et al. (1988) model it is postulated that eastward uplift of the clastic wedge represents isostatic adjustment due to the redistribution of some of the thrustload. This took place by erosion and transport of sediments into more distal or cratonward parts of the basin. As such, the major eastward opening unconformities would record prolonged periods of relative quiescence between tectonic thrusting episodes.

We recognize that other models of foreland basin dynamics would make different, and in some cases, opposite predictions. For example, the elastic model of lithospheric behavior advanced by Jordan and Flemming (1991), would predict that episodes of eastward migration of the basin axis and of accentuated forebulge erosion would actually be times of active thrusting. Westward migration of depocenters would represent infilling of the foreland basin in the interludes following the thrusting episodes. In common with the Beaumont et al. (1988) model, Jordan and Fleming (1991) also postulate that uplift and erosion of the proximal side of the foreland basin will be greatest during times of tectonic quiescence. The main difference between these two models resides in the rate of response of the lithosphere to tectonic loading. In Jordan and Fleming's model the lithosphere is depressed relatively rapidly during tectonic thrust loading. This produces an asymmetrical deepened basin proximal to the thrust load. Conversely, in the viscoelastic model of Beaumont et al. there is a lag time in subsidence in the lithosphere such that the basin tends to contract toward the thrust load after the thrusting is over and during a phase of crustal relaxation. Although our observations are preliminary, they seem to support the views of the Beaumont et al. Clearly, the foreland basin appears to have been driven cratonward during known intervals of thrust loading, such as in the Taconic Orogency (Lehmann et al. in press). These are also intervals associated with increased input of coarse siliciclastics into the foreland basin. In contrast, times of accentuated western forebulge development and of eastward shifting of the basin axis appeared to coincide with intervals with little or no known tectonic activity as in the early part of the Silurian. Tectonic quiescence is also supported by the fact that these times of eastward shift are coincident with a decreasing input of coarse siliciclastic sediments. This might be expected during times of decreased erosion of an inactive tectonic source terrane. We have also made similar observations in the latter part of the Silurian, the early and Middle Devonian (Brett and Baird, this volume).

6.3 0.1 Intersection St. Paul St.; turn right (south)

6.7 0.4 Junction Driving Park Boulevard (or Avenue E), turn right(west) and cross Genesee River on new (1986) Driving Park Bridge

6.95 0.25 At west end of bridge turn left into driveway of YMCA and park. Proceed on foot back onto bridge for overview of Lower Falls.

STOP 1B - LOWER FALLS, GENESEE RIVER

This overlook provides a panorama of the Lower Falls of the Rochester Gorge. The Lower Falls, with a drop of 30 m (97 ft), is capped by resistant sandstones of the Upper Grimsby and Kodak Formations.

The stratigraphic units examined previously (Stop 1A; Queenston Shale to Reynales Limestone) are clearly visible near the base of the gorge beneath and immediately south of the Driving Park Bridge. Higher units of the Clinton Group can also be viewed from this vantage point although their limited accessibility precludes close examination during this trip.

The Reynales carbonates are observable immediately opposite this point near the base of the large water surge tank of the Rochester Gas and Electric Company. They are overlain by an interval of shales which actually constitute two distinct units. The lower 4.4 m (13 ft) of greenish to purple Lower Sodus Shale is unconformably overlain by 2 m (6 ft) of black to greenish-grey, graptolite-bearing Williamson Shale. The contact of these formations is cryptic but actually represents a significant, regional, angular unconformity that bevels successively older units in a westward direction (Lin and Brett, 1988). This erosional surface, which may reflect the eastward migration of a forebulge, is onlapped by a widespread dark shale of the latest Llandoverian Williamson Formation. The Williamson Shale is followed by thin-bedded dolomitic carbonates and grey shales of the Rockway Formation. The overlying Irondequoit Formation is thicker-bedded and contains small bryozoan-algal bioherms up to about 2 m in height. These can be seen to deform strata in the cliff near the deck level of the Driving Park Bridge. They were initiated on a single horizon of crinoidal packstone near the base of the Irondequoit Limestone. In turn, the Irondequoit Limestone is overlain here by about 3 m of brownish-grey, fossiliferous Rochester Shale, which is bevelled and overlain by glacial deposits.

Return to vehicles and retrace route across bridge , turning right on Avenue E.

7.2 0.25 Intersection of St. Paul, turn right and proceed south to Bausch St.

8.3 0.1 Genesee Brewery; look for access road on other south side of plant opposite Ward Street

8.7 0.4 Access Road for Upper Falls Terr Park, turn right and park in lot, walk to Pont de Rennes Bridge

STOP 1C- HIGH (UPPER) FALLS PARK, GENESEE RIVER

This park provides a view of the 25m (80 ft) High Falls, formerly a major source of water power for industry. Rochester developed as a grain milling center in the mid 1800's largely as a result of its location on the Erie Canal and the presence of this source of power (seeGrasso and Leibe article in this guidebook). Nearly vertical cliffs in the Genesee River Gorge below the High Falls display almost complete section of the Rochester Shale. This may be the oldest type locality in North America, the Rochester Shale having been named by James Hall in 1839. The Rochester Shale, here about 30 m (98 ft) thick, contains grey, calcareous shale with numerous thin calcisiltites (Brett, 1983a). The Rochester Shale is moderately to sparsely fossiliferous with an abundance of brachiopods and the trilobite *Dalmanites*. However, it is largely inaccessible at the type section. The Rochester appears to record a slight shallowing of relative sea level, with an increase in the number and thickness of dolomitic carbonate interbeds toward the top as it passes up into the predominantly carbonate DeCew and Lockport Formations. The upper unit of the Rochester Shale, comprising approximately a third of the cliff at this section, is the Gates Member. The Gates, DeCew, and Lockport units will be examined in greater detail at Penfield quarry (Stop 3). Rochester and DeCew strata represent shallow subtidal, muddy marine deposits, as indicated by the abundance of fossils and lack of tidal features.

8.75	0.05	Return to vehicles and turn around; pull out and turn right onto St. Paul and proceed south
8.8	0.05	Railroad overpass
8.9	0.1	Cumberland Ave.; continue south on St. Paul
9.15	0.25	Pleasant Ave.
9.25	0.1	Main St.
9.4	0.15	Broad St; bridge on old aqueduct (see Grasso and Leibe in this guide)
9.6	0.2	Note multiple junctions: Monroe, 490 East, 15 South, 490 East
9.7	0.1	Take left exit to 490
9.9	0.2	Merge onto 490 East
10.4	0.5	Exit 17, Goodman St. View of Pinnacle Hill with radio towers on right
10.9	0.5	Exit 18, Monroe Ave.
11.4	0.5	Exit 19, Culver Rd.
12.4	1.0	Exit 20, Winton Rd. View of Cobbs Hill to south; stay to the right
12.85	0.45	Fork of 1-490 and I-590; bear right onto exit for I-590 North.
13.1	0.25	Former "Can of Worms" area, minor low cuts in upper Penfield and Eramosa formations of Lockport Group formerlly biostromes were exposed near top of Penfield
13.5	0.4	"Can of Worms" ends
13.7	0.2	Exit 6 ; Blossom Road exit
14.2	0.5	Exit 7 ; Browncroft Boulevard (there are low cuts in DeCew dolostone along exit); after passing under Browncroft overpass prepare to merge to the right, entrance ramp coming from Browncroft onto 390 pull right onto entrance ramp
14.35	0.15	Pull off and park immediately before I-590 N sign on shoulder
		Put flashers on vehicles. Walk into the field into the right of the highway and then straight paralleling the highway, down a bank, to ramp road of sewage treatment plant leading down to the level of small creek.

STOP 2 PALMER'S GLEN (TRYON PARK)

This unnamed creek, informally referred to as Palmer's Glen, also serves as an overflow stream for the storm sewer system. The water in the creek is contaminated to some degree and should be dealt with cautiously.

Lowest exposures along the banks of this creek are at the lower part of the Sodus Formation. This interval is slightly above the upper beds viewed at the Genesee gorge along Seth Green Drive. The shales here range from greenish gray to light purplish gray in color. The purple colored shales appear to cap small shallowing cycles within the Sodus. These cycles commence with thin shelly and, in some cases, phosphatic beds and pass upward into greenish gray shales and these latter gray upward to purplish hued shales. Both green and purple shales contain a shallow marine fauna of the classic *Eocoelia* biofacies of (Boucot, 1975) Bedding plane assemblages are dominated by *Eocoelia cf.* E. *hemispherica*. These are mixed with small bivalves (pteriniids), rare *Leptaena*, small tabulate corals, *Tentaculites*, and abundant ostracods of the *Zygobolba decora* zone. At this locality thicker coquina beds of sandy limestone rich in *Eocoelia* also occur near the top of the Sodus.

The Sodus is capped by a major erosion surface, probably the largest unconformity within the Silurian. However, the unconformity is relatively cryptic at this locality, and is overlain by hard black shale of the Williamson Formation. The contact here is marked by a thin yellowish clay, which could be a bentonite. The basal Second Creek bed of the Williamson Shale is a thin phosphatic, skeletal rich sandstone generally less than one centimeter in thickness. Higher beds consist of an alternation of decimeter-scale hard dark gray shale and soft greenish gray shale beds. The black shales contain laminae of fine quartz sand, some of which possess small scale asymmetrical ripples on their tops. The black sandy shales also display well preserved graptolites of the *Monograptus clintonensus-Retiolites venosus* assemblage. These graptolites, together with abundant conodonts obtained from basal lags within the Williamson Shale, indicate a latest Llandovery (or late Telychian) age for these shales. The Williamson here is approximately four meters in thickness. It is overlain at a minor disconformity by the dolomitic limestones and greenish gray shales of the Rockway Formation. A thin bed at the base of the Rockway contains quartz granules and small pebbles as well as black phosphatic nodules, but these are highly scattered and the bed is difficult to observe. In turn, the Rockway is approximately 2.8 m-thick and is overlain by 3.2 m. of dolomitic crinoidal wacke- to packstones of the Irondequoit Formation. These rocks occur high in the bank at this section, and can be best viewed by walking downstream approximately 100 ft to the fork of Browncroft Creek, and walking upstream (south) into the Browncroft fork. Here the Rockway, Irondequoit and basal Rochester Shale can be viewed at a small waterfall section.

Return to vehicles

14.55	0.2	<u>Continue north after stop</u>
14.75	0.2	Overpass Tryon Park Road
14.85	0.1	Town of Irondequoit sign and sign for Exit 8, Empire Blvd.
15.25	0.4	<u>Bear right onto exit 8 for Empire Blvd.</u> (Route 404), to reverse directions. At stop light at junction of Empire, <u>turn left onto Rte. 404 west.</u>
15.65	0.4	Underpass under 590; <u>turn left onto entrance ramp back onto 590 S</u>
15.95	0.3	<u>Merge onto 590 South</u>
16.45	0.5	City of Rochester line

17.10	0.65	Browncroft Boulevard exit; bear right onto Exit 7
17.2	0.1	Stop light at junction Route 286 off exit lane; turn left onto Browncroft Boulevard (Route 286 East)
17.25	0.05	Stop light under bridge at entrance of 590; continue straight on Browncroft Blvd. (Route 286 East)
17.50	0.25	Small stream on right exposes upper portion of Rochester Shale
17.85	0.35	Bridge over Irondequoit Creek, Penfield town line
18.35	0.5	Ascending hill out of Irondequoit Valley; this is the old River Valley of the preglacial Genesee River , now occupied by underfit Irondequoit Creek
19.05	0.7	Stop light at Blossom Road; continue straight on Rte. 286 East
19.25	0.2	Panorama Trail stop light
19.65	0.4	Clark Road stop light ; continue straight on Rte. 286
20.15	0.5	Whalen Road; turn right
20.75	0.6	Bend in Whalen Road at Clark Road junction
20.95	0.2	Quarry entrance on left; enter and check in at office for permission to access quarry.

STOP 3-DOLOMITE PRODUCTS QUARRY, PENFIELD

This large, active dolomite quarry, about 30 m (100 ft) deep, exposes a nearly complete section of the Lockport Group and its contact with the underlying Clinton Group (Grasso and Friedman, 1989; Zenger, 1965). Crushed stone quarried at this location is used primarily for road-paving material. The quarry is famous among amateur mineral collectors as a source for euhedral crystals of saddle dolomite, calcite, sphalerite, galena, pyrite, celestite, selenite, and especially, clear bluish fluorite which occur in 20 to 50 cm. diameter vugs within the dolostone at several horizons. The source and timing of the mineralization is enigmatic but it occurred at temperatures of approximately 150°C. Mineralization occurred selectively within cavities produced during dolomitization, many of which represent former coral and stromatoporoid heads. Dolomitization in the Lockport Group is evidently of replacement type and has resulted in strong recrystallization and obliteration of much of the primary carbonate fabric. Dolomitization may have been produced by seepage refluxion of brines during deposition of the overlying Salina Group or may be of deep burial origin.

The Penfield Quarry was studied in detail by Zenger (1965), who designated this exposure as the type section of his Penfield Member, here comprising 19.2 m (63 ft) of sandy dolostone and dolomitic sandstone. Zenger also identified the upper 5.2 m (17 ft) of the section here as basal "Oak Orchard Member," which is now referred to as the Eramosa Formation. Recent deepening of the quarry and detailed regional correlation permit a reevaluation of this stratigraphy (Brett, et al., in prep.).

The Gates Member of the upper Rochester Shale is the lowest unit exposed in the quarry. It can be observed in a sump pit near the western edge of the quarry. The Gates contains dark grey, sparsely fossiliferous, bioturbated, silty, dolomitic mudstone and argillaceous dolostone. Planar and irregular (hummocky?) cross-stratification common in silty layers; many bedding planes contain the dendritic feeding trace *Chondrites* and oscillation/interference ripple marks aligned roughly NE-SW. Lingulid and rhynchonellid brachiopods and crinoid ossicles are present, a spectacular occurrence of over 1000 complete crinoids (*Dimerocrinites*) was discovered in this portion of the quarry.

The Gates Member is overlain at a sharp contact by about 3m (15') of thick-bedded to massive DeCew Dolostone. The lowest meter is a blocky bed of sandy, pale gray, buff-weathering dolostone with soft sediment deformation and hummocky cross-stratification that is prominently displayed in the main lower floor of the quarry. The hummocks are ellipsoidal, average approximately 1 m in length along their long axes, and are spaced about 0.5-1.0 m apart.

The basal DeCew was probably the lowest unit exposed when Zenger measured the quarry. Subsequent cutting of the sump pit revealed that the quarry floor was very close to the base of the DeCew, (whereas Zenger had apparently considered it the top of that formation). This bed is overlain by about 5 m (15 ft) of dolomitic sandstone and arenaceous dolostone with two prominent partings which display rusty staining due to ground water seepage. The DeCew is sparsely fossiliferous except for scattered stringers of crinoid columnals and occasional orbicerloid brachiopods, but sandy beds contain vertical *Skolithos*-like burrows. The Gates-DeCew interval is interpreted as mixed siliciclastic and allodapic carbonate sands and silts which were rapidly deposited basinward of shallow, winnowed platforms developed north (?) of this area. Gates-DeCew sediments accumulated in shallow subtidal areas below fair-weather wave base but subjected to frequent storm wave disturbance.

Most of the higher quarry wall is composed of brownish grey to medium grey, medium- to thick-bedded, sandy dolostone which displays planar and bidirectional cross-stratification. Layers and lenses of crinoidal grainstone occur in the lower 5 m which terminate at the top of the lower level of the quarry. The meter-thick bed just below the second bench of the quarry (approximately 12.5 m) is highly crinoidal and contains intraclasts as well as rare rugosan and favositid corals. It is overlain by a 9 m (28.5 ft) interval of medium and even-bedded, sandy dolostone, with thin shaley partings, that appears to correlate with the Goat Island and Eramosa formations of the Lockport Group.

The entire 15 m interval, corresponding to the members of the Gasport, Goat Island, and Eramosa formations, is enriched in quartz sand in the Rochester area, and the name Penfield Formation is perhaps useful in emphasizing this facies distinction. However, it should be noted that members, and even certain marker beds, in the Gasport interval can be traced across the facies change. Crowley (1973, unpubl.) emphasized the locally sandy nature of the Penfield and interpreted the unit as representing a shallow water sandy shoal or "Penfield island"; but the persistence of sedimentary cycles and elements of the typical Gasport marine fauna into the Penfield area indicates an environment similar to that of the typical Gasport facies. Crinoidal grainstones of the Gasport and sandy crinoidal dolostones of the lower Penfield represent similar environments, i.e., a shallow wave-winnowed and perhaps tidally-influenced (bimodal cross stratification) shelf with local shoals or bars, close to fairweather wavebase. The increased sand content of this facies in the Penfield area appears to indicate a local source of siliciclastics north of this region. Upper units are thinner-bedded and more argillaceous than the Gasport equivalents and record a transition to somewhat lower energy, probably deeper water environments similar to those in which the Gates-DeCew interval sediments accumulated.

The highest beds exposed in the Penfield Quarry consist of massive, highly fossiliferous and vuggy dolostones that were assigned to the lower 5.5 m (17 ft) portion of the "Oak Orchard Member" by Zenger (1965); recent study demonstrates that the term Oak Orchard is invalid, and we assign these beds to the Eramosa Formation Prominent, 0.5 m thick, light grey-weathering, dolostone beds occur about 2 and 3 m below the top of the quarry. Dark biostromal beds on eiter side of these horizons are rich in poorly preserved, and typically mineralized tabulate corals (*Favosites*, *Cladopora*), and small domal stromatoporoids. These beds contain numerous large vugs which are lined with nodular anhydrite, scalenohedral calcite, pink saddle dolomite, celestite, sphalerite, rare fluorite and sulfides; these vugs are the principal source of the Penfield minerals. Although the vuggy beds are inaccessible in the vertical quarry walls, they can be examined readily in large fallen blocks piled on the higher bench in the quarry.

The upper Lockport strata at Penfield (> 5m) are distinctly less sandy than the lower beds and more highly fossiliferous. The Eramose interval records a general decrease in the input of siliciclastic sediments and the development of coral-stromatoporoid biostomes and associated carbonate sediments in shallow, but relatively quiet water environments.

21.05	0.1	Exit from quarry
21.35	0.3	Five Mile Line Road (jog in Whalen Road); continue on Whalen Road at blinking light
22.85	1.5	Cemetary on right and court house on left
22.95	0.1	Baird Road stop sign; continue straight on Whalen
23.15	0.2	Junction Route 250 (Fairport-Nine Mile Point Road); turn left
24.60	1.55	Junction 286; continue north on Rte 250
26.15	1.65	Junction Plank Road; continue straight on Rte 250; Harris Garden Center on left
27.45	1.3	Junction State Road; stop light village of Webster; continue straight
28.25	0.8	Junction Route 404A; continue straight on Rte 250
28.55	0.3	Entrance ramp to Route 104 East; turn right onto entrance ramp and enter onto Route 104
28.75	0.2	Salt Road Exit
30.85	2.1	Basket Road
31.55	0.7	County Line Road; enter Wayne County, Town of Ontario
31.85	0.3	Switzers Auto Body shop on right; a small outcrop in ditch is the westernmost known exposure of Wolcott Limestone; the unit is truncated between this locality and Rochester.
32.45	0.6	Ontario Center Road intersection
33.45	1.0	Knickerbocker Road; turn left (north) at this light for optional Stop 4. (directions follow)
33.65	0.2	Cross railroad tracks
34.1	0.45	Entrance to Ontario Park on left; turn left (west) onto park road
34.25	0.15	Park in lot near building and walk to the right (north) past end of pond and then left (west) along the north side of the pond onto dump piles of rock dug from the pond

OPTIONAL STOP 4-OLD FRUITLAND ORE PITS PARK

The ponds in this park represent old pits dug in the Furnaceville ematite or iron ore bed for the production of red paint oxide (Gillette, 1947). Although in-place rock is no longer accessible within the flooded pits, dump piles provide an overview of the lithology of the Reynales Formation. Most slabs consist of dolomitic limestone with minor bluish gray chert nodules. Molds of pentamerid brachiopods are common on certain slabs. Pentalobate columns of the newly described inadunate crinoid *Haptocrinus* are relatively abundant. While most of the hematitic band was removed for the paint ore production, occasional blocks and slabs show the highly fossiliferous, oolitic hematite.

Return to vehicles and retrace route to Knickerbocker Road

34.4	0.15	Turn right (south) on Knickerbocker Road.
34.85	0.45	Junction NY 104; turn left (east) and proceed on 104.
35.05	0.2	Furnace Road (named for old blast furnaces for iron production from Furnaceville iron ore).
36.65	1.6	Fisher Road; eastern border of Ontario township
37.35	0.7	Apple Valley Speedway
38.30	0.95	Salmon Creek Road
38.35	0.05	Cross Salmon Creek, type section of the mid Silurian Williamson Shale is slightly upstream (south) of road.
38.55	0.2	Tuckahoe Road; enter village of Williamson
39.05	0.5	Williamson town garage on right; small exposure of yellowish weathering Rockway Shale is in ditch
39.35	0.3	Junction NY 21 (Lake Road)
39.45	0.1	McDonalds and Burger King (possible rest stop); continue east on Route 104.
39.85	0.4	Leaving Williamson
40.25	0.4	Cadbury's Company; Pond Road
41.35	1.1	Seneca Foods on left.
41.55	0.2	Townline Road
41.95	0.4	Very minor low outcrops of Rochester Shale in ditch to south.
42.55	0.6	Redman Road
43.45	0.9	Centenary Road
45.35	1.9	View of large wave modified drumlin on right (south)
45.55	0.2	Junction NY 88 (Alexander Road); continue east on 104. Enter Sodus
46.55	1.0	Maple Avenue (stoplight); eastern edge of Sodus

48.25	1.7	Old Ridge Road (stop light)
49.25	1.0	Barclay Road; Helluva Good Cheese factory on left
49.40	0.15	Cross Salmon Creek (East); Rochester Shale below road.
49.80	0.4	NY 240 (N. Geneva Road; blinking yellow light)
50.00	0.2	Note ditch to north (left) near Wallington town line; yellowish clay is weathered Glenmark Shale (formerly considered as upper part of Rochester Shale); This small exposure has yielded very abundant brachiopods, especially *Nucleospira pisiformis*, and small *Enterolasma* (rugose corals) *Stegerhynchus neglectum*. The Glenmark horizon with its distinctive fauna can be correlated from this small outcrop southward into the central Appalachians of Pennsylvania, Maryland, and West Virginia.
51.00	1.0	Bond Road
51.60	0.6	Junction NY 14; turn left (north) off NY 104 onto 14 toward Alton
51.90	0.3	T-intersection with Old Ridge Road, Rt. 143, Village of Alton; turn right onto Old Ridge and immediately (.025) left (north) onto Shaker Road (this is essentially a jog in the north south road).
52.90	1.0	Bend in road; prepare to stop
53.10	0.2	Park opposite southern fence line separating planted field to south from sheep pasture, on left (west) side of Shaker Road. This is the former Shaker Tract, or Alasa farms; white barns to the north now house the Lake Plains Wildlife Rehabilitation Center Follow along edge of fence back to woods bordering east side of Second Creek; enter woods and turn right following near top of bank for about 0.1 mile to where an old logging road descends to creek level; go downstream (north) for one major bend and walk across floodplain and creek to bank on west side.

STOP 5A-SECOND CREEK AT ALTON

This creek bank exposes a small but excellent section of the Clinton strata. Lowest beds exposed are in the Wolcott Furnace Formation, which consists of alternating greenish gray fossiliferous shales and limestones, containing an undescribed stricklandiid brachiopod, together with chaetetid sclerosponges, crinoid ossicles and small *Enterolasma* rugose corals. These beds also display thin fossiliferous hematitic limestone bands. At the top of this interval is a thin, but distinct bed containing granular black phosphatic material. About 20 cm of overlying greenish gray shale may represent the westernmost occurrence of the Sauquoit Shale, an interval which exceeds 30 m (100 ft) in thickness in east central New York, south of Utica. Here the interval has very nearly been truncated by an erosional surface. This unconformity is marked by a sharp surface and a very distinctive lag conglomeratic bed (the Second Creek phosphatic bed), best developed at this locality. This bed is approximately l to 2 cm in thickness, but in places, it contains clasts of dolomitic limestone, evidently derived from nearby exposures of the Wolcott or other lower Clinton carbonates. The clasts rarely exceed a cm in thickness, but may be up to 10 cm across. In places, these clasts are edgewise and may jut upward into the overlying Williamson Shale a few centimeters. The erosional clasts are packed in a matrix of dark gray, very pyrite- and phosphate-rich crinoidal grainstone. Pebbles of black phosphate and quartz up to 1 centimeter in diameter are present within the matrix. The lower surface of the bed is typically sharp and slightly irregular. Its upper contact is sharply overlain by black or very dark gray Williamson Shale and typically displays an abundance of the brachiopods *Eoplectodonta* and *Atrypa*. Current aligned specimens of the graptolite *Monograptus clintonensus* occur on the contact in some places. The unconformity here is the same one examined at Palmers Glen near Rochester (Stop 2). However, here the Williamson Second Creek phosphate bed unconformably overlies a remnant of the Sauquoit Formation, whereas at Rochester, the Williamson rests on the lower portion of the Sodus, some 10 m lower in the section. Hence, the unconformity is regionally angular and has truncated successively older Clinton strata to the west.

The overlying Williamson Shale includes alternating relatively hard, very dark gray to black graptolitic shale and soft greenish gray claystones. As at the Palmers Glen exposure (Stop 2), graptolites are generally confined to the dark gray layers and may occur on laminae of quartz sand or very coarse silt. Greenish gray layers are generally barren of fossils but may contain bedding plane assemblages of frilly *Atrypa* and *Eoplectodonta*. We will walk upstream along the bank of Second Creek to outcrops of the Rockway Dolostone and shale and the Irondequoit Limestone. The Rockway, well displayed in slabs recently ripped up during flooding from the creek floor, is fossiliferous, somewhat greenish gray silty mudstone, which weathers to a yellowish color. It contains an abundance of brachiopods, particularly *Clorinda,* (considered a classic deeper water, benthic assemblage 4 to 5 indicator), *Leptaena*, *Atrypa,* and others. The shales also contain scattered specimens of the small rugose coral *Enterolasma* and this locality has yielded specimens of the probable green alga *Ischadites.* Thin, muddy, micritic limestones interbedded with the shales contain scattered crinoid-bryozoan debris, including calyces of the tiny inadunate crinoid *Pisocrinus*. The contact with the overlying Irondequoit Limestone at this locality appears conformable, but relatively sharp. The Irondequoit consists of approximately 3 m of crinoid- and brachiopod-rich wacke- and packstone limestone. Small bioherms of bryozoans and possible algal bound micrite are also present at several levels in the creek bed and minor waterfalls here. The Irondequoit is interpreted as a relatively major sea level lowstand or shallowing event that separates the maximally transgressive Williamson and Rockway deposits from the overlying deeper water Rochester Shale. The small patch reefs appear to have grown upward during times of relative deepening as the Irondequoit passed upward into mudstones of the Rochester Formation.

If time permits, we may walk further upstream to sample from the richly fossiliferous lower Rochester Shale which contains a diversity of brachiopods, small corals, and trilobites. Most of these fossils, however, are present as disarticulated and somewhat fragmented material.

		Return to vehicles and continue north on Shaker Road, past buildings of the former Alasa Farms.
55.10	0.2	T-Intersection with Red Mill/Shaker Tract Road; turn left (west) onto Red Mill Road.
55.20	0.1	Pull to right and park in small bay; walk down path to banks of Second Creek

OPTIONAL STOP 5B-SECOND CREEK AT RED MILL ROAD

If time permits, we may look at the section along Second Creek at Red Mill Road. Immediately downstream from the road bridge are bank outcrops of green to purplish colored Upper Sodus Shale, containing a diverse *Eocoelia* brachiopod and ostracod rich *Zygobolba decora* fauna. These beds are sharply overlain by crinoid- and pentamerid-rich grainstones of the Wolcott Formation.

Return to vehicles and pull out to left, reversing direction (east) onto Red Mill Road.

55.30	0.1	Shaker Road intersection and Alasa Farms; continue east on Red Mill Road, which changes name to Shaker Tract Road
56.10	0.8	Good view of Sodus Bay to left (north)
56.60	0.5	Sharp bend in Shaker Tract Road at intersection with Hunter Road; follow Shaker Tract Road around to right (south)
58.00	1.4	Intersection of Old Ridge Road (Rt. 143).; continue on Shaker Tract which here changes name to Brick Schoolhouse Road
59.00	1.0	Intersection with NY 104; turn left (east) onto NY 104 and proceed.
59.50	0.5	Norris Road intersection; prepare to stop
59.60	0.1	Pull off to right and park on shoulder of NY 104; walk down into small gully immediately south of the highway at point where an exposed yellowish clay (Rochester Shale) is visible on bank above gully; you can then continue to the south into woods along the unnamed tributary of Sodus Creek

OPTIONAL STOP 6-UNNAMED TRIBUTARY OF SODUS CREEK IMMEDIATELY EAST OF NORRIS ROAD AND SOUTH OF US ROUTE 104

This small creek provides a relatively complete outcrop of the upper portion of the Rochester Shale. The section begins with the weathered bank parallel to and south of Route 104. This bank yields good specimens of weathered brachiopods, particularly *Eospirifer*, *Rhynchotreta*, and *Whitfieldella*. Upstream, along the main creek, are good exposures of the middle and upper portions of the Rochester Shale. Thick calcisiltite beds occur near the mouth of the stream in some of the lowest exposures. Upstream, a series of low waterfalls or riffles expose highly fossiliferous, bryozoan rich layers corresponding to the upper marker beds of the Lewiston Member (Lewiston E or bryozoan beds in Niagara County, New York).alternating shales and thin biostromal limestones, rich in the small ramose bryozoan *Chilotrypa*. Associated with these layers are abundant articulated and specimens of the brachiopods *Whitfieldella*, *Rhynchotreta*, *Eospirifer*, *Atrypa* and rarewell preserved specimens of the trilobites *Calymene*, *Bumastus* and others..

The change to more sparsely fossiliferous shales above this small falls marks the contact between the Lewiston and Burleigh Hill members. The upper beds, south to the bridge of York Settlement Road, are within the upper part of the Rochester Shale and consist of alternating mudstones, thin shell layers and calcisiltite beds. Some layers particularly near the top are rich in trilobites such as *Dalmanites*, and brachiopods such as *Stegerhynchus*, *Coolina*, *Dalejona* and occasional small rugose corals (*Enterolasma*). The change from bryozoan-rich biostromes to more sparsely fossiliferous shales marks a relatively major flooding surface within the Rochester Shale that can be identified throughout much of New York State and Pennsylvania. The upper shaly interval is more monontonous than the lower portion, reflecting highstand conditions. However, in the uppermost 2 to 3m the shales become increasingly interbedded with dolomitic calcisiltites.

The section is capped by a low waterfalls immediately north of the bridge on York Settlement Road. This falls is capped by a distinctive 40 cm-thick massive dolomitic limestone bed, referred to as the basal Glenmark bed. This bed, which contains small favositid and rugose corals, as well as the distinctive brachiopod *Nucleospira*, marks the approximate position of the DeCew Dolostone of western New York. This bed and the overlying shales, which are rich in distinctive brachiopods, particularly *Nucleospira*, and *Whitfieldella* (cf. *W. marylandica*) represent an important marker interval that is traceable from this area of New York State southward through much of central Pennsylvania Appalachians and into Maryland and West Virginia. This interval, previously unrecognized or treated as simply an upper portion of the Rochester Shale is herein recognized as a distinctive unit referred to as the Glenmark Formation. In nearby Sodus Creek, this 3 m interval is overlain by thinly bedded micritic limestones and very dark grey, fissile shales, as well as thrombolitic and intraformational conglomeratic horizons marking the base of Sconondoa Formation, the lateral equivalent of the lower part of the Lockport Group of western New York. It represents a distinctive shallowing to nearly peritidal or shallow lagunal conditions in the late Wenlock or early Ludlow time.

		Reverse direction on NY 104 by either backing to Norris Road, if traffic is not heavy, or by proceeding 1.7 miles to NY 414 to reverse direction. Mileage on return west-bound log is from Norris Road. Because most landmarks are noted in the eastbound log, only new sites will be noted in the return log
59.00.0		Norris Road
59.50	0.5	Brick Schoolhouse/Shaker Tract Road
60.50	1.0	Pre-Emption Road
61.50	1.0	NY 14 (Alton)
63.10	1.6	Wallington townline sign and Glenmark Shale in ditch
66.60	3.50	Sodus
73.70	7.10	Williamson
81.60	7.90	Monroe/Wayne Co. Line (Rte. 404)
84.60	3.0	Junction NY 250; Webster
86.60	2.0	Five Mile Line Road
87.10	0.5	Maplewood Shale was exposed in ditch between two lanes of Rt. 104
89.00	1.9	Bay Road (a turn left here would lead to Glen Edith Road and outcrops along road leading down to restaurant by Irondequoit.)
89.40	0.4	East side Irondequoit Bay; note slumps of soft lake silt; views of bay and bay mouth bar from bridge
89.90	0.5	West side of Irondequoit Bay
90.60	0.7	Routes. 104 and 590 divide ahead; bear left into entrance of I-590 South.
90.90	0.3	Enter onto I-590 South
91.50	0.6	Cross Densmore Creek; good exposures of Medina and lower Clinton outcrop along the Creek west of the highway to Densmore Road
92.40	0.9	Empire Boulevard exit

94.80	2.4	"Can of Worms," I-490 & I-590 junction; bear right continuing on I-590 South
101.0	6.2	Exit 16A for NY15A; bear right onto exit
101.3	0.3	Junction NY 15A; turn right (north) on 15A
101.4	0.1.	Cross Erie Canal
101.9	0.5	Route 15A forks with South Avenue; bear left on 15A
102.3	0.4	Junction NY Route 15; 15A ends; turn right (north) onto Rte. 15.
102.5	0.2	Junction Elmwood Avenue; turn left (west)
103.2	0.7	Elmwood curves to the right under old railroad overpass marked University of Rochester; follow to right and prepare for righthand turn
103.3	0.1	Junction Wilson Blvd.; turn right.
103.4	0.1	Univ. of Rochester parking lot.

End of Trip.

Corals of the Onondaga Limestone.

1. *Heterophrentis prolifica* 2. *Cylindrophyllum* sp. 3. *Syringopora* sp.

[From Hall, 1843, Geology of the Fourth District, Figure 33, p. 30.]

THE LEROY BIOHERM REVISITED -

EVIDENCE OF A COMPLEX DEVELOPMENTAL HISTORY

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INTRODUCTION

The LeRoy Bioherm is among the best known rock exposures in New York State, having been designated a National Fossil Coral Reef by the U.S. National Parks Service and having been the subject (at least in part) of three past studies (Crowley and Poore, 1974; Coughlin, 1980; Lindemann, 1988). Despite this the uniqueness of the LeRoy bioherm has generally gone unnoticed. It is the only Edgecliff bioherm known to have a mound building community of the very delicate branching tabulate coral (Cladopora) and the only known Edgecliff mound with a thickness greater than about 1.2 meters that was not built by phaceloid colonial rugosans (Wolosz, 1992a). Further, the LeRoy Bioherm is the only bioherm in this area with a well developed coral mound facies, while all other Edgecliff bioherms in western New York State are thicket/bank structures (as defined by Wolosz , 1992a).

The ecological zonation of reef building species both within mounds and across reef tracts due to environmental gradients is a well known feature of both Recent and ancient reefs (Wilson, 1975; James, 1983). Hence, the unusual nature of the mound building community at the LeRoy Bioherm (as compared to the typical rugosan dominated Edgecliff mounds), and the size of this cladoporid dominated mound, suggests that environmental conditions in the area must have differed in some way from those prevalent at the other Edgecliff bioherm locations.

LOCATION AND GEOLOGIC SETTING

The LeRoy Bioherm is located in an abandoned quarry near the southeastern corner of the Byron, N.Y. 7.5' quadrangle, approximately 5 km. NNW of the village of LeRoy, N.Y. The Central bioherm is exposed in the middle of the quarry, while a second bioherm is exposed along the southeast wall of the quarry (Fig. 1).

Coughlin (1980) notes the presence of a third bioherm along the west wall of the quarry, but it is very poorly exposed and will not be discussed here.

Geologically, the bioherm is in the western facies area of the Middle Devonian Edgecliff Member of the Onondaga Formation as defined by Oliver (1976). Here, the basal Edgecliff is a transgressive crinoidal grainstone/packstone deposited on a disconformable surface, with the underlying units ranging from the Lower Devonian Bois Blanc Limestone to the Silurian Akron Dolomite (for detailed stratigraphy see Oliver, 1954, 1956, 1976).

PREVIOUS WORK

The three previous studies of the LeRoy Bioherm lacked a basinwide perspective of the Edgecliff and as a result failed to recognize the unique nature of the LeRoy mound building fauna. Poore's study of the bioherm (Crowley and Poore (1974)) is a detailed micro- and biofacies analysis which subdivided the bioherm into ten distinct facies (Figure 2). Coughlin (1980) and Lindemann (1988) both corroborated the presence of these microfacies. Coughlin's (1980) insightful interpretation placed the bioherm in a shallow water, protected environment. Lindemann (1988) interpreted the bioherm as displaying a tripartite ecological succession related to changes in relative water depth, and suggested a moderately deep shelf to shallow subtidal setting for the bioherm. All three workers envisioned the development of the bioherm as one of continuous coral growth upwards into progressively shallower water. Wolosz (1992a, in press) noted that these bioherms are best referred to as a combination of a coral mound and a "thicket\bank" structure - an Edgecliff bioherm in which a pre-existing crinoidal sand bank was colonized by a single phaceloid colonial rugosan thicket. He interpreted the LeRoy bioherm as an extensively eroded coral mound which was first onlapped and buried by crinoidal sand, and later colonized by a rugosan thicket.

THE CENTRAL BIOHERM - CLASSIC INTERPRETATION

Poore's facies map of the Central bioherm (Figure 2) has been the standard interpretation of this structure for the past 25 years. His Inner Core and Outer Core facies have been interpreted as the growth center of the mound, with the Transitional facies representing a surrounding debris rim. The phaceloid colonial rugosans of the Heliophyllum facies have been considered as predominantly in place colonies with the resulting asymmetry of the mound (the Heliophyllum facies is restricted to the east side of the mound) attributed to an "energy shadow" which allowed the rugosans to grow behind the central mound. The remaining facies are primarily flanking beds (with the exception of the Protocap facies which is a colonial rugosan thicket) which were differentiated on the basis

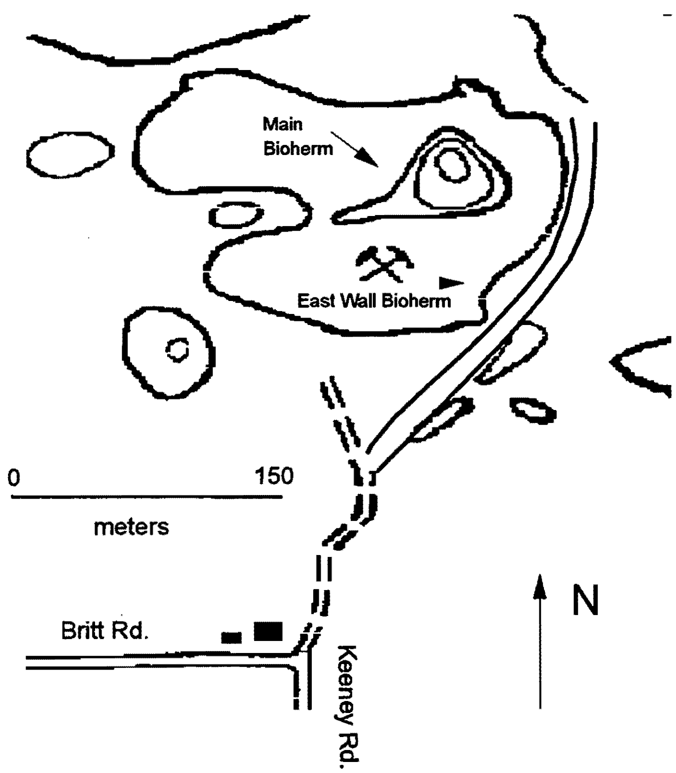

Figure 1: The LeRoy bioherm quarry, showing location of bioherms. Double line on east side of quarry shows current position of road.

of microfacies and/or fossil content, and position - "flank" beds onlap the mound, while "cap" beds cover the mound (for detailed discussion of facies see Crowley and Poore, 1974).

These facies have been interpreted as a record of the transition of the bioherm into shallow water through either simple upward mound growth or due to a shallowing event (Lindemann, 1988), which would not be unusual, especially for Edgecliff bioherms (Wolosz, 1985, 1992b). However, this model fails to explain a number of features which make the LeRoy structure unique among the Edgecliff bioherms.

The organic nature of the bioherm is unusual in two ways. First, the cladoporid mound building community is unique among Edgecliff bioherms. Every other known Edgecliff bioherm has a mound constructed by colonial rugosans, or in the case of some very small mounds, by the tabulate genus Syringopora (Wolosz, 1990, 1992a). Second, the robust branched colonial rugosans (Heliophyllum facies) are interpreted as having grown in a low energy environment ("energy shadow") while the more delicate branched Acinophyllum are found on the high energy side of the mound (Protocap facies) and the mound itself is built by the very delicate and small cladoporids. This would be exactly the opposite of what should be expected based on study of eastern Edgecliff bioherms (Wolosz, 1992b).

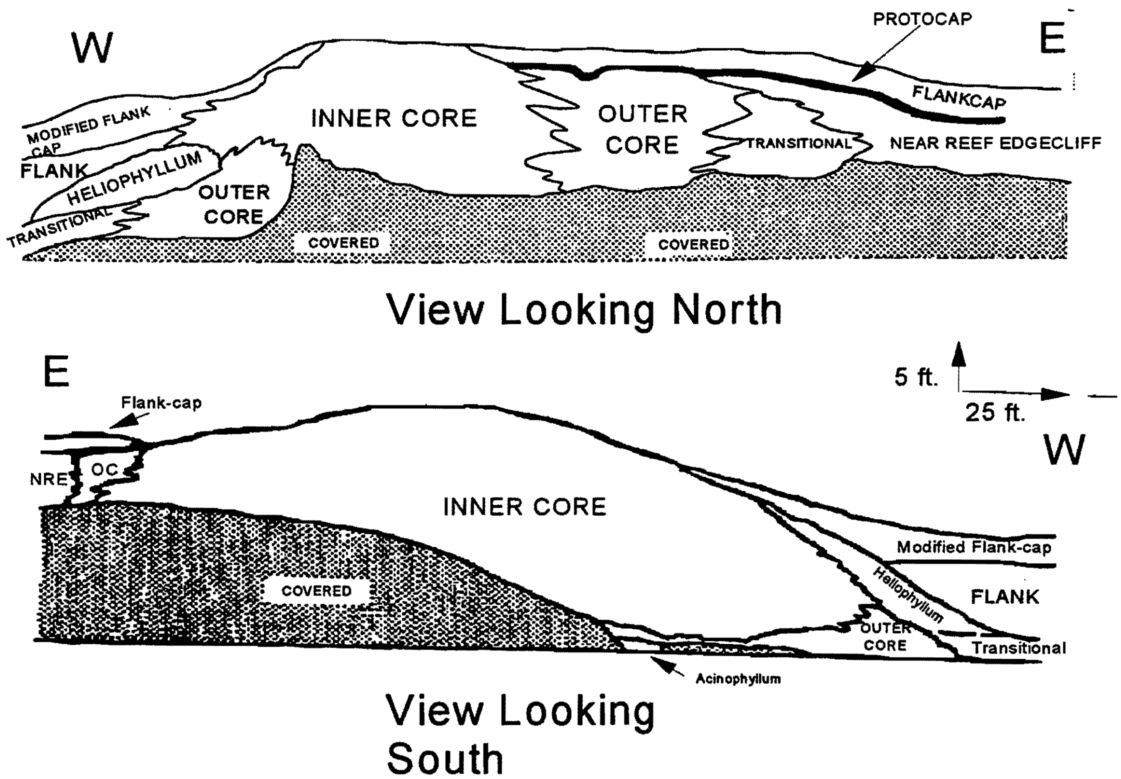

Figure 2: LeRoy bioherm facies as defined and interpreted by Poore. Inner Core, Outer Core (OC) and Transitional facies represent growth center of reef and rubble rim. Heliophyllum facies - in place coral growth in "energy shadow". All other facies are flanking beds differentiated on basis of petrology, fossils and position relative to mound. NRE = Near Reef Edgecliff (for detail see Crowley and Poore, 1974).

The diagenetic history of the bioherm is also unusual. The bioherm core is the only Edgecliff mound which weathers in a "vuggy" pattern. The partial silicification of the cladoporids in the Inner Core facies represents the only case of notable silicification in any Edgecliff mound.

THE CENTRAL BIOHERM - RECENT INTERPRETATION

Re-examination of the Central bioherm and the east wall bioherm were carried out as part of a regional study of Edgecliff bioherms (Wolosz and Paquette, 1988; Wolosz, 1992a). The re-interpretation of the Central bioherm is illustrated in Figure 3.

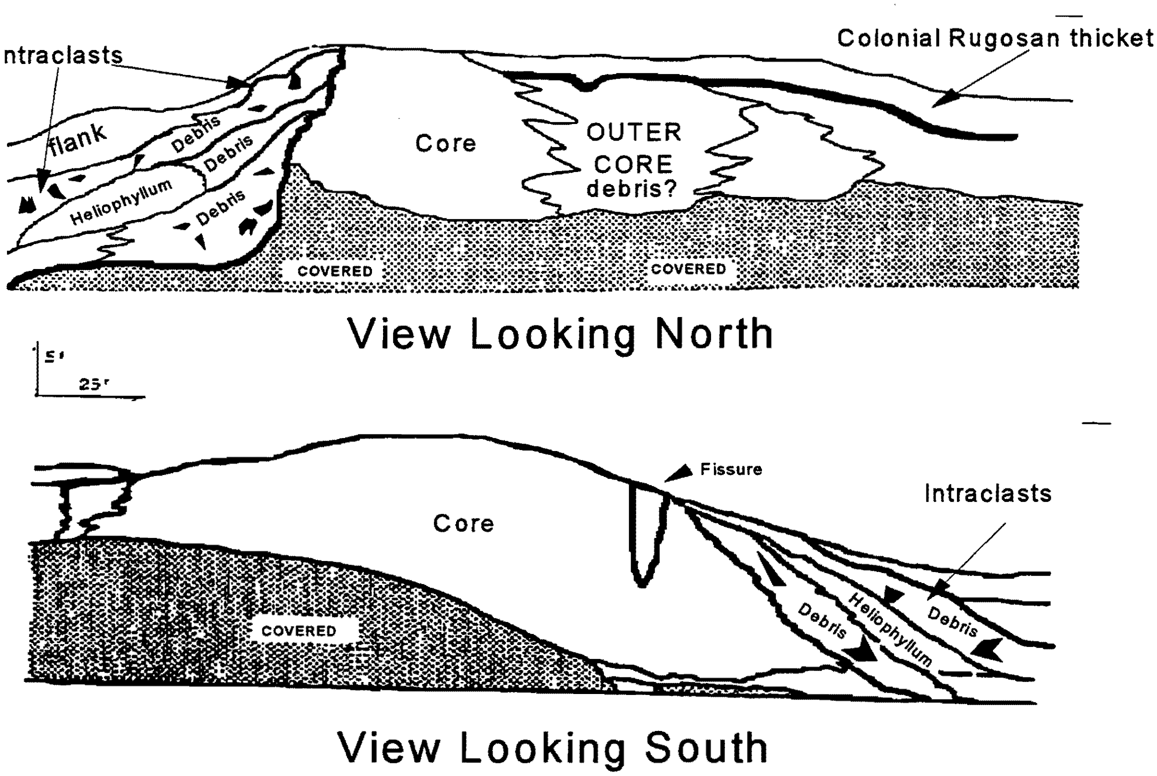

Figure 3: Recent interpretation of the LeRoy bioherm. The western end of the exposure is interpreted as core debris rich flank beds. Large, fine grained intraclasts are common both above and below Heliophyllum facies which is considered to be a coral debris apron. Note presence of a fissure cutting core facies.

Contrary to previous interpretations, evidence indicates that the core was subject to extensive erosion prior to the deposition of the capping beds. In thin-section the contact between the core and the overlying facies is a sharp erosional boundary, with fossils within the cladoporid core cleanly truncated at the contact. It is, however, often difficult to find the exact boundary because the flanking debris beds (much of Poore's Outer Core facies) are made up of cladoporid core intraclasts in a matrix of core debris (calcisilt and cladoporid fragments) with only poor contrast between the clasts and the matrix. On the north side of the core a sharp vertical contact marks the presence of a fissure within the core (Figure 3). This contact is easy to miss because the fissure fill consists of fine crinoidal debris in a calcisilt matrix which weathers to the same hue as the core (dark gray), but upon careful examination a distinct textural difference becomes obvious. The initial fissure filling is a fine, calcisilt rich, crinoidal packstone, which grades into a coarse crinoidal grainstone/packstone. Small intraclasts of core facies are common in the fissure filling facies.

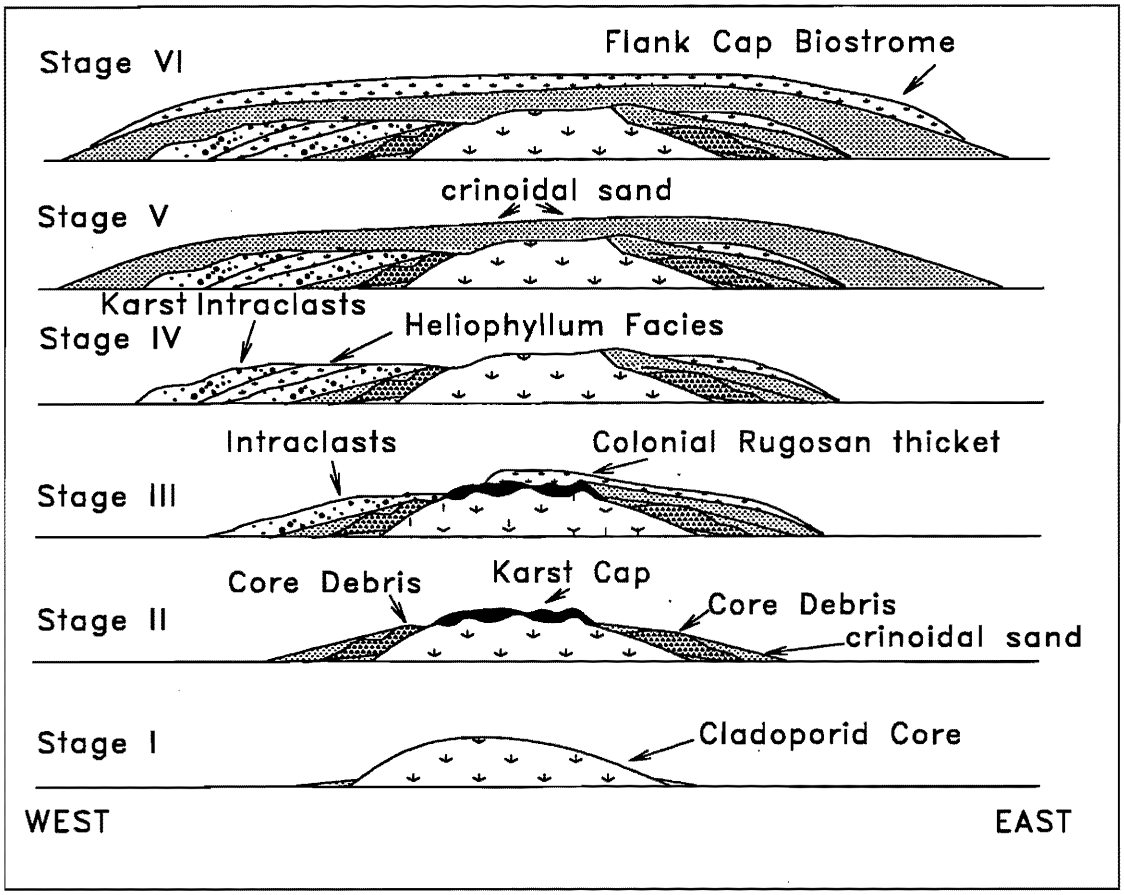

Figure 4: Model for the development of the LeRoy bioherm. Perspective is that of an individual looking north at the central bioherm. "Intraclasts" refers only to intraclasts of karst cap facies. Diagram is not drawn to scale. See text for details.

Large intraclasts are very common in the crinoidal packstones which lap up onto the core on the west side of the exposure. The intraclasts are easily identifiable in outcrop by their light gray, fine-grained, non-fossiliferous appearance. These intraclasts are the only preserved evidence of a mound-capping facies which has been totally removed from the central bioherm exposure. In thin section these intraclasts are composed of fine calcisilt, either massive or weakly laminated, and in some cases containing small fractures filled with smaller intraclasts of the same fine calcisilt. One large fracture contains abraded crinoid grains ranging from nearly complete to fine fragments. These intraclasts, in light of the fissures described above, are interpreted as remnants of a thin karsted cap which has been eroded from the upper portions of the mound.

Finally, the large phaceloid rugosans (Heliophyllum facies) are interpreted as colonies which had occupied the crest of the mound (probably a continuation of Poore's Protocap facies), but which were later displaced into the back-mound area. It is important to note that the fine grained intraclasts can be found both above and below the colonial rugosan horizon indicating continued erosion of the mound following growth and destruction of the rugosan thicket.

EAST WALL BIOHERM.

The east wall bioherm is not well exposed, but does lend support for this new interpretation of the Central bioherm.

Core facies of the east wall bioherm extend to the top of the quarry wall, but an exposure recessed back from the quarry edge among the trees displays a capping colonial rugosan thicket. The thicket does not directly overlie the core, being separated from it by roughly 2 meters of packstone. Further to the north, along the quarry wall, another thicket is exposed which is separated from the top of the core facies by only about 0.2 meter. This second horizon would, based on its relationship to the core facies, be equivalent to Poore's (Crowley and Poore, 1974) Protocap facies. The stratigraphically higher thicket is equivalent to a dense rugosan/favositid biostrome which caps Poore's Flank Cap facies along the eastern edge of the Central mound exposure, and is separated from the Protocap facies by roughly 2 meters of grainstone and packstone.

The presence and position of these colonial rugosan thickets proves that the cladoporid core was at least partially buried by carbonate sand and silt prior to growth of the first thicket and was totally buried prior to development of the second. Hence, the LeRoy bioherm is best described as a thicket\bank structure (Wolosz, 1992, in press). Further, any removal of colonial rugosans from the crest of the mound would form a debris apron. The thickness of the Heliophyllum facies, its onlapping position, and its interfingering with mound debris clearly supports its interpretation as a coral debris apron.

DEVELOPMENTAL HISTORY

This new data leads to a reinterpretation of the developmental history of these bioherms, which can now be seen as a five stage process (Figure 4), and not a simple community succession due to growth into shallow water. Also, this reinterpretation answers some of the questions noted previously.

Stage I: Core growth. The cladoporid core at LeRoy represents growth in shallow, wave agitated water, with conditions generally similar to those described by

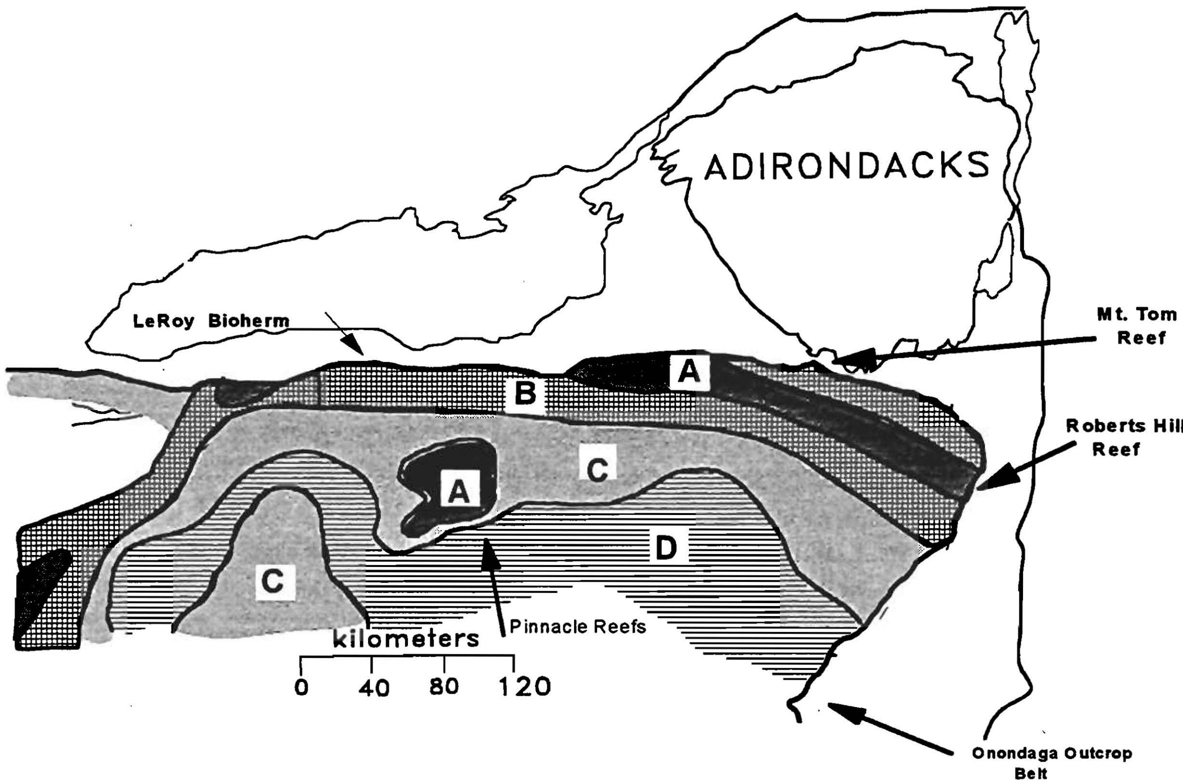

Figure 5: Subsurface facies map of the basal Edgecliff Member of the Onondaga Formation. A = Grainstone; B = packstone; C = wackestone; D = mudstone or nondeposition. Note that presence of pinnacle reef area appears to result in an embayment in carbonate facies to the west. (Redrawn from Cassa (1980)).

Turmel and Swanson (1976) for Rodriguez Key, Florida (see also, James, 1983, p.373-374). The delicate branching nature of the cladoporids precludes high energy conditions, and the lack of any major development of colonial rugosans other than the initial Acinophyllum horizon (Figure 2) also argues for quiet water (see Wolosz, 1992b).

Stage II: Lowering of sea-level results in initial erosion of mound, and deposition of debris flanks, followed by subaerial exposure and the development of a thin karst cap. Minor silicification of cladoporids and early diagenetic modification of the core (resulting in its "vuggy" appearance) occurs at this time. Diagenetic processes are similar to those described by Moore, et al. (1980) for a Cretaceous carbonate beach.

Stage III: Rise in sea-level and overall environmental energy causes partial erosion of karst cap (intraclasts deposited in back-reef). Continued sea-level rise results in colonial rugosan thicket formation (Protocap facies of Crowley and Poore, 1974) which covers the eastern side of the mound.

Stage IV: Second (minor) lowering of sea-level results in scouring of colonial rugosans from the crest of the mound and deposition as rubble apron (Heliophyllum facies). Continued lower sea-level results in renewed erosion of the mound following erosion of karst facies remnants (intraclasts found above Heliophyllum facies).

Stage V: Shallow depth conditions continue, leading to burial of the mound by shallow water crinoidal sand.

Stage VI: Slight sea-level rise results in development of coral biostrome at central bioherm (upper portion of Flank Cap facies at eastern end of exposure) and second rugosan thicket at east wall bioherm.

Note that major sea-level fluctuations are not required by this developmental model. Transition from mound growth to erosion only requires a change from a protected to an open environment with greater wave energy. Following subaerial exposure, erosion and burial of the mound could have been accomplished by fluctuation in sea level of a few meters or less.

DEPOSITIONAL ENVIRONMENT AND CORRELATION WITH OTHER EDGECLIFF REEFS

Coughlin's (1980) assessment of the environment of deposition of the LeRoy core as being a protected, quiet water environment, possibly in the lee of islands of Silurian rock, exhibited great insight. However, it is possible that the protected environment necessary for this unique mound building community may have been provided by the topography of the basin itself, rather than erosional remnants of pre-existing limestone. Cassa (1980) compiled a subsurface facies map of the Edgecliff which presents some interesting data (Figure 5). This map depicts the Edgecliff pinnacle reefs as having been initiated on a grainstone base roughly 80 to 110 km. southeast of LeRoy, but her facies map also appears to depict a type of embayment. The presence of shoals to the southeast of LeRoy would have damped any open ocean waves and resulted in an environment subject only to wind-driven waves near LeRoy. Later subsidence which led to the growth of the pinnacle reefs would have removed much of the original barrier and resulted in a more open environment around the bioherm.

CONCLUSIONS

Current evidence leads to a re-evaluation of the development of the LeRoy bioherm. The classic interpretation of simple upwards mound growth into shallower water is no longer tenable. The current model interprets the development of the bioherm as a five stage process which includes the growth of a quiet water cladoporid mound followed by subaerial exposure and erosion and finally reimmersion in shallow waters. This new model accounts for the unusual mound building community, the positioning of the colonial rugosans of the Heliophyllum facies, and the diagenesis of the mound.

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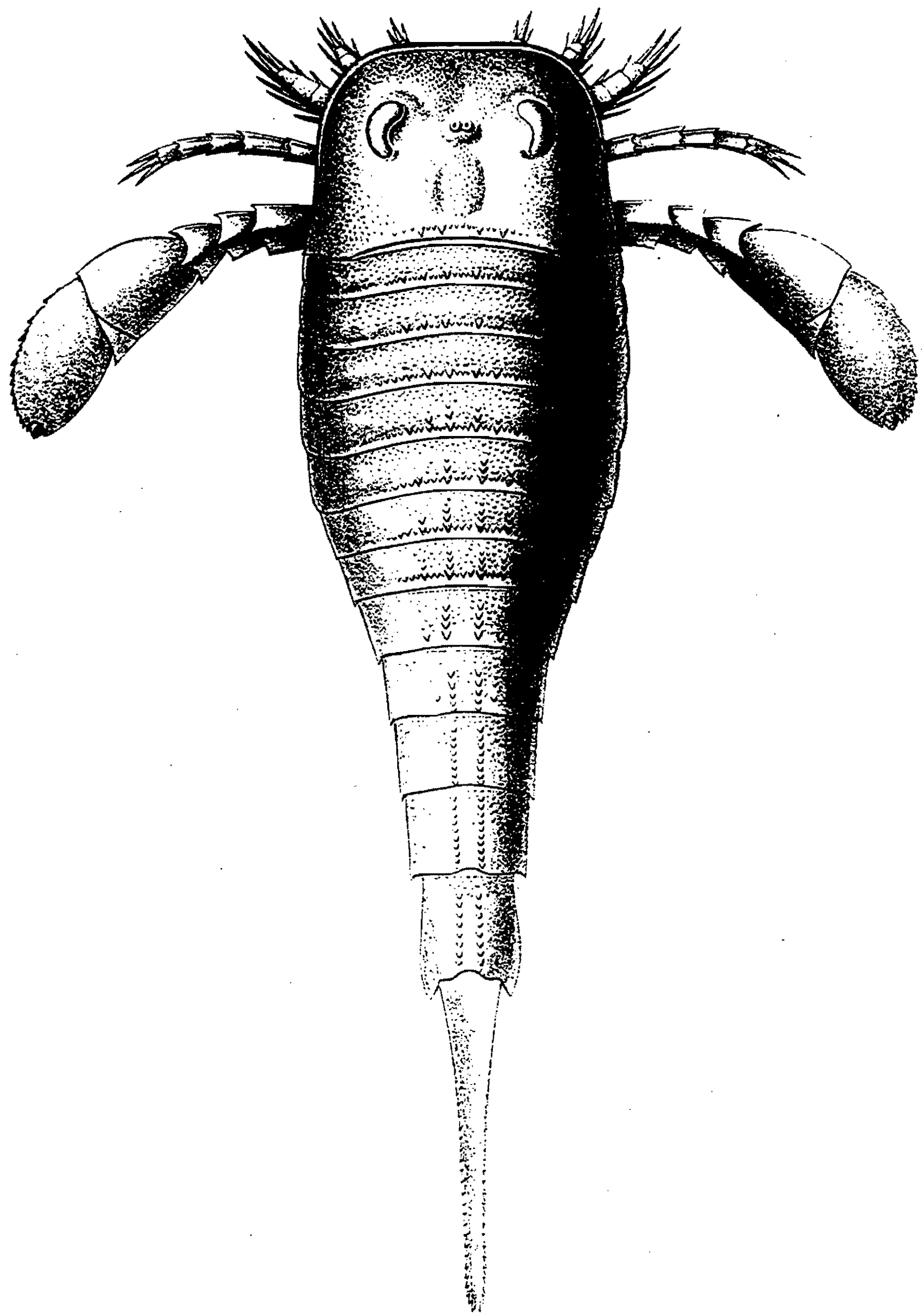
ROAD LOG

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0	0	Intersection of Route 5 and 19 in village of LeRoy, N.Y. Proceed north on Route 19.
2.1	2.1	Left turn (west) onto Richmond Road
3.8	1.7	Right turn (north) onto Keeney road.
4.1	0.3	Road turns due west, becoming Britt Road (see Figure 1 in text). Stop at curve, road leading into quarry is overgrown. Walk past east side of house on north side of road to dirt road. Follow dirt road into quarry.

NOTE: PLEASE ASK PERMISSION AT HOUSE BEFORE ENTERING PROPERTY.

LEROY BIOHERM

The most direct entrance into the quarry is down the talus slope which covers much of the east wall bioherm. From the top of the quarry wall the central bioherm can clearly be seen in the center of the quarry pit. The core facies of the bioherm are massive, weathering to a dark gray and can be easily differentiated from the bedded flanking facies which are very light gray to almost white limestone (see text for description). The quarry walls consist of highly fossiliferous Edgecliff shallow marine facies.



Eurypterus remipes remipes

[From Clarke and Ruedemann, 1912. Eurypterida of New York. N.Y.S. Museum Memior 14, Plate 2.]

LATE SILURIAN SEDIMENTATION, SEDIMENTARY STRUCTURES AND PALEOENVIRONMENTAL SETTINGS WITHIN AN EURYPTERID-BEARING SEQUENCE [SALINA AND BERTIE GROUPS], WESTERN NEW YORK STATE AND SOUTHWESTERN ONTARIO, CANADA

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INTRODUCTION

One of the most fascinating aspects of Late Silurian sedimentation in the Appalachian Basin is the occurrence of thick sequences of evaporites and associated sediments. In western New York and southwestern Ontario, Canada, these sequences comprise the Salina and Bertie Groups and have been, or are economically important sources of both halite and gypsum.

Our particular interest in this evaporite sequence stems from the peculiar occurrence of complex eurypterid assemblages distributed, cyclically, throughout the sequence. If there is any hope of understanding the true habitat of the eurypterids, it will come from a more detailed examination of all lithofacies and contained sedimentary structures. We continue in this paper, and associated field trip, to call attention to the great variety of rock types, sedimentary structures, unusual fossil assemblages, and their stratigraphic and geographic distribution.

In the following text, it will be noted that eurypterids are mentioned throughout. This is not only an attempt to illustrate the importance of the eurypterid faunas found throughout the region, but to show their intimate association with part of the sedimentological record reviewed. Furthermore, because eurypterids are structurally complex; because they molted frequently during their growth; and because they often became disarticulated; eurypterid parts were subjected to fragmentation, redistribution by sorting and orientation, and even comminution as bioclasts. We suggest that these processes are important sedimentological features worthy of observation and study.

Likewise, "algal" stromatolites, those moundlike to laminar structures so often encountered in some of the eurypterid beds, are seen herein not only as important biological structures that affected the local sedimentological regimes, but as contributors to the regional sediments as particles and pieces ripped from the biostromes and redistributed, along with eurypterid debris, into flat-pebble conglomeratic accumulations within, primarily, the waterlimes (e.g. Ellicott Creek Breccia) and the finely-crystalline dolostones (e.g. Victor Dolostone) of the Fiddlers Green Formation (Bertie Group).

STRATIGRAPHY

Late Silurian sedimentation in western New York generated a great variety of interesting litho- and biofacies. Lithofacies consist of redbeds, black shales, green shales and dolomitic mudstone, basinal and marginal evaporites, limestones representing biostromal and biohermal accumulations and their "inter-reef" sediments, and the peculiar *waterlimes* of New York with their contained bizarre arthropods – the eurypterids, pseudoniscids, phyllocarids and scorpions.

We concur with Fisher (1960) and Ciurca (1973; 1978, p. 227) with group status for Salina and Bertie sequences that represent the two important phases of sedimentation that followed deposition of the underlying Lockport Group limestones, dolomitic shales and dolostones. Subsequent to Fisher's Silurian Correlation Chart, Rickard (1975) downgraded the Bertie Group to formation status and included all units in an expanded Salina Group. Ciurca (1990) has further revised and expanded the Bertie Group to include previously unknown, but stratigraphically younger strata (Figure 1).

We interpret the Salina Group to be a sequence consisting primarily of thick redbeds (Vernon-Bloomsburg), evaporites (mostly halite and gypsum or anhydrite), limestones and dolostones (Syracuse Formation), and an upper thick sequence of shale and dolomitic mudstones with evaporites (Camillus Formation). Black-shale units are included primarily within the lowest Salina Group.

In contrast, the Bertie Group, a thinner sequence, consists of massive dolostones with their characteristic intercalated waterlime units; minor shale and mudstone units; and minor evaporites (some gypsum and relict evaporites). The Bertie Group contains the two well-known and important eurypterid assemblages - the *Eurypterus remipes remipes* Fauna (older) and the *Eurypterus remipes lacustris* Fauna (younger). The expanded Bertie Group includes the Moran Corner Waterlime, the youngest *Eurypterus*-bearing formation of the group in western New York.

SALINA GROUP

In western New York, the Salina Group consists of three thick formations. At the base are redbeds and green shales and dolomitic mudstones with intercalated eurypterid-bearing black shales. The black shales are the most studied portions in the area because they contain a rich fauna consisting primarily of eurypterids and other arthropods, bivalves, rare articulate brachiopods, and abundant inarticulates (mostly *Lingula* sp.).

Recent observations (Ciurca, 1990) show a multiplicity of recurrent eurypterid horizons within the lower Vernon Formation. The principal biofacies are the *Eurypterus* Biofacies and the *Hughmilleria* Biofacies. They are overlapping biofacies but certain horizons are completely restricted to one or the other genus.

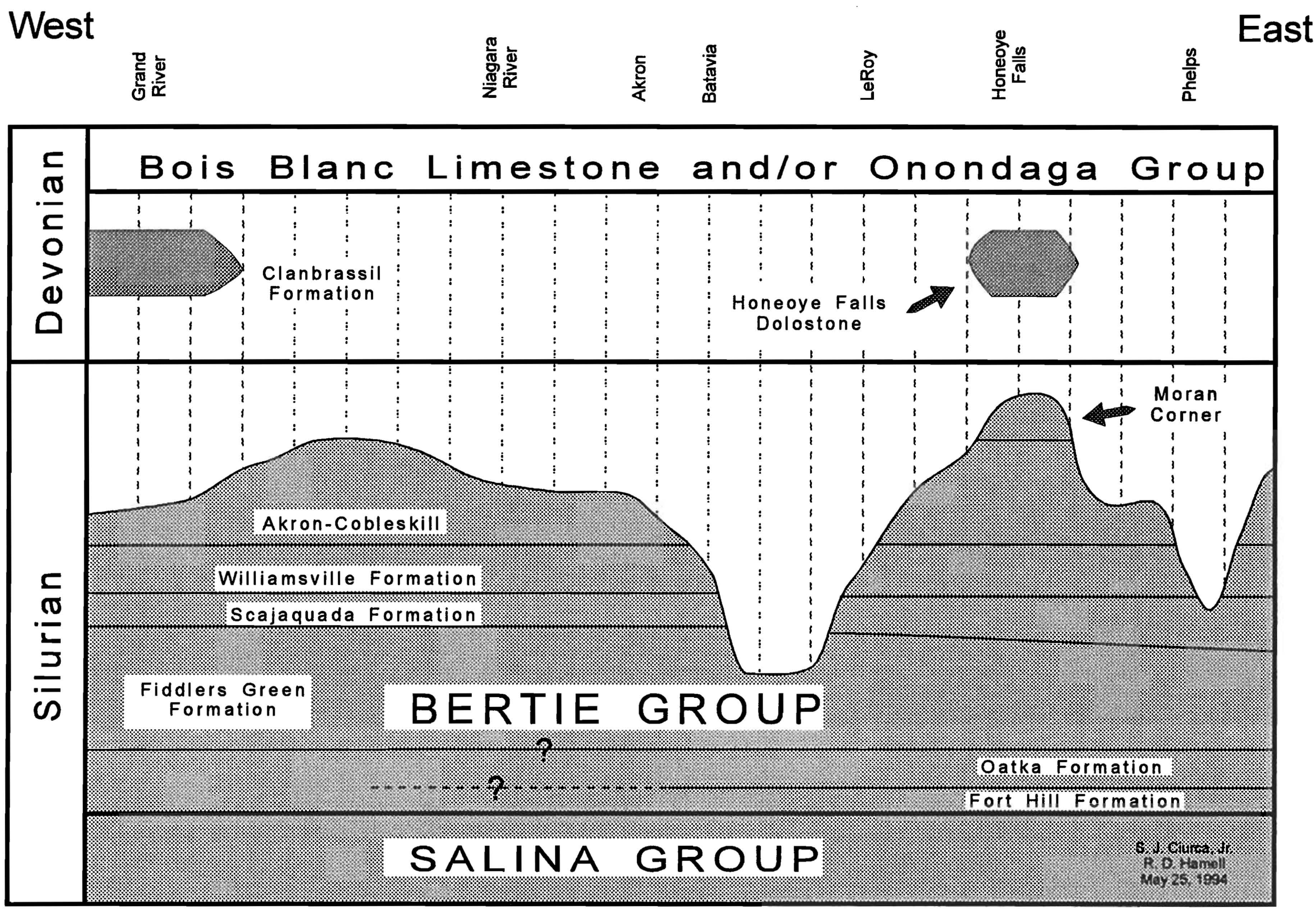

FIGURE 1 Correlation of the Redefined Bertie Group, Western New York and Niagara Peninsula, Ontario, Canada. (Modified from Ciurca, 1990, p. D13; Figure 5).

The middle portion of the Salina Group consists of thin to thick-bedded dolostones and limestones, some argillaceous, usually dolomitic, beds and evaporites that together are referred to as the Syracuse Formation. A transition zone at the base (see Leutze, 1956) bears a significant faunal zone, the *Waeringopterus* Biofacies that has been traced through the region as far west as Batavia and across the border to the Welland Canal in Ontario, Canada (Leutze, 1961, Ciurca, 1990 p. D12). In contrast, the upper portions of the Syracuse Formation contain stromatolitic beds with an *Eurypterus* fauna. Gypsum beds are locally prominent.

The Camillus Formation constitutes uppermost strata of the Salina Group in this part of the state. Shaly dolostone, thin-bedded dolomitic mudstones and evaporites are prevalent. Fossils are exceedingly rare and no important horizons are known in the area.

While exposures of the Salina Group are few and far apart, the divisions can generally be recognized across the region. Important localities exhibiting various portions of the Salina Group are noted below.

<u>Erie (Barge) Canal at Pittsford and eastward</u>

Vernon Formation (redbeds) can be observed along the canal walls at Pittsford, New York (near Main Street Bridge) when the canal is drained for the winter. No natural outcrops of the important eurypterid-bearing units (Pittsford, Monroeav, Barge Canal Beds) are known, all occurrences being excavations for highways, buildings, and the canal.

The Vernon Formation occurs at Fairport in the Erie Canal and was recently reexacavated on the northeast side of the bridge over the canal in the village. Sporadic exposures of Vernon Fm. can also be observed along the canal banks all the way to Newark, New York.

<u>Oatka Creek at Garbutt and the Oatka Trail, Oatka Valley</u>

The Oatka Valley contains important exposures of the Syracuse Formation (Garbutt) and higher beds including much of the Camillus Formation (Oatka Trail and side roads to NY 19). Gypsum was formerly mined at the Garbutt sites and good exposures of resistant Syracuse limestone and dolostone, underlying the gypsum bed, can be observed in the floor of and along the banks of Oatka Creek both upstream and downstream of the bridge over the creek at Garbutt.

The contact of the Salina Group with the Bertie Group occurs along the NY19 roadcut just north of LeRoy where the Camillus Formation is directly overlain by the type Fort Hill Waterlime, the basal formation of the Bertie Group. This contact is also excellently revealed at nearby Buttermilk Falls.

Black Creek at NY33 east of Batavia, New York

Exposures along the west side of Black Creek contain about 20 feet (7m) of the Syracuse Formation and consist of thin to thick-bedded, fine-grained dolostones within the *Waeringopterus* Zone.

I990 Roadcut north of Buffalo, New York

An excellent portion of the Syracuse Formation (Salina Group) can be observed in the I990 exposures. The section consists of argillaceous dolostone and thin to medium-bedded dolostone with intercalated gypsum.

BERTIE GROUP

Overlying the Salina Group in western New York is an assemblage of litho and biofacies quite distinct from much of the strata below. The resistant beds are ledge-formers and consequently a number of waterfalls in the area are supported by Bertie dolostones. Some of the more important Bertie localities are noted at the end of this section.

The term "Bertie" was originated by Chapman (1864) for a sequence of strata in the Niagara Peninsula of Ontario, not far from the Canada-United States border. Originally a formation, Fisher (1960) wisely regarded the rocks as a group consisting of named formations (Oatka through Williamsville). A selected history of Bertie nomenclature is shown in Figure 2). See also Ciurca, 1982.

Chapman undoubtedly included the Akron Dolostone in his interpretation of the Bertie as he gives a thickness of about fifty feet for the sequence west of the Niagara River. Recent measurements of the type Bertie indicate that 45-50 (15-18m) of strata are present, inclusive of the Akron Dolostone. The Akron Dolostone (Akron-Cobleskill strata) consists of very fine-grained rocks in the Canadian sections and approaches a mottled waterlime in character. Because the rocks are so similar to the typical Bertie strata below, especially to the Victor Dolostone, and because the unit is generally thin (3-5 meters), the Akron-Cobleskill here, as well as the newly discovered and stratigraphically higher waterlime (Moran Corner Waterlime, Ciurca, 1990, p. D17), have been included within the Bertie Group (Figure 1). Inclusion of these units assembles all the recurring eurypterid-bearing waterlimes into a useful package (i.e. Bertie Group), presumably completely of Late Silurian age (essentially the *Eccentricosta jerseyensis* Zone in the Appalachian Basin).

Oaks Corners Quarry southeast of Phelps, New York

Exposures in quarry walls exhibit the very irregular contact of the Middle Devonian Onondaga Limestone with the underlying Late Silurian Akron Dolostone. The Williamsville

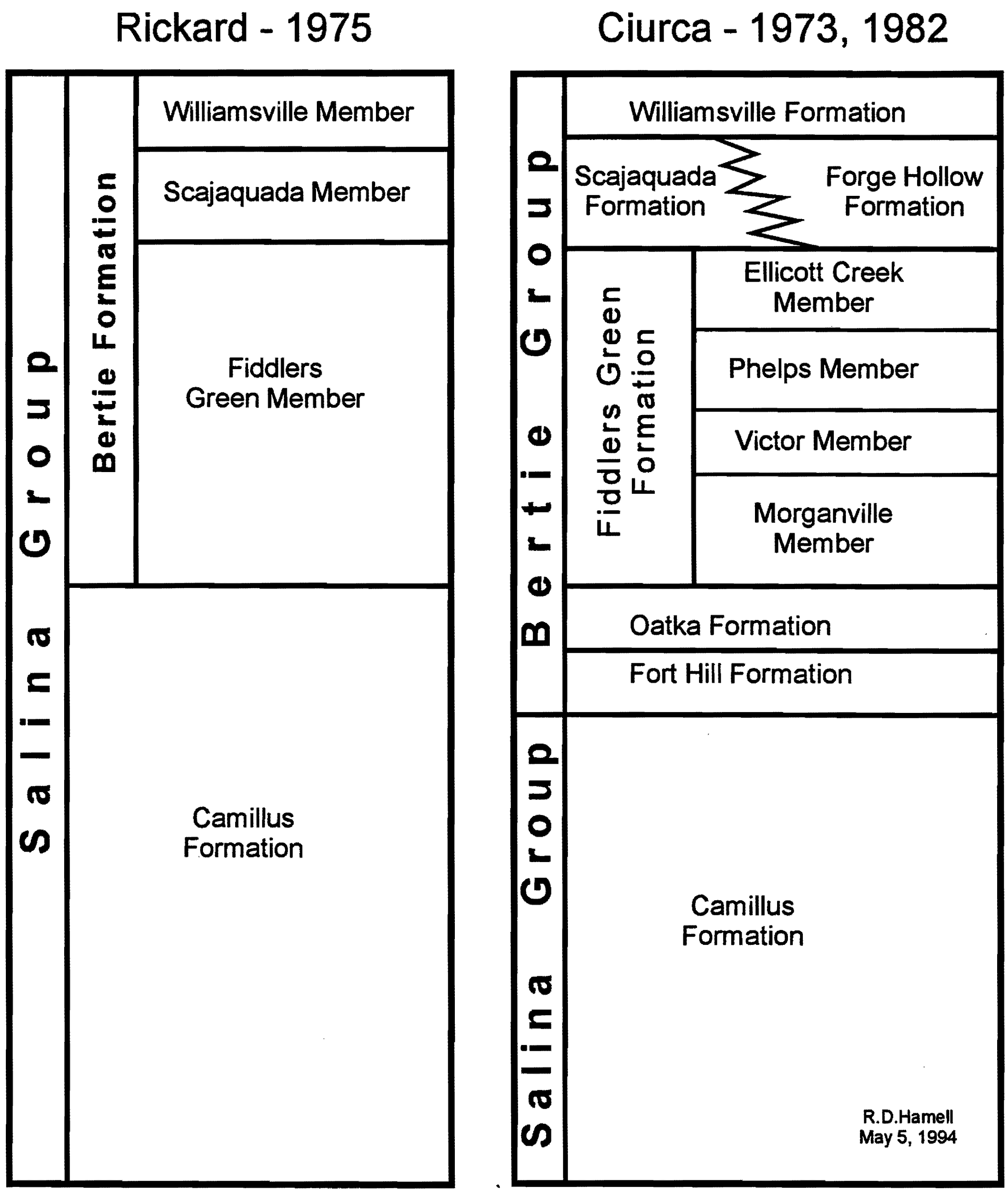

FIGURE 2 Evolution of Stratigraphic terminology for part of the Late Silurian sequence in New York State.

Waterlime is exposed on the quarry floor at the east end with several feet of the underlying Scajaquada Formation usually covered by water at the lowest portion of the quarry.

NYS Thruway and NY88 roadcuts at Phelps, New York

Roadcuts just northwest of Phelps expose Middle Devonian Onondaga Limestone (sandy at base) resting upon only a small portion of the Scajaquada Formation. Most of the Scajaquada Fm., the overlying Williamsville Waterlime and the Akron Dolostone have all been removed by an erosional event prior to the deposition of the Devonian units.

Both roadcuts reveal an almost entire sequence of the Fiddlers Green Formation, including the type Phelps Waterlime Member with a horizon of mudcracks at the top. Underlying units include the Oatka Shaly Dolostone, the Fort Hill Waterlime, and uppermost Camillus Formation (Salina Group). The Silurian-Devonian contact in this region varies considerably in position (See Figure 1).

Mud Creek at East Victor

Just south of NY96, along Mud Creek, is a small waterfall formed of resistant Morganville Waterline (basal Fiddlers Green Formation) with large scale conchoidal fracturing (conchoids) upon weathering. Upstream, the Victor Dolostone (type section) is intermittently exposed and bears a fauna of brachiopods, ostracods and eurypterids.

All higher units, including most or all of the Akron-Cobleskill, are well displayed much farther upstream. Both the *Eurypterus remipes remipes* Fauna and the *Eurypterus remipes lacustris* Fauna are well represented in these exposures

Buttermilk Falls north of LeRoy, New York

A large waterfall, capped by resistant cherty Onondaga Limestone, exposes the entire Fiddlers Green Formation and about 20 feet (7m) of underlying units (Camillus through Oatka Fms.).

The Scajaquada, Williamsville and Akron Formations have all been removed prior to deposition of the overlying Devonian limestones. Nearby, at the Genesee Country Museum Nature Trail, these units reappear beneath the Onondaga Limestone.

Indian Falls west of NY77

A magnificent waterfall, comprised of the very resistant Victor Dolostone (middle Fiddlers Green Fm.) occurs along the creek just west of NY77. These massive beds of Victor Dolostone (fine to medium-grained, finely crystalline dolostone with *Whitfieldella)* occur from the waterfall eastward to the bridge at NY77. Uppermost layers contain salthoppers.

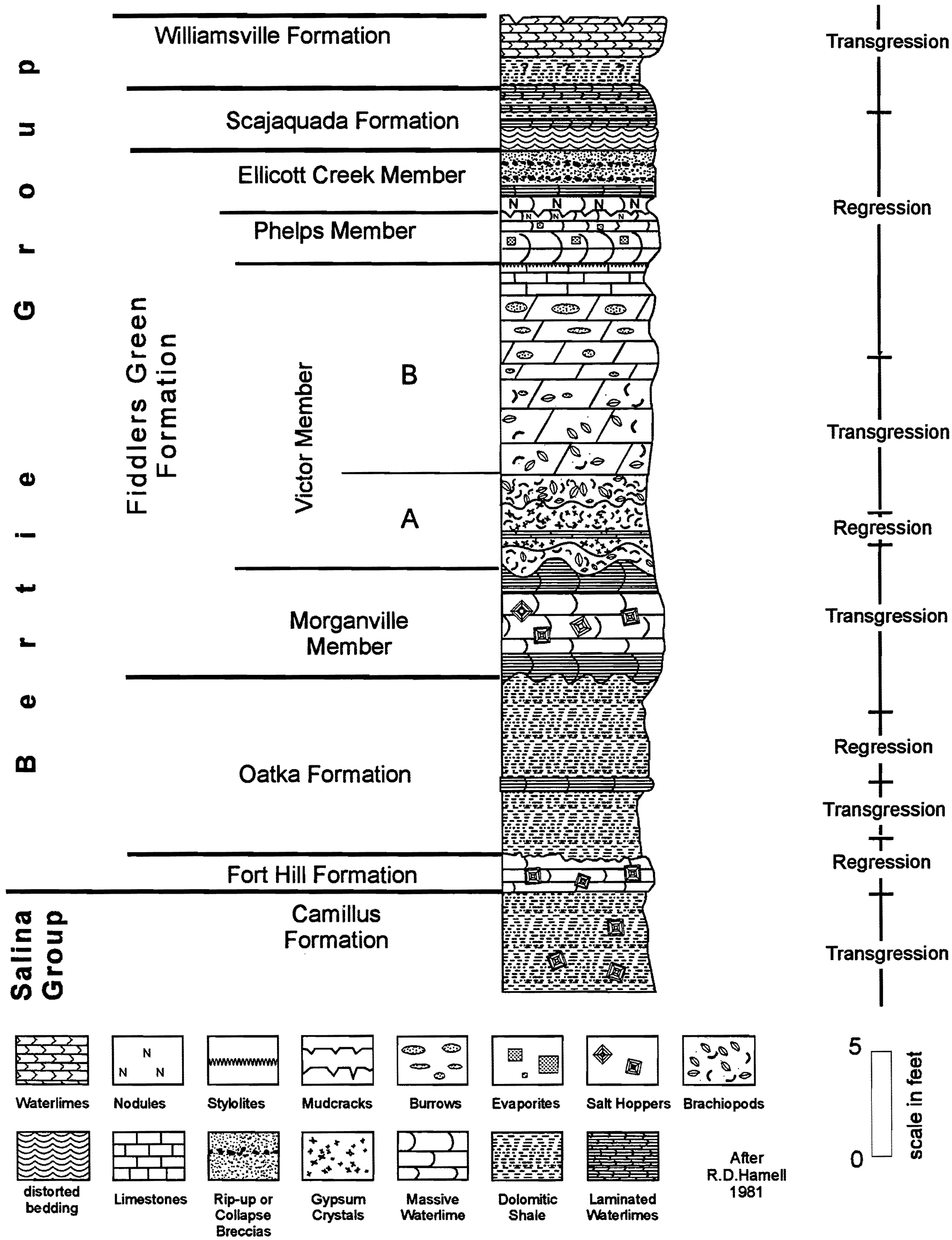

FIGURE 3 Stratigraphic column for part of the Late Silurian sequence in Western New York State. [from Hamell and Ciurca, 1986]

Thick beds of Morganville Waterlime occur in the face of the waterfall and are underlain by the Oatka Fm. followed by the Fort Hill Waterlime. The lowest unit exposed is the Camillus Formation consisting of shaly dolostone and dolomitic mudstone.

POST-SILURIAN SEDIMENTATION

It has been suggested previously (Ciurca, 1982, p. 115, Figure 6) that an unconformity separates Silurian Bertie Group sediments from the subsequent Devonian Manlius Group sediments. This view is reinterpreted in Figure 1 which shows the absence of the Moran Corner Waterlime and part of the Akron-Cobleskill sequence in the Grand River region of the Niagara Peninsula in Ontario.

Biostratigraphically, the hiatus is supported by the abrupt appearance of the genus *Erieopterus* which replaces the "*Eurypterus Beds*" scattered across New York state and the Niagara Peninsula of Ontario. *Eurypterus* is everywhere associated with faunal elements suggestive of a Silurian age (*Cystihalysites, Eccentricosta jerseyensis,* etc.).

The Manlius Group, in contrast, bears *Erieopterus* associated with *Howellella vanuxemi* and grades eastward into a sequence bearing a diverse fauna that is generally interpreted to be Early Devonian in age. Correlatives of the Manlius Group (Honeoye Falls Dolostone, Clanbrassil Formation) are lithologically distinct. The distribution of the Late Silurian and Early Devonian eurypterid-bearing waterlimes of western New York and southwestern Ontario has been illustrated previously (Ciurca, 1990; Figure 5).

SEDIMENTARY STRUCTURES

Many interesting sedimentary structures are preserved throughout the Salina and Bertie Groups. Here we concentrate on those of particular interest in reconstructing the paleoenvironments of the various subdivisions of the Bertie Group. A detailed stratigraphic column indicating the distribution of many kinds of structures is provided in Figure 3. The cyclic nature of the sequence shown is quite evident and has been discussed previously (Ciurca, 1973, 1978; Hamell and Ciurca, 1986).

Salt Crystal Structures

Almost ubiquitous throughout the Salina and Bertie Groups are casts and cavities formed by the dissolution of crystalline halite and other evaporites. Some of the largest relict halite crystals, including salt hoppers, occur within the upper Fiddlers Green Formation of the Bertie Group. Salt hoppers measuring 12 inches (30 cm) on a side are common, particularly along the outcrop belt from Phelps, New York westward to near Buffalo, New York. Though known from the days of the earliest New York geological surveys (see Figure 4), there has been little interest in their distribution geographically and stratigraphically within the evaporite

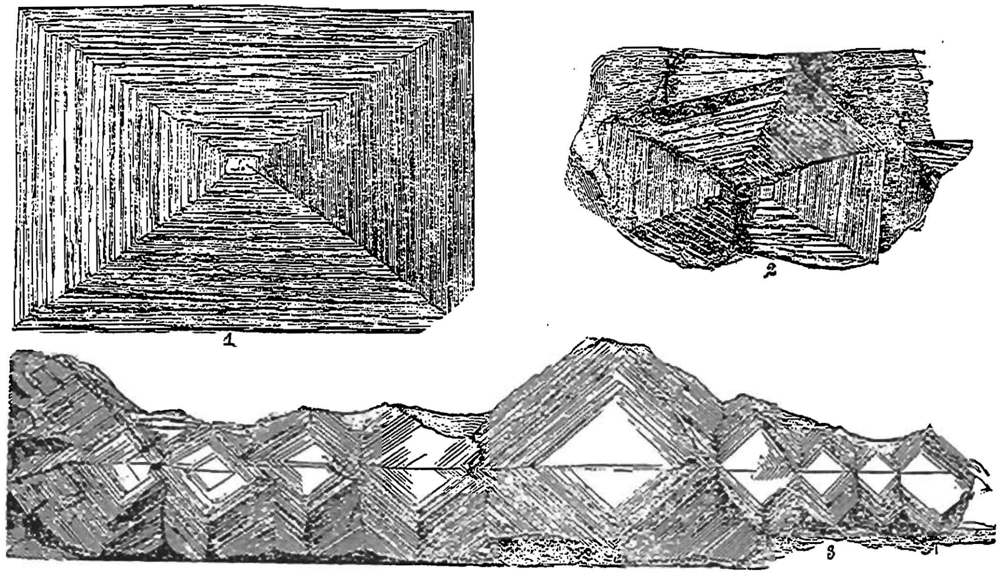

FIGURE 4 "Hopper-shaped crystals from the marl of the Onondaga salt group, town of Lenox, Madison County" (Hall, 1834)

sequence that constitutes much of the Salina and Bertie Groups. Ciurca (1990, p. D6) has recently suggested that hypersaline conditions were actually initiated in upper Lockport Group time within New York State. And just above the Lockport Group, in the basal Vernon Formation, salt hoppers become abundant in association with the eurypterid biofacies distributed in recurrent beds mostly in the Pittsford, New York vicinity (e.g. Barge Canal Bed)..

Figure 5A. *Eurypterus remipes remipes* (carapace) with "halo" of micritized brachiopods (*Whitfieldella*) preserved on a conchoid of upper Victor Dolostone (Fiddlers Green Fm.) (X1).

B. Slab of flat-pebble conglomeratic waterlime. Dark clasts (small arrows) are believed to be of "algal" origin, ripped away from nearby biostromes. Note salt hopper on clast marked H near bottom of figure. Eurypterid fragments (large arrow near top of figure) are present including this pretelson probably belonging to *Eurypterus laculatus*. Specimens: Ellicott Creek Breccia (Fiddlers Green Fm.), Ontario, Canada (X0.75).

C, D. Fossil plants. Figure D is *Cooksonia*, believed to be one of the earliest land plant, from the Williamsville Fm., i.e. the *Eurypterus remipes lacustris* Zone of western New York. Note preservation of terminal sporangia and the fragmentary nature of the specimens (C-X1.2, D-X1.25). Specimens: Ciurca Eurypterid Collection.

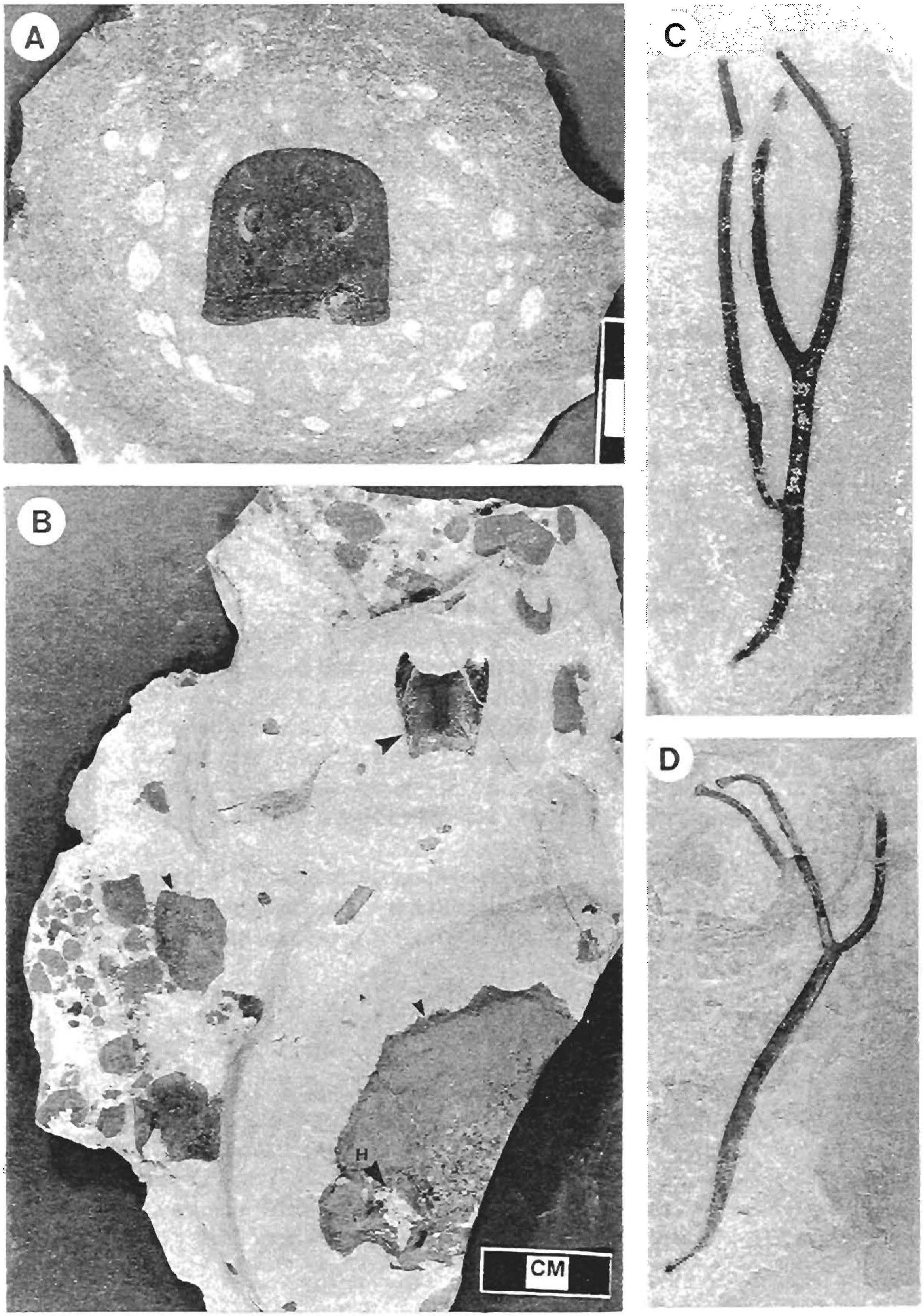

FIGURE 5 Some Fossils and Sedimentary Structures, Bertie Group.

Salt hoppers occur commonly on eurypterid parts, especially carapaces, in the Bertie Group rocks (Ellicott Creek Breccia in Ontario, Canada) and on *Waeringopterus* in the Salina Group rocks. Recently, salt hoppers have also been observed on what are interpreted to be ripped up clasts of algal mats (see Figure 5B).

One of the interesting aspects of the occurrence of salt hoppers throughout the series is replacement mineralization. Pseudomorphs after halite vary considerably. Commonly, dolomitic muds (waterlime) infill the halite crystals or their hopper faces, upon dissolution, and preserve the crystals as molds and casts. Within, particularly, the Ellicott Creek Breccia, the salt hoppers are delineated by red surfaces, presumably an alteration product (hematite or limonite) of iron sulfide. In strata not particularly exposed to weathering, pyrite has been observed to fill in the spaces, especially those along the hopper surfaces.

Crystalline (metallic) hematite replaces the halite within certain zones of the Vernon Formation (Salina Group). White, pearly dolomite replaces halite in the Fiddlers Green Fm. of the Auburn, New York area. And in the Syracuse Fm. at Camillus, New York, halite is replaced by quartz.

Several years ago, we had an opportunity to examine "outcrops" along the new "road" then being constructed at the Retsof Salt Mine (AKZO) from the main level (about 1180 feet beneath the surface) to a lower salt horizon some ninety feet below. In the argillaceous rocks exposed in the “roadcut”, presumably Vernon Fm. (Rickard, 1969), we were able to observe the halite "still in their hopper form". While much of the salt in the mine is of the granular type, clear masses of halite filled the hoppers. When such rocks are exposed at the surface, along the outcrop belt, the halite, of course, is quickly removed by water. Left behind are relict "crystals" or their cavities commonly infilled by various minerals as noted above.

Many of the stromatolites found at various horizons throughout the Salina-Bertie grew in hypersaline environments as indicated by their intimate association with relict halite. Indeed, stromatolites in the Syracuse Fm. (Fayetteville, N.Y.) grew atop salt hoppers, the hoppers having been later replaced by selenite.

Laminations

Some of the waterlimes present very straticulate bedding. The fine laminae are varvelike, and are probably seasonal and intimately associated with "algal" growth in nearby stromatolite tracts. We must be careful, however, in ascribing all finely laminated carbonate rocks to "cryptalgal" activity. In the cases observed (Ciurca, 1982, p. 103) the laminated sediments are generally associated with stromatolite mounds and are "inter-reefal" in nature. They are areally limited due to rapid facies changes. Examples include the upper Victor Member, Fiddlers

Green Fm., exemplified in Bertie Township, Canada. These beds are finely laminated and contain abundant *Eurypterus laculatus* and *Eurypterus remipes remipes* associated, essentially always, with *Whitfieldella* and abundant ostracods ("*Leperditia*"). The laminae grade laterally into well-developed shell beds, the beds becoming coarse-grained dolostone with irregular black, presumably carbonaceous, bedding planes. Figure 5A shows a carapace of *Eurypterus remipes remipes* surrounded by a "halo" of micritized brachiopods (*Whitfieldella*) and is from the upper beds of the Victor Dolostone.

The Ellicott Creek Breccia, especially in its upper portions ("topographic waterlime" beds) is also unusually distinctively straticulate. Close examination, however, showed an intimate association with occasional stromatolitic mounds and, again, eurypterids are found in these intermound strata.

Part of the Williamsville Fm. (Bertie Gp.), with its characteristic *Eurypterus remipes lacustris* Fauna, is also finely straticulate, but no "algal" mounds have yet been identified. On the Niagara Peninsula, a ripplemarked surface at the contact of Williamsville A and B Submembers is interpreted as indicating deposition in very shallow water. Just to the east, at Ellicott Creek and Akron Falls, New York, mudcracks occur in the sequence (Ciurca, 1982, p. 107) and indicate that portions of this waterlime, as in the case of many other cyclically formed waterlimes, were deposited in shallow basins susceptible to subaerial exposure and erosion.

In eurypterid-bearing sequences, laminated strata are typically present, and an excellent example is provided by the well-known Kokomo Formation of Indiana and its equivalent on the Bruce Peninsula of Ontario, Canada ("Eramosa Fm." of authors). These occurrences are all illustrative of a general setting of restrictive marine environments, usually with accompanying preservation of relict evaporite structures and other indicators of shallow-water deposition of the dolomitic lime muds as discussed in previous sections.

Ripplemarks

Ripplemarks appear to be a rarely preserved structure within the Salina-Bertie rocks. There are important horizons in the Victor Dolostone near Jerusalem Hill and in the massive Victor dolostone as displayed in Black Creek, Morganville, New York. At the latter site, the Devonian Onondaga Group rests unconformably upon the Victor Dolostone, all overlying Bertie Group units having been eroded from the site before Middle Devonian sedimentation commenced.

Recent studies at sites within the type area of the Bertie in Ontario, Canada (Ciurca, in progress) have shown the presence of ripplemarks within the Williamsville Waterlime at or just above the top of Williamsville A Submember, the important eurypterid-bearing unit of the Williamsville Formation in the region.

The position of known ripplemark horizons, in association with many other sedimentary structures present in these rocks, especially well-developed mudcracks (suncracks of many authors), establishes the shallow-water nature of most of these sediments (see the following section on paleoenvironmental settings).

The extensive fossil windrows, i.e. widespread current-generated accumulations of plants and animals (see Ciurca, 1978, p. 226), are a peculiar type of ripplemark intimately associated with the eurypterid-bearing waterlimes of the Salina-Bertie sequence.

In addition to the "algal" bioclasts, dislodged and transported into the accumulating lime muds of the quieter interbiostromal regions, and represented by the recurrent waterlime beds of the Bertie Group sequence, a few typically marine forms were introduced. They are generally found in clusters (e.g. small high-spiraled gastropods) or in linear clusters, i.e. windrows (e.g. various cephalopods including *Hexameroceras*). Also obviously introduced were fragments of plants including algae of various kinds, and some of the early "land plants". Recently discovered plants (Ciurca, in preparation), found in the Williamsville Waterlime, and illustrated in Figure 5C and Figure 5D, show the fragmentary nature of the specimens preserved. *Cooksonia*, generally regarded as an early "land plant" bears terminal sporangia (Figure 5D).

Chert Nodules

Small, very round chert nodules are characteristic of the Scajaquada Fm. (Bertie Gp.) from at least Victor, N.Y., westward into Bertie Township on the Niagara Peninsula, Ontario, Canada. Within this formation, and in transitional beds of the Ellicott Creek Breccia, are horizons of an oolitic texture, perhaps silicified oolites (i.e. chert; Hamell, 1981). Petrographic evidence for this type of replacement (gypsum-anhydrite) is the length-slow nature of the chalcedony (Friedman, 1964; West, 1964, 1973; Folk and Pittman; 1971). In the Bertie Gp. occurrences, sphalerite is sometimes associated with the chert replacement.

Evaporite structures (halite) occur in the Scajaquada Fm. at Phelps, N.Y. and the gypsum beds of the Forge Hollow Fm. appear just to the east (Cayuga Lake, Auburn areas). These observations support the interpretation above, that the chert is likely pseudomorphic after gypsum, being intimately associated with an evaporite sequence that divides the Bertie Group (redefined) into two parts of about equal thickness.

Chert beds constitute much of the Cobleskill Fm. in the Marcellus Falls, New York area, but these beds are not believed to be associated with gypsum deposition. They occur, rather, in a more normal marine sequence adjacent to possible Cobleskill "reefs" of corals and stromatoporoids (somewhat analogous to sections of the Lockport Group).

Soft Sediment Deformation And Other Structures

Cross-bedding is generally so fine that it escapes notice. However, many of the eurypterid-bearing waterlimes, especially those that exhibit straticulate bedding, show evidence of cross-bedding, scouring, slumping, reworking of sediments including rip-up clasts and redistribution of sediment. Included within the sedimentation processes are the eurypterid remains which were wrinkled and otherwise distorted as they were transported to the accumulating carbonate muds.

Extensive windrows of oriented fossils in the waterlimes are due to currents that accumulated the hard parts (i.e. exoskeletons) and disarticulated structures into arrays of clustered debris (see Ciurca, 1978, Figure 1, p. 226). Microfaulting is commonly observed (e.g. Phelps Waterlime at Sphon Hill sites east of Cedarville, New York) where dolomitic muds were disturbed by current activity and subjected to subaerial exposure. Sometimes the integument of eurypterid parts are offset by microfaulting. During subaerial exposure, disarticulated eurypterid parts were washed into opened cracks before being sealed by lime muds (Ciurca, in progress).

Brecciation is particularly well-developed in the Ellicott Creek Breccia and was generated by rip-up currents associated with stromatolite "reefs" and their intermound channels. The laterally extensive zone of waterlime breccia (over 100 miles) may represent a seismite deposit. While strong currents undoubtedly ripped-up clasts, both dolomitic mud and "algal mats", from the "algal reef" tract, it may have been a particularly extensive phenomena (e.g. earthquake, tsunami) that dislodged and carried lithoclasts and bioclasts shoreward to form part of the Ellicott Creek Breccia. No other waterlime unit exhibits such a degree of brecciation. As previously noted (Ciurca, 1982, p. 103), the Ellicott Creek Breccia is a tripartite unit with stromatolite mounds particularly important in the middle and brecciated waterlimes particularly evident in upper and lower portions, when present.

Stromatolites

Stromatolitic and thrombolitic structures are among the most important elements distributed through the eurypterid-bearing sequences in New York and these structures continue into the Niagara Peninsula of Ontario (see discussion in Ciurca, 1990). These structures, forming extensive biostromes and bioherms, were initiated upon deposition of much of the Lockport Group (Zenger, 1965; Ciurca and Domagala, 1988).

Within the Salina Group, stromatolite distribution has been little studied. Stromatolites are important within certain zones, most notably the *Waeringopterus* Zone near the base of the Syracuse Formation (Ciurca, 1990, p. 82, p. 111). Within this zone, the stromatolites extend from at least Syracuse westward to the Welland Canal in Ontario, Canada. At the latter site, a bed of gypsum occurs just below the horizon. At Garbutt, New York, stromatolites occur along Oatka Creek in upper beds of the Syracuse Formation. Little is known, however, of the associated fauna in this area.

The recent discovery of *Eurypterus pittsfordensis* near the Niagara Gorge (Ciurca, 1993) was made in a sequence of large-scale stromatolitic mounds in the upper Lockport Group (see Zenger, 1965, p.114, Figure 22). The thin-bedded micritic beds found adjacent to the stromatolites are similar in many ways to the eurypterid-bearing waterlimes of the Salina and Bertie Groups. Ciurca (1990, p. D6) noted distinct stromatolite mounds in a transitional interval at the North Chili site where *Eurypterus pittsfordensis* is particularly abundant in a "*Chondrites* facies." See Figure 6.

Stromatolites are also widely distributed within the Bertie Group from the Sphon Hill area near Cedarville, New York westward throughout the state and into the Niagara Peninsula of Ontario. Particularly noteworthy are the occurrences at the Neid Road Quarry (north of LeRoy, NY) where eurypterids are found in the waterlimes surrounding the stromatolites (presumably Ellicott Creek Breccia). The same observations have been made at the Ridgemount Quarry near Fort Erie, Ontario, Canada within the Ellicott Creek Breccia (uppermost Fiddlers Green Fm.). The *Eurypterus remipes remipes* Fauna is abundantly represented at this site and the remains are mostly concentrated in the dolomicrite (waterlimes) that underlie, overly or surround the stromatolites.

Thrombolites (see Cys and Mazzullo, 1978) are peculiar "algal" mounds, resembling outwardly the shapes of many types of stromatolites, but having a clotted structure rather than a laminated appearance. Impressive thrombolitic biostromes also occur within the Victor Dolostone at the Ridgemount Quarry Site (see Figure 7) and again well-preserved eurypterid remains are often encountered in the interbeds, in this case the *Whitfieldella* beds which dominate much of the Victor Member throughout it extent. One of the most characteristic eurypterids of the Victor Dolostone is *Eurypterus laculatus* (Figure 8). Note although widely misidentified, particularly by Copeland and Bolton (1985), *Eurypterus laculatus* is part of a group termed the "Ontario eurypterids" by Ciurca (1990) because of their widespread occurrence in a belt paralleling Lake Ontario and that continues into the province of Ontario, Canada. *E. laculatus* is intimately associated with the thrombolites in Ontario and is also unusually abundant at Morganville, New York (Victor Dolostone) and also occurs in the Neid Road Quarry associated with stromatolites and other eurypterids. It is also notably a common form in the Victor and Ellicott Creek Members of the Fiddlers Green Formation of southwestern Ontario, Canada (Ciurca, in preparation)

An example of Devonian eurypterid/stromatolite association was given for the Thacher Limestone at Clockville previously (Ciurca, 1978, p. 24). The eurypterid is *Erieopterus* and the stromatolite horizon was illustrated by Rickard (1962, p. 5, Figure 12). *Erieopterus* was also shown "swimming" in a *Howellella* maze among thrombolites (Chrysler Formation, H Member). See Ciurca, 1978, p. D24, Figure 6.

Recent examples of stromatolitic structures, and their modification of the local environment, are rare. For an unusually impressive seascape showing rippled sands flowing between stromatolite mounds in the shallow waters of the eastern Bahamas. See the article by Gore (1989, p. 674-675).

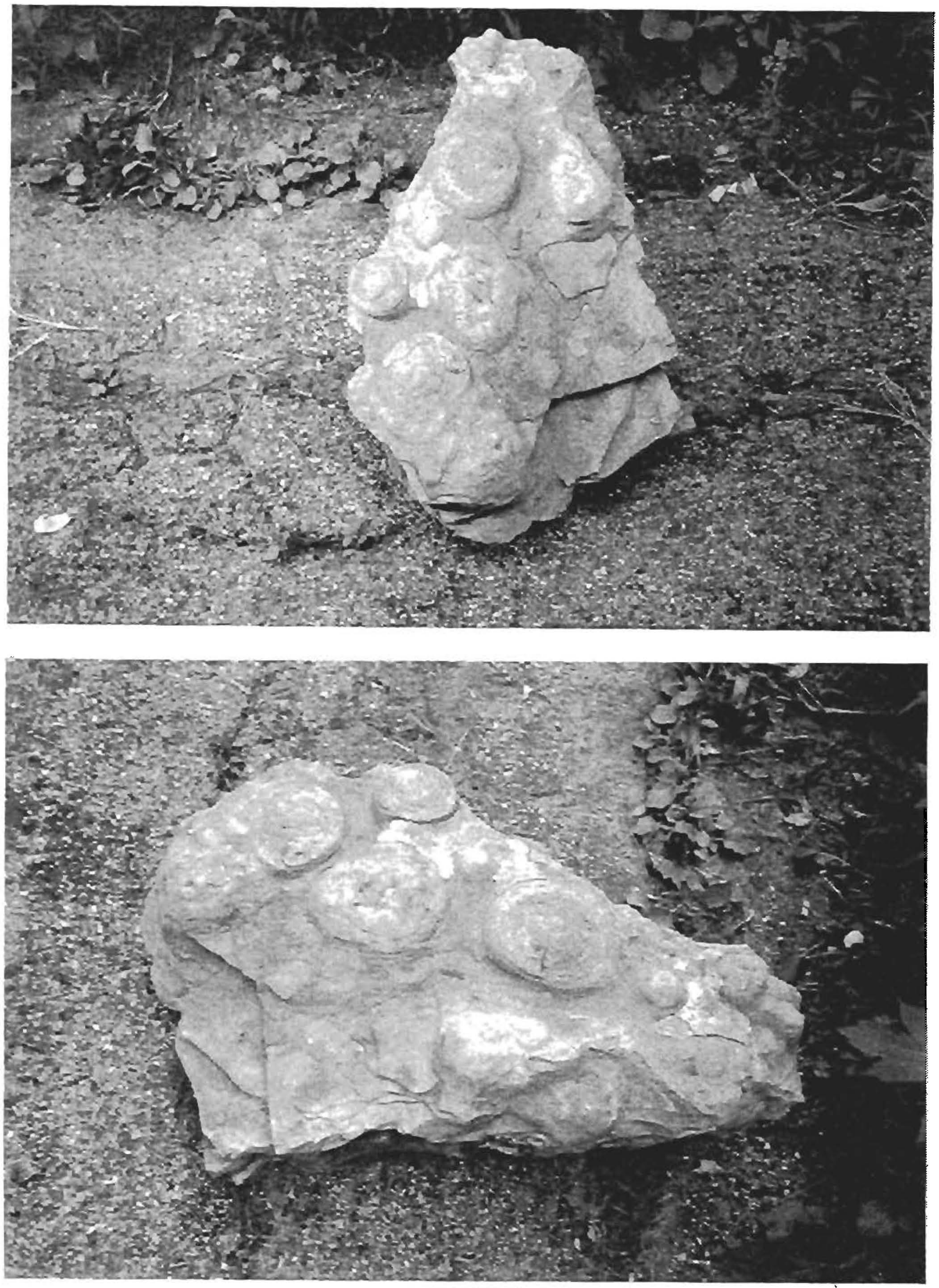

FIGURE 6 Massive stromatolite complex presumably constructed by cyanobacteria ("algal" mound) recovered from ~20 feet below the surface in transitional Lockport/Salina Groups. Irregular top portion was covered by black shaly partings. Sequence is within *Eurypterus pittsfordensis* Zone. NY 33 at North Chili, New York. Stromatolite block is 64 cm along the long axis. Specimen from the Ciurca Eurypterid Collection.

FIGURE 7 Large thrombolite in a block of Victor Dolostone, Fiddlers Green Formation. Note irregular contact of the mound with the overlying waterlime (lighter color) "fused" to its surface. The waterlimes contain the eurypterid fauna. Interthrombolite sediments contain abundant brachiopods, e.g. *Whitfieldella* sp., Ridgemount Quarry, Bertie Township, Ontario, Canada.

FIGURE 8 *Eurypterus laculatus*. Specimen is one inch in length and is from the Ellicott Creek Breccia, Fiddlers Green Formation, Bertie Group, Ridgemount Quarry, Ontario, Canada.

FIGURE 9 Examples of "Boomerangs", peculiar sedimentary structures recently exposed on quarry floor at contact of Williamsville A Submember with Williamsville B Submember, Bertie Group, Ridgemount Quarry, Ontario, Canada. Pointer in photo is 0.5 meters and points north.

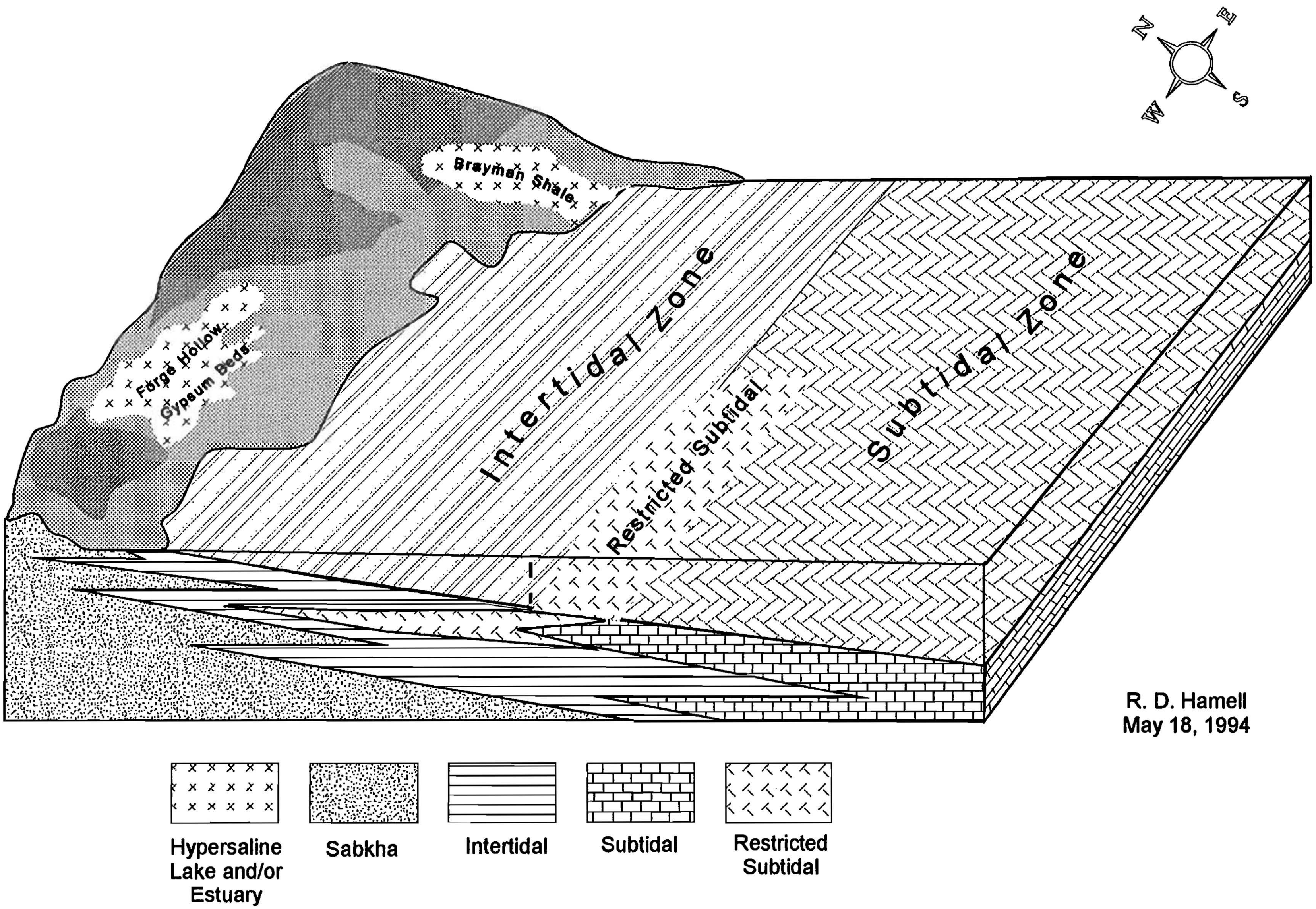

FIGURE 10 Schematic reconstruction of the paleoenvironmental setting during the deposition of the Bertie Group lithologies. [After Hamell, 1981]

Trace Fossils

Trace fossils are relatively rare in the waterlimes of the Bertie Group. Some very peculiar and large structures, about the size and shape of a boomerang, were observed by Ciurca (and recently witnessed by C. Brett) in the Spring 1994. The "boomerangs" were observed on the quarry floor at the Ridgemount Site and occur at or just above the contact of Williamsville A and B Submembers. Samples (or casts) of these unusual structures are currently under study. These structures (Figure 9) may actually prove to be due to some kind of extraordinary current activity.

Exceedingly abundant trace fossils occur within the Victor Dolostone and the Akron-Cobleskill. Undoubtedly the mottling, so characteristic of both units, is due to extensive bioturbation and this in turn appears to have controlled dolomitization of these units. Bedding planes are profusely covered by rodlike structures presumably made by unknown burrowers. Bedding planes in these units are generally irregular and covered by black (carbonaceous) shaly material. Only ostracods and a limited variety of brachiopods are present. Note: Eastward (and southward) the Akron-Cobleskill becomes progressively more fossiliferous. Biostromes with stromatoporoids and corals are encountered, and the brachiopods become more diversified. For a description of Cobleskill Formation stromatoporoids, see Stock (1979); and for brachiopods, see Berdan (1972).

PALEOENVIRONMENTAL SETTINGS

Eight different paleoenvironments are recognized in the Bertie Group (Hamell, 1985). From onshore to offshore they are: 1) sabkha, 2) hypersaline lake, 3) lower supratidal, 4) upper intertidal to lower intertidal, 5) lower intertidal, 6) restricted subtidal, 7) non-restricted subtidal, and 8) semi-restricted lagoon-estuary. A schematic reconstruction of the paleoenvironmental settings during deposition of the Bertie Group is shown in Figure 10. Further details can be found in Hamell and Ciurca, 1986).

Sabkha

Sabkhal sedimentation is characterized by relatively barren dolomitic muds. Carbonate based shells (i.e. brachiopod-*Whitfieldella*) were leached out during periodic flooding. Evaporites are common such as gypsum-anhydrite nodules and halite (casts). This depositional setting is reflected in the sediments of the Camillus Formation (Salina Group) and the Oatka and Scajaquada Formations of the Bertie Group.

Hypersaline Lakes - Estuaries

The gypsum deposits of the Forge Hollow Formation in central New York are considered to be analogous to the present day sedimentation of gypsum reported by Lucia (1968) as occurring in shallow hypersaline lakes on Northern Antilles island of Bonaire. These restricted sabkhal-hypersaline lake deposits are separated from the sea by a barrier composed of permeable sabkhal sediments. The permeability of these sediments permits the influx of sea water to the hypersaline lakes so as to replace water lost to evaporation. Seasonal changes in recharge and evaporation rates result in the laminated bedding of these barren and thick gypsum beds.

Supratidal

Sediments in lower sabkhal to upper supratidal environments are indicated by collapse and rip-up breccias of the Ellicott Creek Member of the Fiddlers Green Formation. Deposition (rip-up breccias) is the result of high spring tides and major storm surges. In the interim, during prolonged subaerial exposure, evaporites form. Periodic influx of water in the sediment, from precipitation and/or flooding, dissolves the evaporites causing the subsequent collapse breccias.

Intertidal

Discussion of the Brayman Shale of eastern New York is important in presenting a complete interpretation of Bertie environmental facies. Past studies by Belak (1980) and Treesh (1972) have inferred that the Brayman Shale and the Phelps and Williamsville Waterlimes were deposited in an intertidal environment. Fisher and Rickard (1953) noted that the dissimilarity of the Brayman and the Forge Hollow Fms. was most likely the result of slight facies changes. The high content of pyrite and black shales in the Brayman suggests a restricted estuarine environment where sulfate-reducing bacteria inhibited the accumulation of gypsum which is found in the Forge Hollow. Morris and Dickey (1957) described a similar depositional environment in Peru.

Intertidal deposition of waterlimes is represented by portions of the Fort Hill Formation, Morganville and Phelps Members of the Fiddlers Green Formation and the Williamsville Formation. Lowermost sabkhal to uppermost intertidal zones are characterized by mudcracks and laminated bedding. Subaerial exposure and evaporation resulted in the generation of hypersaline water in these sediments and subsequent deposition of salt hoppers, reticulate halite structures. Fossils are sparse, cryptalgal structures and an eurypterid fauna being the most common. Ghost structures of cephalopods and gastropods have also been found. Their poor preservation is the product of leaching (dissolving) of calcium carbonate based shells, whereas chitinous exoskeletons of eurypterids and phyllocarids are relatively unaffected. Such selectivity of preservation has been documented in modern sediments of intertidal marsh environments.

Subtidal

The Victor A Submember is a lithographic limestone that contains the small brachiopod *Whitfieldella*. The central portion of this unit contains a one-foot thick layer of tightly packed gypsum crystals. The upper 2.5 feet of the Victor B Submember has a similar deposit of gypsum. These units represent conditions of hypersalinity in a restricted upper subtidal environment.

The rock units of the lower Victor B Submember range from having an undulating (wavy) laminated bedding to mottled and thick-bedded dolostones. When freshly broken the rocks emit a strong petroliferous odor. The mottling of this facies is due to bioturbation and is typical of modern shallow intertidal environments (unrestricted). Complete specimens of the brachiopod *Whitfieldella* and the common ostracod *Leperditia* are a common constituent of this zone.

The Victor B Submember, a dark, slightly argillaceous dolostone with brachiopods horizontal burrows, indicates a deeper subtidal facies. Although this environment represents a more seaward setting, the small size of *Whitfiedella,* along with a low faunal diversity and absence of typical marine organisms, indicate a depositional environment under hypersaline conditions.

SUMMARY

The Bertie Group is a carbonate sequence that accumulated during multiple oscillations (e.g. sabkhal-intertidal-sabkhal; Camillus-Fort Hill-Oatka Formations, see Figure 3) of the Late Cayugan Sea. Numerous waterlimes, noted for the eurypterid assemblages they contain, occur cyclically in this sequence and are associated with numerous structures indicating a shallow-water origin for much of the Bertie Group.

The Fiddlers Green Formation records the major transgressive-regressive cycle within the lower Bertie Group. Farthest offshore facies are represented by the Victor B Submember, a fine-grained fossiliferous limestone that contains well-developed horizontal burrows and mottling due to extensive bioturbation.

The waterlime breccias of the Ellicott Creek Member indicate supratidal conditions. Intraclasts formed from lime muds and ripped-up algal mats characterize the unit from the Niagara Peninsula, Ontario, Canada, eastward to at least Phelps, New York. Supratidal to intertidal sedimentation is displayed by the Morganville and Phelps Waterlime Members, as well as other waterlime units. Key indicators include mudcracks, cryptalgal structures, and laminated sediments.

Subtidal deposition, under conditions approaching near-normal salinity, are recorded in the thickest and most fossiliferous member of the Fiddlers Green Formation, i.e. the Victor Dolostone, and also the Akron-Cobleskill of the region. Restricted subtidal deposition of bladed gypsum crystals is represented in the Victor B Submember.

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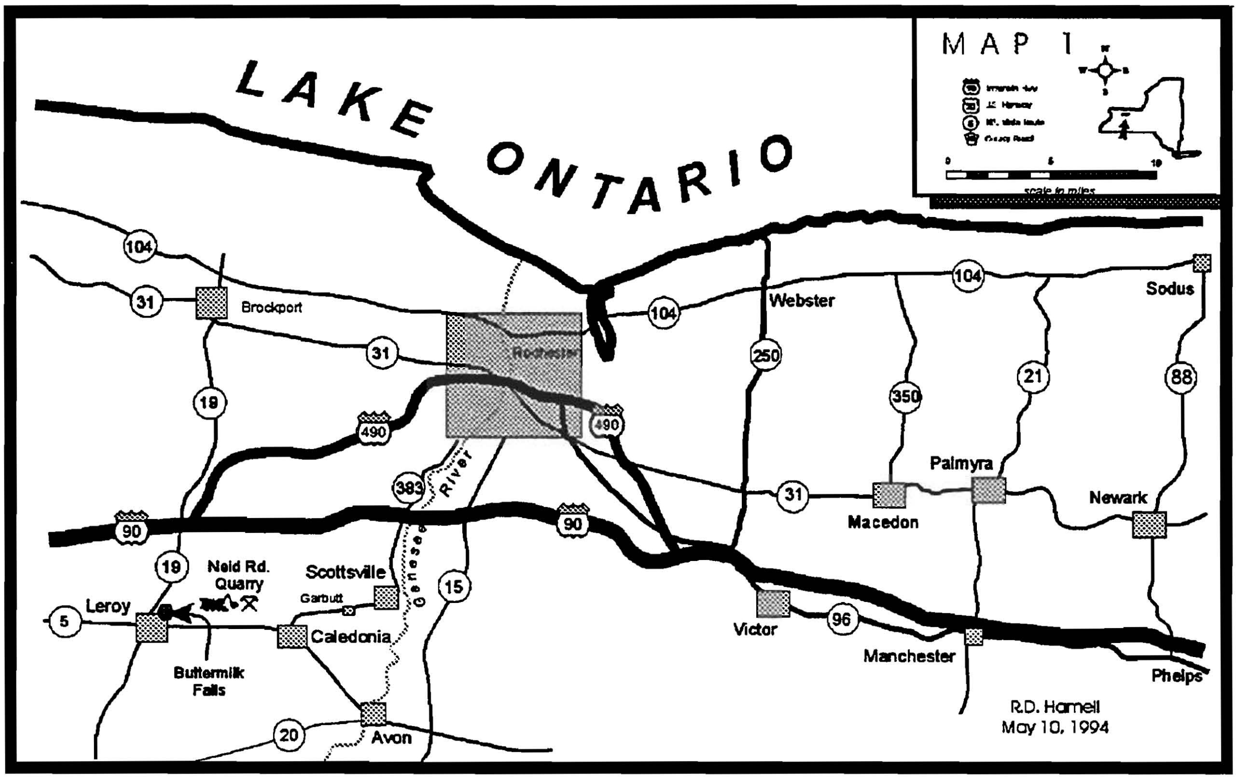


FIGURE 11 Field Trip Map. Follow NY383 South to Garbutt, formerly the site of a gypsum mining industry.

ROADLOG

CUMM. MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
		Depart University of Rochester Campus at Wilson Blvd.
0.00	0.00	Roadlog begins at the intersection of Wilson Blvd. and Elmwood Ave. TURN RIGHT.
0.40	0.40	Genesee River on right. BEAR LEFT. Go south on NY383.
2.60	2.20	Jct. 252A. Continue south on NY383.
3.90	1.70	Jct. NY252 (Jefferson Road).
8.50	4.60	NYS Thruway (I-90) overpass.
9.80	1.30	Village of Scottsville. Continue south on NY383.

11.00	1.20	Jct. NY386.
12.80	0.10	TURN LEFT on Union Street. Cross railroad tracks. Turn RIGHT into the parking lot of Oatka Creek Park.

STOP 1 - OATKA CREEK AT GARBUTT

Excellent exposures of the Syracuse Fm. (Salina Gp.) occur along Oatka Creek both upstream and downstream of the Union Street Bridge over the creek. Over 50 feet of strata have been measured along the south side of the creek and the entire section appears to represent the Upper Syracuse Formation. Included are argillaceous, thin-bedded (flaggy) dolostones, massive limestones and dolostones (particularly along the banks of the creek), thin waterlime units and evaporites. The site was extensively mined for gypsum and the ruins of some of the buildings that were constructed are still visible on the north side of the creek just upstream from the bridge. Note the buildings were constructed with glacial erratics and slabs of Syracuse Formation. An account of the gypsum mining activity here, and in nearby areas, can be found in Newland (1929).

Very few fossils have been described from the Syracuse Fm. of the region. The eurypterid *Waeringopterus* was traced into western New York but has not been found along the Oatka Creek Valley (Ciurca, 1990, p. D10). *Eurypterus* has been found in a waterlime bed immediately upstream from the bridge. The exposures in the wall on the south side of Oatka Creek have yielded cephalopods and *Medusaegraptus*. An algal stromatolite horizon occurs near the top of the section.

13.00	0.10	Leave parking lot. TURN LEFT (north) on Union. Proceed (back) to NY383.
13.20	0.20	Jct. NY383. TURN LEFT.
17.10	3.90	Jct. NY 36 TURN LEFT (south) to Mumford.
17.50	0.40	Flint Hill Road. TURN RIGHT.
18.90	1.40	Genesee Country Museum on left.
20.20	1.30	Entering Genesee County.
21.10	0.90	Jct. Neid Road. TURN RIGHT.
21.30	0.20	Park along roadside adjacent to large boulders.

STOP 2 - VICTOR DOLOSTONE ERRATICS

Massive Victor Dolostone (middle member of the Fiddlers Green Fm.) can be observed in the waterfall at nearby Buttermilk Falls. The large erratics at the driveway exhibit the finely crystalline character typical of the Victor Dolostone. Bedding planes are often covered with the small brachiopod, *Whitfieldella.*

		Continue north on Neid Road.
21.50	0.20	Jct. road to the Town of LeRoy Refuse Area. TURN LEFT.

STOP 3 - THE NEID ROAD QUARRY

This quarry exposes a fine section of the Late Silurian Bertie Group overlain by massive beds of the cherty Onondaga Group. The erosional Silurian-Devonian unconformity is well-displayed here. The quarry floor is comprised of the upper waterlimes of the Fiddlers Green Fm. and includes an eurypterid fauna consisting of *Eurypterus* and *Dolichopteris*. The irregularity of portions of the floor is due to algal stromatolite mounds. Salt hoppers and associated relict halite structures are common at this level.

Brecciated waterlimes (Ellicott Creek Breccia) are present just above the quarry floor, particularly at the north wall of the quarry. Immediately overlying this is 5-8 feet of the Scajaquada Formation, an argillaceous, mostly thin-bedded dolostone containing several chert nodule horizons.

The Williamsville Formation, here a particularly drab gray to cream colored dolostone, is exposed along the north wall of the quarry. Fossils are exceedingly rare in this unit. Only phyllocarid telsal spines and a few articulate brachiopods have been found here.

The Akron-Cobleskill is nearly missing from the section. At nearby Buttermilk Falls, the Scajaquada, Williamsville and Akron Fms. are all missing, the Onondaga Limestone rests directly upon the Fiddlers Green Formation.

		Return to Neid Road Jct. with Flint Hill Road. TURN RIGHT.
22.30	1..50	Park along the roadside.

STOP 4 - STEAM LOCOMOTIVE AND SHOVEL

Turn-of-the-century (1902) steam shovel used quarrying on right.

23.20	0.90	Jct. of Circular Hill Road. TURN RIGHT
25.00	1.80	Jct. with Oatka Trail. TURN LEFT.

25.80	0.80	Jct. Parmelee Road. TURN LEFT.
26.10	0.30	Jct. of NY19. TURN LEFT (south). Park along shoulder.

STOP 5 - FORT HILL WATERLIME

Uppermost Camillus Fm. is well-displayed in the roadcuts along both sides of NY19. At the top is a thin waterlime unit, the Fort Hill Waterlime. The overlying Oatka Formation is concealed by overburden. The Fort Hill Waterlime is considered the base of the Bertie Group and bears an *Eurypterus* Fauna (Ciurca, 1973, p. D3). It is also accessible near the base of nearby Buttermilk Falls.

		Continue south on NY19.
26.70	0.60	Railroad overpass (abandoned)
26.80	0.10	Jct. North Street. TURN LEFT.
27.00	0.20	Park on right shoulder before the bridge over Oatka Creek.

STOP 6 - BUTTERMILK FALLS

Follow the abandoned railroad east to Oatka Creek. A trail to the base of Buttermilk Falls is located on the west side of the valley, north of the railroad bridge over the creek. **Note: walls of the gorge are steep**. The safest trail to the bottom enters the gorge about 1/4 mile north of the falls.

Buttermilk Falls is capped by resistant Onondaga cherty limestone. The underlying Bertie Group forms the reentrant at the unconformable contact with the Onondaga Group.

At creek level is uppermost Camillus Formation (Salina Group). The Bertie Group here starts with the Fort Hill Waterlime, a complete section of the Oatka Formation, and most of the Fiddlers Green Formation. The Scajaquada, Williamsville and Akron Formations are missing due to erosion prior to the deposition of the Middle Devonian Onondaga Limestone.

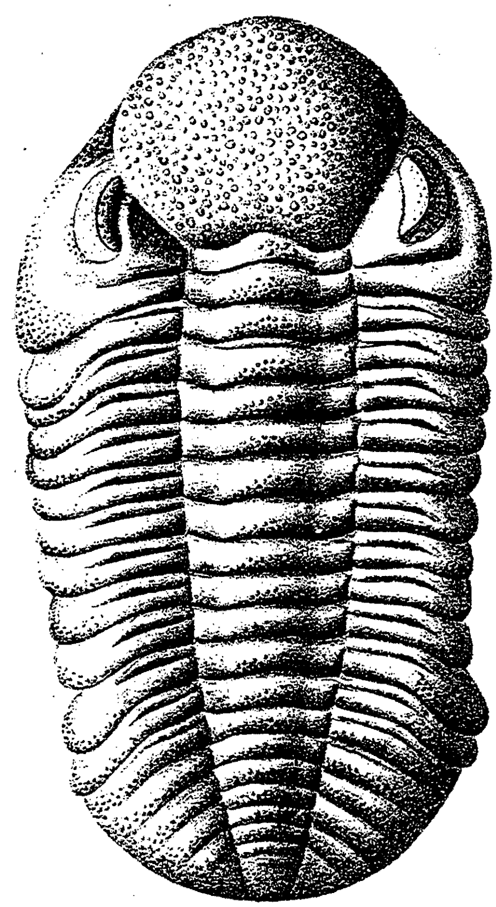
		Proceed on North Street into the Village of LeRoy.
29.00	2.00	Jct. with NY5. TURN RIGHT (west).
		Village of LeRoy, renowned for the creation of **J-E-L-L-O**
29.10	0.10	Bridge over Oatka Creek

STOP 7 - McDONALD'S (on left)

Note: To return to Rochester take NY19 (just west of McDonald's) north to the NYS Thruway (I90) or to I490.

END OF ROADLOG

Late Silurian Estuarine Landscape, Western New York. [Pen and ink drawing by Kay O'Connell, 1980 under the direction of R. Hamell]



Phacops rana

[From Hall, 1888, Natural History of New York: Palaeontology, Vol. VII, Plate I, Figure 10]

FIELD STUDIES OF THE MIDDLE DEVONIAN LUDLOWVILLE-MOSCOW SEQUENCE IN THE GENESEE AND SENECA VALLEYS, NEW YORK STATE

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INTRODUCTION

Numerous authors have studied and described in detail the unique depositional environments and the excellent fossil preservation found in the Middle Devonian Hamilton Group sediments of New York State (Grabau, 1899; Cooper, 1930; Grasso, 1973; Brett and others, 1986; Savarese and others, 1986). The advancement of high resolution stratigraphic mapping has permitted detailed correlations of facies divisions in the uppermost Ludlowville Formation across western and central New York (Mayer, 1989; Mayer and others, 1994). The focus of this field trip is to examine the spatial and temporal sedimentary relationships as well as the abundant and well preserved fossils in the Upper Ludlowville-Lower Moscow Formations of the Hamilton Group; in particular, the uppermost Wanakah Shale, Jaycox Shale and Tichenor Limestone Members in the Genesee and Seneca Valleys (Figure 1).

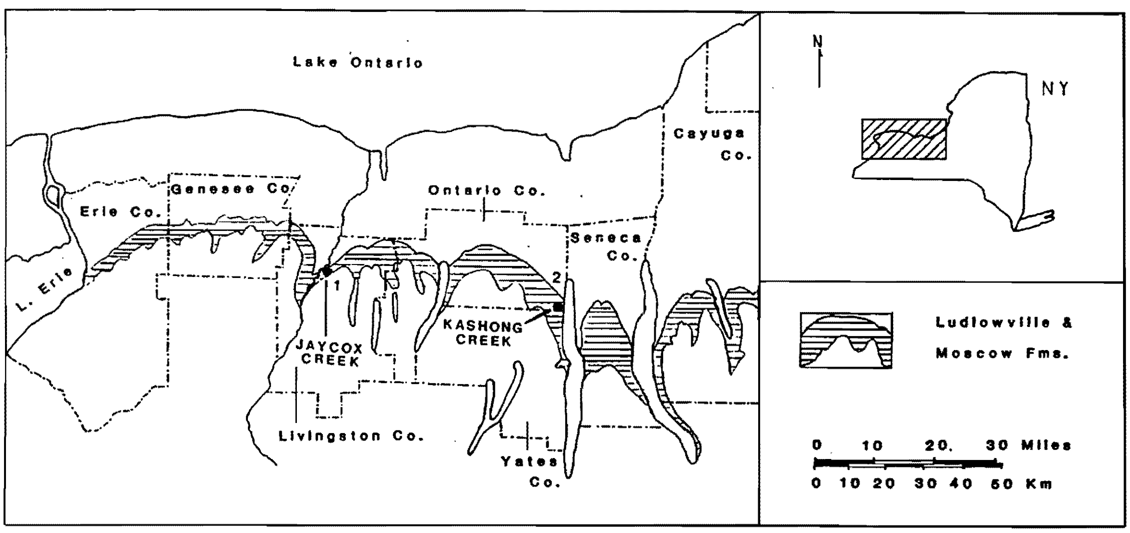

Figure 1. Key Ludlowville - Moscow outcrops in the Genesee Valley and Seneca Valley region. Modified from Baird (1979).

GEOLOGIC SETTING

The sediments of the Hamilton Group were deposited as part of the prograding Catskill Delta Complex at the northern margin of the Appalachian Basin in North America. Reconstruction of North American paleolatitudes, based on paleomagnetic studies, have placed the continental landmass across the paleoequator with the Hamilton deposition occurring at approximately 0-30^{o} S latitude (Oliver, 1976; Ettensohn, 1985; Van der Voo, 1988). The sediments comprising the Ludlowville-Moscow Formations accumulated across a shallow epeiric sea. The seafloor extended across a western and eastern shelf separated by a central subsiding basinal trough, centered around the present-day Seneca Valley (Figure 2).

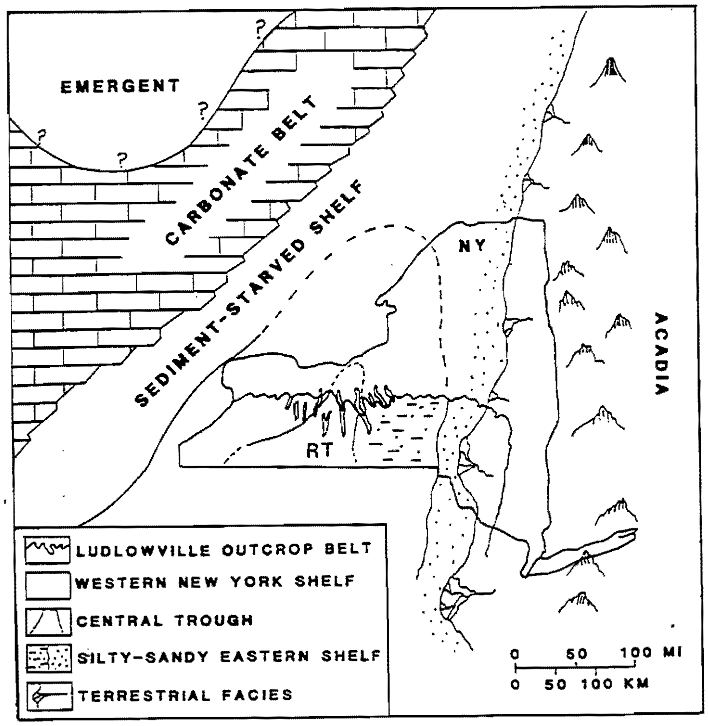

Figure 2. Generalized depositional setting and paleogeography of New York and adjacent areas during Middle Devonian (Givetian) time. RT= Romulus Trough. Modified from Mayer (1989).

The Jaycox Shale Member is an eastwardly thickening sequence of fossiliferous mudrock extending from Erie County to Seneca County and reaching its maximum thickness in the northern Cayuga Valley. However, the Jaycox discernably thins southeastward across Cayuga County with apparent litho- and biofacies changes. In western Onondaga County, Skaneateles Lake region, the Jaycox has been directly correlated to the Owasco Siltstone while the subjacent shales are directly correlated to the Spafford Member (Mayer and others, 1990,1994).

Stratigraphy of Jaycox Shale Member, Jaycox Creek Type Section, Genesee Valley

The type section for the Jaycox Shale Member is at Jaycox Creek, 2 miles north of Geneseo, New York. Here the stratigraphy was originally defined as shown in Figure 3. The Jaycox is bracketed at the base by the Hill's Gulch Bed and at the top by the Tichenor Limestone. The Jaycox strata contain two prominent coral horizons as well as variably fossiliferous mudstone intervals.

Downstream along the floor and in the banks of the creek, dark grey shales of the Wanakah Member are exposed. The uppermost layers contain a diminutive fauna of the brachiopod *Ambocoelia umbonata* var. *nana.* These fossils occur in such great numbers that they impart a granulated texture to the bedding planes and form a particularly distinctive marker horizon which is traceable throughout the region. In addition, chonetid brachiopods, the bivalves *Nuculites* and *Paleoneilo*, as well as the gastropod *Palaeozygopleura*, which is typically highly compressed laterally due to sediment compaction, occur in the uppermost Wanakah shales.

Directly overlying and gradational with the upper Wanakah Shale, is a 32 cm thick, medium grey fossiliferous silty mudstone originally designated the Limerick Road Bed (Mayer, 1989). Within the lower 7 cm of the Limerick Road Bed, cephalopods encrusted by the tubuliporate bryozoan *Reptaria stolonifera* commonly occur and form an important marker horizon throughout the study area. The Limerick Road bed also contains a conspicuous fauna of robust brachiopods including *Mediospirifer audaculus, Athyris spiriferoides*, and *Mucrospirifer mucronatus*. These fossils are preserved uncrushed and in life positions in many cases. Also, bivalves and the coral *Pleurodictyum* comprise part of the faunal assemblage.

The Limerick Road Bed is, in turn, overlain by a 45 cm thick interval of medium grey, silty shale which lacks fossils. The Limerick Road Bed and overlying barren shales are the westernmost equivalent of the Spafford Shale Member of the Owasco and Skaneateles Lake Valleys in central New York (Mayer and others, 1990).

The Hill's Gulch Bed (originally designated by Kloc, 1983, for exposures at Hill's Gulch Creek) marks the base of the Jaycox Shale Member. At Jaycox Creek, the unit is a 30 cm thick silty limestone. This unit forms the upper lip of the second falls along the creek. Fossils are abundant and include numerous species. Representative brachiopods are *Athyris spiriferoides, Devonochonetes coronatus, Tropidoleptus carinatus, Douvillina inequistriata, Mediospirifer audaculous*, and *Mucrospirifer mucronatus*. Additional fossils include the corals *Amplexiphyllum hamiltoniae* and *Stereolasma rectum*, the trilobite *Phacops rana*, the gastropod *Mourlonia* sp., the bivalves *Cypricardella bellistriata* and *Modiomorpha concentrica,* as well as various fenestrate bryozoans.

The Hill's Gulch Bed correlates directly with the Owasco Siltstone Member of central New York. (Mayer and others, 1990). The unit was mapped across the thick shale deposits of the central trough region of the Seneca and Cayuga Valleys and into the thinner deposits of the eastern shelf region of the Owasco and Skaneateles Valleys.

The Hill's Gulch Bed is overlain by a 50 cm thick interval of thinly bedded, grey mudstone, which is, in turn, overlain by the Green's Landing Coral Bed. This interval contains a shelly fauna but the fauna is not as diverse as the overlying coral bed fauna. Brachiopods, bryozoans, and pelmatozoans are common; however, corals are rare and tend to first appear just below the Green's Landing Coral Bed. The *Tropidoleptus-Longispina* pavements observed further east in the Seneca Valley originate in this interval, but at Jaycox Creek, these brachiopods are scarce.

The Green's Landing Coral Bed, the first of two coral-rich units in the Jaycox, is exposed along the banks and floor of this creek. The 32 cm thick, grey, crumbly mudstone contains a high diversity of corals, brachiopods, bryozoans, pelmatozoans, bivalves, gastropods, and trilobites, all of which can be readily collected. Conspicuous corals include *Heliophyllum halli, Heliophyllum confluens, Eridophyllum subcaespitosum, Favosites hamiltoniae* and *Favosites argus*. Dominant brachiopods include *Douvillina inequistriata, Elita fimbriata, Pentamerella pavillionensis, Parazyga hirsuta*, and

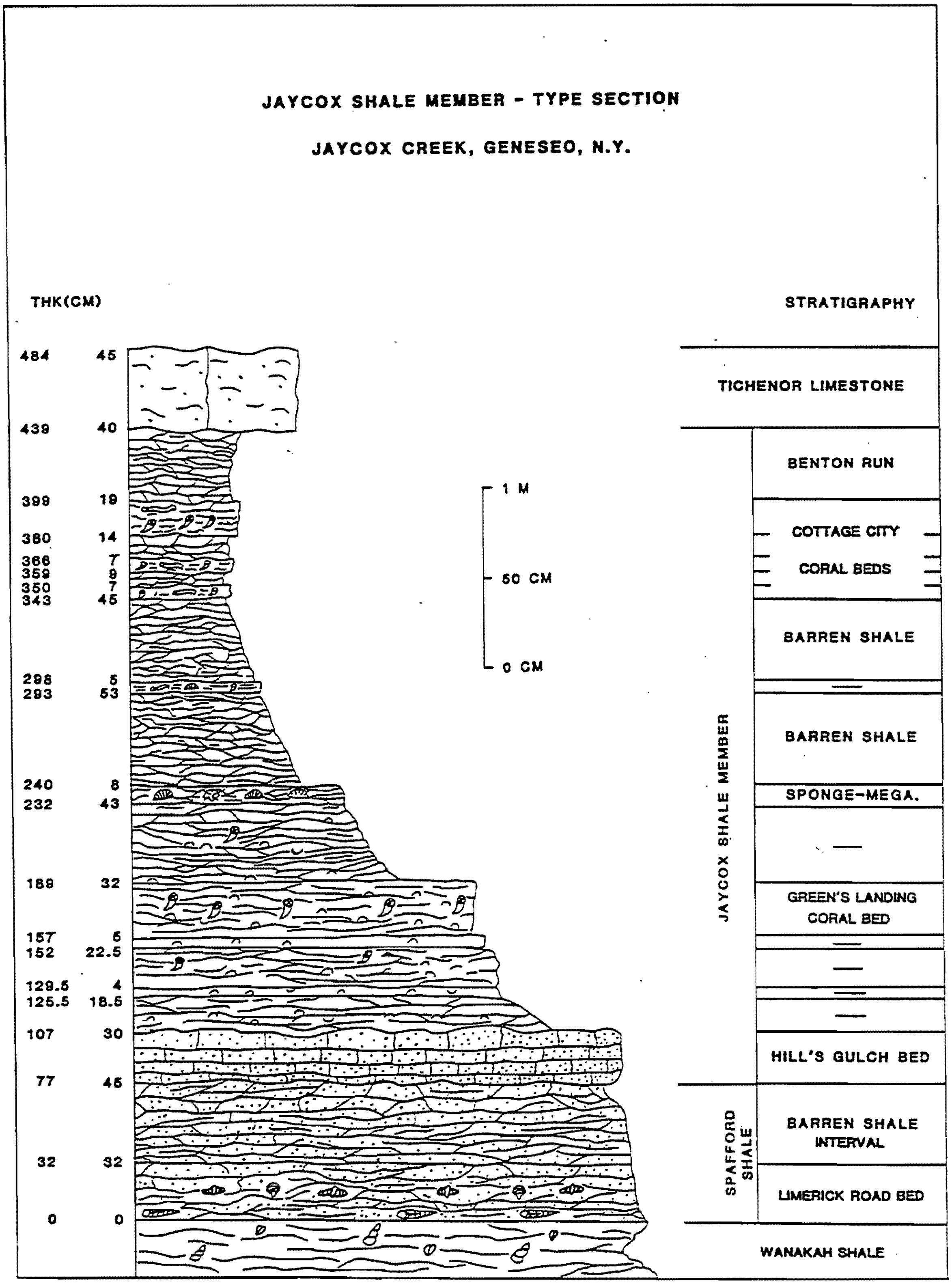

Figure 3. Stratigraphy of the Jaycox Shale Member, Genesee Valley.

Rhipidomella spp. Also common are fistuliporoid and fenestrate bryozoans, large crinoid holdfast systems and camerate crinoids such as *Dolatocrinus liratus* and *Megistocrinus depressus*, bivalves including *Modiomorpha concentrica* and *Plethomytilus oviformis*, gastropods such as *Naticonema lineata* and various platyceratids, as well as the trilobite *Phacops rana*.

The Green's Landing Coral Bed is overlain by a 43.5 cm thick interval of sparsely fossiliferous grey mudstone. This mudstone, in turn, grades upward into an 8 cm thick, hard, slightly calcareous, grey mudstone designated the Sponge-*Megastrophia* Bed (Mayer, 1989) due to a conspicuous fauna of unidentified "lithistid" demosponges and the brachiopod *Megastrophia concava*.

The Sponge-*Megastrophia* Bed is overlain by two approximately 50 cm thick intervals of unfossiliferous calcareous shales separated by a thin (5 cm) fossiliferous calcareous mudstone bed. This intervening bed contains organisms similar to that of the Green's Landing Coral Bed; however, fossils are less abundant.

The Cottage City Coral Beds, the second of the two coral-rich units in the Jaycox, immediately overlie these barren shales (Mayer, 1989). The Cottage City Coral Beds are 56 cm thick and comprise three distinct fossiliferous calcareous mudstone layers; each bed is overlain by a layer lacking fossils and each bed contains a greater abundance of fossils than the preceding bed. The Cottage City Coral Beds contain a diverse fauna very similar to the Green's Landing Coral Bed; however, the coral *Heliophyllum confluens* has never been observed in the Cottage City Coral Beds.

The uppermost division of the Jaycox is a 38-40 cm thick, light grey, calcareous shale that contains a very sparse fauna of brachiopods. These may include *Mesoleptostrophia*, *Mucrospirifer*, *Athyris*, and chonetids. The Jaycox upper contact with the Tichenor Limestone is sharp, irregular and erosional.

The Tichenor Limestone is the basal unit of the Moscow Formation (Baird, 1979). At Jaycox Creek, the upper falls are formed by the hard Tichenor limestone. The unit is a massive, calcarenitic, crinoidal grainstone. The Tichenor contains a diverse fauna including the rugose corals *Heliophyllum halli* and *Eridophyllum subcaespitosum*, the tabulate coral *Favosites hamiltoniae*, and the brachipod *Meristella*. Large holdfast systems, many over 12 inches long and one-half inch in diameter, as well as camerate crinoids including *Dolatocrinus* are plentiful.

The Tichenor Limestone grades quickly upward into the Deep Run Shale Member. The Deep Run is a medium grey-blue silty shale with numerous individual fossiliferous beds similar to the Jaycox. Well preserved fossils are abundant in the lowermost Deep Run strata, however; fossils become less plentiful stratigraphically upward. Fossils include the corals *Heliophyllum halli*, *Eridophyllum subcaespitosum*, and *Favosites* sp. as well as the brachiopods *Douvillina*, *Elita*, and *Parazyga*. Large fistuliporoid bryozoan mounds and large specimens of the trilobite *Phacops rana* occur in the Deep Run beds. Moreover, platyceratid gastropods attached to crinoid calyces have been collected from the floor and banks of the creek.

The Deep Run Shale Member grades stratigraphically upward into the Menteth - a very hard, rubbly, fossil-poor limestone. The Menteth is exposed furthest upstream and is the first rock ledge visible in outcrop.

Stratigraphy of Jaycox Shale Member, Kashong Creek, Seneca Valley

The beds of the Jaycox Shale Member as exemplified at the type section have been correlated into Yates County and can be observed directly at Kashong Creek (Figure 4) in the Seneca Valley. The Jaycox thickens significantly from 3.62 m at Jaycox Creek to 11.51 m at Kashong Creek. Both lithologic and faunal changes accompany this thickening. Jaycox shales contain more silt and are less calcareous in the Seneca Valley than in the Genesee Valley. Faunas are less diverse and represent those species more tolerant of higher sedimentation rates (see Mayer, 1989; Mayer and others, 1994).

The Limerick Road Bed remains sharply defined at Kashong Creek where the unit contains bioencrusted hiatus concretions. These concretions form a mappable horizon throughout the Seneca and Cayuga Valleys. The barren shale interval separating the Limerick Road Bed from the Hill's Gulch Bed thickens to approximately 2 m at Kashong Creek. From Seneca Lake eastward the unit contains unidentified discoidal fossils that can be traced into the Spafford Shale Member of the Skaneateles Lake region.

The lithology of the Hill's Gulch Bed grades laterally across a facies spectrum from a limestone on the western shelf in Erie County, to a silty limestone in the Genesee Valley, and to a calcareous siltstone in the trough region in the Seneca and Cayuga Valleys. As the unit is correlated out of the depocenter and onto the eastern shelf in the Skaneateles Valley, it exhibits the siltstone lithology of the Owasco Member. At Kashong Creek, the Hill's Gulch Bed is 24 cm thick and caps the second falls downstream from the Tichenor falls. Biofacies present in the unit also grade laterally from communities containing the corals *Favosites hamiltoniae* and *Heliophyllum halli* in Erie County to communities containing the brachiopods *Tropidoleptus carinatus* and *Mucrospirifer mucronatus* and lacking most corals in Yates County. These biofacies change further east to communities dominated by the brachiopod *Allanella tullius* indicative of the Owasco Siltstone facies in the Skaneateles Valley.

The mudstone interval above the Hill's Gulch Bed at Jaycox Creek extends east to the Seneca Lake Valley, forming a unique marker interval rich in the brachiopods *Tropidoleptus carinatus* and *Longispina mucronatus*. These brachiopods occur concentrated along bedding planes. Also, large specimens of the trilobite *Dipleura dekayi* occur associated with these brachiopods.

The Green's Landing Coral Bed undergoes litho- and biofacies changes eastward of Canandaigua Lake. The unit grades from a crumbly mudstone in the Canandaigua Valley to a silty mudstone in the Seneca Valley. At Kashong Creek, the bed is approximately 36 cm thick. Although, the bed still contains an abundant and diverse fauna, the corals *Heliophyllum*, *Eridophyllum*, and *Favosites* are absent and instead have been replaced by the corals *Amplexiphyllum hamiltoniae* and *Stereolasma rectum*. Also the brachiopods *Productella spinulocosta* and *Tropidoleptus carinatus* are present in the fossil assemblage.

The Sponge-*Megastrophia* Bed observed at Jaycox Creek also grades laterally eastward from a calcareous mudstone to a silty mudstone in the Seneca Valley. At Kashong Creek, the bed is approximately 51 cm thick. The demosponges are absent and *Megastrophia concava* is rare while *Tropidoleptus* and *Mucrospirifer* are more abundant.

The middle Jaycox "barren shale" interval progressively thickens eastward from less than 1 m in the Genesee Valley up to about 2 m in the Canandaigua Valley.

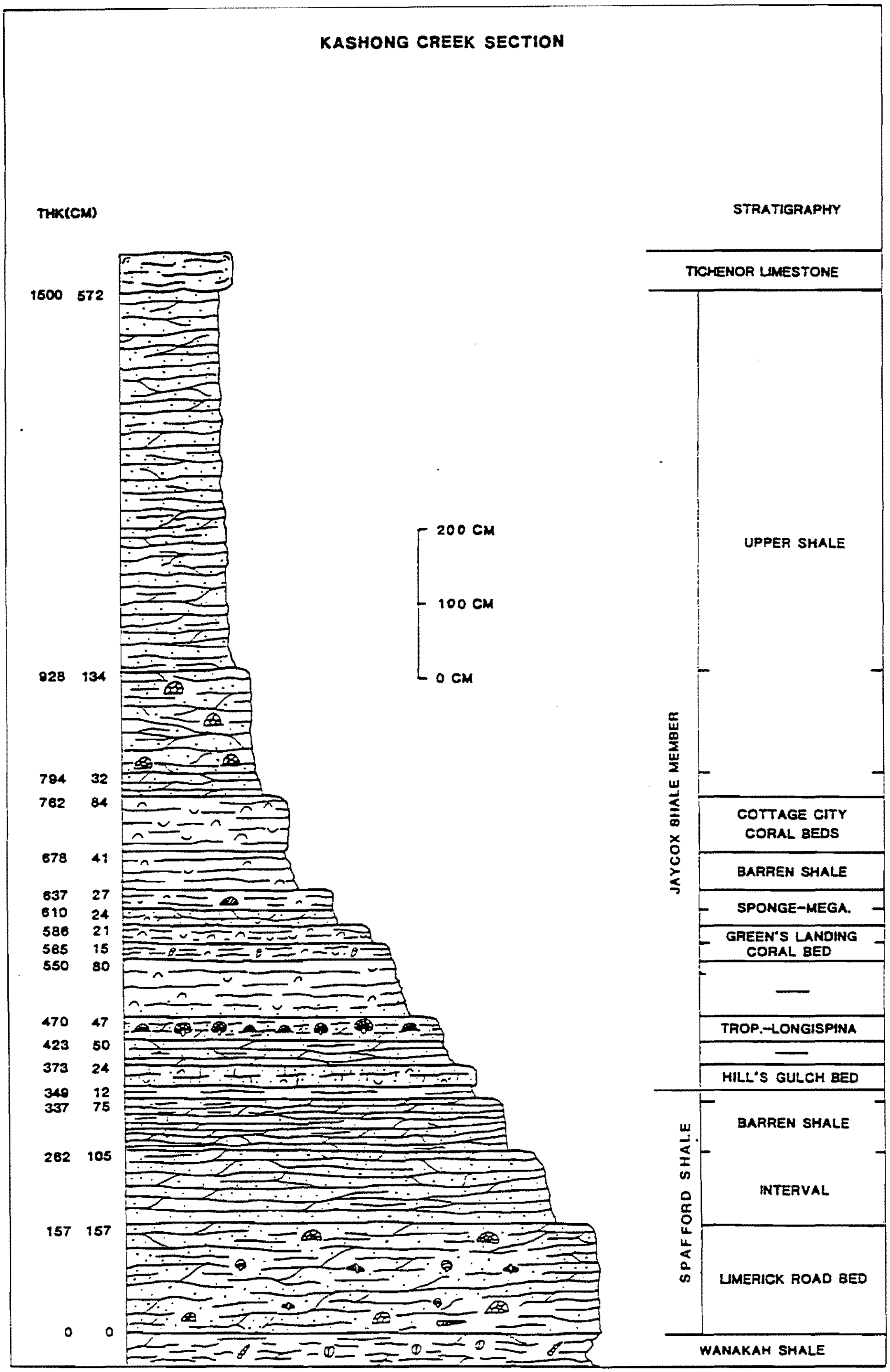

Figure 4. Stratigraphy of Jaycox Shale Member, Seneca Valley.

However, in the Seneca Valley, these shales thin to less than one-half m. The shales in the Seneca Valley contain more silt and are less calcareous than in the Genesee Valley. The interval still contains a very low diversity of fossils including *Tropidoleptus*, a few bivalves and the trilobite *Greenops boothi*.

The Cottage City Coral Beds also undergo eastward facies changes similar to the Green's Landing Coral Bed. At Kashong Creek, the unit is approximately 80 cm thick. The lithology changes from a calcareous mudstone to a silty mudstone. The faunal assemblage remains abundant and diverse in the Seneca Valley; however, the large rugosans and tabulates are absent and have been replaced by *Amplexiphyllum* and *Stereolasma*. Also *Productella* and *Tropidoleptus* are part of the biota and occur with *Parazyga*, *Pentamerella*, *Elita*, and *Douvillina*.

The uppermost shale division of the Jaycox has thickened exponentially from 40 cm at Jaycox Creek to 740 cm at Kashong Creek. The division is composed principally of *Zoophycos*-swirled calcareous siltstone. Bedding planes typically contain a sparse fauna of brachiopods consisting primarily of *Mesoleptostrophia*, *Athyris*, *Mucrospirifer*, *Mediospirifer* and chonetids. Also, large *Pleurodictyum* corals and articulated *Greenops boothi* trilobites occur in an interval in the lower part of this division.

The upper contact of this division with the Tichenor Limestone of the overlying Moscow Formation is sharp and irregular. The uppermost Jaycox shale thins dramatically both northward and westward. The Tichenor Limestone is predominately a calcareous siltstone in the Seneca Lake Valley but still contains very large crinoidal holdfast systems, as well as brachiopods and other fauna.

The Tichenor grades upward into the overlying Deep Run Shale Member. Like the Jaycox in the Seneca Lake Valley, the Deep Run is a slightly calcareous, silty mudrock containing shell-rich layers separated by unfossiliferous zones. Perhaps through additional research, the stratigraphy of these beds can be further refined and correlated westward.

DISCUSSION

Key fossiliferous marker beds facilitated detailed correlation of the Jaycox Shale Member across western and central New York (Figure 5A). Important correlative units include the Hill's Gulch Bed and the Green's Landing and Cottage City Coral Beds. Species diversity is high in Jaycox coral and shell beds but decreases basinward from both eastern and western shelves probably reflecting increased turbidity in the Middle Devonian trough setting. Rapid, episodic, storm-induced deposition is believed to explain the excellent fossil preservation and the occurrence of many organisms in life position in these beds. The Green's Landing Coral Bed seems to represent many rapid burial events. Except for smothered bottom assemblages, fossil assemblages reflect the biota as it persisted through time; that is, they represent "time-averaged" communities (*sensu* Walker and Bambach, 1971). Moreover, these assemblages or biofacies can be tracked and compared to the paleoecological model for Ludlowville biofacies put forth by Brett, Baird and Miller (1986) which relates biofacies to inferred depths, turbidity conditions and rates of sedimentation (see Mayer, 1989; Mayer and others, 1994).

The sparsely fossiliferous "barren shale" intervals within the Jaycox are heavily bioturbated and display abundant spreiten of *Zoophycos*. Fossils are rare and scattered but typically well preserved. It is believed that this facies represents rapid deposition of muds as a result of storm activity. These muds were probably winnowed from high

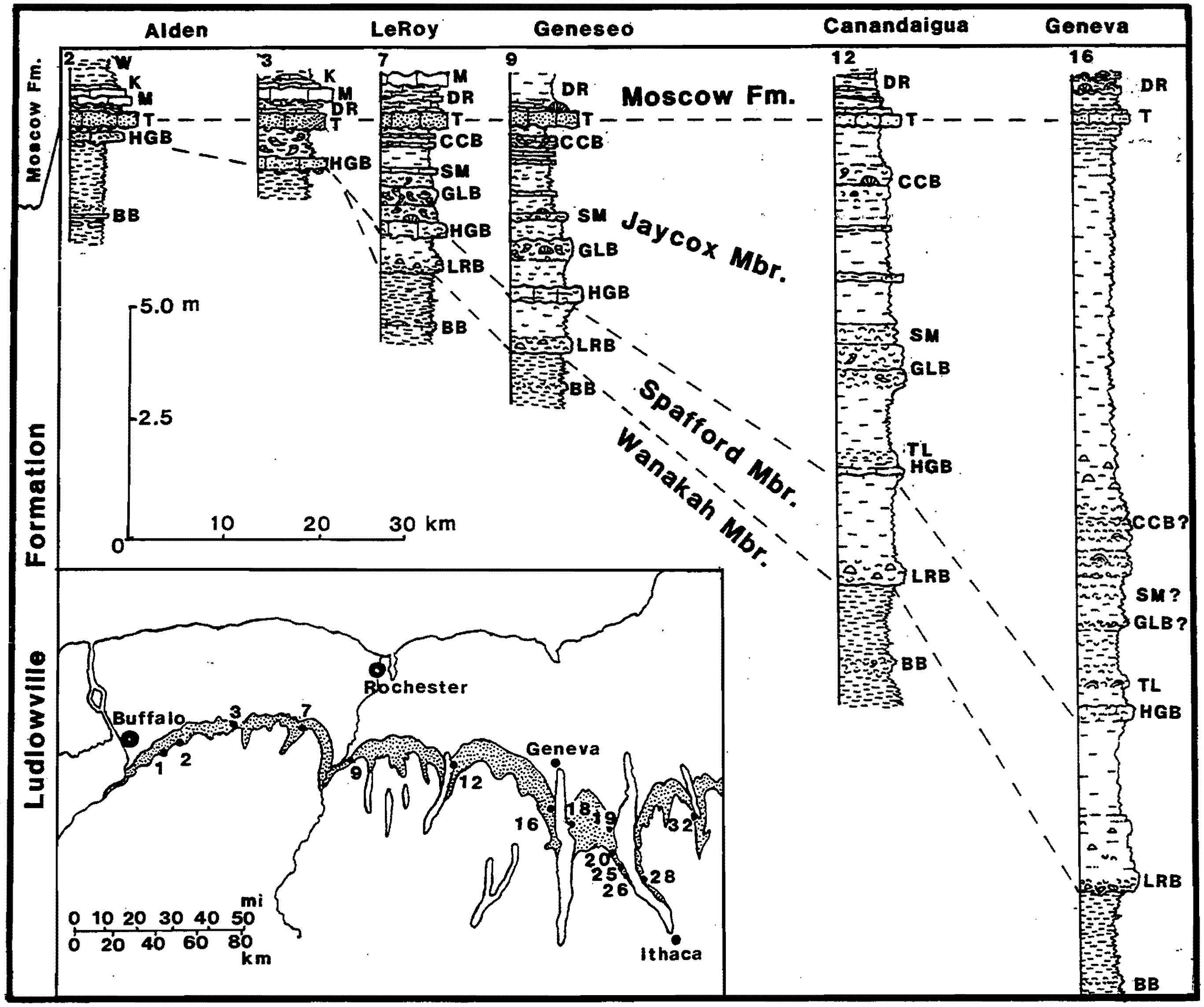

Figure 5A. Correlation of facies divisions of the Upper Ludlowville - Lower Moscow Formations across western New York. Units include BB=Bloomer Creek Bed; LRB=Limerick Road Bed; HGB=Hill's Gulch Bed; TL=*Tropidoleptus-Longispina* mudstone interval; GLB=Green's Landing Coral Bed; SM=Demosponge-*Megastrophia* Bed; CCB=Cottage City Coral Beds; T=Tichenor Limestone; DR=Deep Run Shale Member; M=Menteth Member; K=Kashong Member; and W=Windom Member. Locality numbers can be referenced to Mayer and others (1994).

energy shallow shelf regions and were transported into lower energy settings, with subsequent bioturbation destroying most primary bedding (Brett and others, 1986).

Moreover, Aigner's (1982, 1985) model of tempestite proximality allows one to make inferences of depositional environments. Ideally, proximal storm beds (generally shallow water) are thick, bioclast-dominated amalagamated beds with erosional bases; conversely, distal beds (generally deeper water equivalents) are thin, mud-dominated discrete beds with non-erosional bases. These characteristics are exhibited by the Hill's Gulch Bed. The unit shows evidence of multiple winnowing and reworking of fossil and sediment accumulations as a result of proximal storm deposition on the western shelf as well as evidence of less reworking of lithoclasts and bioclasts distally or basinward. Furthermore, deposition of the Hill's Gulch Bed on the western shelf probably was characterized by high energy direct wave impingement on the seafloor as evidenced by scour-and-fill structures resulting in a highly irregular basal surface of the bed (Brett and Baird, 1985; Brett and others, 1986). Conversely, deposition of the Hill's Gulch Bed within the basinal trough was probably characterized by low energy sediment accumulations.

The upper contact of the Jaycox with the Tichenor is sharp and irregular at most localities. Although westward thinning of the Jaycox is due in part to sedimentary condensation, more importantly it is due to sub-Tichenor erosional overstep of the Jaycox. This erosional truncation removed upper portions of the Jaycox, superimposing the Tichenor on the Hill's Gulch Bed or uppermost Wanakah in western Erie County. However, the Jaycox consists of thick deposits in the Seneca and northern Cayuga Valleys, where sediment accumulations were greatest. To the southeast of the northern Cayuga Valley, Jaycox equivalent beds become markedly condensed and are erosionally truncated; the Tichenor disconformably overlies the Owasco Siltstone in the Owasco and Skaneateles Lake regions (Figures 5A-B). These regional stratigraphic patterns reveal a northeast-southwest trending structural belt of differential subsidence and deposition, (Romulus Sag) that was centered in the Seneca Valley and northern Cayuga Valley region during Ludlowville-Moscow time (Mayer and others, 1994).

Jaycox Creek and Kashong Creek are two key outcrops of the Upper Ludlowville-Lower Moscow strata exposing the Wanakah, Jaycox, Tichenor, and Deep Run Members. Numerous other outcrops were studied to reveal the stratigraphic trends of the Jaycox-Spafford deposition. This research is presented in greater detail in a recent paper by Mayer, Baird and Brett (1994).

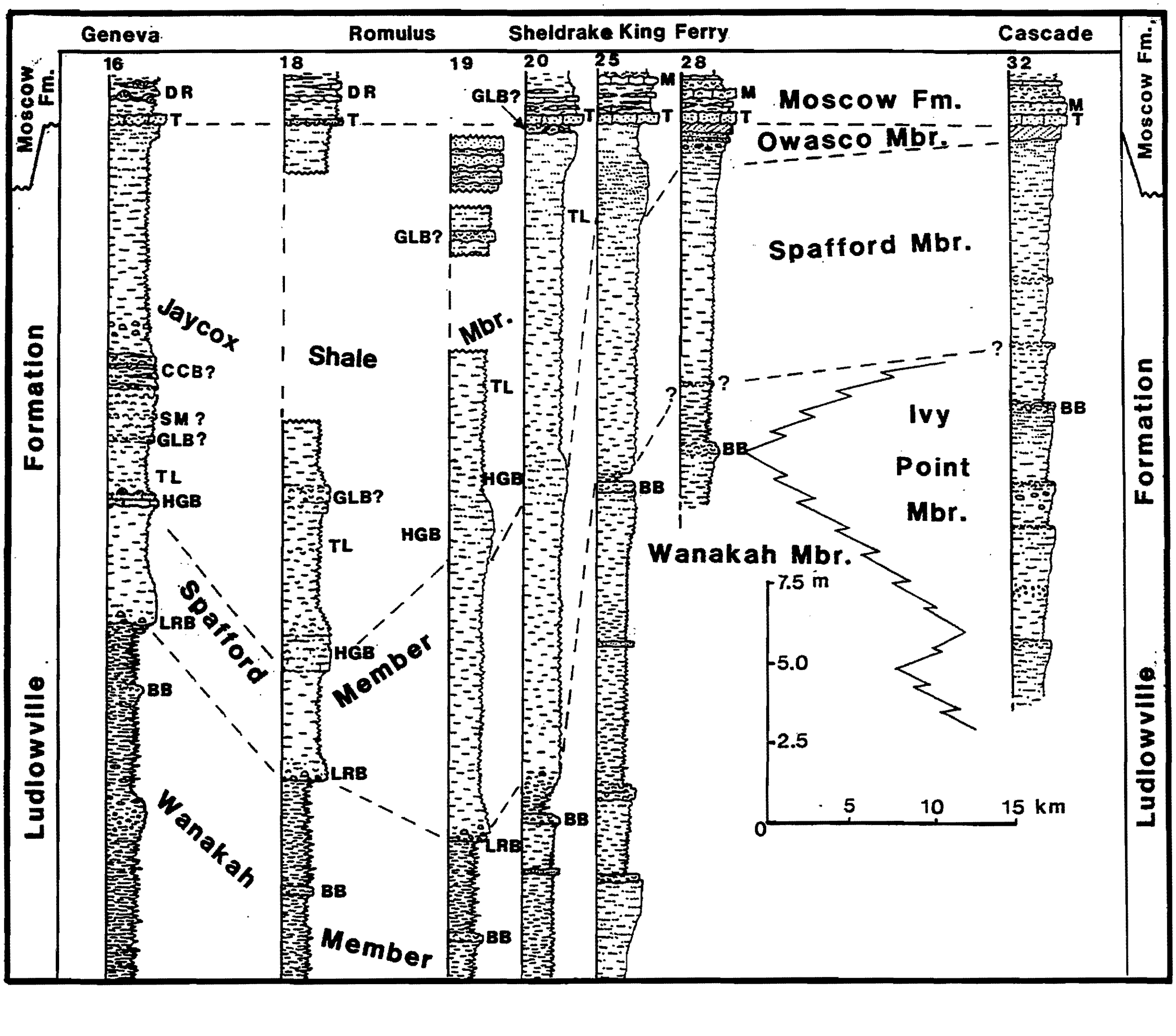

Figure 5B. Correlation of facies divisions of the Upper Ludlowville-Lower Moscow Formations across west-central New York. For abbreviations see Figure 5A.

ACKNOWLEDGMENTS

I would like to sincerely thank Dr. Gordon Baird for initially suggesting this detailed stratigraphic and paleontologic study of the Upper Ludlowville-Lower Moscow sequence. I would also like to thank Dr. Carlton Brett for his willingness to help throughout all phases of this research. Together they have provided valuable discussions both in and out of the field for the completion of this project. Dave Lehmann also reviewed this manuscript. Acknowledgment is also made to the Petroleum Research Fund of the American Chemical Society and to the National Science Foundation (Grant EAR-8313103) for support of field reseasrch.

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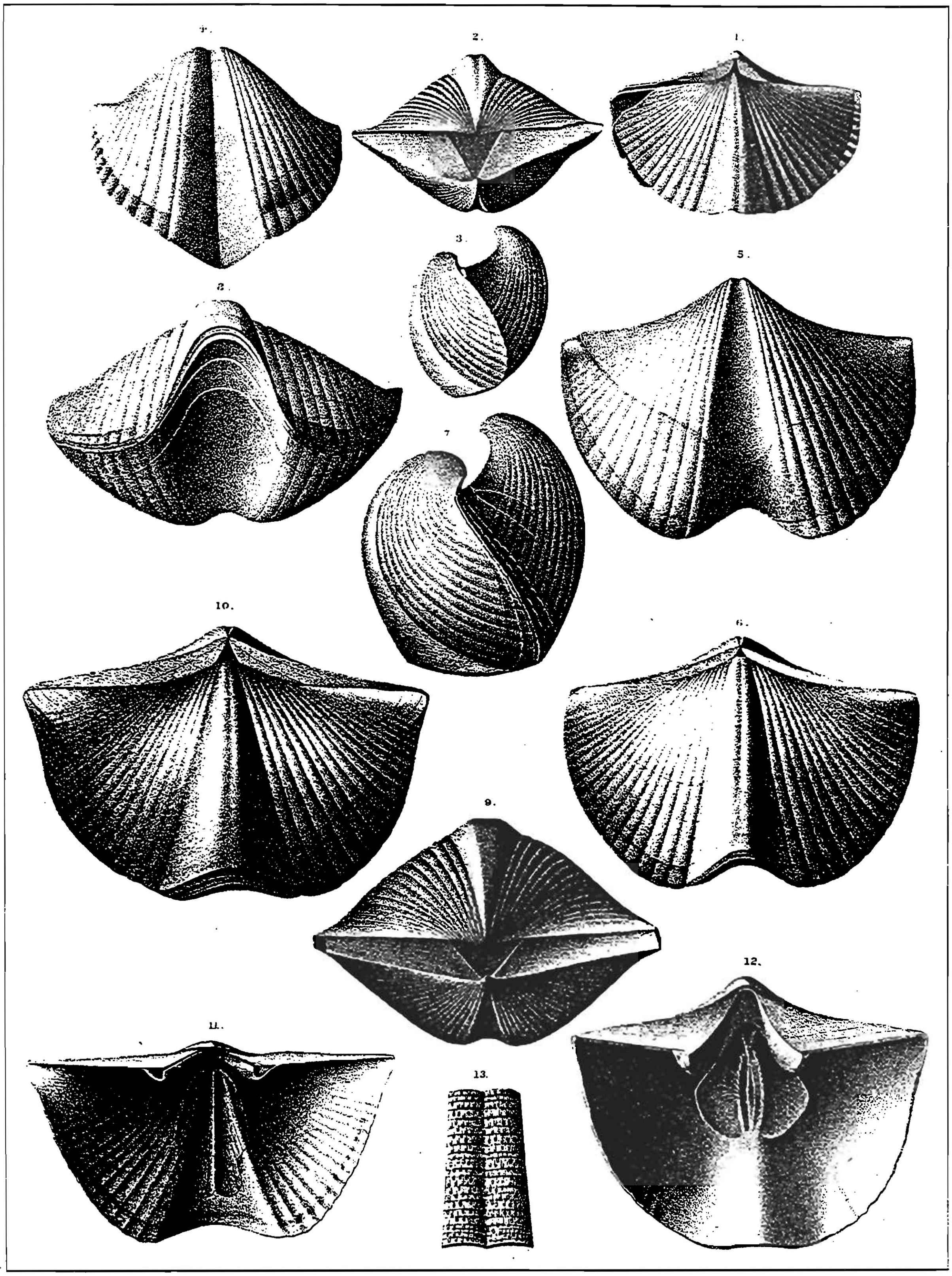

Spinocrytia granulosa. Hamilton Group.
From Hall, 1867, Plate 36.

DEPOSITIONAL SEQUENCES, CYCLES, AND FORELAND BASIN DYNAMICS IN THE LATE MIDDLE DEVONIAN (GIVETIAN) OF THE GENESEE VALLEY AND WESTERN FINGER LAKES REGION

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INTRODUCTION

Late Middle Devonian (late Givetian) deposits in the Genesee Valley - central Finger Lakes region are represented in ascending order by the Moscow Formation of the Hamilton Group, the Tully (Limestone) Formation, and the lower third of the Genesee Formation. This succession records a dramatic facies transition from fossiliferous subtidal shelf mudstones in the Moscow Formation through outer lagoonal-to-shelf carbonate deposits in the Tully Limestone, into black shales recording anoxic basinal conditions in the Genesee Formation. The Tully-Geneseo succession records a dramatic shift from carbonate platform conditions to a deep-water, anoxic foredeep setting due to inferred flexural loading of the craton by a pulse of orogenic thrusting (Ettensohn, 1985, 1987).

The Moscow-lower Genesee stratal succession yields abundant and diverse fossils ranging from robust, shallow water forms to delicate taxa that lived under conditions of near-anoxia. Moscow and Tully biotas are highly diverse and have been well known since the publication of James Hall's "fourth district" report in 1839. Well preserved fossils can be collected at all of the field stops.

Strata discussed herein were deposited as muds in a foreland basin setting linked to a key phase of the Acadian Orogeny, a collision event involving the "docking" of one or more microcontinents in what is now New England and the central Atlantic states (Faill et al., 1978; Woodrow 1985; Ettensohn, 1985, 1987). Orogenic uplift and overthrusting in the tectonic settings to the east and southeast of New York State led to flexural downbending of the craton in eastern North America (Quinlan and Beaumont, 1984; Beaumont et al., 1988). This loading produced a foreland basin characterized by greater regional subsidence and often greater water depths (Figure 1). Erosion of orogenic highlands in the east led to the formation of progradational complexes ("Great Catskill delta") which filled in the Devonian shelf seas associated with the foreland basin, thus producing the Devonian stratal record in New York State, Pennsylvania, Maryland, and Virginia. In actuality, several inferred collision and overthrusting events are believed to have produced tectophases of overthrust-induced basin deepening and progradational basin-filling (Ettensohn, 1987; Ettensohn and Elam, 1985). A mainly quiescent phase at the end of Tectophase II is recorded by the upper Hamilton Group (Moscow Formation). However, the Tully-Genesee transition records a grand deepening and progradational pulse associated with the onset of Tectophase III.

The Devonian ("Catskill") clastic wedge provides a classic example of an ancient deltaic complex. Because orogenic events took place to the east,

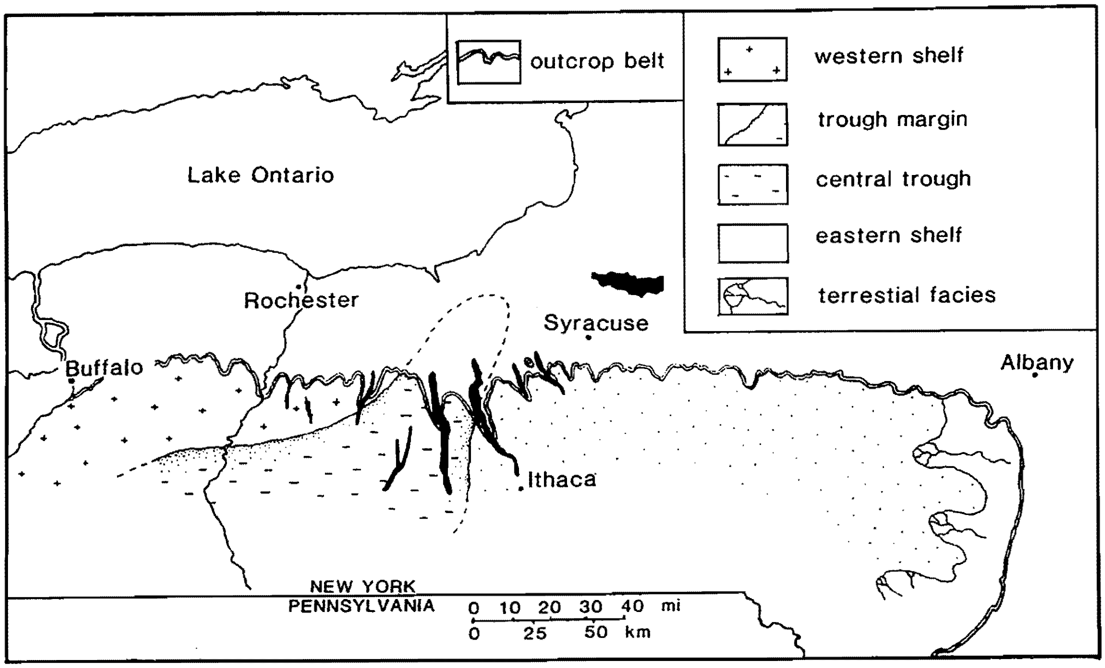

Figure 1. Geological setting of the northern Appalachian Basin during the Middle Devonian Givetian Age. Note the position of the Finger Lakes through (area of maximum subsidence) relative to the present outcrop belt. Dashed lines indicate conjectural positions of facies belts north of the outcrop belt.

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sediments supplied to the basin came from the eastern Acadian mountain belt. This is typically, though not always, reflected in the eastward-thickening of sediments within the Catskill clastic wedge. It is also reflected by west-to-east facies changes (limestone-to-terrigenous sediment, shale-to-siltstone, siltstone-to-sandstone, etc.) at almost all stratigraphic levels. These diachronous facies shifts are well displayed in the Devonian time-rock stratigraphic chart for the state (Rickard, 1975). The Catskill Delta is also a story of numerous isochronous events: evolutionary and immigration (epibole) events, volcanic eruptions, and sea level changes can be traced stratigraphically over long distances.

In this paper, we will focus on the record of sea level oscillations which have left a detailed record of sedimentary cycles during the Late Givetian Age (approximately 380-375 million years ago). Numerous thin limestones, marine erosion surfaces, and stratigraphically condensed intervals record sediment-starvation events associated with transgression events; these are well displayed in strata discussed herein. Recent advances in our understanding of Catskill Delta depositional processes involve the use of principles of sequence stratigraphic mapping.

The purpose of the present paper and associated field trip is fourfold: first, we document a rich assortment of shelf-to basin facies which is characterized by a broad array of Devonian fossils; second, we illustrate advances in stratigraphy which have been developed with the past few years, particularly involving application of sequence stratigraphy concepts to Paleozoic foreland basin deposits; third, we illustrate patterns of differential subsidence and migration of the foreland basin centers; finally, we illustrate several excellent Late Givetian sections within two hours' drive of Rochester which yield abundant, diverse, and well preserved fossils for the Devonian enthusiast.

We have chosen to describe Moscow subdivisions in some detail in the present paper in order to provide the details upon which our interpretations are based and as a guide to the various fossil bearing levels to aid paleontologic studies. Beacuse most horizons are laterally traceable for tens of kilometers we have followed the practice of assigning many bed names for those divisions. This is not intended to overwhelm the reader with detail but to facilitate unambiguous communication about the many levels. The more general reader may wish to bypass these sections and read the sections on sequence and cycle interpretation.

DETAILED STRATIGRAPHY AND SEQUENCE INTERPRETATION OF THE MOSCOW FORMATION, UPPER PART OF HAMILTON GROUP, WEST CENTRAL NEW YORK STATE

The Middle Devonian Moscow Formation is the uppermost of five major packages or formations in the Hamilton Group (Figure 2). The term "Moscow" was first applied to bluish gray shales (mainly Windom Member) exposed along Little Beards Creek in the town of Moscow (later renamed Leicester), Livingston County, New York, by James Hall (1839). In western New York, the Moscow consists of gray to black shales; calcareous mudstones and thin limestones. The Moscow Formation can be construed as a single major depositional sequence, representing perhaps 1.5 to 2 million years of geologic time and bounded above and below by relatively important and regionally angular unconformities. The basal sequence boundary coincides with the erosion surface below the Tichenor Limestone. The upper boundary is the disconformity below the Tully Limestone, or, where that unit is absent, below the Genesee Formation black shales. However, within a sequence context, the Moscow Formation can be subdivided into two unequal packages or fourth order sequences which constitute groups of previously named members. The lower portion, which might be termed the Portland Point "subformation," consists of the Tichenor, Deep Run, Menteth and Kashong members, as defined by Cooper (1930, 1933). In the Cayuga Lake region and, again, in Erie County, members thin dramatically and merge into a thin limestone-rich interval previously termed the "Portland Point Member" (Cooper, 1933); however, Baird (1979) demonstrated that individual limestone members can still be recognized within the Portland Point. Overall, this is a retrogradational or deepening-upward succession that involves several internal cycles ("fifth order cycles in the terminology of Busch and Rollins, 1984). The upper and thicker portion of the Moscow Formation consists of a single member in western New York, the Windom Shale and its lateral equivalent, the Cooperstown Siltstone in central New York. In the Central Finger Lakes region another thin, silty, fossil-rich interval, herein informally termed the "unnamed member" is interposed between the Kashong and Windom members (Figures 2,3). This interval consists primarily of shales and thin concretionary limestones in western New York, but farther to the east, it can be subdivided into coarsening-upward mudstone to siltstone packages, which have been mapped in some detail by Zell (1985). The Windom Shale overall is interpreted as the highstand portion of the Moscow sequence. Uppermost regressive portions of the sequence (or late highstand

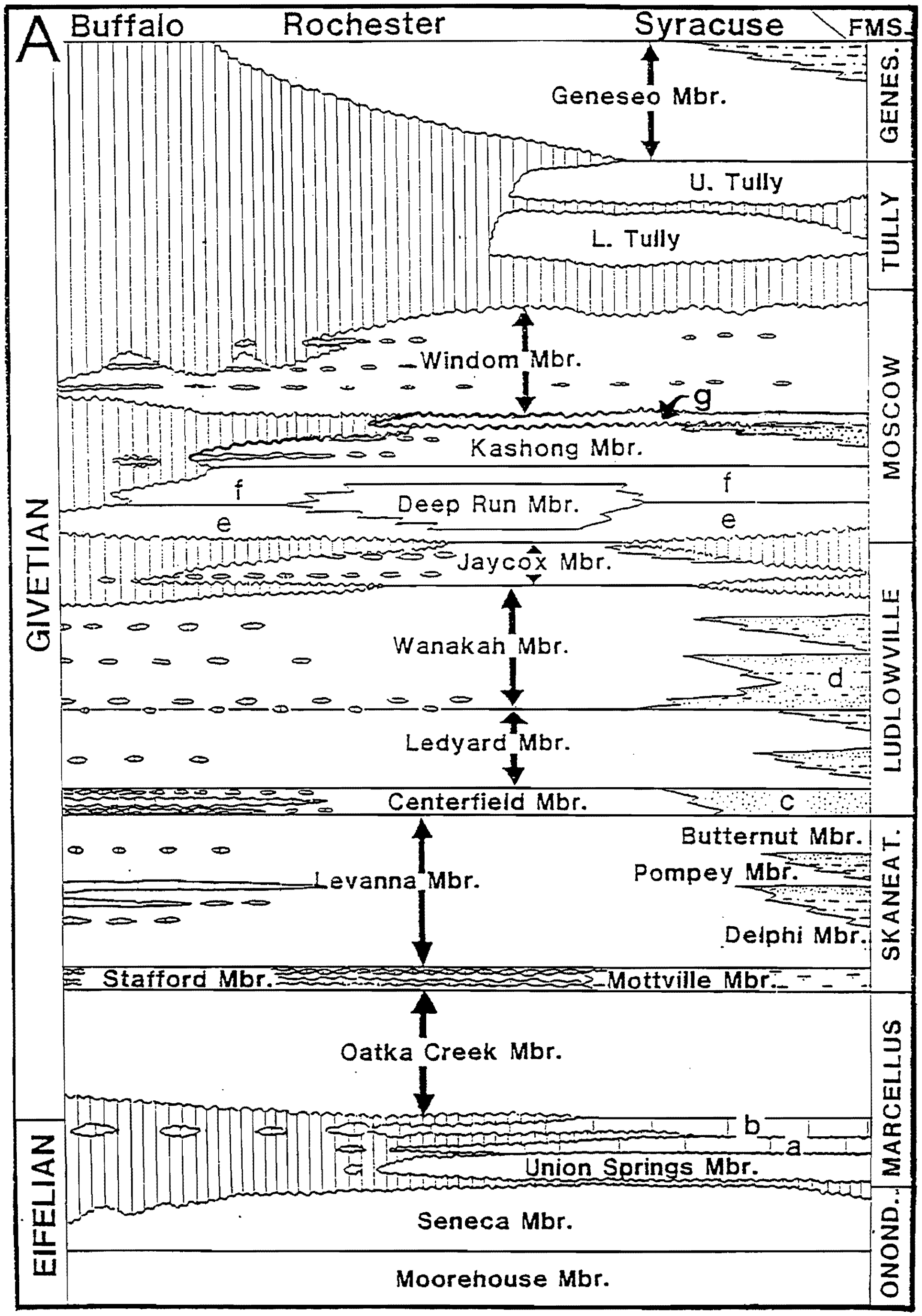

Figure 2A. General stratigraphy of the Middle Devonian Hamilton Group and under and overlying units in western New York. Lettered units include: a) Chestnut Street beds; b) Cherry Valley; c) Chenango Sandstone; d) Ivy Point Siltstone; e) Tichenor Limestone; f) Menteth Limestone; g) unnamed member (Barnes Gully bed to Geer Road bed, see text).

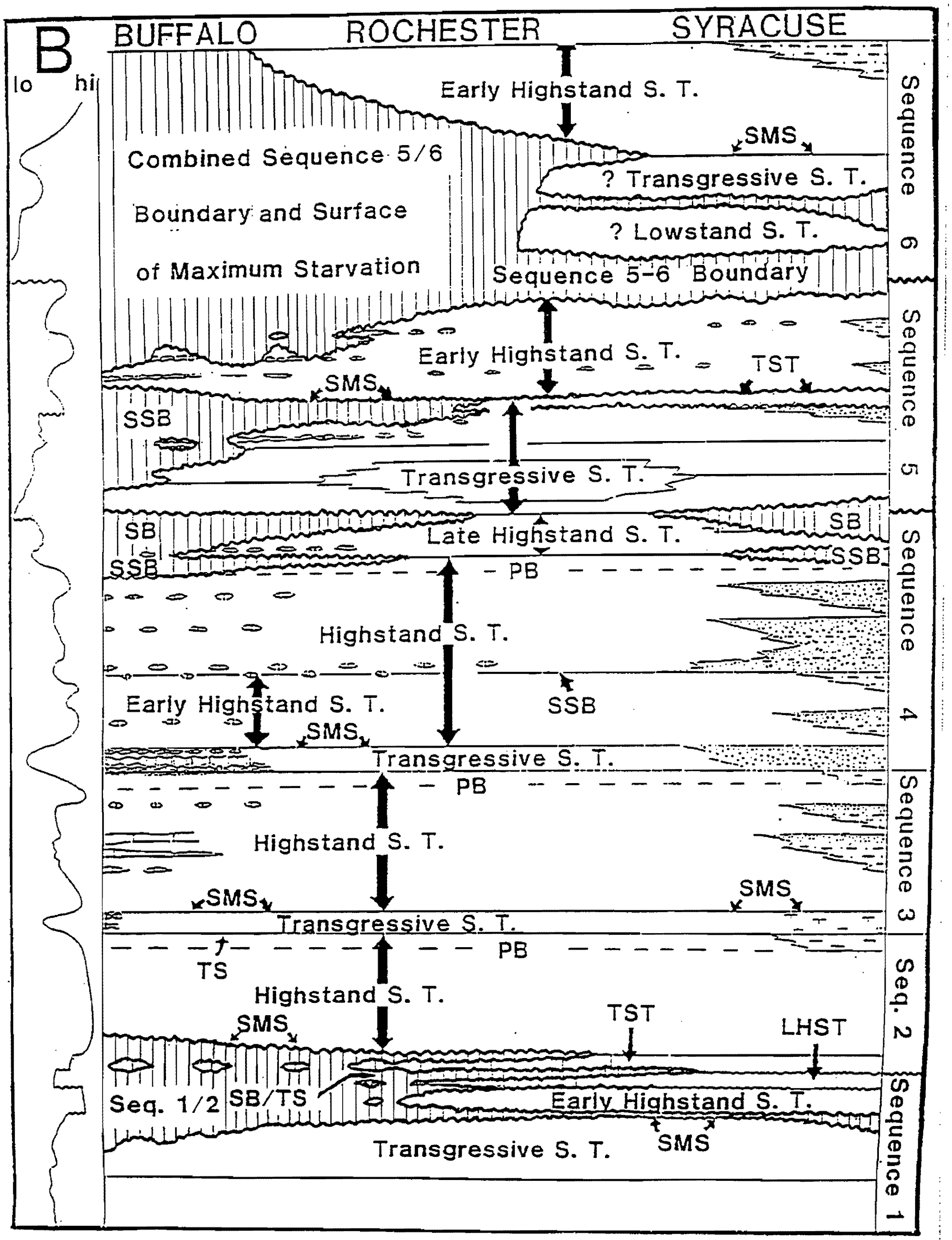

Figure 2B. Sequence interpretation of the Middle Devonian. Curve on left illustrates inferred sea level fluctuations. Abbreviations. SB = sequence boundary; SSB = subsequence boundary; TS = transgressive surface; PB = precursor bed; ST = systems tract; SMS = surface of maximum starvation.

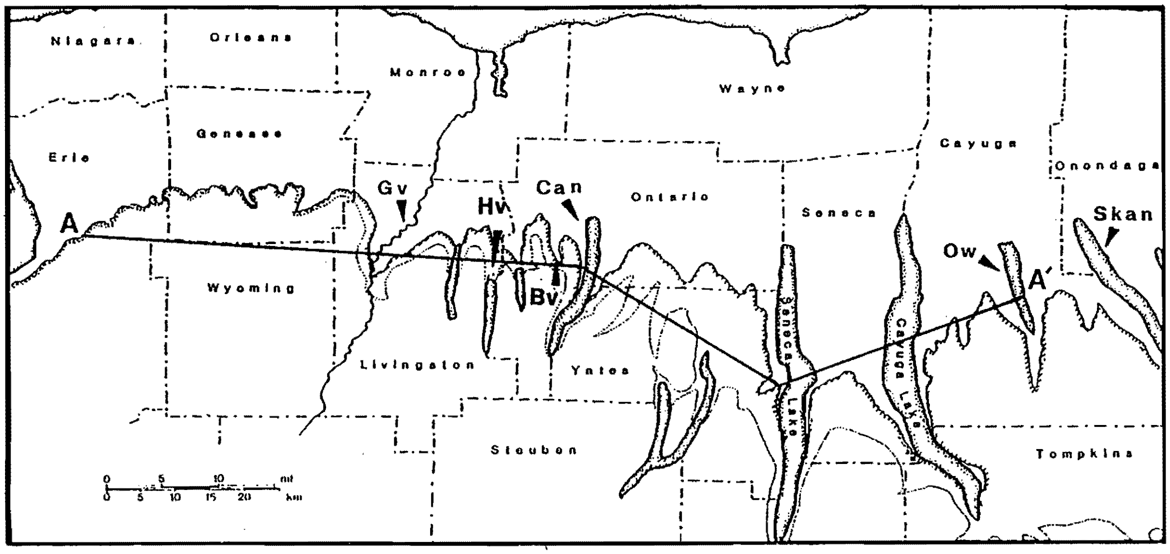

Figure 3. Map of western New York State showing location of key geographic features referred to frequently in the text. Counties are named; abbreviations include: Bv, Bristol Valley; Can, Canandaigua Lake; Gv, Genesee Valley; Hv, Honeoye valley; Ow, Owasco Lake; Skan, Skaneateles Lake. Cross section of Figure 5 is indicated at line A-A'.

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progradational succession), representing the Windom-into-Tully facies transition, have been removed by erosion at the sub-Tully unconformity. Details of various increments of this sequence are described in the following sections. For location of counties and other features referred to in these sections, see Figure 3.

Tichenor Limestone

Description: In west-central New York State the Tichenor Limestone is a thin, 30-50 cm- (1 to 2 foot-) thick compact limestone bed; its western facies seen primarily in Erie to Livingston counties consist of coarse crinoidal skeletal grainstone. However, eastward in the vicinity of Canandaigua Lake the Tichenor becomes slightly thicker and considerably finer-grained, represented by a styliolinid and crinoidal packstone (Griffing, 1994).

Interpretation: The Tichenor is considered to represent the first transgressive limestone of the Moscow sequence (Figures 2,4). It overlies a distinct disconformity which bevels the underlying Ludlowville Formation in both a westward, and probably an eastward, direction from the central Finger Lakes region. In the vicinity of Seneca and Cayuga Lakes where the boundary becomes most nearly conformable, the Tichenor still sharply overlies upper silty beds of the underlying Jaycox Shale Member (Figures 2,4). Indeed, an upper transitional coarsening-upward cycle culminating in medium- to coarse-grained siltstone deposits has been located at Big Hollow Creek just west of Cayuga

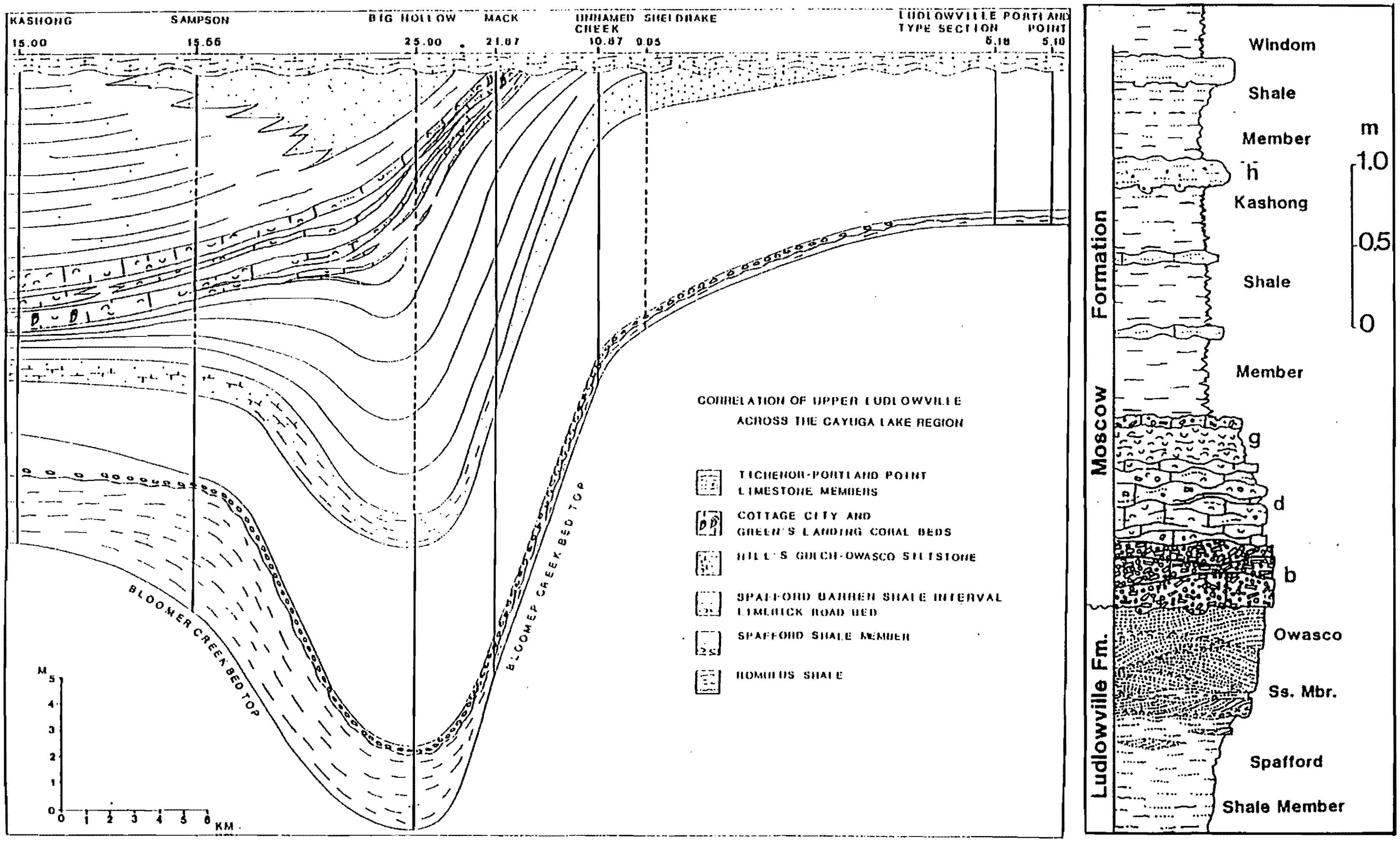

Figure 4. Schematic northwest-southeast cross section of the uppermost Ludlowville Formation and overlying Tichenor Member of the Moscow Formation; position of cross section in Seneca-Cayuga Lake region is shown in Figure 3. Note depocenters of upper Ludlowville in vicinity of Big Hollow Creek. Inset shows stratigraphic section of upper Ludlowville and Lower Moscow Formations of Portland Point. Symbols include: b, Tichenor Limestone; d, Deep Run-Mentith; g, condensed lower Kashong Member; h, Barnes Gully phosphatic bed. Modified from Mayer et al., (1994).

Lake in the Town of Ovid. This latter may be interpreted as the late highstand or progradational package of the underlying Ludlowville sequence (Figures 2,4).

To the west, the basal sequence boundary of the Moscow Formation (base-Tichenor disconformity) progressively bevels beds of the middle to lower Jaycox Member; in western Erie County the Tichenor rests locally on the basal Hills Gulch limestone bed of the Jaycox Member, and even on the underlying Wanakah Shale Member where the Hills Gulch Bed has been removed by erosion. A comparable pattern of progressive erosive overstep is observed beneath the Tichenor southeastward from the Ovid area both across and along Cayuga Lake (Figure 4). The upper silty beds of the Jaycox Member have been progressively removed and the Tichenor comes to be juxtaposed sharply, but with a welded contact either on the Owasco Siltstone (lateral equivalent of Hills Gulch bed), or, where that has been removed, on the underlying Spafford Shale (Mayer et al., 1994; Figure 4, herein). This "mirror image" unconformity on either side of the Cayuga Seneca Lake region defines a trough of most active subsidence during early Moscow deposition in which lowstand siltstones apparently were dumped into the basin at the same time that erosion took place along the margins. The sub-Tichenor lowstand was sufficiently strong that in most places at least a minor disconformity underlies the Tichenor.

The Tichenor, in turn, is readily interpretable as a transgressive, sheet-like, lag deposit of crinoidal, coral, and other fossil debris, except near the basin center (McCave, 1969, 1973; Brett and Baird, 1985, 1990; Griffing, 1994). The Tichenor grainstone was deposited in relatively clear, shallow water environments during an initial transgression that took place after maximal sea level lowstand.

The upper contact of the Tichenor in most localities is relatively abrupt, although apparently gradational and conformable with the overlying Deep Run mudstones. At Jaycox Run near Geneseo small mounds of fistuliporoid bryozoans occur at this surface and extend upward into the mudstone. This appears to reflect a minor sea level rise, during which time these minature bioherms built upward. Tichenor thus records two scales of process. On the one hand, it represents the basal transgressive carbonate of the overall Moscow sequence; however, at a smaller scale, the Tichenor, in many localities, appears to represent an abbreviated small-scale cycle (parasequence) which is capped by a marine flooding surface at its contact with the overlying Deep Run mudstone.

In western New York a second crinoidal packstone ledge occurs somewhat above the main coarsely crystalline Tichenor ledge. Eastward, this bed appears to splay outward, being separated from the Tichenor proper by up to a meter of Deep Run mudrock. This then is interpreted as the capping bed of a second parasequence which eventually merged together with those of the underlying minor cycle in up-ramp basin margin sections.

Deep Run Member

The Deep Run Member was named by Cooper (1933) for exposures on Deep Run Gully just south of Kipp Road on the east side of Canandaigua Lake. At this locality, the Deep Run attains a thickness of approximately 18 meters, this being one of its thickest sections. A somewhat lower thickness of 11 meters occurs at Kashong Glen in the Seneca Lake Region. The interval thins markedly both to the southeast along Cayuga Lake, where the Deep Run (or

Deep Run-equivalent) Interval ranges from 4.0 to 0.5 meters, as well as west of the Canandaigua Lake area where the interval rapidly thins to approximately 2.5 meters in the Genesee Valley, and less than one meter in central Genesee County (Figure 5). Hence, the overall geometry of the Deep Run is that of a large-scale lens in cross section with its maximum thickness in the west central Finger Lakes Region and thinning both to the east and west to a feather edge (Figure 5).

The Deep Run consists of hard bluish gray, calcareous, and typically highly bioturbated (*Zoophycos*-bearing) silty mudstones. In many areas, particularly in the eastern outcrops, the Deep Run becomes a calcareous silty mudstone or siltstone with well preserved *Zoophycos* spreiten. The basal 0.5 to 1.5 meters of the Deep Run is somewhat richer in fossils, particularly the branching coral *Heliophyllum proliferum*, which is diagnostic of this interval, as well as large camerate crinoid columns and fenestrate and cryptostome bryozoans. Calyces of the crinoids *Dolatocrinus* and *Megistocrinus*, platyceratid gastropods, the brachiopods *Pentamerella, Protodouvillina*, and *Elita*, and relatively large specimens of the trilobites *Phacops*, *Greenops*, and *Monodechenella* are also typical of these basal beds. As noted above, in many localities a 20 to 30 centimeter, compact, crinoid coral-rich limestone occurs some 0.5 to 2 meters above the Tichenor Limestone. This latter bed resembles the Tichenor and has sometimes been included within that unit, although it is separated from the typical Tichenor by the interval of more characteristic Deep Run mudstone.

The upper beds of the Deep Run are relatively homogenous in appearance and contain only scattered fossils. However, well preserved crinoids, brachiopods, large bivalves and large trilobites may be encountered rarely. The Deep Run appears to become somewhat more silt-rich (coarser) upward. Narrow prod like structures (burrows or gutter casts?) filled with laminated silt occur sporadically within these beds. Near the basin center, the Deep Run appears to grade upward into the overlying silty Menteth Limestone; however, towards the basin margins, near Cayuga Lake and in western New York the contact is sharp.

Menteth Limestone

Description: The Menteth Limestone consists of about 30 to 40 centimeters of thoroughly bioturbated (*Zoophycos*-churned), hard, silty, calcareous mudrock or very silty limestone. This uniformly thin limestone bed separates the Deep Run and Kashong members (Figure 5). On the whole, the unit appears to be sparsely fossiliferous. Thin sections show abundant sponge spicules in some areas (Griffing, 1994). In western New York, the Menteth displays very scattered small bluish-gray chert nodules.

The fossils in the Menteth are dominated by brachiopods and relatively large trilobites, some of which may be silicified. Large specimens of *Spinocyrtia*, *Mucrospirifer*, *Tropidoleptus*, *Athyris*, and *Nucleospira* are abundant; this fauna resembles that of the overlying Kashong Shale, except for the abundance of *Spinocyrtia*. The upper surface of the Menteth typically displays a somewhat richer biota that may include occasional rugose corals and fairly abundant coralla of the branching tabulate coral or *Thamnoptychia* (= *Trachypora*) as well as large camerate crinoids.

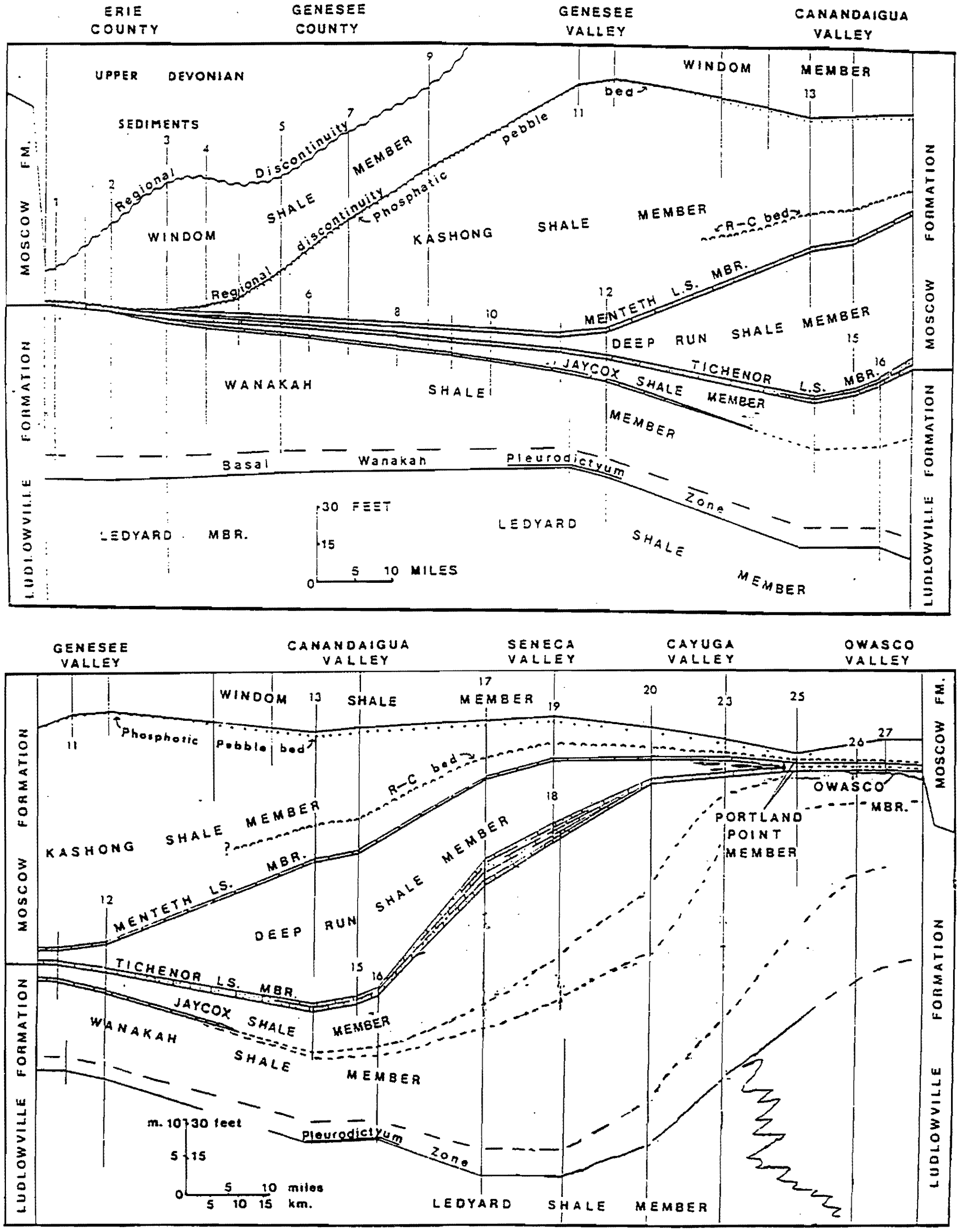

Figure 5. Regional east-west cross sections of upper Ludlowville and lower Moscow Formations. A) Western panel, Erie County to Canandaigua; B) Eastern panel, Genesee Valley to Owasco Lake. See figure 3 for approximate cross section lines. Numbered locations are described in Baird (1979).

Interpretation of the Deep Run and Menteth Members: The Deep Run-Menteth succession is interpreted as representing two intermediate-scale, shallowing-upward cycles. The first begins above the Tichenor Limestone flooding surface and culminates in the lower Deep Run unnamed limestone. This interval, contains sparsely fossiliferous, calcareous mudstone at its base and passes upward into fossiliferous, calcareous mudstone and limestone. It is interpreted as a small-scale, shallowing-upward succession or parasequence. The second cycle commences with the flooding surface above the unnamed lower Deep Run limestone and passes upward into the Menteth Limestone. Shallowing was accompanied by an increased influx of muds and silts from eastern source areas. The Menteth Member, which caps the cycle, is locally a calcareous siltstone.

There is some ambiguity to the interpretation of the Menteth capping beds. In particular, a sharp contact which separates the limestone from underlying Deep Run mudstone on both eastern and western margins of the central Finger Lakes trough or basin suggests that the carbonate should not be thought of as occurring in the shallowing or progradational phase of a cycle, but rather subsequent initial deepening phase. The carbonate overlaps a regionally, slightly angular beveled surface, which therefore more closely resembles a boundary of a small-scale sequence than a marine flooding surface of a parasequence.

Kashong Shale Member

General Description: The Kashong Shale Member was recognized by Cooper (1933) for exposures along Kashong Glen east of Bellona and about 1 kilometer west of Seneca Lake. More recently, Lukasik (1984) provided a detailed, updated study of Kashong member stratigraphy, sedimentololgy and paleoecology (Figures 5,6). At its type locality, the Kashong is a soft, bluish gray mudstone with some thin but regionally widespread limestones and nodular concretionary horizons. The Kashong interval thickens to a maximum of about 25 meters (80 feet) in the Genesee Valley and thins westward to a feather-edge in Erie County; the Kashong also thins towards the southeast and east (Figure 6). The Kashong somewhat resembles the Deep Run in lithology, but it is a softer, less calcareous, and typically less silty mudstone.

The Kashong mudstones carry a very distinctive fauna dominated by the concavo-convex orthid brachiopod *Tropidoleptus carinatus,* as well as *Mucrospirifer*, *Nucleospira concinna*, *Devonochonetes coronatus*, the tabulate coral *Pleurodictyum americanum* relatively large bivalves such as *Orthonota undulata*, and the trilobite *Dipleura*. Component limestone beds contain a much more diverse biota, dominated by small twig-like bryozoans such as *Taeniopora* and *Sulcoretepora*, crinoids as well as a variety of brachiopods, especially rhynchonellids.

Lower Kashong: The Kashong Member can be subdivided into three intervals, of which the lower usually constitutes a third or less of the total thickness of the Kashong (Figure 6). The lower division of the Kashong Shale consists of about 0.5 to 3 meters of bluish-gray mudstone with abundant small concretions in its upper third. The base of this interval sharply overlies the Menteth Limestone, but the basal 30 centimeters. contains a highly diverse fossil assemblage, including abundant large crinoids, bryozoans, brachiopods, and bivalves. Overall, the fauna closely resembles that seen in the more fossiliferous beds of the older Deep Run Member. The remainder of the lower division is more

The Fisher Gully interval displays a reciprocal thickness pattern with the underlying Bear Swamp beds--thinning to the southeast as the latter interval thickens (Figure 9). For example, the Fisher Gully interval is 7 m-thick at Bloomer Creek but only 3 m (10 ft) at Willow Point, along the west side of Cayuga Lake. Fisher Gully beds display a sharp contact with the underlying beds, which, as noted above, is locally an erosional unconformity that has removed substantial portions of the Bear Swamp beds. This basal contact is typically overlain by a thin, cryptic shell hash bed that displays a concentration of varied fossils reworked from underlying shales, hiatus concretions, and even reworked tubular pyrite clasts in at least one locality. This reworked debris is variably commingled with the typical fauna of the Amsdell Bed which includes specimens of the ambocoeliids *Emanuella praeumbona*, *Ambocoelia umbonata*, and the leiorhynchid *Eumetabolotoechia multicosta*. The overlying 1 to 2 m of shale are typically quite dark gray and platey and range from nearly barren to highly fossiliferous on certain bedding planes. The latter are typically crowded with nearly monospecific assemblages of *E. multicosta*. However, toward the upper portion of the interval, thin beds rich in *Emanuella praeumbona* appear at several levels, together with a somewhat more diverse fauna that may include chonetids, small *Athyris*, rare *Mediospirifer*, various nuculid and modiomorphoid bivalves, the cephalopod *Spyroceras* and various gastropods. The trilobite *Phacops rana* may be present as well. One or two beds near the middle of the interval are also rich in diminutive specimens of the brachiopod *Allanella tullius*. In the eastern Finger Lakes area, this interval becomes silty medium gray and may contain small ellipsoidal concretions. To the west, near Canandaigua Lake and the Genesee Valley, the interval again tends to be medium gray, loses its platy nature, and contains more abundant *Emanuella* and other brachiopods and fewer specimens of *E. multicosta*.

The upper boundary of the Fisher Gully beds varies from apparently gradational in areas around Seneca and Cayuga Lakes and eastward to very sharply defined in the Canandaigua to Genesee Valley area where the Fall Brook coral bed rests with sharp and apparently unconformable contact on the underlying *E. praeumbona*-rich shales.

No trace of the *E. praeumbona*-rich interval has been found in outcrops in Genesee or eastern Erie Counties. However, the apparently corresponding interval reappears in western Erie County (Figure 8). Here, the *Emanuella praeumbona*-rich zone is contained within an interval, up to about 2.5 m in thickness of medium gray, light weathering, highly calcareous and somewhat petroliferous mudstone. Very argillaceous limestone occurs at the base of the interval and near the top of the Windom Member in outcrops between Cazenovia Creek and the Lake Erie Cliffs near Hamburg. This bed, previously termed the "*Praeumbona* bed" by Grabau (1898, 1899), and more recently the Amsdell Bed (Brett and Baird, 1982), is a highly distinctive marker in Erie County. Its lithology somewhat resembles the basal part of Unit 11 in the Canandaigua-Genesee Valley area. The Amsdell bed is believed to have been formerly coextensive with the lower *E. praeumbona* dark shales of the Finger Lakes area (Figure 8b). However, the absence of this bed in Genesee and eastern Erie counties is accounted for by the presence of two unconformities. In previous work, we have documented the northeastward removal of the Amsdell bed in Erie County as a result of the Windom-Genesee unconformity (Brett and Baird, 1982). This contact was demonstrated to be regionally angular, truncating not only the Amsdell bed, but also underlying mid Windom shale units in the eastern Erie County and Genesee outcrop area. However, this disconformity surface does not account for the absence of the Amsdell bed in eastern Genesee and western Livingston County areas west of the Genesee Valley. In these areas, the Fall Brook bed (Unit 12) is still present below the Windom-Genesee unconformity. Consequently, it is necessary to postulate that a lower erosion surface beneath the Fall Brook coral bed has locally beveled out the Fisher Gully strata (Unit 11) in the area of northwest of the Genesee Valley.

Indian Creek Bed and "Willard Channel": A highly distinctive crinoidal pack- and grainstone bed, up to 20 cm-thick, occura along Indian, South Indian, and Simpson Creeks in the town of Willard on the east side of Seneca Lake. For this bed, the informal term Indian Creek bed is proposed herein. At these localities, especially along South Indian Creek, the bed is compact, highly fossiliferous, pyritic, packstone layer containing extremely abundant, corroded crinoid ossicles, *Stereolasma* (rugose corals), and brachiopods, as well as reworked, although non-encrusted, concretions, derived from the underlying shales. This compact bed, which locally resembles the Tichenor Limestone, is obviously lenticular. No similar thick bed has been located beyond the east Seneca Lake region. At South Indian and Simpson Creeks, the base of the bed is razor sharp and overlies an erosion surface that has beveled out units 4, 5 and 6 and cut down into concretionary levels at the top of the *D. coronatus* interval (Figure 9). Indeed, the upper concretions protrude into the bed and have been reworked as isolated clasts within the base of the skeletal pack- and grainstone. This is the lowest of three important mid-Windom erosion surfaces which display complex relationships with respect to one another. However, the erosion at this level was relatively localized, evidence being limited to the area of thick development of the Indian Creek bed around Willard.

The Indian Creek bed is overlain in the Willard area by 2.1 to 3.9 m (7 to 13 ft) of sparsely fossiliferous, medium gray shale with a distinctive satiny sheen. The lowest 0.5 m of the shale contains thin stringers of crinoid debris; upper beds carry chonetids and rare *Eumetabolotoechia* and *Emanuella*. The interval is capped by a thin brachiopod-rich bed that may be coextensive with the lag bed at the base of the Fisher Gully beds near Cayuga Lake.

Detailed measurements of this shale indicate that it is lenticular and probably fills a channel-like depression cut into the lower Windom Shale (Figure 9). The stratigraphic affinities of this shale are somewhat enigmatic; we offer two possible alternatives in Figure 9. The first model (Figure 9a) assumes that the Indian Creek bed is correlative with the mid-Windom conodont bed and that the shell bed above the "satiny shale" is the base of the Fisher Gully interval. Hence, the "satiny shale" is a local expression of the Bear Swamp shale.

The alternative correlation (Figure 9b) implies that the Indian Creek bed is a local thickening of the basal Fisher Gully shell lag; the channel-like erosion surface would be coextensive with the discontinuity beneath the Fisher Gully beds documented in the vicinity of western Cayuga Lake. By this interpretation, the "satiny shale" would be a locally expanded facies of the Fisher Gully beds and the shell bed at the top would be a minor bed within the Fisher Gully interval. Resolution of this issue will require still more detailed study.

Unit 12. Fall Brook Coral Bed: The Fall Brook coral bed was defined and described by Baird and Brett (1983) for an interval of approximately 0.5 to 2 m in thickness overlying the *E. preaumbona*-rich dark gray (Fisher Gully) shales in the Genesee-to-Canandaigua and Seneca Lake area (Figure 8). This interval is characterized by an abundance of large rugose corals, especially *Cystiphylloides*, *Heliophyllum*, and *Heterophrentis*. An extremely diverse associated fauna, comprising over 70 species of brachiopods, bryozoans, small-to-medium sized tabulate corals, mollusks, trilobites, and echinoderms was also discussed in detail by Baird and Brett (1983) and a regional gradient of faunal change was also noted in that publication. The Fall Brook bed contains the greatest number of corals and displays the highest proportion of skeletal debris in the more western sections. To the east of the Genesee Valley, the Fall Brook bed displays a marked decrease in the abundance of larger rugose corals, but an increase in overall faunal diversity.

The easternmost locality at which the Fall Brook bed can be readily recognized as a discrete interval is along Perry Ravine on the west side of Seneca Lake south of Dresden, New York. Here it is a somewhat thicker interval of fossiliferous mudstone (up to 3 m) that contains abundant small corals, such as *Amplexiphyllum*, as well as a high diversity of brachiopods, including forms such as *Tropidoleptus* which are rare or absent in the western sections. Also, accompanying an increase in thickness and decrease in density of fossils is the appearance of greater numbers of the trace fossil *Zoophycos* and semi-infaunal bivalves such as *Modiomorpha.* The Fall Brook bed is readily recognizable as a highly fossiliferous zone westward to Linden, Genesee County, where it occurs immediately below the Leicester Pyrite and succeeding Geneseo Shale. Westward of this locality regional truncation at the Moscow-Genesee contact has removed the Fall Brook bed and underlying strata downward locally as far as the upper Bay View beds interval (Figure 8). Although units 10 and 11 reappear beneath the unconformity in southwestern Erie County outcrops, the Fall Brook bed does not reappear in this area (Figure 8). Evidently the highest beds exposed just beneath the unconformity in areas of western Erie County belong to the *Emanuella praeumbona*-rich Unit 11.

It is notable that the latter interval had been removed by an unconformity below the Fall Brook bed in areas of eastern Genesee County prior to erosional truncation of the Fall Brook bed and subjacent shales by the post-Windom unconformity. However, in Erie County the unconformity beneath the Fall Brook bed, if ever present, must have been minor, as evidenced by the reappearance of Amsdell (Unit 11) bed, strata in that area.

To the east, the Fall Brook bed displays a less distinct basal contact, and eventually, near Seneca Lake, the bed appears to be conformable with underlying mudstones. At the same time, the bed loses its compact nature and splays into a series of thin, brachiopod and small coral-rich beds. These beds are exceptionally rich in *Mediospirifer* and *Athyris*, but also contain relatively abundant small corals, including *Pleurodicytum*. Intervening shales are relatively dark gray and may even contain specimens of the leiorhynchid *Eumetabolotoechia multicosta* This is a curious mixture of faunas as it is most unusual for small rugose corals to occur with leiorhynchids in the Hamilton Group. Cleland (1903) referred to this interval as a "transition zone" to his "*Spirifer-Atrypa*" Zone. Grasso (1966) and Zell (1985) traced this interval of *E. multicosta* eastward through the Tully and Chenango Valleys. Moreover, the Fall Brook interval becomes exceptionally pyrite-rich in outcrops in the Cayuga Valley. Relatively finely crystalline but euhedral pyrite occurs as crusts on shell, as irregular lumps and nodules, and occasionally as infillings in fossils; pyritized wood is relatively common in this interval. One or more levels of relatively large septarial concretions are also observed within the Fall Brook interval. Shell beds within the interval appear to become slightly more diverse upward. Beds 1.5 to 3 m above the apparent base of the Fall Brook interval contain rare specimens of the brachiopod *Spinatrypa*, which is generally common to the west in the Fall Brook bed. However, large rugose corals have never been found within this interval east of Seneca Lake.

The upper contact of the Fall Brook bed with the overlying Taunton beds is relatively distinct in the Genesee Valley area, where it is marked by a fairly abrupt change to moderately fossiliferous blue gray mudstones. However, this upper contact, like the lower, becomes diffuse and indistinct towards the east.

Unit 13. Taunton Beds: In the Genesee Valley area, the Fall Brook Bed is overlain by an interval, up to 3 m in thickness, of moderately fossiliferous bluish gray *Zoophycos*-burrowed mudstone with abundant large fossiliferous concretions (up to 30 and 40 centimeters in diameter). This interval has been termed the Taunton beds interval for excellent exposures in Taunton Gully north of Leicester (Baird and Brett, 1983). The lowest portion of the Taunton beds succession, directly overlying the Fall Brook coral bed tends to be blocky, sparsely fossiliferous mudstone, somewhat resembling the older Kashong Shale; like the Kashong the lower portion of the Taunton beds typically contains brachiopods such as *Tropidoleptus* and *Orthospirfer (?) marceyi*. Upward, the Taunton interval displays a general increase in thin (1 to 2 cm-thick) shell and bryozoan-rich beds.

Parsons et al. (1988) were able to correlate these thin shell hash beds at least regionally within sections of the Genesee Valley. Distinctive horizons in which shell layers are locally incorporated into the carbonate concretions, as well as pyritic shell rich beds provide useful regional markers. Crinoid columns and calyces also become abundant within this portion of the Taunton beds. Large camerate crinoids, such as *Megistocrinus*, *Dolatocrinus*, occasional blastoids, and inadunates occur sporadically in this interval. In the medial part of the Taunton interval beds of large (0.5 m) concretions typically display an abundance of crinoid and bryozoan material. This interval is the source of large clusters of crinoids, particularly *Clarkeocrinus*, found in the Bristol Valley and described in detail by Goldring (1923). This *Clarkeocrinius*-rich horizon, typically associated with fistuliporoid bryozoan mounds has now been traced from the Genesee Valley to the east side of Seneca Lake. It appears to be a widespread obrution or rapid burial horizon. It also represents an epibole, in that it displays a considerable abundance of the typically very rare species such as *Clarkeocrinus troosti*.

The upper Taunton beds are rich in the brachiopods *Pseudoatrypa*, and *Mediospirifer*, the small rugose corals, *Stereolasma* and *Amplexiphyllum*, the bryozoan *Sulcoretepora*. This fauna corresponds to Cleland's (1903) "*Spirifer-Atrypa* zone" in the Cayuga Lake region. The highest beds also are exceptionally rich in fenestellid bryozoans. Near Seneca Lake, this interval displays corals such as *Thamnoptychia* and occasional large rugosans such as *Heliophyllum*.

Eastward from outcrops near Seneca Lake, the Taunton interval becomes increasingly silty and displays a coarsening-upward cycle from soft, burrowed mudstones upward to heavily *Zoophycos* churned silty mudstones and/or siltstones. In sections east of Cayuga Lake, the creeks typically display a cascade or waterfall over these more resistant, silty upper Taunton beds.

Although the Fall Brook interval at the base of the Taunton submember becomes extremely diffuse and difficult to distinguish, the upper contact becomes increasingly sharp and is marked by a compact argillaceous limestone carrying abundant rugoans and the large tabulate coral, *Favosites hamiltoniae*. These beds, particularly well displayed in the area around Portland Point Quarry, South Lansing were termed the South Lansing coral bed by Baird and Brett (1983).

This South Lansing coral rich limestone bed displays a relatively abrupt lower and upper contact. This bed appears to correlate with a widespread coral-rich zone within the upper Mahantango (Sherman Ridge) shale in Pennsylvania (Ellison, 1966).

As the Taunton interval becomes increasingly silty, its fossil content decreases and becomes much more sparse, corresponding with a great increase in the abundance of *Zoophycos*. The increase in silt content is also accompanied by loss of the large, rather tabular horizontal concretions so typical of the Taunton, particularly the middle Taunton interval to the west.

The eastern limit of the Taunton beds interval is presently unknown; this highly silty portion of the Windom is distinctive and caps small waterfalls eastward at least to the Chenango Valley. To the west, the Moscow-Genesee unconformity has removed upper Windom beds, and the Taunton interval is missing at this erosion surface west of the Genesee Valley (Figure 8).

Unit 14. Spezzano Gully Beds: The Taunton interval is overlain in the Genesee Valley area by a series of calcareous shales and very argillaceous, somewhat concretionary to tabular limestone beds. This bundle of limestone and shale beds, up to about 1.5 m in thickness, is named the "Spezzano Gully beds" for exposures along Spezzano Gully near the town of Retsof, New York. The Spezzano Gully interval is rich in the small rugose corals, *Stereolasma* and *Amplexiphyllum*, and, in some layers, contains an abundance of partial, enrolled, and complete outstretched specimens of the trilobites, *Phacops rana* and *Greenops spp*. As such, the interval quite closely resembles the older Smoke Creek bed (Unit 6) of the lower portion of the Windom. The Spezzano Gully interval is roughly subdivisable into two parts near the type section. The lower portion contains a series of three tabular very sparsely fossiliferous argillaceous limestones, which can be correlated, at least, within the Genesee Valley Region. These closely resemble a bundle of three tabular micritic limestones found in the mid-Windom (Bear Swamp) interval in Erie County. Previous incorrect assumptions about the stratigraphy had led us to correlate the Spezzano Gully beds with these mid-Windom limestones. Shales immediately above and between these calcareous ledge-forming limestones carry an abundance of *Pseudoatrypa*, *Protodouvillina*, and small *Stereolasma* corals. The lower two ledges, appear to merge into a single calcareous band in the area of Fall Brook (Parsons et al., 1988). The upper portion of the type Spezzano Gully interval not does contain discrete carbonate beds, but is a rather platy, calcareous mudstone containing sparse, but commonly well preserved trilobites, scattered brachiopods, and even rare blastoids. The lower boundary of the Spezzano Gully interval is marked by abundant bryozoan-rich shell hashes which may be the stratigraphic equivalent of the South Lansing coral beds. The upper boundary is marked sharply in most localities by a thin, but very distinctive shell hash bed (Unit 15). To the east, in the central Finger Lakes area, the Spezzano Gully interval loses its discrete carbonate beds. However, it remains a distinctive fossil-rich mudstone unit.

The Spezzano Gully interval in west central New York rarely contains the brachiopods *Pustulatia*. This interval may be correlative with the well known "*Pustulatia* beds" high in the Sherman Ridge Member of the Mahantango Formation in Pennsylvania (Ellison, 1965). Like the Taunton interval below, the Spezzano Gully beds are truncated to the northwest of the Genesee Valley and do not reappear in any sections west of this area.

Unit 15. Simpson Creek Bed: The Spezzano Gully interval is everywhere abruptly terminated at a sharp contact with the overlying dark gray uppermost Windom Shale interval. This contact is overlain by a thin (1 cm-thick), yet highly persistent, shell-rich bed (Parsons et al., 1988). This bed, named for its excellent development along Simpson Creek near the Willard Psychiatric Center on the east side of Seneca Lake, is particularly rich in the small brachiopod *Emanuella praeumbona*. This brachiopod is otherwise seen only in the Fisher Gully (Amsdell) beds interval of the mid-Windom succession (Units 12 and 13). In the Simpson Creek bed, it is commonly mixed with abundant fragmentary brachiopod material, crinoidal hash and other debris. At Willard, the bed contains some pyritized nuculid bivalves and gastropods, including the high-spired forms *Palaeozygopleura* and *Glyptotomaria*. The bed locally carries evidence of being a lag deposit with some evidence of local scouring and minor removal of subjacent beds. This thin bed has been recognized from the Genesee Valley eastward to at least Barnum Creek on the west side of Cayuga Lake.

Unit 16. Gage Gully Beds: The highest strata of the Windom exposed in Western New York are medium to dark gray, typically somewhat petroliferous, chippy shales (Figure 8). These shales are characterized by a very distinctive fauna that includes small specimens of *Ambocoelia (?) nana*, and especially of the spiriferid brachiopod "*Allanella*" *tullius*. Small mollusks, especially nuculoid bivalves, gastropods, the nautiloid *Spyroceras*, and rarely *Tornoceras* are also present to abundant within the Gage Gully beds. In central New York, Grasso (1966) referred to this interval and perhaps the underlying Spezzano Gully beds, as the *Allanella-Pustulatia* zone.

This interval is named for one of its areas of maximum development, along Gage Gully on the east side of Canandaigua Lake. In this area, it is about 5-6 m-thick and consists of dark gray to nearly black, platy shale. To the west, the interval thins towards the Genesee Valley where it is at most about 2 m in thickness and is a medium dark gray mudstone containing an abundance of *Greenops* trilobites, small mollusks, and lesser numbers of *Allanella tullius*. Pyritized nodules, burrow tubes and fossil steinkerns, including those of large *Tornoceras*, occur sporadically in the upper portion of the Gage Gully beds. To the east, the interval is again truncated east of Canandaigua Lake such that it is only 2 meters thick at Kashong Creek at Bellona, less than 0.5 m in creeks along the northwest side of Cayuga Lake and absent altogether in the Owasco Lake-Ithaca region. However, the shale reappears eastward, being about 2 m-thick at Skaneateles Lake and up to 3 m-thick in the Tully-Chenango Valley region, where it is again a relatively dark gray, slightly rusty, weathering pyritic shale. Where present, the Gage Gully beds are everywhere capped by a marked disconformity flooring either the Tully Formation or the younger Genesee Formation, where the Tully is absent. This is a regionally angular bevel-surface which has removed the Gage Gully interval in the central Finger Lakes Region. Hence, the Gage Gully beds are the highest beds ever seen in the Windom within western and west central New York State. However, to the east in the

Chenango Valley area, still-higher gradational silty mudstone units occur at the top of the dark shale equivalents, recording the next shallowing cycle prior to the deposition of the Tully Limestone or its correlatives.

SEQUENCE AND CYCLIC INTERPRETATION OF THE MOSCOW FORMATION

Lower Moscow Formation

Overall, the Tichenor-to-top-Kashong interval which ranges from slightly over a meter on either side of the basin center to approximately 34 meters in the Canandaigua Lake region is inferred to represent a third-order transgressive systems tract (Figure 10). This package displays a generally upward-deepening pattern culminating with the *Ambocoelia*-rich mudstone interval of the basal Windom immediately above the Little Beards Creek phosphatic bed. However, it should be clear that this was a stepwise process and that it involved at least five lesser-scale shallowing-deepening cycles. These cycles, including two in the Deep Run and three in the Kashong Member, range from nearly symmetrical at the depocenter, for any given interval, to markedly asymmetrical toward the basin margins. In the latter areas, the cycles resemble small-scale sequences, in that their bases are sharply defined at erosion surfaces which underlie each of the thin condensed carbonate units, (i.e.; Tichenor, unnamed Deep Run limestone, Menteth, RC bed, upper Kashong siltstone, and Barnes Gully phosphatic bed). The latter beds probably formed at or near maximum lowstand in the basin center, but are slightly diachronous such that on the lateral ramps of the basin the carbonates formed during the immediately ensuing transgression and sediment-starved interval. The carbonate beds are of two types. First are skeleton-rich beds containing an abundance of fossil debris that include corroded brachiopods and corals. This type characterizes the Tichenor, unamed Deep Run limestone, RC bed and the Barnes Gully phosphatic bed. The alternate lithology is highly bioturbated to slightly laminated silty carbonate or calcisiltite represented by the Menteth and the upper Kashong unnamed silty limestone (Figure 10). It should be noted, however, that the latter lithology can assume the more shell-rich facies aspect near the basin center for these various limestones. In particular, the Tichenor limestone comes to take on this facies in the region around Canandaigua Lake.

Each of the limestones also displays a sharp upper contact with overlying shales which is interpreted as a flooding surface. Hence, the limestones appear to be small-scale analogs of transgressive systems tracts for each of the smaller-scale cycles. The overlying shaley beds (for example, the main body of Deep Run shale, lower and upper Kashong mudstones) constitute relative highstand to regressive deposits. The succession of five lesser cycles within the Tichenor Limestone-Barnes Gully bed interval displays a progressive relative deepening pattern with the upward passage of each cycle characteristic of an overall transgression with superimposed minor fifth-order cycles.

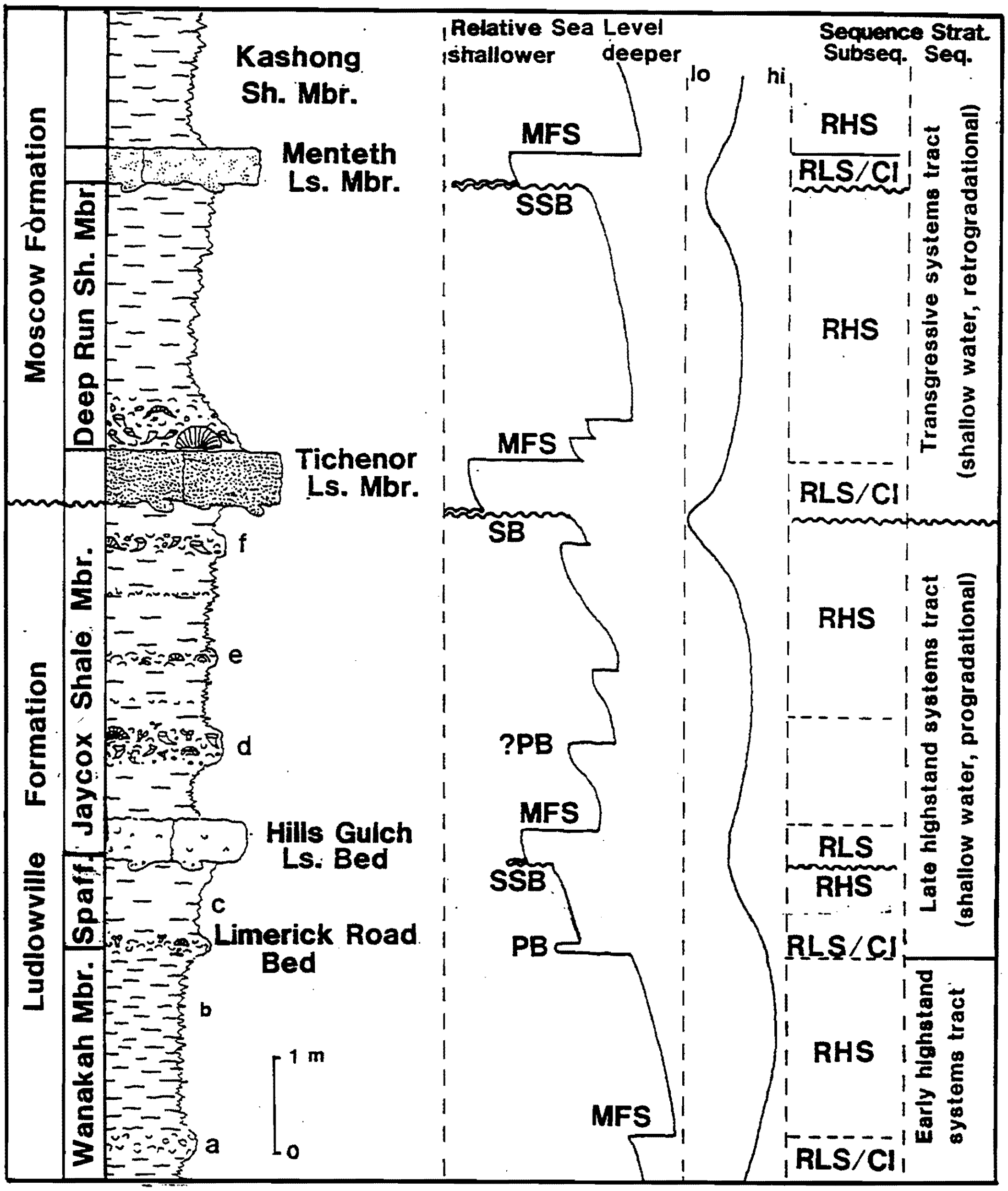

Figure 10. Sequence interpretation of the upper Ludlowville and lower Moscow Formations in the Genesee Valley area. Note sequence boundary at base of Tichenor Limestone. Symbols: RLS = relative lowstand; RHS = relative highstand; CI = condensed interval; MFS = Marine flooding surfaces; SSB = subsequence boundary; PB = precursor bed;

Windom Member Cycles

Parts of the Windom Shale appear to represent some of the deepest-water conditions within the upper part of the Hamilton Group. For this reason, and because it also shows a shoaling-upward pattern, as well as abrupt juxtaposition on shallow-water, somewhat condensed carbonates, the Windom is interpreted in the broadest sense, as the highstand systems tract of the upper Moscow depositional sequence. The informal "Unnamed member" represents a genetically-related transgressive systems tract separated from the Windom proper by a surface of maxiumum starvation (Figures 2,9).

However, it is equally clear that the Windom is subdivisible into a series of four cyclic packages (Figure 11), probably corresponding to fifth order cycles, or cyclothems (Busch and Rollins, 1984). These are, in ascending order: a) dark gray *Ambocoelia*-rich shales which pass upward through the *Devonochonetes coronatus* beds, and into the Bay View coral bed and Smoke Creek beds (Units 1 through 6); b) a previously unrecognized middle Windom cycle in western and central New York State comprising units 7 through 10 (i.e., the Bear Swamp chonetid- and *Ambocoelia*-rich shales) which pass upward to thin, calcareous, shell-rich beds overlain by diminutive brachiopod faunas (Penn Dixie pyritic beds); c) the Fisher Gully dark gray, *Ambocoelia*-and *Emanuella*-bearing shales (including Amsdell beds) upward through the Fall Brook coral horizon and overlying Taunton and Spezzano Gully beds (Units 11-15); d) an incomplete cycle comprising the uppermost Windom dark shales (Gage Gully beds) and hints of a final upward-shallowing package which is largely truncated by the sub-Tully (or sub-Genesee) erosion surface (Units 16, 17).

These cycles are widespread and even appear to be correlative from New York State into the upper Mahantango Formation of central Pennsylvania. For example, at Milesburg, Pennsylvania, exposures along the westbound entrance lane to Interstate 80 display a succession that appears to start in the *Devonochonetes cornatus* beds, continue up through the fossil-rich Bay View beds, through a sparsely fossiliferous, dark gray, middle Windom interval, and finally into beds associated with the Taunton beds-Gage Gully beds succession in New York. As such, this is the most complete Windom succession yet recognized in central Pennsylvania. Future work will attempt to extend these correlations.

Despite differences in detail, each of the Windom cycles shows several common elements. As noted below, these include a) basal lag beds, b) ambocoeliid shales, c) coarsening upward silty mudstones and siltstones, and d) thin, calcareous "trilobite beds" near the cycle top (Figure 11).

Basal Lag-Maximum Flooding Surface: The cycles begin with very thin condensed shell-rich beds: a) Little Beards - Geer Road phosphate, shell-rich interval, b) mid Windom conodont bed, c) unnamed basal lag of Unit 11; and, d) Simpson Creek bed (*E. preumbona*-rich phosphatic lag horizon). These inconspicous, thin (typically < 1 cm-thick) and widespread condensed horizons are interpreted as surfaces of maximum sediment-starvation associated with relatively rapid sea level-rise events. Presumably, these events flooded the shoreline and produced conditions of extreme siliciclastic sediment-starvation in offshore areas. Thin lag beds of variably disarticulated, fragmented, and

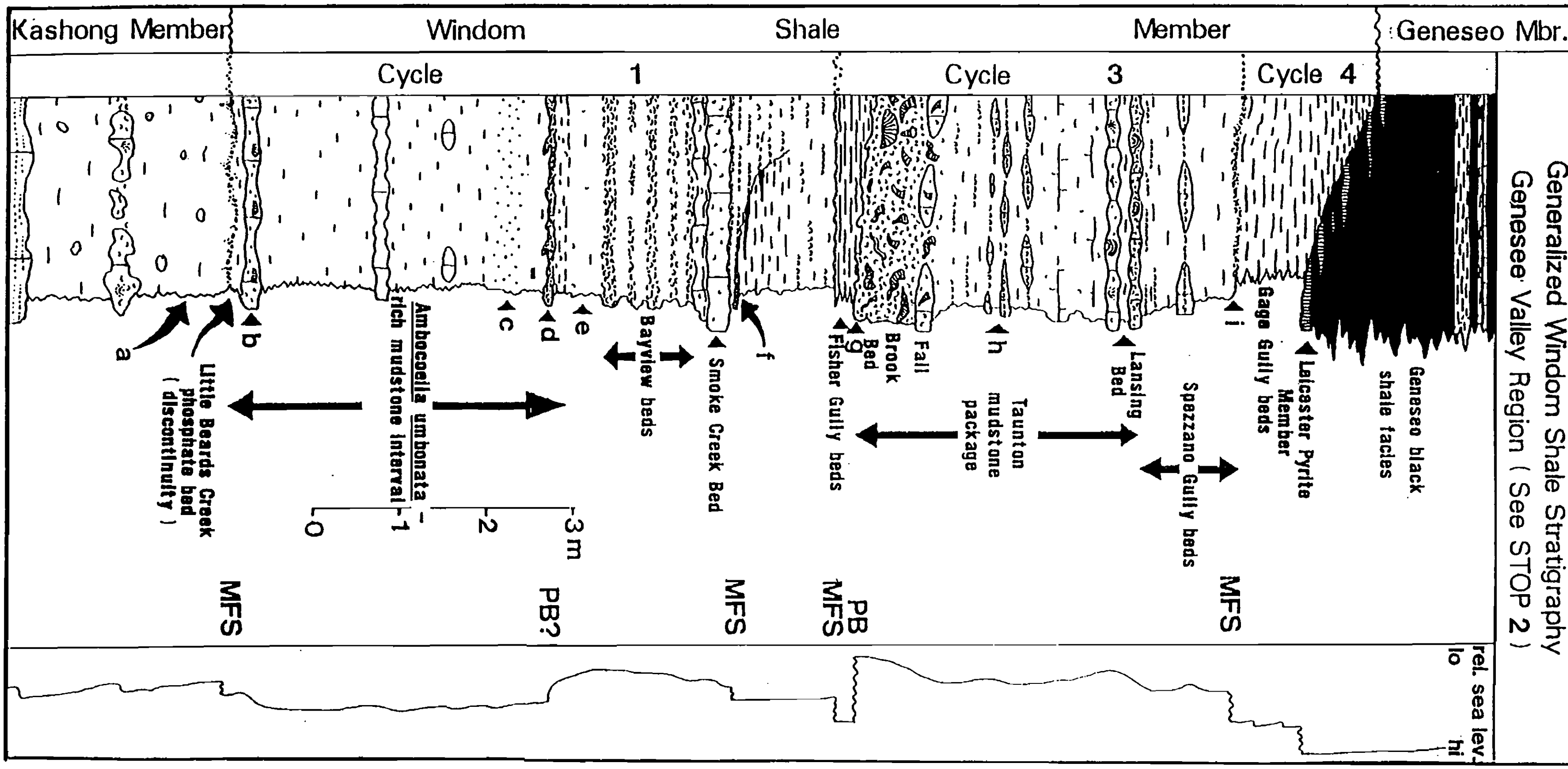

Figure 11. Windom stratigraphy generalized for the Genesee Valley region (Fall Brook, Little Beards Creek, Taunton Gully and Spezzano Gully sections). Inferred sea level changes and sequence stratigraphy units are shown on the right. Windom component cycles are also shown. Lettered units include: a) uppermost Kashong Member (phyllocarid epibole); b, *Phacops*, *Ambocoelia*, and demosponge-rich concretionary limestone bed; c, mudstone interval rich in pyritic nodules, *Phacops*, pyritized *Tornoceras*, *Athyris* and *Ambocoelia*; d, bed composed of concentrated *Ambocoelia* and displaying *Ambocoelia* filled hypichnial burrows; e, mudstone-rich in *Devonochonetes coronatus* and distinctive amplexiphylloid coral ("*D. coronatus* epibole"); f, erosional lag deposit composed of abundant conodonts, bone fragments and shell debris which is associated with discontinuity surface; g, discontinuity at base of Fall Brook Coral bed; h, bed yielding *Clarkeocrinus troosti* ("*Clarkeocrinus* epibole"); i, Simpson Creek transgressive shelly lag deposit rich in *Emanuella praeumbona*. Abbreviations as in Figure 10.

corroded skeletal debris, phosphatic nodules, and/or reworked concretions are mixed with better preserved shells representing a distinctly more offshore, dysaerobic community. Moveover, there is evidence associated with each of these beds for an interval of submarine erosion. The erosion appears to have produced minor discontinuities and in some cases possible furrowing of the underlying muddy sea floor.

Ambocoeliid Shales - Early Highstand Phase: The basal lag beds are overlain by intervals of medium to dark gray shales carrying diminutive brachiopod assemblages. In the lower three cycles *Ambocoelia umbonata* and two to three species of small chonetid brachiopods dominate the assemblages within these intervals; in the upper two cycles *Emanuella praeumbona*, diminutive *Allanella*, *Tropidoleptus*, and *Eumetabolotoechia multicosta* are mixed with less common *Ambocoelia* and chonetids. These intervals appear to represent somewhat more dysoxic conditions than those seen in underlying cycles. The ambocoeliid shales are interpreted to have accumulated under deepest water portions of each cycle.

Silty Mudstone Late Highstand-Progradational Phase: The ambocoeliid-bearing medium to dark gray shaley mudstones are in each case overlain by intervals of calcareous to silty mudstone. These appear to represent progradational shallowing intervals during which coarser siliciclastic sediments were transported further offshore than during most of the rest of the time interval. These silty sediments in the eastern Finger Lakes area generally appear to be transitional upward from the underlying medium gray mudstones.

However, in the case of the third cycle (Fisher Gully to Spezzano Gully beds succession) the appearance of this somewhat shallower water biofacies is abrupt, at least in western New York, and is marked by the highly fossiliferous, condensed Fall Brook coral bed. This interval displays a particularly interesting regional pattern. As noted, it overlies a sharp discontinuity that cuts out portions of the subjacent middle Windom beds in western New York State. However, to the east, the bed splays out into a series of thin and much less spectacular shell horizons and the base becomes transitional.

The Fall Brook bed is representative of a category of somewhat condensed skeletal debris-rich horizons that we have termed "*precursor beds*" (Brett and Baird, 1990). These are condensed beds that occur abruptly at the top of the early highstand shale intervals and at the bases of upward-coarsening parts of cycles. The Fall Brook bed appears at the base of the Taunton beds interval that reflects upward-shallowing conditions. To the east, this interval is capped by resistant *Zoophycos*-bioturbated siltstone beds which bear at their top a coral-rich interval.

The precise causes of sedimentary condensation associated with the beginning of upward-shallowing cycles remains rather enigmatic at this time. Precursor beds may reflect minor-sediment starvation occuring during an initial sea level drop (forced regression) due either to subaerial accommodation processes (i.e., regrading of streams to new equilibrium profiles with consequent reduction in detrital discharge of sediments to offshore regions; Posamentier et al., 1988), or the impingement of occasional deep-storm waves as a threshold of depth was obtained during a relatively rapid shallowing interval. If submarine erosion became prevalent prior to sediment progradation, then truncation of older beds would be expected. One of the curious features of the precursor beds is their tendency to downgrade in terms of faunal diversity and abundance in an upramp (easterly) direction. This eastward splaying of the precursor beds is probably the effect of greater proximity to the siliciclastic source area. Where the precursor beds are most condensed in the west, it is the result of the stacking of thin shell beds, each the cap of a minor shallowing- cycle, onto one another in a sediment-starved setting. Complex amalgamation of shell beds capping small-scale cycles produces the effect of the thin condensed precursor bed.

The precursor beds are overlain by a coarsening upward succession of mudstone to calcareous, highly fossil-rich siltstone deposits of the later highstand or progradational phase of the cycle. These intervals are typically heavily churned by *Zoophycos* and contain scattered lenticular brachiopod- and coral-rich shell debris horizons as well as concretions. The abundance of concretions in the upper part of this interval probably reflects a relative slowdown in sedimentation toward the end of the cycle; this enhanced diagenetic reactions. Each concretion horizon presumably reflects a minor episode of sediment cut-off to the western part of the basin. During such times, burrow peripheries became enriched with bicarbonate and sulfides, leading to growth of carbonate or pyrite concretions. Modern concretions, about 10 centimeters in diameter are thought to form within sediments during times of overall sediment-starvation and stability of the zone of sulfate reduction. (Hallam, 1986; Raiswell, pers. comm., 1993). Such nodules appear to have formed during a 3 to 5 thousand year period of general sediment cut-off.

In the case of Bay View and South Lansing coral beds, diverse brachiopods, bryozoans, crinoids and, in some cases, larger rugose and tabulate corals flourished over wide areas as a result of shoaling into the zone of turbulence, greater sunlight, and oxygen level. These organisms clearly also thrived in a sediment-starved regime associated with initial sea level-rise and or winnowing during peak sea level lowstand.

Trilobite Beds: Transgressive Phase : Finally, in each of the cycles, the early transgressive phases overlying thecapping shell-coral beds are expressed as thin, "trilobite beds" facies. This facies consists of variably fossiliferous intervals of thin bedded concretionary limestone (e.g., Smoke Creek and Spezzano Gully beds; Figure 11). The beds in these intervals are typically enriched in small rugose corals, and/or auloporids, brachiopods, including *Pseudoatrypa*, and trilobites. Trilobites are commonly articulated and may be found in clusters of complete and molted skeletal parts giving rise to the name "trilobite beds" for this facies. These calcareous, trilobite and small coral-rich layers appear to reflect times of minimal input of siliciclastic sediments. Traced in an eastwardly direction they appear to merge laterally into shell beds which overlie major progradational cycle. The enrichment of carbonate in the pore spaces of the sediment probably reflects prolonged stability the sediment-water interface, which permitted development of reducing, nonsulfidic conditions within the sediment.

These trilobite-rich, calcareous beds are one manifestation of marine sediment starvation associated with initial sea level-rise following a lowstand event. The trilobite bed interval represents a transition from regressive maximum to the maximum flooding surface for each cycle. The decimeter-scale nature of the bedding within these intervals is suggestive of a minor, but widespread climatic-eustatic signal. Either these small-scale fluctuations in carbonate content represent oscillation in relative sea level or variations in sediment-supply. The former contention is supported by lateral correlation of these small-scale cycles from western New York, where they are expressed as calcareous to concretionary mudstones alternating with shales, into central New York areas where cycles are manifested as minor (meter-scale) coarsening-upward successions. These latter asymmetric minor cycles are typically capped by siltstones and overlain by flooding surfaces, each marked with a thin shell-rich bed. These flooding surface beds are widespread and appear to correlate into the nodular concretionary carbonates of the western minor cycles (Brett and Baird, 1986).

This widespread nature of "trilobite beds" suggests rather uniform conditions over large tracts of the Devonian sea bottom. The calcareous and shell-rich nature of the beds over long distances further supports the notion of periodic episodes of widespread siliciclastic sediment-starvation. In central New York, these intervals are capped in each case by a sharp discontinuity which is overlain by a condensed, shell hash bed that begins the next larger (fifth order) cycle within the Windom Shale. These thin lag beds are associated with minor erosion surfaces and lie at the base of relative highstand deposits characterized by dark gray *Ambocoelia*-rich shale.

MOSCOW DISCONTINUITIES: PATTERNS AND PROCESSES

Careful stratigraphic correlation has revealed significant discontinuities within the Moscow Formation. The most prominent are associated with the major flooding surfaces, and thus appear to be associated with times of maximum sediment-starvation within the basin. They are overlain by the thin lag deposits at the bases of dark gray to black shale facies. These lag beds occur along sharp contacts at the tops of the underlying transgressive portions of the subjacent cycles. In several instances, portions of underlying cycles have been removed locally by these erosion surfaces. Although the regional geometry of these discontinuities is incompletely understood, it is clear that certain of these erosion surfaces locally cut more deeply than elsewhere and that these may represent broad channel-like scours in the top of the underlying mudstones (Figure 9).

Perhaps the most dramatic, but previously unrecognized example of such an erosion surface lies at the base of the Windom. As noted above, an erosion surface (Little Beards Creek Bed), which directly overlies Geer Road phosphatic bed of central New York, appears to truncate 3 to 4 m of underlying silty shale and concretionary limestone in the western Finger Lakes area (Figure 7).

At least two and perhaps three minor cycles are removed west of Canandaigua Lake by this erosion surface. The erosional truncation is particularly notable between Mud Creek in the Bristol Valley and Frost Hollow, 10 km to the west. Here about a meter of strata of the "Unnamed Member" is removed beneath a very cryptic discontinuity (Figure 7). These include, in descending order: a) the Geer Road bed; b, the "*Longispina-Mucrospirifer*-rich silty mudstone interval;" c, the Curtice Road bed; d, the "*Megastrophia* beds"; and, e) the Barnes Gully beds. This succession of beds, which is present and consistent from the central Finger Lakes west to the Bristol Valley is completely missing at Frost Hollow Creek, where typical *Ambocoelia*-bearing Windom Shale rests directly on Kashong mudstones below the level of the Barnes Gully shell bed. This contact is marked by the Little Beards phosphatic horizon. The source of the phosphatic nodules is apparently the Barnes Gully bed or the upper Kashong strata where the Barnes Gully bed is missing.

We postulate that the erosional process involved a combination of winnowing and dissolution which removed all fine grained siliciclastics and carbonates, but left a residue of phosphatic clasts on the erosion surface. Similar processes of erosive ablation have been described for reworked concretion beds by Baird (1981).

Two phosphatic-pyritic lag units in the middle Windom Member require further explanation. Because these beds occur at the sharp contacts of dark gray, typically dysoxic to anoxic shale facies, and contain materials such as reworked pyrite which would be unstable in a fully oxidizing environment, we have postulated a special mechanism of submarine erosion to account for the discontinuities. In part, the sharp surfaces relate to the high degree of sediment-starvation as evident from the overlying lag beds. For example, the high concentration of conodonts in a cryptic but important discontinuity above the Smoke Creek bed suggests a long interval of repeated winnowing, dissolution, and concentration of only the most stable,geochemically-resistant particles on the sea floor. In the absence of a renewed supply of sediment, deep-flowing bottom currents may well have been the agent of erosion. A variety of mechanisms can be visualized, including deep-storm waves, gradient currents, and contour-following currents. Turbidity currents seem unlikely in that little associated sediment was deposited following erosional episodes. An additional mechanism that may be of importance for explaining discontinuities roofed by dark gray shales has been advanced by Baird and Brett (1986, 1991). We hypothesized that internal waves were generated along water mass boundaries that separated denser, perhaps slightly cooler and/or slightly more saline waters below from the overlying aerobic and also lower density waters at a pycnocline. Similar internal waves have been documented in certain modern environments. We postulate that such waves, breaking against submarine slopes could erode the substrate in a sediment-starved regime.

In any event, the presence of these subtle shale-on-shale discontinuities has considerably complicated the correlation of an otherwise rather predictable layer cake type stratigraphy. For example, in certain areas of western New York, at least three different discontinuities interact to produce complex local patterns of stratigraphic preservation (Figure 12f). The mid-Windom Bear Swamp beds for example are locally very thin or absent in the Livingston County area as a result of a discontinuity underlying the superjacent Fisher Gully beds. However, in nearby Genesee County, the discontinuity surface evidently rises up sufficiently that beds of the Bear Swamp interval reappear. However, in this area, a still higher discontinuity below the Fall Brook coral bed has completely truncated the Fisher Gully horizon, such that the Fall Brook beds rests on the Bear Swamp interval. Still farther west, the Fall Brook bed appears to rise off this discontinuity once again preserving both the Bear Swamp and the Fisher Gully beds of the middle Windom. Moreover, in western New York, the top-Windom discontinuity locally bevels the Fall Brook Coral Bed and even the Bay View bed (Figure 12f). The result of this is a complex mosaic of preserved lenses of formerly widespread strata.

DEPOSITIONAL DYNAMICS AND BASIN AXIS MIGRATION

Compilation of thickness data from the rather precisely bounded packages within the Moscow formation reveals a distinctive pattern of westward migration of depocenters (areas of greatest thickness); study of litho-and biofacies relationships bears out this general pattern and indicates the deepest water areas in the basin also were subject to westward migration during the deposition of the late part of the Hamilton Group. The depocenter for the upper unit of the underlying Ludlowville Formation appears to be in the vicinity of Romulus, specifically on the upper end of Big Hollow Creek where a total of at least 18 m of silty mudstone representing the Jaycox Member have been measured (Mayer et al., 1994; Figures 4,5,10a). This area also appears to show the strongest degree of conformity between the Ludlowville and Moscow formations. In particular, an uppermost siltstone package within the Jaycox Formation, not represented in other localities, is present along Big Hollow Creek (Figure 4). This presumably represents the last major shallowing up cycle of the Ludlowville. Hence, evidence points to the most continuous and thickest deposition of the high Ludlowville in the area between Cayuga and Seneca Lakes.

The basal Tichenor Member of the Moscow Formation displays a change to calcareous silty mudstone in the vicinity of Seneca Lake; it is condensed to a compact pack- or grainstone bed both east and west of this location. The thickest development of the Tichenor appears to be in the vicinity of Sampson State Park on the east shore of Seneca Lake. The overlying Deep Run mudstone Member is among the most distinctly lenticular units in the Hamilton Group (Figures 5,12b). The Deep Run attains a maximum thickness of about 18 meters in the area of the type section at Canandaigua Lake. The Deep Run thins to a feather edge both to the west, in central Erie County and to the east, in the vicinity of southeastern Cayuga Lake. Hence, the Deep Run depocenter is subparallel to that of the uppermost Ludlowville Jaycox Member, but is shifted westward by approximately 35 km (Figures 5,12a,b).

The lower portion of the Kashong mudstone is rather ill-defined in the Genesee Valley, in part owing to the lack of a compact condensed horizon at its top (i.e., RC or TT bed). Consequently it is not possible to specify precisely the area of greatest development of the lower Kashong Shale. However, it is clearly in the vicinity of the Genesee River Valley, where the unit attains approximately 4.9 m in the Retsof drill core (Figures 6,12c). Lower Kashong shale thins to the west being only approximately 1 m (3 to 4 ft) thick in central Genesee County. In this area, it is overlain by the compact fossiliferous bed referred to herein as the *Thamnoptychia-Taeniopora* bed (see Lukasik, 1984).

To the east, the lower Kashong also thins as it is capped by the typical RC bed or *Rhipidomella-Centronella* bed. Again, lower Kashong mudstones appear to persist slightly east of Cayuga Lake before pinching out. In areas near Portland point, for example, the RC bed comes to rest essentially upon the Menteth Limestone in an almagmated series of beds referred to in the past as the upper part of the Portland Point Member (Baird, 1979).

The middle Kashong submember is very poorly exposed in the Genesee Valley region, but it is clearly considerably thicker than in the region of Canandaigua or Seneca Lakes. The unit may attain its greatest thickness near Geneseo where it is approximately 10.6 m in thickness (Lukasik, 1984).

Depocenters within the Windom Member, herein interpreted as the deepest portion of the overall Moscow sequence, are more complex and less readily interpreted. (Figures 8,12d,e). The Windom is not as distintinctly lenticular as are the Deep Run or Kashong. The lack of a well defined lenticular region of maximum thickness within the Windom may perhaps reflect the intensity of sea level rise during this interval which provided abundant accomodation space. Hence, slight topographic highs on the seafloor had relatively little influence upon sedimentation patterns, in contrast to the evident bypass and winnowing that took place along the basin margins during deposition of the Tichenor, Deep Run, and Kashong members. Nonetheless, some patterns are evident within the Windom that may suggest a continuation of the trends seen in underlying units. For example, the first cycle of the Windom Shale, constituting the *Ambocoelia umbonata* beds and overlying Bay View and Smoke Creek Beds, appears to thicken westward from Canandaigua Lake toward Erie County. Near the Erie-Genesee County border the total *Ambocoelia* beds portion of the section is on the order of 7 m in thickness, making this one of the thickest sections of *Ambocoelia* beds measured. However, *Ambocoelia* beds thin dramatically to the southwest along the outcrop belt within Erie County, reaching approximately 2 m (6 ft) at Cazenovia Creek and only about 0.5 m (1.5 ft) near Lake Erie. This westward thinning pattern appears to define a region of local subsidence and a regional north westward directed ramp within Erie County. However, the other margin of the lower Windom basin is not so well defined. In contrast to the lower Moscow Formation patterns, the lower Windom appears to thin to the east of Genesee County to a minimum of about 3 m (10 ft) in the western Finger Lakes (Figure 12d). From this point, the interval thickens somewhat to the east and southeast becoming silty and dividing into a series of coarsening-upward shale to siltstone cycles within the area of Chenango Valley (Zell, 1985). The lower Windom beds attain a thickness of about 25 m in the area of Chenango Valley. The middle portion of the Windom (i.e., beds above the Smoke Creek bed), display a complex pattern of thinning and thickening. The Bear Swamp Creek interval is about 15 m-thick at the type area on Skaneateles Lake. It is not known whether this interval thickens substantially to the east, although it does appear to pick up increasing amounts of siltstone. To the west, the interval thins to a minimum near Mack Creek on Cayuga Lake largely as a result of erosional down-cutting at a discontinuity that underlies dark gray Fisher Gully beds (Figure 8).

Age equivalent strata reappear in western Erie County and appear to thicken to the west of the depocenter for the basal Windom in central Erie County, (compare Figures 12d and 12e; Brett and Baird, 1982). Likewise, the Fisher Gully beds, with their distinctive *Emanuella praeumbona* fauna, appear to thicken both west and east of a region near the Genesee-Livingston County line at which point the Fisher Gully or Amsdell beds have been eliminated by erosion beneath the Fall Brook coral bed. The Amsdell bed thickens into southwestern Erie County, suggesting a local depocenter which has migrated 30 km to the southwest from the vicinity of Genesee County during this late interval. However, the Fisher Gully beds also thicken and become progressively darker and more platey in the vicinity between Canandaigua and Cayuga Lakes, the same region in which the mid Windom Bear Swamp Creek beds become thin or absent. Farther east (for example, at Skaneateles Lake), the Fisher Gully beds thin once again in concordance with the thickening of the underlying Bear Swamp beds (Figures 8,9).

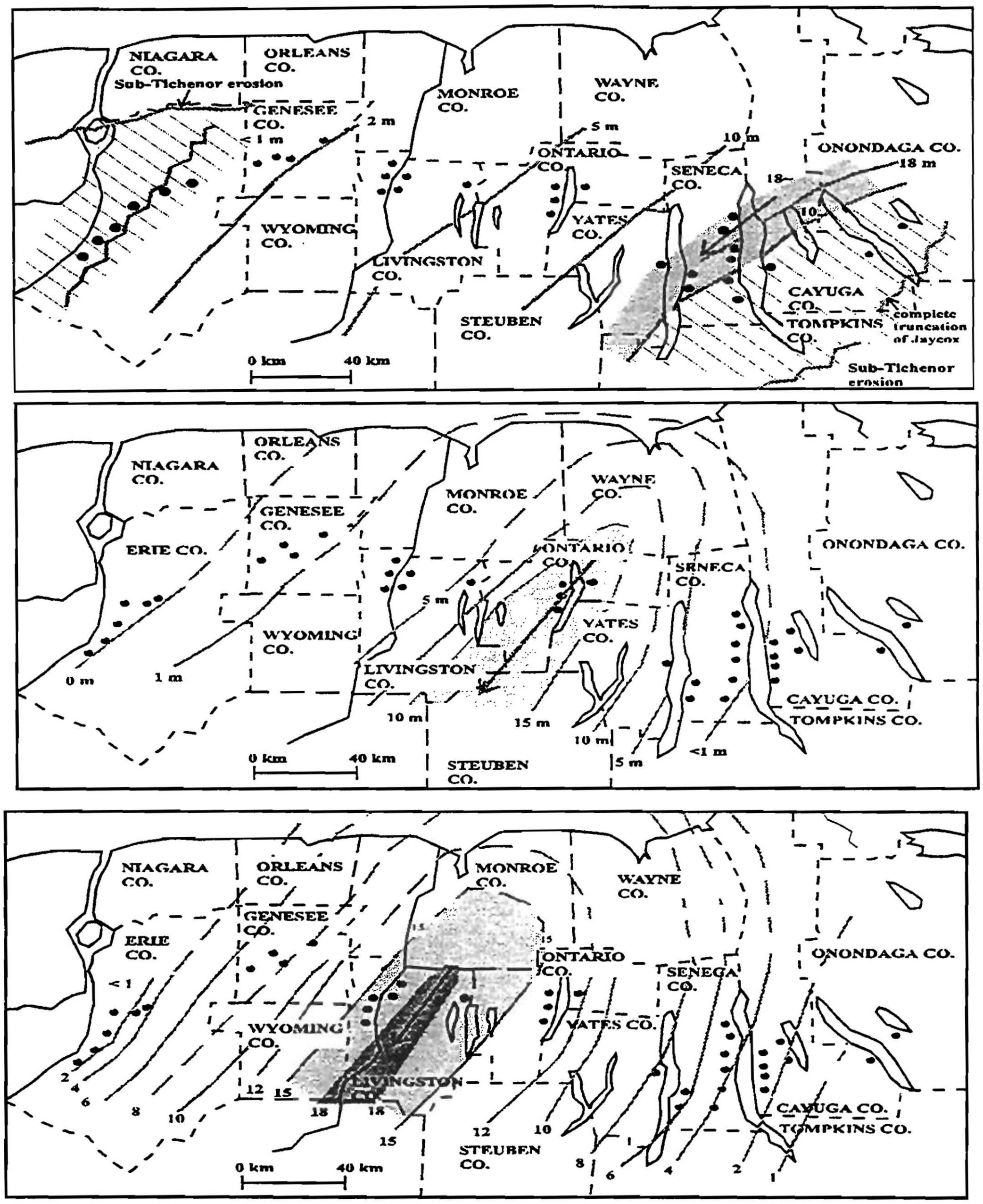

Figure 12. Schematic isopach maps for the Moscow Formation in western New York. Data indicate key localities; depocenter area shaded; approximate location and plunge of basin axis indicated by arrows; isopleths numbered in meters. A) Jaycox Member; B) Deep Run Member; C) Kashong Member. Note general westward shift of depocenters through this interval.

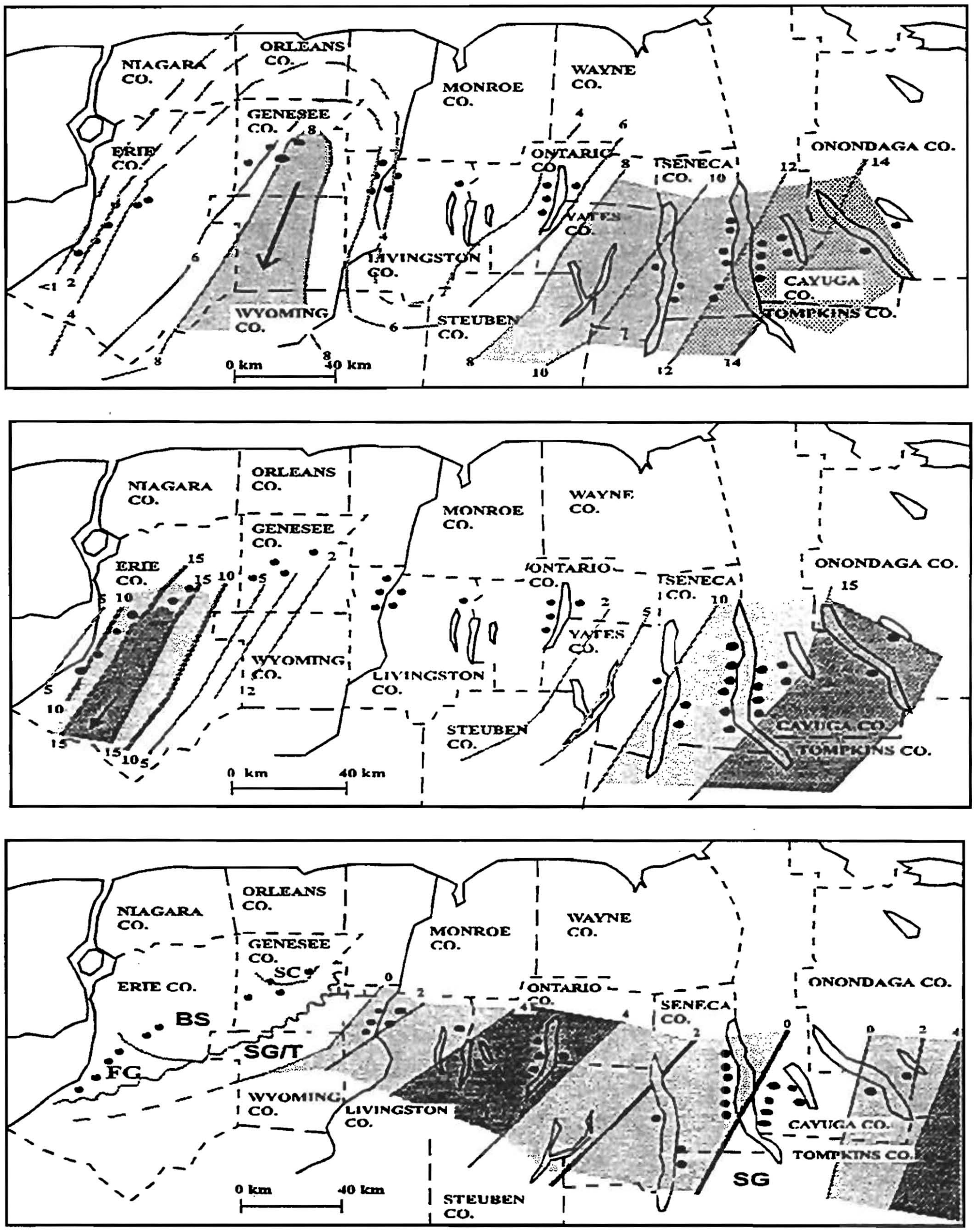

Figure 12 (Continued): D) Isopach map of lower portion of Windom Shale (up to Smoke Creek bed), note presence of two depocenters separated by thin area (arch) in Genesee-Canandaigua region. E) Isopach of middle Windom (Bear Swamp-Fisher Gully beds), note westward shift of western depocenter relative to "D." F) Isopach map of Gage Gully beds; thinning reflects erosional truncation; where Gage Gully beds completely removed lower units exposed beneath Taghanic unconformity. Symbols include: SG, Spezzano Gully; TG, Taunton beds; FG, Fisher Gully beds; BS, Bear Swamp beds; SC, Smoke Creek bed.

The Fall Brook coral bed, likewise, displays a pattern of increased skeletal content, increasing predominance of large rugose corals, and a thinning to the west near the Livingston-Genesee County line. Taken together these patterns suggest that the Windom sea floor carried at least two depocenters separated by a local swell or high region (Figures 12d,e). Perhaps this region became a bypass slope during this time and was subject to submarine erosion during the ensuing interval of sediment starvation. In any event, evidence from the successive more westwardly truncation of the Bear Swamp beds and then the Fisher Gully-Amsdell beds between Canandaigua Lake and west central Genesee County suggests that the local intrabasinal high as well as the sub-basins themselves display a westward migration pattern during the interval of the middle part of the Windom.

Patterns within the upper Windom, primarily the Fall Brook and Taunton cycles are less completely known, particularly due to major erosional removal of this portion of the section west of the Genesee Valley area (Figure 10). However, the evidence from remaining strata suggests that during this interval the eastern sub-basin of the Windom migrated into the vicinity of Canandaigua Lake. Here some of the thickest sections of the uppermost Windom unit, the Gage Gully black shales, are preserved (Figure 8). Correspondingly, the intrabasinal high area appears to have passed through the region of central Genesee County. At this location, the Windom Member, as a whole, displays maximum truncation by the pre-Tully erosion surface. Likewise, the eastern margin of the eastern sub-basin also migrated westward such that maximum truncation of the Windom Member, down to the level of the Taunton or Spezzano Gully beds occurs in the vicinity of Moravia or Portland Point.

At the close of Windom shale deposition, probably three separate centers or depocenters existed, separated by two intrabasinal highs as evidenced by the pattern of erosion of the sub-Tully unconformity (Figure 12f).

The paleogeography during this time would have included, from west to east: a) a relatively shallow western sub-basin centered in southwestern Erie County; b) a broad intrabasinal high region in north central Genesee County, at least partially separating the Erie County sub-basin from; c) a central and perhaps deepest sub-basin, with area of maximum subsidence at or near Canandaigua Lake where the uppermost Windom Gage Gully beds (Unit 17) are best developed as a black dysoxic shale unit (Figure 12f). In turn, this sub-basin was bordered to the southeast by a second low arch or submarine high which crested along a diagonal line approximately from Owasco Lake southeastward toward Portland Point, an area in which none of the Gage Gully beds have been preserved (Figure 12f). Still further to the east a shallow but actively subsiding sub basin existed in the region of Chenango Valley or perhaps further east in which region the upper Windom units become well defined, thick, and relatively silty.

Thus, in a broad overview, the depocenter and axis of greatest subsidence for the Appalachian foreland displays a progressive, albeit complex pattern of east-to-west migration during deposition of the Moscow Formation, the last major sequence of the Hamilton Group (Figure 13). During an interval of no more than 1.5 to 2 million years of geologic time, the lower Moscow depocenter shifted its position some 90 km from the region between Cayuga and Seneca Lakes (upper Jaycox Member), to Seneca Lake (Tichenor Member),

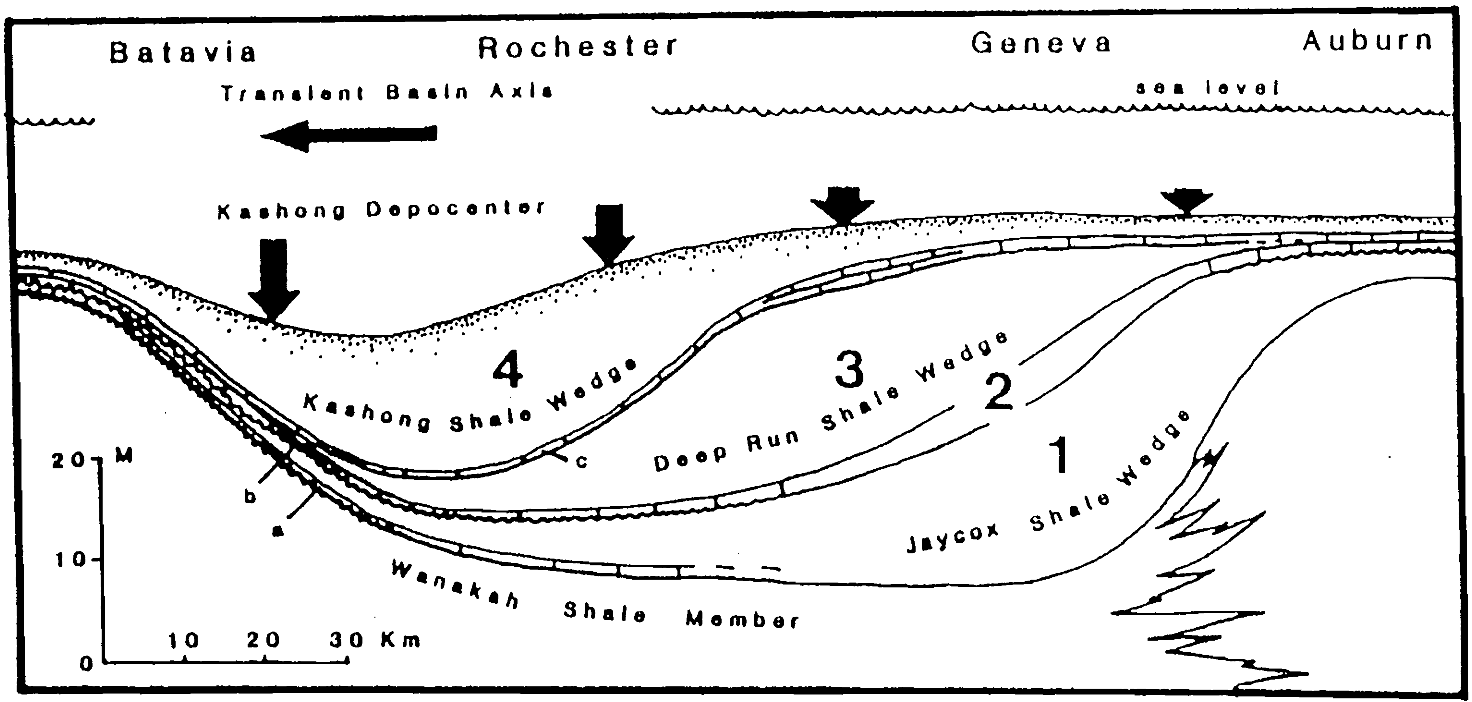

Figure 13. Diagrammatic east-west cross section of the upper part of Ludlowville and lower part of Moscow Formation. Note that successive wedges of mudstone, bounded by through-going carbonates display a progressive westward migration.

Canandaigua Lake (Deep Run Member), Livingston County (Kashong Member), and western Genesee County (lower Windom interval) (Figure 13). However, during deposition of the upper parts of the Windom Member, i.e., during the relative highstand portion of the overall sequence, the simple geometry observed in the earlier phases of the Moscow sequence gave way to a more complex topography in which at least two and perhaps three sub-basins were developed, each bounded by low broad submarine swells or highs. This suite of partitioned sub-basins and local highs also displayed a general westward drift throughout the remainder of the Hamilton deposition.

The minor "rippling" of the Hamilton foreland basin floor is a harbinger of more dramatic folding in the foreland basin that took place during deposition of the overlying Tully Formation. Heckel (1973) thoroughly documented development of a small scale anticline during deposition of the lower Tully and of a down-to-the-east monocline/fault that developed, perhaps along old lines of structural weakness in the region east of the Chenango Valley (Figure 14). This down-to-the-east subsided trough may have provided a clastic trap which starved the basin nearly completely of sililiclastic sediments and permited deposition of the unusual Tully Limestone over much of west central New York State. Furthermore, Heckel (1973) documented a mid-Tully interval of erosion during which the Chenango Valley high or anticline was breached and its center then became a locally subsided depocenter wherein the thickest part of the upper Tully Member was deposited.

The interval of partitioning and rapid westward migration of the basin during the later Moscow and into the pre-Tully interval suggests tectonic instability, perhaps associated with the beginnings of the third and strongest tectophase of the Acadian orogeny. It is notable that similar buckling, arching and erosion of the basin floor is also witnessed at the beginning of Onondaga Limestone deposition (see contribution by Ver Straeten et al., this volume). Ettensohn (1987) has suggesed that periods of docking of the microplates of Avalonia with North America produced tectophases of the Acadian orogeny and that times immediately following the docking were relatively quiesent intervals during which carbonates developed in sheet-like units over extensive areas of the craton, following the development of major unconformities such as the Wallbridge and the sub-Tully or Taghanic unconformity. Following a lag time, active tectonic thrusting from the east produced renewed pulses of major subsidence toward the hinterland. These deep underfilled basins were initially areas of dysoxic black mud accumulation.

STRATIGRAPHY AND INTERPRETATION OF THE TULLY AND GENESEO FORMATION

Sub-Tully Unconformity

The Hamilton Group is separated from the Tully Limestone in the Finger Lakes east of Canandaigua, and from the overlying Genesee black shale from Canandaigua westward, by a significant disconformity that has been termed the Taghanic Unconformity (Johnson, 1970). This disconformity serves to separate the early and later phases (holostromes) of the Kaskaskia megasequence. It displays significant although subtle, regional beveling as documented by Cooper & Williams (1935), Heckel (1973), and Brett and Baird (1982). This erosion surface is complex, having removed substantial amounts of the upper portion of the Windom Shale, particularly in areas of western Livingston, Genesee and eastern Erie Counties. However, the relatively condensed western Windom facies in the vicinity of Lake Erie display a much lesser degree of truncation at the top (Figure 12f). Maximum preservation of Windom strata occurs in the Canandaigua Lake Region, where 5-6 m of upper black to dark gray shale of the uppermost cycle occur beneath the Tully Limestone or the Genesee Formation where the Tully has been completely removed. Farther to the east, as well as to the west of the Canandaigua Lake area, significant amounts of the upper Windom cycle have been removed such that the Tully or Geneseo rests on the middle or even lower cycles of the Windom Member (Figures 8,12f,14).

During the interval between deposition of the highest Windom beds and initiation of the Tully Limestone, a major and apparently rather widespread sea level lowstand produced this erosion surface over much of New York State. Because of the complex topography, different portions of the upper Hamilton Group were preserved in different areas. Local topographic highs were exposed to greater amounts of erosion during the pre-Tully lowstand interval and we infer that regions of maximum cut out of the Windom permit delineation of these submarine highs (Figure 12f).

Farther west, into the mid craton, the sub-Tully unconformity removed a great deal of the upper part of the Hamilton Group. In many areas, Upper Devonian (Frasnian) strata rest directly upon beds equivalent to the lower part of the Hamilton Group in New York State, particularly the Centerfield Limestone or its lateral equivalents in the Beechwood of Indiana and the Boyle of Kentucky. Hence, this is a major erosion surface subdividing the Kaskaskia mesasequence into two portions or holostromes.

In central Pennsylvania, closer to the depocenter of the late Givetian times, the Mahantango Formation of the Hamilton Group is separated from the overlying Tully equivalent calcareous shales by a subtle unconformity. As in New York State, the patterns of units beneath the Tully *Emanuella* and *Hypothyridina*-rich calcareous shales is rather variable and complex. Units equivalent to the uppermost Windom Gage Gully beds downward through the South Lansing coral bed and parts of the Taunton beds may crop out beneath the sub Tully erosion surface in Pennsylvania.

Tully Limestone

General Stratigraphy: The Tully Limestone and its internal divisions have been documented in considerable detail by Heckel (1973). In most of west central New York east of Seneca Lake the Tully is subdivisible into two members separated from one another by a mid-Tully unconformity (Figure 14). The lower Tully consists primarily of micritic limestone (calcilutite) containing an unusual brachiopod assemblage characterized by forms such as *Hypothyridina*, *Emanuella* spp., *Rhyssochonetes aurora*, and others. The occurrence of possible stromatolitic lamination and desiccation cracks near the top of the lower Tully unit at Bellona suggests that these lime mudstones accumulated in shallow water, possibly a broad lagoonal setting and their contained brachiopod faunas reflect a distinctive nearshore biofacies not otherwise present in the Middle Devonian of the Appalachian Basin. In contrast, beds overlying the mid-Tully unconformity record a spectrum of facies closely similar to that of the portions of the underlying upper Hamilton Group (Figure 15). For example, the Taughannock Falls bed possesses a fauna typified by small rugose corals (*Stereolasma* or *Metriophyllum*), and the trilobites *Phacops rana*, *Dechenella* spp. and *Greenops* together with abundant *Pseudoatrypa*, *Protodouvillina* and other brachiopods that characterize the more calcareous portions [see preceding discussion of trilobite beds, (Units 6 and 13) of the Windom Member]. A similar biofacies spectrum is presented in the upper or Moravia bed in the Tully. Sandwiched between these is an intriguing interval displaying the highest Givetian coral bed. The Bellona bed contains a suite of fossils virtually identical to those found in many of the upper Hamilton coral beds, for example, in the Bay View, Fall Brook, and South Lansing coral beds discussed herein (Baird and Brett, 1983; Brett

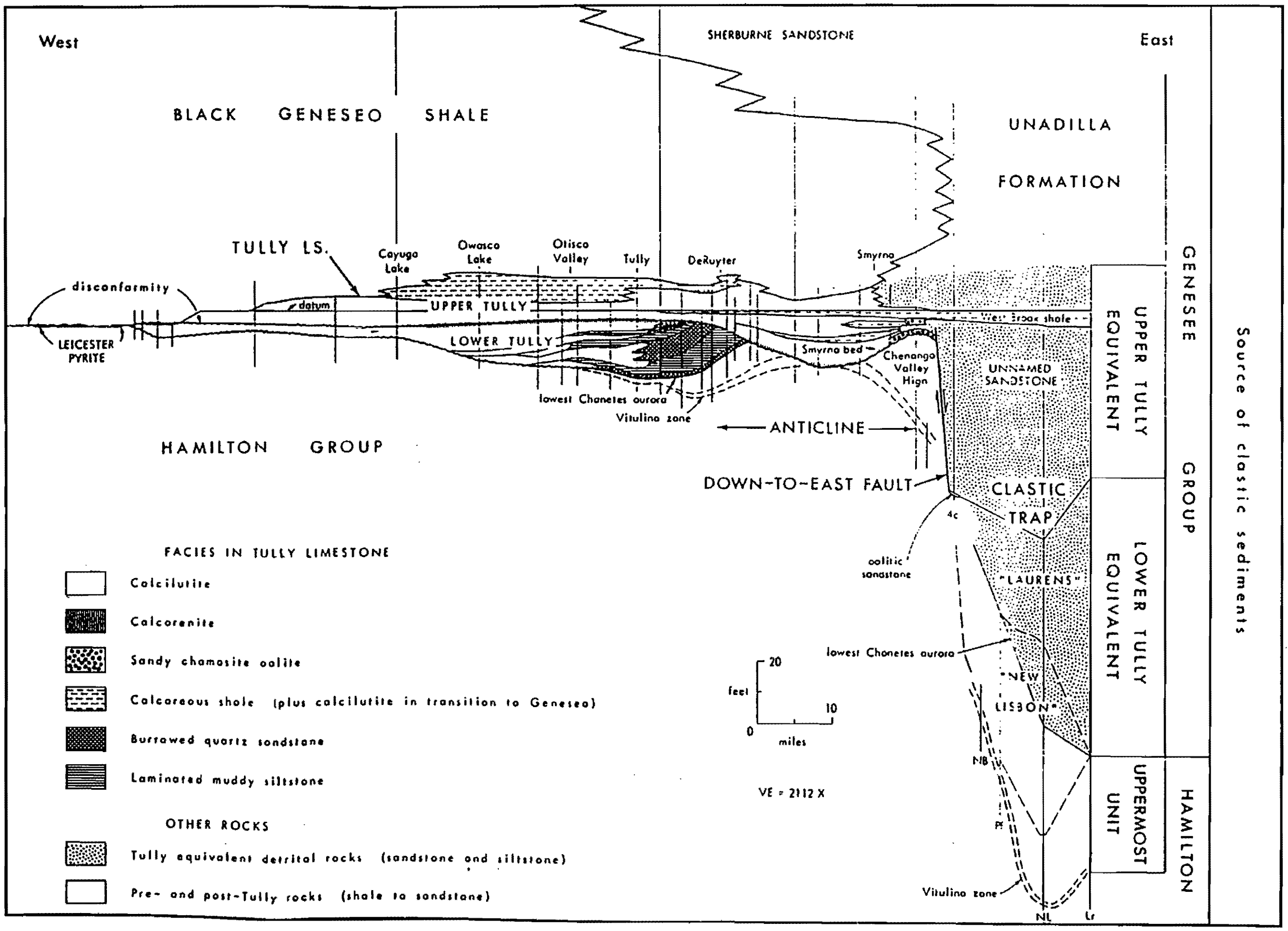

Figure 14. Regional cross section of Tully Limestone in western and central New York. Note truncation of upper Hamilton (Windom Shale) and adjacent units below Tully and, development of anticline which later becomes site of upper Tully depocenter west of Chenango Valley; from Heckel, 1973.

et al., 1983). The Bellona bed appears to reflect lowstand conditions within the upper Tully cycle. Its fauna of *Cystiphylloides*, *Heliophyllum*, favositids, and diverse brachiopods, including *Spinatrypa spinosa* strongly resembles those seen in the Windom beds. To the east, this unit and/or the immediately overlying shale carry a *Tropidoleptus* fauna not unlike that of muddier portions of the shallow water Hamilton Group such as the Kashong or Deep Run members discussed above. Hence the Tully appears to reflect a somewhat more siliciclastic-free version of facies well expressed in the upper Hamilton Group.

Interpretation: The Tully Limestone appears to record a reversal of the general pattern of siliciclastic input within the Appalachian foreland basin, and an abrupt "cleaning" of offshore marine environments. Tully carbonate is relatively siliciclastic free, which raises the significant question as to what processes temporarily shut down the influx of terrigenous sediments to the foreland basin. It may very well be that flexural folding and buckling within the basin, produced an arch-like barrier and sediment trap to the east of the Chenango Valley (Heckel, 1973; Figure 16, herein). On the other hand, the Tully interval is relatively carbonate-rich throughout much of the Appalachian basin suggesting that additional processes may have been at work. In many ways, the Tully (including its eastern equivalents which are quartz rich sand) appears to record a return to an orthoquartzite-carbonate succession characteristic of interval of relatively deep weathering of a diminished siliciclastic supply and, by inference, tectonic quiescence.

It is intriguing that both the Tully and underlying Windom display complex erosion surfaces as well as regionally angular bounding unconformities. Both of these phenomenona point to the presence of some activity within the foreland basin during this supposedly quiescent phase. Ettensohn (1987) has suggested that these effects record the initial docking of Avalon Terrain along the various salients of the North American margin. A broad uplift of arches might be predicted to occur during this time. Another significant factor is the apparent dramatic eastward shift in the depocenter and axis of greatest subsidence within Tully deposition. This amounts to a displacement of over 120 km of the basin center from near Canandaigua Lake during deposition of the black Gage Gully beds (Figure 12f) of the underlying Windom Shale back towards the Chenango Valley Region during deposition of the upper portion of the Tully Formation. The main center of the foreland basin during late Tully deposition is identified on the basis of evidence for most dysoxic facies, and relative thickness. These data tentatively indicate that the basin axis lies along a diagonal northeast, southwest trending line that ran approximately from southern Madison County in central New York through Williamsport, Pennsyvlannia. At the latter locality Tully equivalents exceed 30 m (100 ft) in thickness (Figure 16). In most areas, equivalent Tully strata are no more than 10 meters (25 to 30 ft) thick.

In the axial part of the basin, the upper portion of the Tully (Taughannock Falls bed) is represented by a rhythmic micritic limestone-dark gray shale succession, which may record a succession of carbonate turbidites. However, to the west of this area, the upper portion of the Tully thins into more highly fossiliferous shallow-water facies.

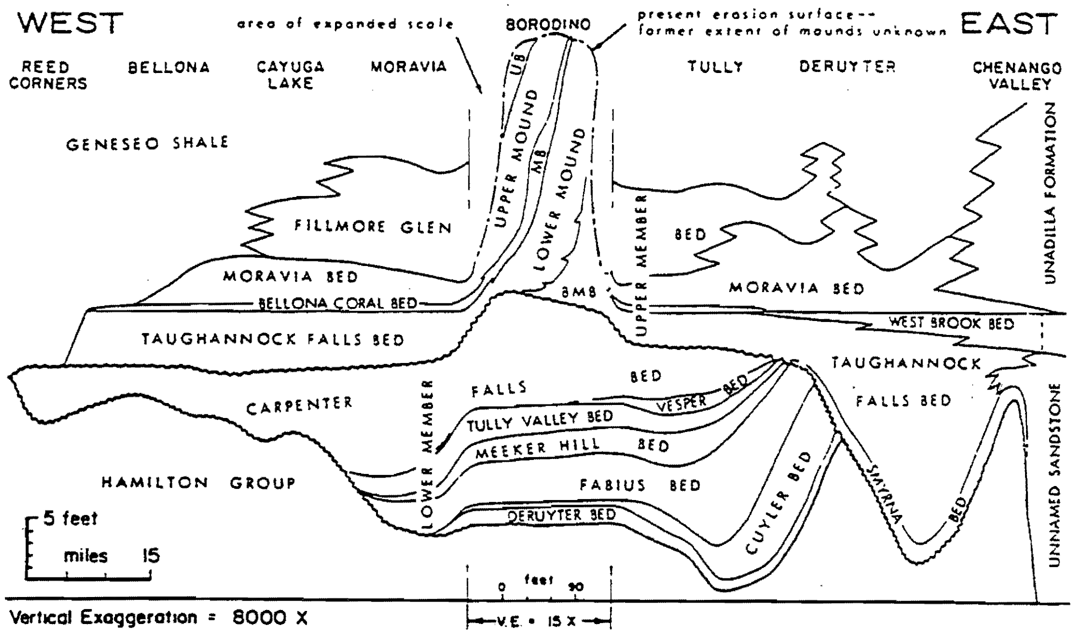

Figure 15. Stratigraphic units of the Tully Limestone, from Heckel (1973).

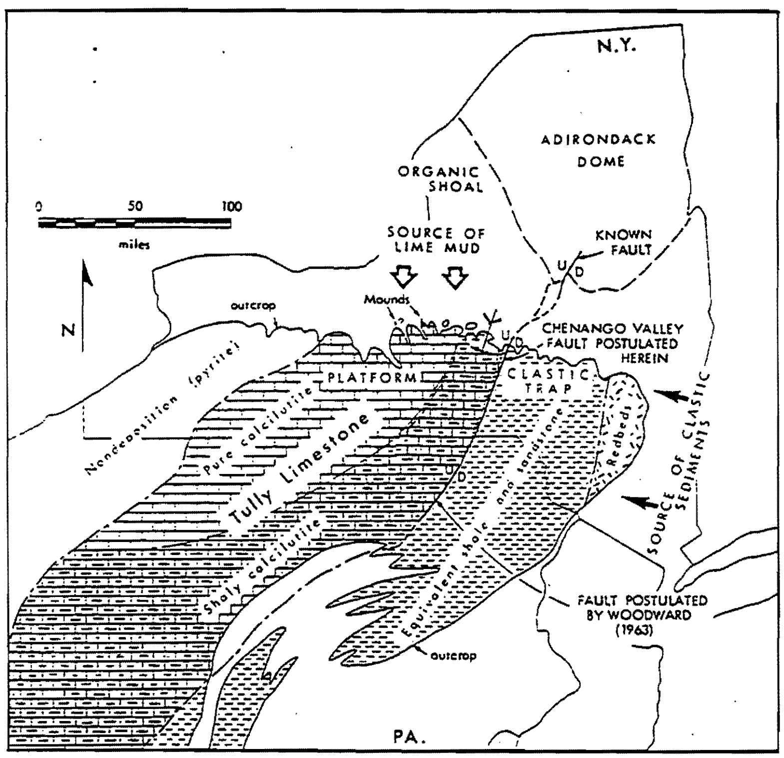

Figure 16. Map of structural and sedimentary features controlling Tully Limestone deposition, and resulting contemporaneous facies in New York and Pennsylvania. Note depocenter of upper Tully in region marked "shaley calcilutite;" from Heckel (1973).

Tully-Genesee Unconformity and Leicester Pyrite

In the area of southern Cayuga Lake, the uppermost Tully submember, i.e., Fillmore Glen beds, consists of argillaceous micritic limestones that alternate with black or dark grey calcareous shales. These grade upward into more and more sparsely fossiliferous dark shales. A sharpdiscontinuity (surface of maximum starvation) separates the upper Tully Fillmore Glen beds from the basal Genesee Formation black shales. In sections around Taughanock Falls, this surface of maximum starvation is demarcated by a thin (millimeters) lag of corroded fossil material and phosphatic diaclasts. The surface is merely a diastem between more calcareous, dark gray mudstones below and overlying black platey less calcareous shales above. However, to the west, between Cayuga and Seneca Lakes, this contact becomes sharp and erosive. The upper portion of the Moravia bed is successively beveled to the northwest from the region of Taughanock Falls in the southern area of Cayuga Lake northwestward to Seneca Lake (Figure 15). At Bellona (west side of Seneca Lake), black shale of the Geneseo Formation rests directly upon the Bellona coral bed in one area or on an erosional remnant of the Moravia bed in different banks of the Kashong Creek. This local scour surface at the base thus truncates the upper parts of the Tully and the nearly conformable successions seen in the southern and eastern Finger Lakes region passes into an important unconformity.

The Tully Limestone or the upper Windom Shale are abruptly overlain by black Geneseo shale west of Seneca Lake. In places this contact is marked by thin lines of reworked pyrite clasts, fish bones and conodonts, termed the Leicester Pyrite Member of the Geneseo Shale (Baird and Brett, 1986, 1991). In the vicinity of Gorham, between Seneca and Canandaigua Lakes, the sharp contact between black Geneseo Shale and the Bellona coral bed is marked by a thin lag of pyritic and crinoidal debris. This would appear to represent the easternmost locality at which the Leicester Pyrite can be recognized. Farther to the northwest, along the east side of Canandaigua Lake, even the lower Tully appears to be beveled beneath the sub-Geneseo unconformity. In the upper end of Gage Gully, small lenses of Leicester Pyrite have been observed overlying an erosional knob or remnant of the lower Tully Carpenter Falls bed. West of Canandaigua Lake, the Tully is missing altogether as a result of corrosion or erosion at this unconformity. In its place, are somewhat larger lenses of pyritic fossil and nodular debris, up to a meter across and in places 30 cm in thickness, occur along the otherwise knife-sharp contact between the Geneseo and the underlying Windom Shale (Figure 17). These pods of Leicester Pyrite interfinger with the black Genesee Shale and not with the Windom below (Baird and Brett, 1986).

Although the pyritic clasts are apparently derived from the erosion of the upper Windom beds, the Leicester Pyrite lenses are interbedded with laminated black shales of the basal Geneseo Member. We have previously argued (Baird and Brett, 1986, 1991) that the pyritic clasts derived from erosion of the underlying Windom muds were concentrated into windrows on the dysoxic sea bottom by internal waves or currents associated with a rising pycnocline. Hence, the pyrite and associated erosion surface at the base of the Geneseo muds provide an example of the processes associated with sedimentation during maximum marine flooding. In contrast to the basal Tully lowstand erosion surface, the disconformable upper contact of the Tully with the overlying black Geneseo Shale is of lesser magnitude and represents a highstand rather than a

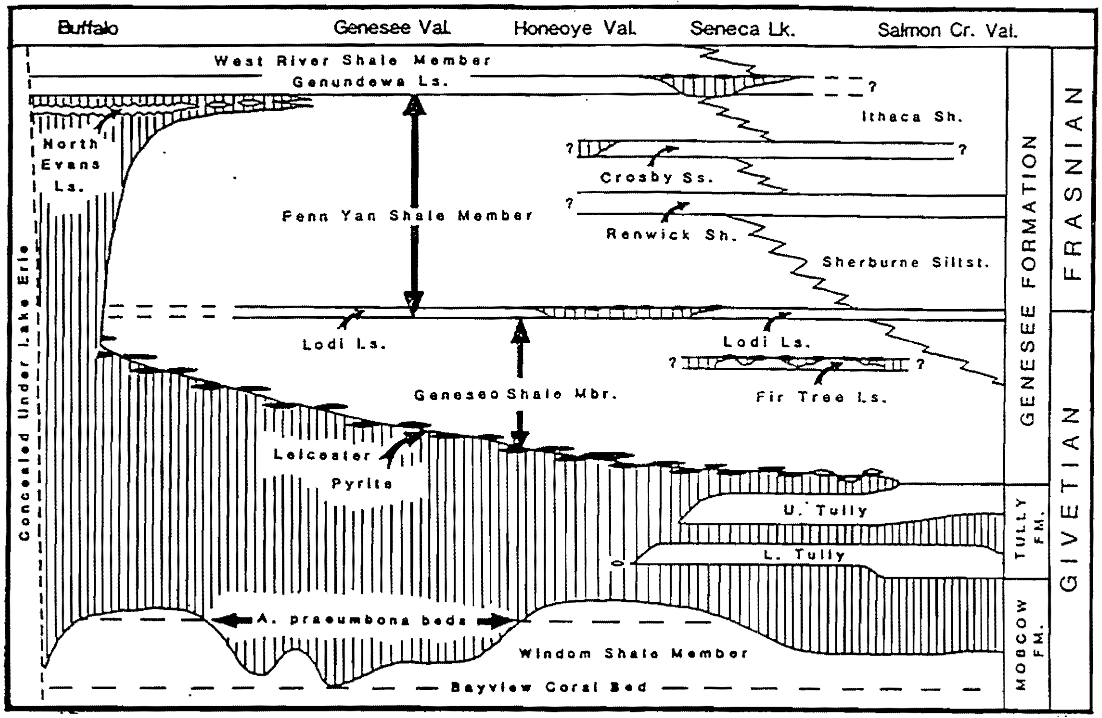

Figure 17. Chronostratigraphic cross section of lower Genesee Formation and subjacent Moscow Formation (Windom Shale Member). Note positions of the thin and locally beveled Fir Tree and Lodi Limestone submembers. Large hiatus below the Genesee Formation marks the position of the Taghanic Unconformity; lenses of detrital Leicester Pyrite are derived from this erosion but were deposited through a long period of diachronous overlap of Geneseo black muds from this discontinuity. From Baird et al., 1988.

lowstand related unconformity. In previous papers (Baird and Brett, 1986, 1991; Brett and Baird, 1990), we have argued that these types of surfaces were developed at oxic-anoxic water mass boundaries and are related, in part, to intervals of sediment-starvation associated with maximum flooding surfaces. In areas to the west of Canandaigua Lake, where the Tully Limestone has been removed below this erosion surface, the discontinuity between the Hamilton and the Geneseo black Shale is a composite contact which has overstepped both the mid-Tully discontinuity and the sub-Tully erosion surface.

Removal of the Tully Limestone in the areas west of Canandaigua Lake juxtaposes the post-Tully, sub-Genesee unconformity on to the sub-Tully unconformity (Taghanic unconformity). However, the sub-Tully erosion surface occurred under fully oxygenated conditions. The basal surface of the Tully is characterized by casts of mega-burrows excavated into the still soft, but over-compacted muds of the Windom Member. Such distinctly burrowed irregular firmground contacts characterize lowstand sequence boundaries in parts of the Hamilton Group, as well. In contrast, the contact between the Tully limestone and the overlying Geneseo is associated with maximum flooding, the beginning of the highstand is a surface of maximum sediment starvation even where close to conformable near the basin center and the Cayuga-Owasco Lake region.

On the basin margin, erosion at this time, perhaps as a result of deep storm waves or internal waves (see discussion above under internal Windom erosion surfaces) produced beveling along a knife-sharp, gently furrowed surface lacking any type of burrow structure. In areas where the Tully was exposed on the sea floor during times of prolonged sediment starvation, submarine corrosion or dissolution may have been a critical factor in removal of the older carbonate sediments. Farther to the west, once the Tully had been breached, physical erosive processes must have worked to flatten and smooth the old pre-Tully erosion surface into a nearly planer or very gently furrowed surface. Furthermore, current velocities must have been sufficient to erode and aggregate pyrite from the underlying Windom beds, forming the Leicester Pyrite lenses. Hence, the basal Genesee contact in the area of the Genesee Valley and westward to Erie County is an excellent example of a combined or composite sequence boundary formed by lowstand erosion prior to the Tully and by erosion under deep marine conditions during a period of maximum sediment starvation at the base of the Genesee highstand.

The abrupt transition from Tully Limestone to Genesee black shale appears to record an interval of rapid subsidence of the Appalachian foreland associated with the influx of a large wedge-like body of black shales. This pattern, is closely comparable to the situation seen Tectophases I and II of the Acadian Orogeny in the Appalachian basin. One curious anomaly, however, is that in the case of the Tully-to-Geneseo transition, no bundle of thin K-bentonites (altered volcanic ash beds) is recognized. In the other examples of tectonic-flexural deepening (e.g., Oriskany-Esopus and Onondaga-Marcellus transitions), a distinctive bundle of bentonites occurs near the top of the carbonates. These clustered ashes suggest the onset of renewed orogenic activity. The change to Geneseo which undoubtedly records the strongest pulse or Tectophase III of the Acadian Orogeny is not apparently accompanied by the input of ash beds. However, this may reflect the absence of a concerted search for such layers.

Geneseo Shale

The third and uppermost portion of the Givetian in New York State is represented by post-Tully clastic sediments of the lower part of the Genesee Formation (Figure 17). The lower Geneseo Black Shale, assignable to the *hermanni cristatus* conodont zone upward to the *disparalis* zone (Kirchgasser et al., 1988, this volume), records maximum deepening within the Appalachian basin for the Middle Devonian interval. Black laminated shales dominate the facies in the study area between the Genesee Valley type section and the Seneca Lake Region; however, we have recognized two significant, although subtle, cycles of shallowing within the Geneseo package (Baird and Brett, 1986; Baird et al., 1988). These cycles are capped by impure, nodular argillaceous concretionary limestones with abundant styliolinids, auloporid corals and goniatites. These two shallowing events are significant because the base of the lower *asymmetricus* conodont zone occurs at or near the top of the second nodular concretionary limestones cap. Thus, the top of this unit (Lodi Limestone) corresponds to the Givetian-Frasnian Stage boundary (Kirchgasser et al., 1988, this volume). The nodular limestones capping the two cycles are significant for a second reason; they are analogous to nodular cephalopod limestone facies described by Ver Straeten et al. (this volume) for the Cherry Valley limestone which characterizes a shallowing pulse associated with the earlier Tectophase II black shale depositional interval.

The major transgression, recorded by the change from Tully carbonate into basinal Geneseo black shale facies, is associated with a westward advance (onlap) of basinal deposits over the composite disconformity into Ohio and Ontario. It is referred to as the Taghanic Onlap or IIA transgression event (Johnson, Klapper and Sandberg, 1985), and it records the combined effects of eustatic and tectonic processes. Detailed stratigraphic relationships indicate that the transgressive record (Tully Formation) of this onlaps has been completely removed in the west by deep submarine erosion and corrosion prior to deposition of the overlying siliciclastics.

In certain ways, the abrupt deepening observed at the top of the Tully and the infilling of a local foreland trough with carbonate turbidites and dark shales is reminiscent of patterns observed in earlier Devonian tectophases. The Fillmore Glen beds at the top of the Tully are similar to thin bedded carbonates and dark shales in the upper Onondaga and lower Marcellus transition at the beginning of tectophase II at the base of the Hamilton Group (Ver Straeten et al., this volume). In the Tully example, as in the other cited instances, the carbonate appears to represent the transgressive deposits (systems tract) of a megasequence and the abrupt change to black shales at the top of the Tully mirrors that seen at the top of the Onondaga Limestone. As with the other cases, a transitional section of thinly-bedded carbonates and dark shales occurs in the area of maximum subsidence. Towards the margins of the basin, the carbonate to black shale contact is knife sharp and is marked in places on analogous surfaces (i.e., upper Onondaga and upper Tully), by a lag of phosphatic, conodont, bone and/or pyrite-rich debris. The Leicester Pyrite is one such thin lenticular lag deposit which accumulated under anoxic conditions in a deep-water setting (Baird and Brett, 1991).

DISCUSSION AND CONCLUSIONS

The highly predictable ("layer cake") carbonate-rich stratigraphy that characterizes the upper part of the Hamilton Group and the Tully Formation suggests an interval of relative quiescence between the second and third tectophases of the Acadian Orogeny (Ettensohn, 1975; Dennison, 1985). Widespread general lowering of relative sea level combined with diastrophic local flexural uplift within the foreland basin produced two relatively large and several lesser erosion surfaces that underlie widespread transgressive carbonate units. The most significant erosion surface within the Hamilton Group occurs at the base of the Tichenor Limestone which marks the sharp basal contact of the Moscow Formation: the Tichenor and overlying condensed carbonate deposits have been recently interpreted (Brett and Baird, 1990) as an initial transgressive systems tract of the Moscow third-order depositional sequence. Erosion at the sub-Tichenor surface-level occurred on both eastern and western ramps of the foreland basin producing a nearly symmetrical pattern of truncation of beds in the upper portion of the Ludlowville Formation both east and west of the central Finger Lakes (Mayer et al., 1994).

The Moscow Formation records a general deepening pattern from the basal skeletal grainstones of the Tichenor Limestone upward into dark gray to black shales of the Windom member which have been interpreted as highstand deposits. However, at least two fourth order sequences and a number of smaller scale cycles (fifth or sixth order cycles in the terminology of Busch and Rollins, 1984) are superimposed upon this general deepening pattern. Several member- or submember-scale shallowing-upward cycles, each capped by a thin silty to calcareous, bioclast-rich horizon have previously been documented. In ascending order these are: a) the Deep Run Member, calcareous mudstones which are capped by a silty, bioturbated Menteth Limestone; b) the lower Kashong Member blue gray mudstone overlain abruptly by the "RC" shell bed; c) the middle Kashong mudstones, capped by a Menteth-like unit; d) an upper Kashong cycle capped by silty shell rich and phosphatic Barnes Gully carbonate bed; e) two poorly developed presently unnamed cycles which range from shales to silty mudstones or siltstones depending upon location; the upper is capped by the Geer Road phosphatic shell bed; f) the lower *Ambocoelia*-rich shales of the Windom Member capped by Bayview shell and coral beds; g) the middle Windom ambocoeliid and chonetid bearing shales capped by unnamed shelly beds and h) dark grey Fisher Gully beds capped by the Fall Brook coral-rich bed and the overlying Taunton calcareous mudstones; and, i) the upper Windom dark gray shales which typically contain diminutive brachiopods, especially small spiriferids (juvenile ? "*Allanella*").

The dark gray to nearly black shales of the highest Windom Member appear to reflect some of the deepest water facies of the upper Hamilton Group, as a whole. This is rather enigmatic given the fact that they are overlain by a widespread erosive disconformity. Presumably, a still higher cycle of more profound shallowing originally occurred within uppermost Windom mudstones, but the upper, transitional, shallowing portions of this cycle have been removed everywhere throughout the New York outcrop belt by a sub-Tully unconformity of considerable magnitude.

Detailed mapping of the individual shallowing-deepening cycles within the upper part of the Hamilton Group has revealed a significant and intriguing pattern. The depocenters, areas of greatest thickness, as well as the areas of apparently deepest- water facies for each successive unit, display a displacement to the west relative to those of the underlying unit. This westward migration of the depositional axis of the basin has resulted, in part, from progradation of small siliciclastic wedges into the basin with consequent reduction of accommodation space progressively further to the west. However, the relative thinness of these wedge-like sedimentary bundles suggest that sediment loading was not a particularly significant effect. The pattern of westward migration seen progressively within the upper portion of the Hamilton Group appears to have been abruptly reversed during deposition of the overlying Tully Limestone.

Within the Moscow interval, relative lowstand erosion surfaces are overlain by two distinct types of somewhat condensed calcareous beds. The first type, well represented by the Tichenor limestone, the RC-TT bed of the Kashong Shale, Barnes Gully bed at the base of the unnamed member, and precursor beds, such as the Fall Brook Bed within the Windom. The beds rest sharply on the underlying mudstones with little or no hint of a shallowing upward cycle beneath them, and they are composed of extremely skeletal rich highly condensed lag deposits. These beds also overlie some of the more major erosional surfaces within the Hamilton Group. They are best interpreted as first transgressive lag beds that develop during initial sea level rise following a "forced regression" i.e., a time of relatively rapid sea level drop not associated with major progradation. This could be caused by either eustatic sea level fall or regional uplift; the latter would require that major portions of the northern Appalachian both east and west of the depocenter were more or less simultaneously uplifted. We suggest that eustatic sea level fall is the more likely alternative.

The second type of compact bed that overlies at least minor erosion surfaces, is exemplified by silty limestones and calcareous siltstones, such as the Menteth Limestone, unnamed calcisiltite beds in the upper Kashong shale, the Curtice Road bed of the unnamed member and perhaps capping siltstone beds within the Windom Member east of the Finger Lakes (Zell, 1985). These beds appear in depocenter areas to be gradationally conformable with underlying silty mudrocks (e.g. Deep Run, upper Kashong shales, *Megastrophia* beds silty mudstones). Towards basin margins, these beds display distinctly burrowed basal surfaces that evidently cut slightly into underlying mudstones, becoming slightly disconformable. Two factors seem particularly relevant in the interpretation of these beds and their associated discontinuities. First, they typically seem to alternate in position with the Type 1 lag beds (e.g., Tichenor vs. Menteth, RC bed vs. upper Kashong calcisiltites, Barnes Gully bed vs. Menteth Gully bed, Fall Brook precursor bed versus upper Taunton siltstone or Lansing bed). Secondly, these silty limestone beds do not show particularly dramatic lateral gradients. Indeed, they appear to run rather uniformly across the tops of wedge-like or lenticular bodies of silty mudstone which do show dramatic thickness changes, where as the silty beds display little or none. To that degree, these beds appear to be associated with times of "leveling out" of basin topography. More notably, the beds appear to separate episodes of basin axis shifting. For example, the Menteth Limestone separates large scale lenticular mudstone bodies (Deep Run and Kashong) which are offset from one another by ten of kilometers. Furthermore, the type 2 compact limestones are silty and heavily bioturbated, typically with swirly spreiten of *Zoophycos*, but they are not particularly shell rich, although they may display minor shelly basal lag deposits and/or thin shell hash beds at their tops.

We suggest the following model for the interpretation of the Type 2 (Menteth-like) silty limestones or calcareous siltstones. First, these beds developed only during intervals of generally shallow conditions over the northern part of the foreland basin. The shallowing episodes are probably represented by the Type 1 limestones or, more precisely, by the discontinuities that underlie the Type 1 limestones. Following shallowing by forced regression (relative sea level drop) minor sea level rise and/or development of a local subsiding basin provided accommodation space for the deposition of silty muds which were winnowed from basin marginal areas. These muds prograded into the most actively subsiding portions of the basin until those regions were filled to a threshold level, perhaps near the lower depth of storm wave base. At this depth, processes of winnowing and bioturbation became dominant, producing relatively compact deposits of heavily bioturbated silty and in some case limey sediment. Finer grained muds bypassed into more axial regions of the basin, presumably to the southwest into the western Pennsylvannia. Following the kinds of level filling of local basins, new episodes of subsidence, developed westward of the previous subsided basin axis and provided more accommodation space for a new package of mudstones. Note, for example the shift in position between the upper Deep Run and basal Kashong mudstones on either side of the Menteth Limestone. Minor sediment starvation associated with the deepening following deposition of the silty limestones produced thin shelly lags. It appears that shifts in the depocenters were also timed with intervals of true relative sea level rise which created widespread shelly lags. In some cases, the leveled out topography, created by the progradation of clastic wedges into the basins, persisted following the development of the silty capping limestones. In these instances, thin widespread calcareous beds showing relatively little lateral facies change also developed during the transgressive sea level rise interval; these include the widespread "trilobite beds" mentioned above.

Particularly strong pulses of relative sea level rise, with or without intervals of major shift in subsidence, produced the distinct, but very thin, phosphate, conodont- and/or bone-rich lag beds that lie at the bases of dark gray to black shales. We infer that these beds mark surfaces with maximum sediment starvation in the basin. A third type of erosional surface occurs beneath these maximum flooding surface lag beds. These erosion surfaces are typically channelized, and may remove major portions of the underlying transgressive sediments. We infer that this erosion in relatively deep water settings, was caused by basinward flowing currents and/or internal waves at water mass boundaries. Perhaps the most important factor in the development of highstand erosion surfaces was the absence of sediment input in offshore areas due to the trapping of siliciclastics near the source area. In the absence of new sediment input, even minor basinward flowing currents have been enough to initiate erosion, and over extended periods of time these currents could remove substantial amounts of the underlying sediment.

Hence, the stratal patterns of the upper portion of the Hamilton Group appear to reflect a mixture of actual relative sea level fluctuations and tectonic processes. The latter is recorded by the generally westward drift of the basin axis. This process appears not to have been steady, but rather episodic on a relatively long time scale, such that new basin depocenters were produced by local subsidence and largely infilled prior to their shifting to new and more westerly positions.

Finally, the processes which are recorded by subtle facies shifts and discontinuities in the upper part of the Hamilton Group were played out in a more extreme fashion during the deposition of the overlying Tully limestone and Genesee formations. A relatively major sea level drop or forced regression toward the end of the Windom deposition created a major unconformity that terminated the Hamilton Group. Moreover, it was not evidently preceded by a major or long-lasting shallowing-upward interval in the high parts of the Hamilton Group. Indeed, the beds that underlie this erosion surface represent some of the deepest water facies in the Moscow Formation. This erosion surface may have been produced, at least in part, by a relative uplift of the seafloor, and clearly somewhat more active diastrophic arching of the foreland basin floor did precede the sub-Tully erosion surface. This regionally complex pattern of uplift produced a complex paleotopography which was beveled during the erosive interval. Furthermore, development of anticline barriers on the seafloor, coupled with very low, but rising sea level may have produced one of the most highly sediment starved intervals during the Middle Devonian. This resulted in the deposition of clean, shallow water carbonates of the lower portion of the Tully over substantial areas of western New York, Pennsylvannia and perhaps as far south as Virginia. The basal Tully appears to represent a more extreme version of the Type 1 shallow water transgressive limestones discussed above. We agree with Ettensohn (1987) that this limestone and/or the erosion surface that underlies it may be in part the indirect result of renewed collision along the eastern margin of North America.

The Tully internal disconformity may represent another episode of forced sea level drop which eroded the lower Tully to some degree. A marked eastwardly migration of the basin axis is apparent in the deposition of the overlying upper part of the Tully. This eastwardly migration reverses the general pattern observed within the Hamilton Group below. It may signal either a time of renewed thrust-loading in the hinterland or an interval of load relaxation immediately following a thrusting episode (cf. Quinlan and Beaumont, 1984, Beaumont et al. 1988). The transition from the Tully into the Geneseo appears to initiate the third and most pronounced Devonian tectophase (Ettensohn, 1987). Again, as in the Moscow, but on a larger scale, the sediment starvation produced during this major relative deepening event is also associated with erosion. Westward beveling of the Tully Limestone beneath the Geneseo Black Shale is associated with development of the distinctive Leicester pyritic lag deposits. These lags were developed under stratified basinal conditions associated with the deposition of the initial black shales of the Catskill wedge. The abrupt pulse of deepening in the later portion of Tully deposition and into the overlying black shales of the Geneseo Formation is probably of both eustatic (Johnson et al., 1985) and tectonic (Ettensohn, 1987) origin. It is intriguing that precisely the same combination of probable eustatic sea level rise and enhanced basin subsidence is witnessed at each of the two earlier Devonian tectophases (i.e., Oriskany-Glenerie) into Esopus Shale and Onondaga Limestone into Marcellus black shale (Ver Straten et al., this volume).

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ROAD LOG FOR DEVONIAN MOSCOW-GENESEE FIELD TRIP

Mileage:		Instructions:
0.00	0.00	Leave University of Rochester parking lot, turn left onto Wilson Blvd.
0.10	0.10	Turn right onto Elmwood Avenue and cross Genesee River
0.30	0.20	Bear left at Scottsville Road
		Underpass under I-390
0.95	0.65	Turn right onto exit for I-390 south
1.05	0.10	Merge onto I-390
1.25	0.20	Crossing Genesee River
2.75	1.50	Overpass of Rt. 15 over I-390
3.25	0.50	Overpass of Rt. 15A over I-390
3.55	0.30	Cross Erie Canal
3.65	0.10	I-390/I-590 fork; bear right following I-390 South
4.15	0.50	Erie Canal on right; Monroe Community College to the west
8.55	4.40	Exit for NY State Thruway (I-90); continue south on I-390
9.05	0.50	Overpass of Thruway over I-390
13.35	4.30	Cross Honeoye Creek in approximate pre-glacial valley of Genesee River
16.95	3.60	Livingston County Line
19.05	2.10	Exit 8 for U.S. Rt. 20/Avon
19.30	0.25	Underpass under Rt. 20
20.55	1.25	Cross tributary of Conesus Creek
20.85	0.30	Minor hill slope exposures of Middle Devonian of Jaycox Shale; middle Devonian; excavation for I-390 in 1979 yielded a prolific fossil assemblage at this site.
22.15	1.30	Exit 9 for Lakeville; bear right onto exit
22.30	0.15	Junction Rt. 15; turn left (south)

22.40	0.10	Cross I-390
22.70	0.30	Railroad underpass over Rt. 15
23.25	0.55	Junction Triphammer Road (on right); <u>turn right onto Triphammer Rd.</u>
23.35	0.10	Cross Conesus Creek (swampy stretch)
23.95	0.60	Overpass over I-390
25.75	1.80	Side tributary of Conesus creek
25.95	0.20	Farm on right side borders Conesus Creek at Triphammer Falls, about 0.25 miles north of Triphammer Road; this is a classic locality for the Middle Devonian Centerfield Limestone
26.15	0.20	Junction Rt. 39; <u>turn left (south) onto 39</u>
27.10	0.95	Cross north branch of Jaycox Creek
27.90	0.80	Yellow barns on right are opposite Wheeler Falls of North Jaycox Creek. For optional stop pull off on shoulder of Rt. 39. Walk back to path along fence to west banks of Jaycox Creek

OPTIONAL STOP A. WHEELER FALLS
(See Description for Jaycox Creek)

Continue on Rt. 39

28.20	0.30	<u>White Devon Farm on left, pull off on right shoulder near gateway</u> into pasture, walk straight west from gate through pasture to clump of oak trees along Jaycox Creek, south branch for Stop 1. **Note: Visitors must obtain permission from the Wadsworth Estate,** Geneseo to study this locality.

STOP 1: JAYCOX RUN

Locality

Exposures along the south (main) fork of Jaycox Creek, on the property of William P. Wadsworth (White Devon Farm), 0.4 km (0.25 mi) west of NY Rt. 39 about 0.8 km. (0.5 mi) north of Nations Road, Geneseo, Livingston County, N.Y. (Geneseo 7.5 ft Quadrangle, 1950).

References

Cooper (1930), Grasso (1973), Baird (1979). The following description is from Brett and Baird (1981).

Descriptions of Units

Jaycox Creek exposes a nearly continuous section of the Ludlowville (Centerfield, Ledyard, Wanakah, and Jaycox members) and lower Moscow (Tichenor, Deep Run, Menteth, and basal Kashong members) formations (Figure 10). Exposures to be examined during the field trip are those in the upper portion of the Jaycox ravine associated with a series of three low waterfalls held up the thin limestone bands; a low upper falls is formed by the Menteth Limestone, a middle falls by the Tichenor Limestone, and the highest, lower falls by a calcareous band in the Jaycox Member.

The Jaycox through Menteth interval will be examined in detail. Faunal lists for these units are given by Grasso (1973). Units are described in ascending order as follows.

Ludlowville Formation

Wanakah Shale Member: The upper 4 m (13.2 ft) of the Wanakah Shale Member are visible below the lowest falls in Jaycox Creek. This constitutes an undifferentiated interval of medium gray fissile, friable shales with a low diversity fossil assemblage. *Ambocoelia umbonata*, *Mucrospirifer*, chonetids, and *Athyris*, the small rugose coral *Stereolasma*, trilobite remains (*Phacops* and *Greenops*), and various bivalves are common. No calcareous bands or concretions are present in this portion of the Wanakah Shale.

Jaycox Shale Member: Baird (1979) selected this creek exposure as the type section for his Jaycox Shale Member, the uppermost member of the redefined Ludlowville Formation. At this locality, the Jaycox consists of about 4 m (13.2 ft) of soft, bluish-gray shale with abundant thin fossil-rich horizons (biostromes) and calcareous bands (see article by Mayer, this volume).

The basal Hills Gulch bed of the Jaycox Member (Unit A) is a bioturbated, silty, argillaceous limestone band 30-35 cm (~1 ft.) thick, which grades upward from the calcareous uppermost Wanakah Shale (Figure 11). This blocky band which caps the lower falls of Jaycox Creek is notably rich in brachiopods (*Mucrospirifer*, chonetids, and *Mediospirifer*), bivalves (*Actinopteria*, *Modiomorpha*, *Paleoneilo*, and *Cypricardella*), and gastropods (*Mourlonia* and *Palaeozygopleura*). This bed has yielded exceptionally well preserved bivalves, including rare specimens in burrow positions. The bed contains numerous spreiten of *Zoophycos.* Westward this bed merges into a large coral-bearing unit resembling the overlying Tichenor limestone. Coral beds which occur at several levels within the Jaycox Member (Units B-C) contain assemblages very similar to those of older Centerfield Limestone, including a diversity of corals such as *Eridophyllum*, *Heliophyllum halli*, and the colonial form *Heliophyllum confluens* which is essentially restricted to this member. Large favositids and *Thamnoptychia* are also common at several levels within the Jaycox Member. Fossil layers contain abundant fenestellid bryozoans and large runner-like columns of crinoids modified as holdfasts (stolons). Diagnostic brachiopods of the Jaycox include *Parazyga hirsuta* and *Pentamerella pavillionensis*. These brachiopods are restricted to coral rich facies and recur only in association with biostromes (Centerfield Limestone, Jaycox Shale, Tichenor Limestone, and Windom coral beds).

Moscow Formation

Tichenor Limestone: The Tichenor Limestone (Unit D) constitutes a 30 cm ledge-forming band of medium gray, buff- to rusty weathering biomicrite (pack- or grainstone). As elsewhere, the Tichenor is an encrinite with abundant rugose and tabulate corals. The basal surface of the Tichenor Limestone represents a minor disconformity (Baird, 1979). Undersurfaces of projecting ledges, which can be examined in detail at this locality, show an irregular, sharp contact with the underlying Jaycox locality, show an irregular, sharp contact with the underlying Jaycox Shale. Very large hypichnia, probably crawling furrows similar to *Cruziana*, occur abundantly on the undersurface of the Tichenor.

The upper contact of the Tichenor Limestone at Jaycox Run is gradational through 15 cm of calcareous shale into the overlying Deep Run Member.

Deep Run Shale Member: Between the middle and upper falls, Jaycox Creek bed exposes 2.5 m (8.25 ft) of medium gray, calcareous mudstone of the Deep Run Member. These mudstones are distinctly harder (more calcareous) than most of the Jaycox Shale and break with irregular fractures due to the lack of fissility. Lower Deep Run shales (Unit E) contain a fauna similar to that of the Jaycox coral biostromes, including *Heliophyllum*, *Eridophyllum*, (but *H. proliferum* rather than *H. confluens*) and large favositid heads. Small mounds of fistuliporoid bryozoans project upward some 50 cm into the base of the Deep Run Member; these are exposed in the floor of Jaycox Creek. Associated with these mounds are abundant large stems and stoloniferous holdfasts of camerate crinoids; calyces of *Dolatocrinus*, *Megistocrinus*, and other crinoids occur rarely in the basal Deep Run. The Deep Run Member is particularly noted for the large size of its trilobites. Specimens of *Phacops rana*, *Greenops boothi*, and *Dechenella* sp. are abundant and are typically very large relative to those found in underlying and overlying shale units.

The upper beds of the Deep Run (Unit F) are sparsely fossiliferous bluish gray shale mudstones containing very abundant limonitized (pyritized) sinuous burrows. Occasionally, vague Zoophycos traces can be observed. These sparsely fossiliferous shales contain scattered stringers and lenses of large crinoid columnals and fenestellid bryozoans suggestive of storm lags. The Deep Run Member is gradational into the overlying Menteth Limestone Member.

Menteth Limestone Member: The Menteth Limestone (Unit G) which caps the uppermost falls in Jaycox ravine, consists of approximately 28-30 cm of light gray, buff-weathering irregularly bedded, strongly bioturbated calcisiltite (biomicrite). The lower contact of the Menteth is gradational, but the upper surface is sharply defined and hummocky suggesting a slight discontinuity or burrowed omission horizon. The Menteth contains a very abundant and often well preserved traces (*Zoophycos*) and scattered body fossils including spiriferid brachiopods, rare corals, and large trilobite fragments.

Kashong Shale Member: Only the lowest 30 - 50 cm of bluish gray, calcreous Kashong Shale (Unit H) crop out in Jaycox Run. These basal shales contain abundant fossils including the ramose tabulate coral *Thamnoptychia*, various bryozoans, brachiopods, camerate crinoids, and large trilobites.

Discussion

The basal Moscow Formation in the Genesee Valley represents a slightly different spectrum of environments. This sequence begins with the Tichenor Limestone which, as in most western N.Y. localities rests with marked disconformity on the Ludlowville Formation (Baird, 1979). Presumably, a portion of the upper Jaycox interval has been removed by submarine erosion prior to the deposition of the Tichenor. The Tichenor Limestone was probably deposited in a high energy, above wave base setting, as indicated by winnowing of fines, abrasion of fossil fragments, vague cross lamination, and overturned favosited coral heads. This unit is inferred to represent the shallowest water conditions in the entire sequence exposed at Jaycox Run. Gradation of the Tichenor into the overlying Deep Run calcareous mudstone indicates a return to lower energy, mud depositional conditions. The Deep Run thickens rapidly eastward into a lentil of calcareous mudstone which attains a maximum thickness of about 18 m. (59.4 ft) near Canandaigua Lake and then thins again eastward toward Cayuga Lake where the Deep Run pinches out (Baird, 1979). This lenticular configuration suggests active subsidence of a localized basin following the Tichenor deposition. The Deep Run locally contains cross laminated calcisiltite and calcarenite lenses which were probably transported into the depositional basin from nearby shoal regions during severe storms. Hence, the unit was deposited below normal wave base, but within reach of storm waves.

Deep Run mudstones grade upward into the Menteth Limestone, a uniformly thin (30-40 cm) but very widespread carbonate unit. The Menteth is a thoroughly bioturbated calcisiltite and is inferred to represent the terminal phases of infilling of the Deep Run depositional basin. Processes of sediment winnowing, probably aided by biological activity, became dominant over deposition of clays, resulting in a thin, lag deposit of skeletal debris and carbonate silt. Locally, coral beds developed on the upper surface of the Menteth. Menteth Limestone, in turn, grades upward into Kashong mudstones which faunally and lithologically somewhat resemble portions of the Deep Run. The entire Deep Run-Kashong interval (equals Portland Point of earlier workers; Baird, 1979) is inferred to represent deposition during a generally regressive interval. The sequence is bracketed by units which yield evidence of deposition under shallow water conditions. Minor oscillations of environmental conditions may have been the result of regional subsidence or downwarp of a shallow basin or trough.

Return to vehicles and <u>continue</u> <u>south</u> <u>on</u> <u>Rt.</u> <u>39</u>

28.70 0.50 Nations Road; to south is Wadsworth estate with ancient white oak trees

29.80	1.10	Village of Geneseo; Geneseo High School on right
31.00	1.20	Junction Rt. U.S. 20A; off south side of Geneseo, turn right onto 20A
31.35	0.35	South end of Geneseo State College dorms on right
31.70	0.35	Junction Rt. NY 63 on right, continue on 20A
31.95	0.25	US 20A turns to right; 63 goes straight. Follow 20A - turn right
32.25	0.30	Beginning of Dewey Hill roadcut; Genesee Formation (mainly Penn Yan dark gray shales)
32.55	0.30	Pull to right and park at guard rail for Stop 2. Walk across highway and enter pasture along Fall Brook at gateway. Proceed on foot upstream to exposures.

STOP 2. FALL BROOK, GENESEO

Locality

Exposures along the bed and banks of Fall Brook 0.2-0.6 km (0.1-0.37 mi) east of U.S. 20A and 39 and below Fall Brook Falls, Geneseo, Livingston County, N.Y. (Geneseo 7.5 ft Quadrangle, 1950).

References

Cooper (1930), Baird (1978), Baird and Brett (1983). The following description is from Brett and Baird (1981).

Description of Units: This stretch of Fall Brook exposes the upper portion of the Middle Devonian (Givetian) Moscow Formation (Windom and Kashong Shale Members) as well as the overlying Upper Devonian (Frasnian) Genesee Group (Figure 12).

Low cut banks along Fall Brook near U.S. 20A expose the upper 1.7 m (5.6 ft) of the Kashong Shale Member. This unit is comprised of sparsely fossiliferous, bluish gray mudstone with abundant pyritic tubular burrows. Body fossils include large ramose bryozoans, the brachiopods *Lingula* and *Tropidoleptus*, rare *Dipleura* trilobites, and phyllocarid fragments.

The disconformable contact with the overlying Windom Shale Member is exposed here and consists of a mudstone band approximately 10 cm-thick. This unit contains an admixture of reworked Kashong and Windom fossils and small phosphatic steinkerns and nodules (Little Beards phosphate bed). This heavily bioturbated sediment has been interpreted as a biogenically blurred unconformity or "stratomictic discontinuity" (Baird, 1978). The zone of phosphatic pebbles in the upper Kashong has been traced from Erie County, eastward to Owasco Valley, and is inferred to represent a widespread disconformity within the Moscow Formation (Baird, 1978). The informal "Unnamed Member" is absent here, having been cut out by erosion beneath the Little Beards bed.

Overlying the phosphatic pebble bed is typical basal Windom Shale: soft, medium gray shale with abundant specimens of *Ambocoelia umbonata*. A calcareous bed rich in trilobites occurs about 20 cm. above the Little Beards phosphatic nodule bed. Aside from the basal 30-50 cm, the lower 4.5 m (14.9 ft) of the Windom, are largely covered with slumped talus and are not accessible at Fall Brook. However, a complete Windom thickness of 15.63 m (51.6 ft) was measured at this locality.

The Windom section continues upstream at the base of a high cliff face on the northern side of Fall Brook. The lowest unit exposed here is soft shale rich in *Ambocoelia* brachiopods which may occur in small prod-like concentrations. The overlying Bay View bed consists of medium gray calcareous and very fossiliferous mudstone. Abundant fossils include *Pseudoatrypa*, *Mucrospirifer consobrinus*, *Protodouvillina*, the rugose corals, *Stereolasma* and *Amplexiphyllum*, crinoid columnals, and abundant *Phacops* trilobites. This unit is gradationally overlain by a band of medium gray calcareous shale and argillaceous limestone, approximately 40 cm-thick, which forms a falls in the creek bed. This is gradational above and below into less calcareous mudstone and contains abundant crinoid columnals in stringers, the small rugose coral *Stereolasma*, and fragmentary and complete specimens of the trilobites *Phacops rana* and *Greenops boothi*. Trace fossils in form of pyritic, sinuous burrows are very abundant in this zone. This bed is traceable into the "Smoke Creek bed" of Erie County, NY (Brett, 1974; Baird and Brett, 1983). The unit, in turn, is gradationally overlain by gray, calcareous and relatively fossiliferous mudstone with a fauna resembling that of Unit A which grades upward into darker, less fossiliferous shale.

A band of darker gray, relatively hard, fissile shale, forms a slight projection in the bank in some areas about 8 m (26.4 ft) above the base of the Windom. This interval (the Fisher Gully beds) contains abundant *Emanuella praeumbona*, large specimens of *Ambocoelia umbonata*, and occasional *Leiorynchus* and *Spyroceras*. It is overlain abruptly by a bed of crumbly medium gray mudstone packed with fossils including *Cystilphylloides*, *Helioplyllum*, and other large rugose and tabulate corals, atrypid brachiopods (*Pseudoatrypa* and *Spinatrypa*), *Mediospirifer*, and at least 55 other species of invertebrate fossils. Unit E, which occurs 8.57 m (28.3 ft) above the base of the Windom at Fall Brook, has been designated the Fall Brook coral bed closely resembles that of the slightly older Bayview coral bed and indicates a recurrence of nearly identical environmental conditions under which coral biostromes developed widely in western N.Y. The Fall Brook bed has been traced from Central Genesee county, eastward to Seneca Lake, a distance of approximately 90 km (56 mi). Large rugose corals are restricted to a 30-40 cm-thick band, however, other fossils including small rugose corals (*Amplexiphyllum* and *Stereolasma*), *Pseudoatrypa*, *Mediospirifer*, *Douvillina*, abundant crinoid columnals, and bryozoans persist upward for approximately 1.5 m (5 ft) into the overlying mudstone. This richly fossiliferous, but non-coral bed assemblage has been termed the Taunton beds by Baird and Brett (1983). The upper portion of the Taunton interval bears three zones of large, calcareous, nonseptarian concretions. Fossil bands run continuously through concretions, which indicates possible relationship between carbonate diagenesis and organic remains.

A thin (30 cm-band) of argillaceous limestone containing abundant small rugose corals (*Stereolasma*) and a few atrypid brachiopods Spezzano Gully bed occurs about 11 m (36.3 ft) above the base of the Windom Shale at Fall Brook. This unit grades upward into dark gray, slightly calcareous, petroliferous shale with abundant pyritic sinous burrows and occasional nodules of pyrite and scattered fossils including *Stereolasma* and *Pseudoatrypa*.

The upper Windom Shale at Fall Brook consists of very dark gray fissile, pyritic shale (Gage Gully Bed). This shale is generally sparsely fossiliferous, but some layers contain abundant *Allanella tullius*, small *Ambocoelia*, *Devonochonetes*, and rare *Leiorhynchus* and *E. praeumbona*. A zone of pyritized fossils near the top of this interval has yielded nuculid bivalves, the trilobite *Greenops boothi*, the nautiloid *Spyroceras*, and rare large specimens of the goniatite *Tornoceras* cf. *T. uniangulare*. The highest beds of the Windom are black, platey shale with *Allanella tullius*. The contact between these beds and the overlying Geneseo Black Shale is often difficult to recognize in the Genesee Valley area because their lithologies are similar. However, the Geneseo is somewhat harder, weathers rusty and is more petroliferous than the upper Windom. The contact, here, as elsewhere in western New York, is sharp and unconformable.

Small lenses of Leicester Pyrite generally less than 10 cm-thick, occur at the Windom/Geneseo contact. The Leicester in this area contains abundant tubular pyrite, apparently reworked burrow fillings, as well as small brachiopods, nuculid bivalves, and rare nautiloids and goniatites from the erosionally truncated Windom (Brett and Baird, 1982). Return to vehicles, proceed straight to turn around.

32.65	0.10	<u>In driveway, reverse route back up Dewey Hill</u>
33.35	0.70	Junction NY 63/39 at Fall Brook fruit stand. <u>Turn right (south) on NY 63.</u>
33.80	0.45	<u>Pull off at wide area along shoulder</u>; walk up bank to right and proceed down dirt path to brink of Fall Brook Falls.

STOP 2A

Scenic overlook of Fall Brook Gorge: Dark shales on upper creek banks are West River Member of Genesee Formation falls is capped by Genundewa Limestone (20 in-thick). Caution: Stay back from edge, vertical drop of over 100 feet (Possible lunch stop).

Return to vehicles and proceed straight to turn around in farm drive.

34.30	0.50	Cross Fall Brook

34.40	0.10	Turn around at driveway.
34.90	0.50	Return to U.S. 20A junction; proceed straight on Rts. 20A/63/39 and continue to Geneseo.
35.90	1.00	Fork of Rts. 39/20A at edge of Geneseo. Proceed straight on US 20A.
37.10	1.20	McDonalds on right (possible rest stop).
39.10	2.00	Upper end of Jaycox Creek tributary; Genundewa Limestone in exposed south of road.
40.70	1.60	Entrance ramp to I-390 (northbound) on right; bear right onto ramp.
40.90	0.20	Merge onto I-390
43.20	2.30	Cross Conesus Creek; outlet from Conesus Lake flows into the Genesee River.
46.80	3.60	Exit for NYS-U.S. 20; Avon, Lima; bear right onto exit.
47.10	0.30	Junction; Rts. 5 & 20; turn right (east)
50.60	3.50	Village of Lima
51.10	0.60	Junction Rt. 15A; continue on Rts. 5-20.
53.85	2.75	Valley of Honeoye outlet creek.
54.85	1.00	Route 65; West Bloomfield
57.30	2.45	View of Bristol Hills to south
59.80	2.50	Route 64; East Bloomfield
62.80	3.00	Rt. 64 to right leads south into Bristol Valley and classic Windom localities; near this location the "Unnamed member" is becoming truncated
64.00	1.20	Village of Centerfield; type locality of the widespread Centerfield Limestone is on Schaeffer Creek north of Rt. 20
64.60	0.60	Cross Schaeffer Creek
66.80	2.20	Junction Rt. 5-20 bypass around Canandaigua; turn right onto bypass
67.50	0.70	Junction 21 south; (goes to Naples); continue on Rt. 5-20.

69.10	1.60	Junction 21 north (into Canandaigua) <u>continue on Rt. 5-20</u>
69.40	0.30	Driveway to McDonalds on right (possible rest stop).
70.30	0.90	Junction NY 364 (to east shore of Canandaigua)
70.95	0.65	Creek to left exposes Levanna black shale.
72.65	1.70	Cross Hopewell Ravine; creek just below 20 exposes Jaycox and Tichenor members, small but excellent outcrop.
75.30	2.65	Village of Alonquin
77.40	2.10	Village of Flint
77.60	0.20	Cross Flint Creek
78.40	0.80	Large landfill site on right
81.80	3.40	Commercial strip on western fringes of Geneva
82.50	0.70	Junction 14A-245 to Penn Yan; <u>continue on Rt.5/20</u>
82.60	0.10	Junction Pre-Emption Road (County Rt. 6) to Bellona; <u>turn right (south)</u>; Pre-Emption line is an old western boundary line of Massachusetts territory.
85.30	2.70	Snell Rd. on left
85.40	0.10	Crossing Wilson Creek; downstream is a large waterfall over Centerfield Limestone; top phosphatic bed is exposed near the Pre-Emption Road bridge.
87.00	1.60	Benton Run on left exposes nearly complete Ludlowville Formation.
88.85	1.85	Junction Kashong Rd.; <u>turn left (east)</u>.
89.40	0.55	<u>Pull off in gravel parking area to right</u>. Proceed on foot down path on bank to Kashong Creek.

STOP 3. KASHONG CREEK, LOWER FALLS

Locality

Exposures along the bed and banks of Kashong Creek just south of Kashong Road and 1.5-2.0 km (0.9-1.2 mi) west of NY 14, township of Geneva, Seneca County, NY (Geneva South 7.5 ft Quadrangle, 1953).

References

Cooper (1930), Baird (1978, 1979).

Description of Units

This creek section is, arguably, one of the most complete upper Hamilton sequences exposed in the Finger Lakes Region; it is a classic reference section and the type locality for the Kashong Shale Member. Nearly the entirety of the Ludlowville and Moscow formations are exposed on Kashong Gully. We will examine the Moscow Formation in detail, commencing with the Tichenor Limestone which caps the highest wateralls (10 m, 33 ft) in the glen; a somewhat lower (8 m, 26 ft) waterfall 400 m upstream from the latter, shield up the Menteth Limestone Member, and a third very low falls (approximately 2 m, 6.6 ft high) is formed by the resistant upper phosphatic shell bed at the top fo the Kashong member and the overlying basal siltstone beds of the Canandaigua Member. A short distance upstream from this uppermost falls the section ends and the creek flows on glacial till.

The Hamilton beds in this area are gently deformed by Alleghenian (?) folding and faulting. The creek cuts most perpendicularly through a small anticline-syncline pair near the Kashong shale type section; the axis of the syncline is approximately at the low falls over the Kashong/Windom contact beds (discussed above); this contact is exposed again about 100 m upstream on the opposite (western limb) of the syncline, at the position of the old bridge across the creek. The bulk of Kashong section is exposed downstream from the small falls on the strongly west dipping eastern limb of the syncline. The crest occurs just downstream (east) of the high Menteth waterfalls. Below that falls the creek flows down the gentle east limb dipslope on the lower Deep Run beds for about 100 m (330 ft) before descending over the high Tichenor capped waterfalls. A relatively large down-to-the east normal fault with about 20 m (66 ft) displacement and minor splay faults occur near the mouth of Kashong Creek, but will not be examined on this trip.

The Tichenor through basal Windom sequence will be described and examined in detail; lower stratigraphic units at the base of the high Tichenor waterfalls are accessible by a steep (dangerous!) pathway down the south side of the gully or by walking upstream from the mouth of the Kashong Glen; however, we will simply observe these units from the top of the waterfalls. Stratigraphic units are described below in ascending order.

Ludlowville Formation

Upper Wanakah Shales: The lowest stratigraphic units observable in this part of the gorge are medium to dark gray shales on the upper Wanakah Member. These shales outcrop below a smaller waterfall 100 m (330 ft) downstream from the large Tichenor-capped falls; this latter falls is capped by silty, calcareeous mudstones of the basal Jaycox Member (correlative with Unit A of the Jaycox run description; Stop 10, page 86). Upper Wanakah at the base of this falls consists of relatively fossiliferous, gray shale characterized by an abundance of small *Pleurodictyum* corals, brachiopods (such as *Athyris*, chonetids, and *Ambocoelia*), and *Phacops* trilobites. This sequence is capped by a fossil has bed, identified by Baird (1981) as the Bloomer Creek bed. Overlying this layer in the face of the small waterfall are 2 m (6.6 ft) of dark gray to nearly black shales, bearing *Leiorhynchus*, small *Tropidoleptus* and small ambocoeliids.

Jaycox Member: The face of the high waterfalls and the creekbed down stream to the cap fo the next small falls expose a light bluish gray mudstone of the Jaycox Member. The lower half of this sequence consists of alternating calcareous mudstone and fossil rich limestone layers; these latter have a diverse fauna of abundant fenestrate and fistuliporid bryozoans, corals, especially *Trachypora*, brachiopods such as *Tropidoleptus* and *Longispina*, andabundant crinoidal debris. The upper Jaycox beds (upper half of falls face) consist of sparsely fossiliferous, hard, blocky, calcareous mudstone containing scattered very large speciments of the tabulate coral *Pleurodictyum*. The uppermost few centimeters of the Jaycox beds are visible at the top of the waterfalls.

Moscow Formation

Tichenor Limestone: This is a very widespread, thin (0.5-1 m, 1.7-3.3 ft) carbonate interval, traceable from Lake Erie, eastward at least to Chenango Valley; east of Cayuga Lake. (Baird, 1979; Griffing, 1994). The Tichenor is sometimes referred to as the basal bed of the "Portland Point" Member (Baird, 1979) At Kashong Glen this member, which crops out just upstream from the brink of the highest waterfalls, consists of thin-bedded argillaceous, crinoidal packstone. The Tichenor is highly fossiliferous, characterized by large crinoid columnals and bryozoans with a few solitary, large rugose and tabulate corals.

In the Seneca Valley, the Tichenor does not form a compct, sharply based ledge, as it does at localities both east and west of this area. In fact, the basal disconformity which characterizes the Tichenor at all other locations here is either very crptic or absent, and the lithology closely resemble that of more fossiliferous beds in the under-and overlying Jaycox and Deep Run members.

Deep Run Member: Overlying the Tichenor is about 10 m (33 ft) of hard, blocky, calcareous mudstone referred by Cooper (1930) to the Deep Run abundant, poorly preserved *Zoophycos* trace fossils. However, beds occurring about 1-2 cm above the Tichenor Limestone are highly fossiliferous, biostromal, argillaceous limestones resembling the Tichenor. The creek flows on dip slope of these units exposing relatively large bedding planes; these display small mounds of fistuliporoid bryozoans and crinoid columns (up to half meter in length); some of these columns form rhizomatous ("runner") holdfasts. Other common fossils include fenestrate and sulcoreteporid bryozoans, varied brachiopods, bivalves, large trilobites and well preserved (but difficult to extract) calyces of crinoids and blastoids. The large size of many fossils (especially trilobites) indicates favorable environments in terms of food supply, as might be expected in shallow muddy areas; however, the rarity and scattered to patchy distribution of fossils in most of the Deep Run Member suggests that colonization of the sea floor by benthos was sporadic and probably limited by soft, unstable substrate and/or high turbidity near the sea floor.

Note the considerably greater thickness of the Deep Run at Kashong Glen than at Jaycox Run (Stop 1). The Tichenor and Menteth limestones converge and eventually merge both east west of this area.

The Deep Run wedges out abruptly both east and west of the Finger Lakes. This unit which closely resembles the calcareous mudstone facies of the older Mottville, Centerfield, and Jaycox carbonate intervals apparently characterizes deposits of a shallow rapidly subsiding trough bordered by very shallow (within wave base), crinoidal sandy shoals. The Deep Run facies is considered to represent a wedge of rapidly-deposited, fine-grained sediments which were winnowed from adjacent shallow tichenor-type shoal environments that bordered a central trough. Interfingering of Tichenor-like lithologies with the barren mudrock indicates close proximity of the two types of facies.

Menteth Limestone Member: This is compact, 30 cm-thick silty; siliceous, heavily bioturbated limestone which caps the moderately high waterfall west (upstream) of the Tichenor falls. This limestone is particularly characterized by an abundance of *Zoophycos* burrows, *Mucrospirifer*, *Spinocyrtia*, and large specimens of *Phacops*.

The undersurface of the Menteth in some areas appears gradational with the underlying Deep Run; elsewhere, as here, it is undulating and displays large scoured out burrow fillings indicating some erosion prior to its deposition.

This unit appears nearly identical throughout its outcrop despite thickening and thinning of synjacent shales (see Jaycox Run, Stop 1, descriptions). We have interpreted this unit as a winnowed, bioturbated calcisiltite reflecting reworking of Deep Run muds, following infilling of a subsided trough.

Kashong Shale Member: The Kashong Member abruptly overlies the Menteth. It consists of bluish gray mudstones which are somewhat softer, less calcareous and silty than the Deep Run Member, with horizons of small irregular ovoid to pipelike carbonate concretions, silty limestones and highly fossiliferous calcareous mudstones. Fossils are moderately abundant in the bluish gray mudstones and include, particularly, brachiopods (*Tropidoleptus*, "*Spirifer*" *marceyi*, and *Mucrospirifer*), the tabulate coral *Pleurodictyum*, bryozoans, bivalves, large trilobites, including *Dipleura*, phyllocarids, and crinoid stems.

The Kashong at this location, displays repetitive, cyclic motifs, consisting of: (a) sparsely fossiliferous blue gray shale, (b) zones of grotesquely-shaped carbonate concretions, capped by, (c) thin beds of burrowed calcisiltite resembling the Menteth Limestone, and/or (d) complex condensed shell beds. There are two complete cycles and two thin incomplete cycles within the member. These are interpreted as minor regressive, shallowing upward hemicycles, the complex shelly phosphatic pebble beds record sediment starvation and/or winnowing associated with cycle reversals.

A widespread, complex shell bed occurs within the Kashong Shale about 2.6 m (8.6 ft) above the base. Informally named the "*Rhipidomella*-

Centronella" bed (or R-C bed) this unit contains a great abundance and diversity of fossils including the nominal brachiopods and the large spiriferid *Spinocyrtia* (commonly highly corroded), *Douvillina*, varied bryozoans, bivalves, gastropods, trilobites, and crinoid debris. This bed, which locally approaches the appearance of the Tichenor Limestone (and was formerly considered to be the top of the Portland Point Member; Cooper, 1930; Baird, 1979) appears to represent a widespread interval of winnowing and shell concentration, perhaps associated with a minor shallowing cycle. It is overlain by bluish gray shale resembling the lower Kashong.

Unnamed Shale Member: Only the basal 2 m (6.6 ft) of this previously unrecognized member of the Hamilton Group are exposed at this section. The unit commences with the Barnes Gully phosphatic pebble bed, a slight discontinuity may exist. At eastern localities a tongue of typical Kashong mudstone with *Tropidoleptus* is present between the phosphatic bed and the Curtice Road bed. This bed resembles burrowed calcisiltites of the Menteth, and consists of *Zoophycos*-burrowed calcareous silt, with a fauna of *Spinocyrtia*, *Mucrospirifer*, chonetids, *Pholidostrophia* and other brachiopods.

Also, the basal "unnamed" unit locally displays a shell bed especially rich in strophomenid brachiopods and containing rare rugose corals. The silty bed shows a gradational top, passing upward into silty shales with a deeper water biofacies of small chonetids which then passes up into soft medium gray shales with very abundant *Longispina*. This reflects a deepening event, stronger than those of the lower Moscow, which resulted in a return of lower aerobic to dysaerobic, deeper water enviornments in western New York. This transgression apparently explains the recurrence of litho- and biofacies closely similar to those of the lower Hamilton formation. We presume that such facies coexised with the shallower water facies of the Moscow interval, but that they were displaced southwestward into the basin axis and do not appear anywhere within the New York outcrop belt during deposition of the Jaycox to Kashong interval.

Discussion: The upper Hamilton Group in the vicinity of Kashong Glen displays several differences from outcrops of the equivalent interval to the west (Genesee Valley, see Stop 1). Several units (e.g., upper Ludlowville, Tichenor, and Deep Run shales) show subtle differences indicating deeper water in the Kashong Glen area than at sites either east or west of this region. For example, the Wanakah and Spafford-equivalent shales which are represented by fossiliferous, gray mudstones at Genesee Valley, and by comparable but siltier facies at Cayuga Lake, here are composed mainly of dark gray, fissile shale with a low-diversity, *Leiorhynchus*-dominated fauna; this seems to indicate deeper, more dysaerobic water in an area centered in the Seneca Valley than in the bordering areas. Similarly, the Tichenor-Deep run units are differentially thicker here than at Genesee Valley or the Cayuga region; they are also less carbonate rich; in the case of the Tichenor the limestone is a pack- or wackestone in the Seneca Lake region but a grainstone both to the east and west. It is less rich in large corals and crinoids and appears to be nearly conformable above and below whereas to the east and west it is bounded by an erosional unconformity at the base and, locally, on the top. Again, this seems to indicate an area of differentially deeper, quieter water near the Seneca Lake meridian.

		Return to vehicles; reverse route on Kashong Road (west)
89.95	0.55	Junction Pre-Emption Road; turn left (south)
90.40	0.45	Rice Road; enter Bellona Village
90.70	0.30	Cross Kashong Creek; turn left to parking area adjacent to Old Mill building; cross bridge on foot and proceed east on path to foot of Falls over Tully Limestone

STOP 4: KASHONG CREEK AT BELLONA

Locality

Exposures at waterfalls along Kashong Creek 0.1 km east (downstream) from Pre-emption Road and upstream to 0.2 km west of road, town Bellona, Yates County, N.Y. (Stanley 7.5 ft Quadrangle, 1952).

Reference

Heckel (1973)

Description

This outcrop provides an excellent section of the top of the Hamilton Group (Windom Shale), the unconformable overlying Tully Limestone, and the basal Genesee Group (Geneseo Shale). This classic section was critical in Heckel's (1973) subdivision of the Tully Formation into the lower and upper members separated by an unconformity.

Moscow Formation

Windom Shale: Lowest units exposed in this portion of Kashong Creek are fossil-rich medium bluish gray shales of the upper Windom Member ("Taunton Beds"); this 7 m (23 ft) interval contains many thin, lenticular brachiopod-, rugose coral -, bryozoan-bearing shell beds, mutually separated by sparsely fossiliferous mudstones. Thicker beds probably record relatively long spans of slow deposition during which shells accumulated and were reworked by occasional strong storm waves; overlying mudstones apparently represent rapid influxes of mud that terminated shell buildup.

The uppermost 2 m (6.6 ft) of the Windom display marked change to dark gray or nearly black shale with diminutive brachiopods. This records a relatively strong transgressive (deepening event) that occurred near the close of Hamilton deposition.

These beds display a sharp, erosional contact with the overlying Tully limestone. An interval of diastrophic upwarp and erosional truncation, perhaps associated with subaerial exposure prior to Tully deposition.

Nonetheless, despite this evident break both the Windom and the lower Tully belong to the same conodont subzone (middle *P. varcus* subzone).

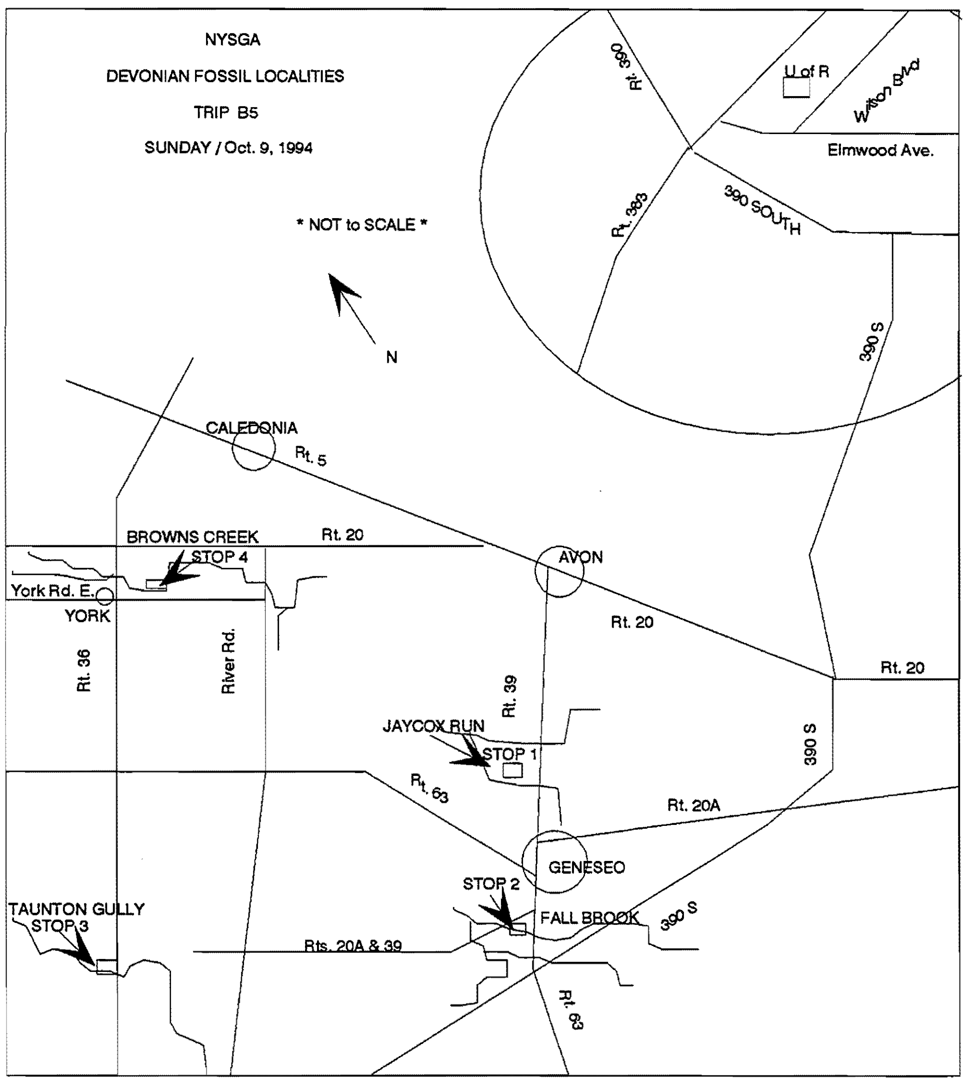
Tully Limestone: At the falls the light-gray, micritic Carpenter Falls bed (0.6 m-thick) of the Tully Lower Member, overhangs the Windom Shale; its basal surface displays large elongate burrows up to 8 cm. across, that were produced in overcompacted Windom muds and subsequently infilled with Tully matrix. The Carpenter Falls bed contains the index brachiopod *Hypothyridina*, as well as atrypids, small rugosans, trilobites, and crinoid debris. At the upper surface of this bed Heckel observed a widespread mid-Tully erosion surface, and, at this locality (upstream from Pre-emption Road Bridge), he noted possible truncated stromatolites at this surface. Hard fossiliferous wackestones, of the Taughannock Falls bed (Tully Upper Member) overlies this intraformational disconformity. The Taughannock Falls bed is capped by a thin, shaley bed with very abundant large rugose corals (*Heliophyllum* and *Cystiphyllolides*) and *Favosites* termed the Bellona coral bed for this excellent exposure. Upstream (west) from Pre-emption Road, the large corals are clearly visible on a large bedding plane on the creek floor. The Bellona coral bed has proven to be a key mappable unit. The Bellona bed is abruptly overlain by a thin (0.6 to 1.0 m) remnant of the Moravia bed; the upper Moravia bed and Fillmore Glen bed are absent here.

The top of the Moravia bed is marked by an erosion surface that, in this area, shows a thin layer rich in crinoidal debris, conodonts, fish bones, pyrite and glauconite; a few thin stingers of this lithology recur in the overlying black Geneseo Shale. This lag debris is believed to be a lateral equivalent of the Leicester Pyrite which is know to overlie the erosional top of the Tully about 12 km (7 m) northwest at Gorham. This bed has yielded conodonts diagnostic of the upper *hermanni-cristatus* subzone, whereas the underlying Moravia bed is of the middle *varcus* Zone age. Evidently, a considerable hiatus separates these units, during which time the upper Tully units were erosionally bevelled, prior to Geneseo deposition.

Geneseo Shale: The erosional upper surface of the Moravia bed, or, locally, the Bellona bed, is sharply overlain by platey, black Geneseo Shale, which is fossiliferous, except for the stringers of reworked crinoidal debris, men tioned above. This unit records major deepending of the Taghanic onlap. It is presently considered of the latest Middle Devonian (late Givetian) age.

		Return to vehicles
		<u>Reverse route and proceed north on Pre-Emption Road</u>
97.60	6.90	Junction US20 NY5, but <u>continue straight north on Pre-Emption Road</u>
102.80	5.20	Entrance to Oak corners Quarry on left. This is an excellent Onondaga Limestone section (Ver Straeten et al. trip - this guidebook).
103.70	0.90	Junction NY 96; <u>turn left (west)</u>
105.70	2.0	Town of Phelps

106.20	0.50	Cross Flint Creek; outcrops of upper Silurian dolostone
107.00	0.80	Junction NY88; a right turn here leads up to cuts below thruway in Bertie\Bois Blanc\Onondaga
107.10	0.10	Outcrops of Onondaga Limestone in Flint Creek left side
107.50	0.40	Driveway to left leads to outcrops of Nedrow shale along Flint Creek; (see Ver Straeten, this volume).
114.70	7.20	Cross Canandaigua outlet creek.
115.00	0.30	NY21; turn right (north)
115.05	0.05	Turn left into entrance to NY State Thruway (Manchester)
115.80	0.75	Merge onto I-90 (Thruway) westbound from entrance
117.20	1.40	Field exposure of Camillus Shale on left
124.60	7.40	Rising onto moraine; rest stop at top
126.30	1.70	Exit 45; I-490 eastern Rochester exit
137.50	11.20	Exit 46; I-390 north; exit onto 390 N and continue to university


NYSGA
DEVONIAN FOSSIL LOCALITIES
TRIP B5
SUNDAY / Oct. 9, 1994
* NOT to SCALE *
N
Rt. 390
U of R
Wilson Blvd
Elmwood Ave.
Rt. 383
390 SOUTH
390 S
CALEDONIA
Rt. 5
BROWNS CREEK
Rt. 20
STOP 4
AVON
York Rd. E.
YORK
Rt. 20
Rt. 20
Rt. 36
River Rd.
Rt. 39
JAYCOX RUN
STOP 1
390 S
Rt. 63
Rt. 20A
GENESEO
STOP 2
TAUNTON GULLY
STOP 3
FALL BROOK
390 S
Rts. 20A & 39
Rt. 63

DEVONIAN FOSSIL LOCALITIES IN WESTERN NEW YORK

STEPHEN PAVELSKY AND JAMES NARDI

NOTE: The accompanying text for this field trip has been incorporated into the articles for field trip B-6, following.

0.0 University of Rochester River Campus, Parking Lot B, outside Hutchison Hall; depart campus via Wilson Blvd.; turn right on Elmwood Avenue. At second light on Elmwood Avenue, turn left on Scottsville Road (Route 383)

1.0 Turn right on 390 South (just after overpass)

3.5 Stay right on 390 South and head toward Corning

25.0 Exit 8 at Geneseo; turn right and head west on Route 20A

30.1 Intersection of Routes 39 East and 20A; turn right on Route 39 East and proceed through the Village of Geneseo

33.0 **STOP #1. JAYCOX RUN**

Turn right into driveway of "White Devon Farm" (watch for sign). Park along left side of driveway. Be careful to leave ample room for horse trailers and traffic.

Entrance to the Wadsworth property and Jaycox Run is through a farm gate immediately across the road. Walk across the field to Jaycox Run (NW corner). Last person through gate MUST secure and fasten!

Locality: Exposures along the south (main) fork of Jaycox Creek, on the property of William P. Wadsworth (White Devon Farm), 0.4 km (0.25 mi) west of N.Y. Route 39, about 0.8 km (0.5 mi) north of Nations Road, Geneseo, Livingston County, New York.

References: Cooper (1930), Grasso (1973), Baird (1979). The following description is from Brett and Baird (1981a).

Description of Units: Jaycox Creek exposes a nearly continuous section of the Ludlowville (Centerfield, Ledyard, Wanakah, and Jaycox members) and lower Moscow (Tichenor, Deep Run, Menteth, and basal Kashong members) formations. Exposures to be examined during the field trip are those in the upper portion of the Jaycox ravine associated with a series of three low waterfalls held up by thin limestone bands. A low upper falls is formed by the Menteth Limestone, a middle falls by the Tichenor Limestone, and a highest, lower falls by a calcareous band in the Jaycox Member.

The Jaycox through Menteth interval will be examined in detail. Faunal lists for these units are given by Grasso (1973). For futher detail of unit descriptions see Brett and Baird in Miller, M.A. (ed.), 1986.

35.9 Head back through Village of Geneseo on Route 38 West to junction of Routes 39 and 20A West. Turn right onto 20A West and stay to the left

36.9 Turn right on Route 20A and Route 39 West

37.1 Roadcut through the Penn Yan shale as you descend the hill

37.3 **STOP #2. FALL BROOK.**

Turn around in driveway of the "Christiano Alfalfa Milling Co." (mustard colored sign on left) and park along guardrail facing uphill. Entrance to Fall Brook is at the uphill end of the guardrail

Locality: Exposures along the bed and banks of Fall Brook 0.2-0.6 km (0.1-0.37 mi) east of U.S. Route 20A-N.Y. Route 39 and below.
Fall Brook Falls, Geneseo, Livingston County, New York (Geneseo 7.5' Quadrangle, 1950)

References: Cooper (1930), Baird (1978), Baird and Brett (1983). The following description is from Baird and Brett (1981b).

Description of Units: This stretch of Fall Brook exposes the upper portion of the Middle Devonian (Givetian) Moscow Formation (Windom and Kashong Shale members) as well as the overlying Upper Devonian (Frasnian) Genesee Group.

37.4 Head back uphill to 20A; turn left towards Geneseo

38.3 Turn left on Route 63 North

39.5 Cross bridge over Genesee River

41.9 Continue on Route 63 North through Village of Piffard and pass Akzo Salt Mine on left

44.0 Junction of Route 63 North and Route 36 South; turn left on Route 36 South

46.4 **STOP #3. TAUNTON GULLY**

Pull off road just past small white house on right; park along roadside next to small field adjacent to house, before guardrail.

Walk back along edge of field, cross railroad tracks immediately behind the house and continue to far left-hand corner of field. Access to Gully is on small path through woods.

Locality: Gully exposures approximately 0.1 miles west of Route 36, 2.4 miles south of intersection of N. Y. Routes 36 and 63

References: Kirchgasser (1973; "Stop 6"), Patchen and Dugolinsky (1979; "Stop 35")

Description: About 25 feet of uppermost Windom Shale is exposed upstream from the railroad. There are a number of concretionary horizons occuring with layers of pyrite nodules. The shales below the first waterfall are blue-grey, shaly mudstones becoming darker and more compact further up in the section.

Faunal diversity is greater in the lower shales, consisting of bryozoan beds, stereolasmatid corals, trilobites, crinoids, and a rich brachiopod fauna. Above the first falls the fauna consist of most *Phacops* fragments and stereolasmatid coral horizons.

The Leicester Pyrite overlies the Windom. It is 4-6 inches thick and poorly exposed. Heavy collecting is discouraged to preserve the extent of the exposure.

The Geneseo and Penn Yan shales are exposed above the Leicester Pyrite (60-70 feet) and capped by one foot of Genundewa Limestone at the upper falls.

48.8 Return to Junction of Routes 36 and 63 and head north on Route 36

51.6 Continue north on Route 36 to Town of York. Turn right onto York Road East (center of village

52.0 **STOP #4. BROWN'S CREEK**

Proceed 0.4 miles to large, grey, metal pole barn on left just after bend in road (split-rail fence; parking on left side of road on either side of fence). Take wooden stairwell down to house and creek exposure.

Locality: Stream excavation 0.1 mile north of York Road East, about 0.4 mi east of Town of York on property of Jim Nardi

References: None for Ledyard Shale at this locality as it is a relatively new exposure

Description of Units: Until recent excavations, the Ledyard Shale was only poorly exposed upstream from the classic Centerfield/Levanna locality. The horizon studied on this trip is an excavation (widening) of the creek bed approximately 15 feet above the Ledyard/Centerfield contact. Of special interest are concretionary horizons and a horizon containing large *Phacops* trilobites. These are generally dark grey shales with many *Ambocoelia* brachiopods.

52.4 Return to Village of York. Turn right on Route 36 to Route 20. Turn right on Route 20 (east). Route 20 intersects with Route 390 North just east of the Village of Avon

Albany	(518) 899-7491
Buffalo	(716) 649-8110
Groton	(607) 898-5881
Rochester	(716) 359-1960
Middlesex	(908) 764-9676

Huntingdon

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